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Long Term Water Budget in a Grassland Catchment in Ireland

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Abstract

This paper describes some of the results of a one year hydrometeorological field experiment of a grassland catchment near Cork, Ireland (from November 1996 to November 1997). The objective of the experiment was to compare actual evapotranspiration from the annual water balance of a small catchment with Penman-Monteith reference evapotranspiration. In addition, Class A pan measurements are compared over a shorter period of 4 months with daily resolution. Actual evapotranspiration is 98% of Penman-Monteith reference evpotranspiration. In the investigated period (July to October 1997) Penman- Monteith reference evapotranspiration is 73% of Class A pan evaporation. The annual actual evapotranspiration from the water balance is 29.27 % of precipitation from the grassland. This suggests that for the humid Irish climate Penman-Monteith is a good estimate of actual evapotranspiration (for grasslands).

Introduction

The hydrologic cycle affects many environmental, physical, chemical and biological processes. It is integral to the understanding of climate, weather, biochemical cycles, and ecosystem dynamics. Therefore it is necessary to appreciate and define the driving forces of the different components within the hydrologic cycle. There have been many studies on the water balance at different catchment scales and climate conditions (e.g. Famiglietti et al. (1992), Famiglietti and Wood (1991), Milly (1994), Blackie (1993), Hudson and Gilman (1993), Tiktak and Bouten (1994), Parlange et al. (1996))which indicate the uncertainties of evapotranspiration rates as the most crucial components of the catchment annual water balance. However, the annual water balance should be able to provide realistic (actual) evapotranspiration rates when accurate measurements of precipitation input (P) and runoff (R) are available and storage effects e.g. change in soil moisture ΔS and deep seepage ΔG can be neglected. This is a legitimate assumption when the integration period is at least one year. Allen et al. (1989) described a high correlation (0.93 to 0.99) between lysimeter evapotranspiration measurements and calculations with the Penman-Monteith calculations over a wide range of climate locations and landuse. The reference evapotranspiration is often calculated with the Penman-Monteith approach using commonly measured meteorological variables. The difference between reference evapotranspiration ET_{ref} and actual evapotranspiration ET_{act} is used to determine a mean crop factor (Allen et al., 1989, Wright, 1981 a. 1982). In high rainfall climates there is little opportunity for significant soil water stress to develop and therefore the Penman-Monteith approach should produce reasonable values for estimating actual evapotranspiration. Brutsaert (1991) stated that several field experiments have shown that over longer periods (of order 1 year) pan evaporation is highly correlated with evapotranspiration from the surrounding vegetation under conditions of full cover and unlimited water supply. In humid climate conditions it seems possible to use pan data to estimate actual evapotranspiration. In this paper we investigate evapotranspiration from a grassland humid catchment using the annual water balance, Penman-Monteith and class A pan. This study is part of a larger investigation of hillslope processes on a small (14 Ha) catchment in Southern Ireland (Pahlow and Kiely, 1998).

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Site and experimental description

The research area is sited 25 km northwest of Cork, in the south of Ireland. It encompasses a 14 Ha grassland subcatchment (elevation 210 MoD) of the

Dripsey catchment which is in turn a subcatchment of the river Lee watershed. Research records began in November 1996. The site is agricultural grassland, typical of the landuse and vegetation in this part of the country. The grassland type can be described as high and moderately high quality pasture and meadow whereas the dominant plant species is perennial ryegrass (Collins and Cummins, 1996). The bedrock geology is Devonian Sandstone. The soil profile is characterised by a top 5 cm humus layer overlying a dark brown A horizon of sand texture to a depth of 20 cm. The yellowish-brown B horizon of sand texture grades into a brown gravely sand parent material at about 30 cm. The site is gently sloping to a stream. The investigated area of 14 Ha drains into a nearby stream. The field instrumentation relevant to the paper include: two rain gages, a class A evaporation pan, two water level recorders, an automatic weather station (Campbell Scientific Inc.), including a net radiometer, an air temperature probe, a relative humidity probe, a barometric pressure sensor, an infrared surface temperature sensor, four thermocouples for soil temperature, two soil heat flux plates and one monitor for wind speed and wind direction.

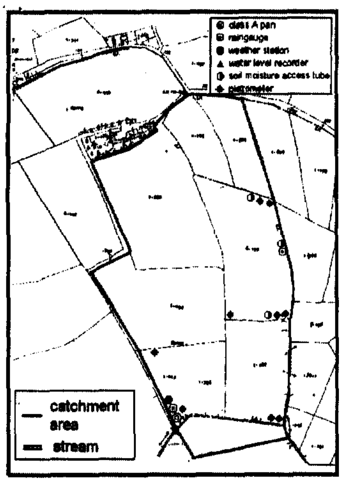

Fig. 1: Site description and instrumentation

The climate is temperate and humid influenced by the warm Gulf Stream in the North East Atlantic Ocean. Mean annual precipitation in the Cork region is about 1200 mm. The rainfall regime is characterised by long duration events of variable intensity and total depth, which occur at any time of the year. Short duration, high intensity events, occurring mainly in summer. Over the 12 month period (Nov. 1996 to Nov. 1997) the maximum daily rainfall was 75 mm on August 3, 1997 and the peak hourly intensity was 11.6 mm on August 26, 1997.

Atmospheric forcing factors

To understand the major influences on the evapotranspiration process we look closer at the specific climate conditions of our experimental site. The atmospheric forcing factors are mainly the net radiation, wind speed, relative humidity and vapour pressure deficit. The parameter are shown in Figure 2 for 132 days of 1997. We are interested in how and to what extent these parameters influence the evapotranspiration process. Figure 2 shows that evapotranspiration is generally correlated to the net radiation and the vapor pressure deficit ($\Delta e = e_s - e_{act}$ in kPa, where e_s is the saturated vapor pressure and e_{act} the actual vapor pressure).

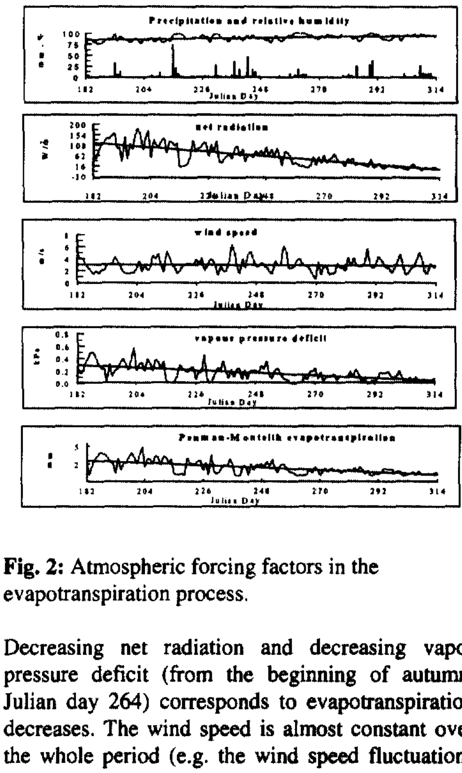

Fig. 2: Atmospheric forcing factors in the evapotranspiration process.

Decreasing net radiation and decreasing vapor pressure deficit (from the beginning of autumn, Julian day 264) corresponds to evapotranspiration decreases. The wind speed is almost constant over the whole period (e.g. the wind speed fluctuations are almost in the same range over the period). The relative humidity trendline shows an increase over time and exceeded 75% most of the time. The aerodynamic resistance r_a in [sec/m] is mainly dependent on the wind speed. The relative humidity indicates the potential amount of water, which can be absorbed from the air.

Evapotranspiration

As described in Allen et al. (1989) a common procedure for estimating evapotranspiration is to first estimate a reference evapotranspiration from a standard surface. 'Reference evapotranspiration' represents evapotranspiration from a known crop, such as grass, when enough water is available for active growth (assuming homogeneity and uniformity). Allen et al. (1989) showed that the Penman-Monteith equation (1) with aerodynamic and canopy resistances had the lowest values of standard errors of estimate (SEE) and the highest correlation coefficients in comparison with lysimeter measurements for different climate conditions. Rana et al. (1994) also emphasised the primary role of canopy resistance in the evapotranspiration process. We use the physically based Penman-Monteith equation to estimate the reference evapotranspiration ET_{ref}. The equation can be written (Monteith, 1965):

$$ET_{ref} = \frac{\delta \cdot R_n + c_p \cdot \rho \cdot (e_s - e_{act}) / r_a}{[\delta + \gamma \cdot (1 + r_s / r_a)] \cdot L_v} \quad (1)$$

where R_n is the net radiation in [W/m^2], e_s is the saturated vapour pressure in [kPa], e_{act} is the actual vapour pressure in [kPa], ρ is the density of the air in [kg/m^3], c_p is the heat capacity of the air [J/kg], r_a is the aerodynamic resistance in [s/m], r_s is the canopy resistance in [s/m], δ is the gradient of the saturated vapour pressure curve in [J/kgK], γ is the psychometric constant in [Pa/K] and L_v is the latent heat of vaporization in [J/kg].

We measure net radiation (Rn), saturated vapor pressure (e_s) and actual vapor pressure (e_a) with the automatic weather station, where each parameter is measured every 10 seconds and averaged over 20 minutes. Because of the large influence of temperature in the thermodynamically process of evaporation the results are improved when the latent heat ,the density of air, the psychometric constant and the gradient of the saturated vapour pressure curve were calculated as functions of temperature. The relevant equations are described below.

The gradient of the saturated vapour pressure curve δ was calculated with the measured values of saturated and actual vapour pressure, using the Clausius-Clapeyron-equation (Brustaert, 1992) in [J/kgK]:

$$\delta = \frac{de_s}{dT} = \frac{0.622 \cdot L_v \cdot e_s}{R_w \cdot T_a^2} \quad (2)$$

In this equation T_a is the air temperature in [°C] at 2 m.

The density of the air ρ was calculated dependent on the air temperature with the ideal gas law in [kg/m^3]:

$$\rho = \frac{1}{V} = \frac{m}{V} = \frac{P}{Rw \cdot T_a} = \frac{10^5}{461.9 \cdot (273 + T_a)} \quad (3)$$

The latent heat of vaporization was computed from the Clausius-Clapeyron-equation as a linear function of air temperature(for water temperatures between 0°C and 100°C) in [J/kg] (Stephan and Mayinger, 1974):

$$L_v = (3161.8 - 2.43 \cdot (273 + T_a)) \cdot 10^3 \quad (5)$$

The psychometric constant γ in [Pa/K] was calculated with the heat capacity of air c_p in [J/kgK], the latent heat L_v in [J/kg] (Brustaert, 1991). We used an hourly resolution for the atmospheric vapour pressure p_a in [Pa] as well. The equation can then be written as

$$\gamma = \frac{c_p \cdot p_a}{0.622 \cdot L_v} \quad (5)$$

The aerodynamic resistance r_a in [s/m] by Feddes et al. (1978) is:

$$r_a = \frac{0.622 \cdot \left(\frac{\rho_{air}}{\rho_{atm}} \right)}{1.15 \cdot FL \cdot v^{0.75}{}_{2m}} \quad (6)$$

where the wind speed v is measured at a reference height of 2 m. This assumes a surface layer under neutral atmospheric stability, where the wind is active and the solar radiation reaching the Earth's surface is limited by clouds (the most common stability condition for Cork).

When the buoyancy forces are negligible it is appropriate to describe the vertical profile of the mean wind speed with the well-known logarithmic wind profile (Parlange et al. 1996).

Table 1: Values of the factor FL from Feddes et al. (1978)

Leaf Area Index	FL
LAI < 1	$0.164 \cdot 10^{-7}$
1 < LAI < 20	$0.164 \cdot 10^{-7}\ LAI^{0.59}$
LAI > 20	$0.3704 \cdot 10^{-7} \cdot LAI^{0.2827} \leq 1.3 \cdot 10^{-7}$

We used the logarithmic wind profile equation (Brustaert, 1991) to correct the wind speed measured at a height of 2.85m to the reference height of 2m:

$$v(z=2m)=\frac{u_*}{k}\cdot\ln\left(\frac{z}{z_0}\right) \tag{7}$$

where u_* is the friction velocity (calculated with the rearranged equation (7) and wind speed at 2.85m height), k (= 0.4) is the Karman constant, z is the height above ground. z_0 (=2.0 mm) is the surface roughness for grass (10-50 cm height) (Brustaert, 1991).

In equation 6 FL depends on the leaf area index (LAI). The values of FL are shown in Table 1.

LAI varies with time, grass height and agronomic practices. In estimating the reference ET from a grass surface, the major variable related to leaf area is height, although many types of grasses can differ significantly in physiological composition and structure (Allen et al. (1989)). We used the approximation (Allen et al., 1989) :

$$LAI = 5.5 + 1.5\cdot\ln(h_c)\ for\ h_c > 0.03\ m \tag{8}$$

where h_c is the mean canopy height in [m]. The logarithmic relationship describes stem extension with less leaf development with increasing height. The weekly measured values of the canopy height h_c were transformed to hourly values with linear interpolation.

Canopy resistance r_s is dependent on climate, weather (radiation, vapor pressure deficit, aerodynamic resistance), agronomic practices (irrigation, grass cutting) and time scale (hour, day). Allen et al. (1989) present an equation for r_s as daily average, but pointed out the weakness of using daily averages due to diurnal variations of wind. The approach of Feddes et al. (1978) was used to calculate the canopy resistance r_s in [sec/m].

$$r_s = r_{s\min}\cdot\left(\frac{r_{s\max}}{r_{s\min}}\right)^{\frac{E_{pot}-0.7}{1.1-0.7}} \tag{9}$$

In this investigation r_s is presented as a function of the potential evaporation E_{pot}. It was calculated for hourly values. The values for r_{smin} is 30 [s/m] and r_{smax} is 100 [s/m], respectively when $0.7 < E_{pot} > 1.1$ (with E_{pot} in [m/s]). This potential or free water evaporation E_{pot} was also determined with the „resistance concept". Therefore the canopy resistance r_s in the reference Evapotranspiration equation ET_{ref} was set to zero. In this equation the water has to pass only the aerodynamic resistance r_a. It can be written as:

$$E_{pot}=\frac{\delta\cdot R_n + c_p\cdot\rho\cdot(e_s-e_{act})/r_a}{(\delta+\gamma)\cdot L_v} \tag{10}$$

Water Balance

The water balance seems to be simple and readily understandable in principle, but it is still rather difficult to measure in practice. To describe the various soil-water flow processes as separate phenomena in very detailed balances requires high temporal resolution measurements. Therefore the initial objective in this study was to look at the longer (annual) water balance equation

$$\overline{ET}_{act} = \overline{P} - \overline{R} \tag{11}$$

In this equation $\overline{ET}_{act}$ is the cumulative actual evapotranspiration in [mm], $\overline{P}$ is the cumulative precipitation in [mm] and $\overline{R}$ is the cumulative runoff per unit area from the basin in [mm] (the overbars means average over the year). We assume on a annual basis that the changes in soil moisture ΔS and the deep seepage ΔG are negligible. The annual water balance can be used to determine the annual actual evapotranspiration and compare it with the Penman-Monteith method. The ratio of actual and reference evapotranspiration can then be expressed as a crop factor.

Results

The class A pan data were compared on a daily increment with the computed evapotranspiration ET_{ref} (Penman-Monteith) over a period of four months (July 1997 to October 1997). The correlation was 0.73 and is shown in Figure 3. This seems to be a reasonable result for the conditions of full grass cover (height 0.03m to 0.15m) and unlimited water supply (the summer was wet as seen in Figure 2). The pan evaporation multiplied with 0.73 closely approximates the Penman-Monteith reference evapotranspiration ET_{ref}.

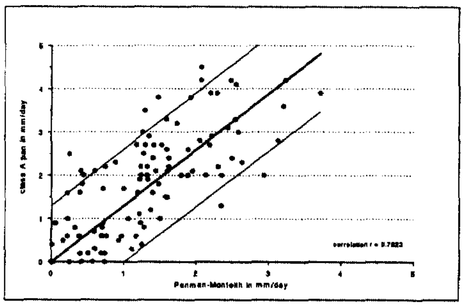

Fig. 3: Correlation between Penman-Monteith evapotranspiration and class A pan measurments

The results of the fitting with the reduced pan evaporation are included in Figure 4 (0.73*Class A pan).

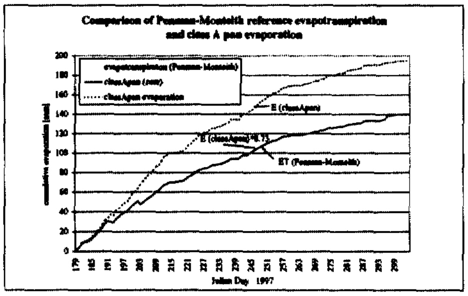

Fig. 4: Cumulative evapotranspiration of Penman-Monteith evapotranspiration and class A pan measurments

It is seen from Figure 4 that the cumulative evaporation values of the pan evaporation are very close to the computed fluctuations of Penman-Monteith reference evapotranspiration ET_{ref} (as described in Brustaert (1991) and Penman (1948)). The results of the free water evaporation E_{pot} are shown in Figure 5. To determine the transpiration rate in Figure 5, the evaporation E_{pot} was subtracted from the reference Evapotranspiration ET_{ref}. Over the 4 month period the evaporation rate is 47.17 % and the transpiration rate is 52.83 % of the total amount. The results are comparable to Baumgartner (1990). He mentions that 50% transpiration and and 50 % evaporation occur over full covered grassland for humid areas.

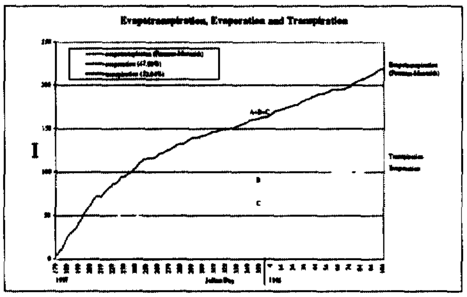

Fig. 5: Cumulative evapotranspiration of Penman-Monteith evapotranspiration and evaporation with r_c =0

Figure 6 show that the evaporation and evapotranspiration values tend to approach each other and remain constant as we move into the winter months (i.e. the grass growth diminishes). This suggests that wind speed and the net radiation mainly control the evaporation, while the transpiration is mainly controlled by the vapor pressure deficit.

The relative humidity, which is generally very high (>75) increases in winter and so the transpiration rate decreases and evaporation and transpiration approach each other.

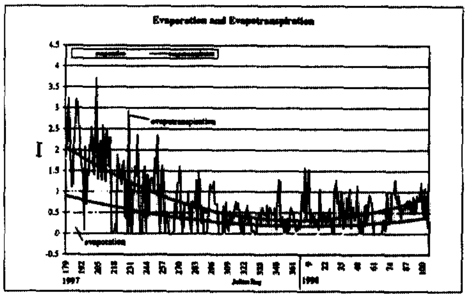

Fig. 6: Approximation of Evapotranspiration and Evaporation with the beginning of autumn

Figure 7 shows the water balance components and the computed reference evapotranspiration ET_{ref}.. From the water balance, we computed the cumulative actual evapotranspiration $\overline{ET}_{act}$ to be 29.27% of precipitation for the year From Penman-Monteith we computed the reference evapotranspiration $\overline{ET}_{ref}$ to be 31.73 % of annual precipitation. This result shows the influence of humid conditions in the evaporation process as the evapotranspiration rate is generally assumed as 60% of the precipitation (Dingman, 1994, Szilaggio and Parlange, 1998).

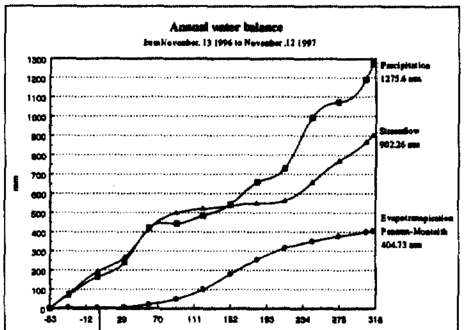

Fig. 7: Annual water balance for the calendar year 1997

The reference evapotranspiration $\overline{ET}_{ref}$ is higher than the actual evapotranspiration $\overline{ET}_{act}$ over the period of one year by 11% of the total amount

($\overline{ET}_{act}/\overline{ET}_{ref}$ is 0.92). The results seem to be reasonable as Allen et al. (1989) and Wright (1981,1982) found corresponding results ($\overline{ET}_{ref}/\overline{ET}_{lys} \Rightarrow$ 0.90 to 0.99) . The Penman-Monteith evapotranspiration overestimate by 6.4 % the „water loss" in the water balance (Table 3).

Conclusions and discussion

Water balance (precipitation and streamflow) measurement and meteorological station measurements on a small humid grassland catchment were carried over one year (November 1996 to November 1997). Actual Evapotranspiration from the water balance was 373.42 mm , reference Evaporation from Penman-Monteith was 404.73 mm. The actual Evapotranspiration from the water balance is 92 % of Penman-Monteith reference Evapotranspiration. The Penman-Monteith reference Evapotranspiration is 73 % of class A pan evaporation and

It is of interest to determine the ratio of actual and reference evapotranspiration for shorter time scales than one year. The combination of water balance and energy balance might show the main influences of the evapotranspiration process and the diurnal variations (first and second stage evapotranspiration). Several field studies and investigations (i.e. Allen et al. (1989), Brustaert and Chen (1996), Salvucci (1997), Malek (1992), Parlange et al. (1992)) showed remarkable results when considered two stages of drying (day and night time). It was observed in these investigations that the parameters of vapour pressure deficit, wind speed, net radiation and air temperature are in reality often significantly out of phase with one another during the course of a diurnal cycle. Thus, the two stages of drying is a useful concept, when calculating the evapotranspiration at a daily or hourly time scale.

From the perspective of precipitation, it is necessary also to use high resolution data (20 minutes intervals) because of the daily variation (Bormann et al., 1996). Another aspect is the determination of the correlation of soil moisture changes, precipitation and runoff (Parlange et al. (1996)). However, with the water balance we assume that the energy supply is not limited and the energy balance calculates evapotranspiration when vapor transport is not limited. Thus, the combination with the energy balance provides the actual evapotranspiration which can be used in two cases: (1) computation of the ratio of actual and reference evapotranspiration at a shorter time scale. (2) a detailed water balance which allows the calculation of the deep seepage ΔG when the change in soil moisture can be neglected, e.g. choosing a time period with equal soil water contents (Feyen et al., 1996), or measured directly, e.g. with TDR's.

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References

Allen, R.G., Jensen, M.E., Wright, J.L. and Burman, R.D., Operational Estimates of reference Evapotranspiration. *J. of Agron.*, 81, 650-662, 1989.

Blackie, J.R., The water balance of the Balquhidder catchments. *J. of Hydrol.*, 145, 239-257, 1993.

Bormann, H., Diekkrueger, B., Richter O., Effects of data available on estimation of Evapotranspiration. *Physics and chemistry of the Earth*, special issue, 21, 3, 171-175, 1996.

Brutsaert, W., *Evaporation into the atmosphere, Theory, History and Application*, Kluwer academic publishers, Dordrecht, pp.41-42, 1991.

Baumgartner, A. and Liebscher, H.-J., *Lehrbuch der Hydrologie*, Band 1 Allgemeine Hydrologie, Verlag Gebrueder Borntraeger, Stuttgart, 334-335, 1990.

Collins, J.F., Cummins. Th., *Agroclimat Atlas of Ireland*, 56-58, AGMET, 1996.

Dingman, S.L., *Physical Hydrology*, Macmillan Publishing Company New York, 56-57, 1994.

Famiglietti, J.S., Wood, E.F., Sivapalan, M. and Thongs, D.J., A catchment scale water balance model for FIFE. *J. of Geophysic. Research*, 97, 18.997-19.007, 1992.

Famiglietti, J.S. and Wood, E.F., multiscale modeling of spatially variable water and energy balance processes, *Water Resour. Res.*, 30, 3061-3079, 1991.

Feddes, R.A. , Kowalik, P.J. Zaradny, H., Simulation of field water use and crop yield. *Simulation Monograph*, PUDOC, Wageningen, 1978.

Feyen, H., Leutenberger, J., Papritz, A., Gysi, M., Fluehler, H., Schleppi, Runoff processes in catchments with small scale topography. *Physics and chemistry of the Earth*, special issue, 21, 3, 177-181, 1996

Hudson, J.A. and Gilman, K., Long-Term variability in the water balances of the Plynlimon catchments. *J. of Hydrol.*, 143, 355-380, 1993.

Milly, P.C.D., Climate, soil water storage, and the average annual water balance. *Water Resour. Res.*, 30, 2143-2156, 1994.

Monteith, J.L., Evaporation and environment. *Symp.Soc. Exp. Biol.*, 19:205-234, 1965.

Pahlow M., Kiely, G., Moehrlen, C., Field experiments and mathematical modelling of soil moisture in a grassland catchment. Submitted to *J. of Hydrol.*, 1998.

Parlange, M.B., Eichinger, W., Albertson, J., Evaporation and the Atmospheric Boundary Layer. *Reviews of Geophysics*, 1996.

Parlange, M.B., Katul, G.G., Cuenca, R.H., Kavvas, M.L. Nielsen, D.R., Mata, M., Pysical basis for a time series model of soil water content. *Water Resour. Res.*,28, 2437-2446, 1992.

Penman, H.L., Natural evaporation from open water, bare soil and grass. *Proc. Royal Society London Series A*, 120-145, 1948.

Penman, H.L., Estimating evaporation. *Trans. Amer. Geophysic. Union*, 37, 43-46, 1956.

Rana, G., Katerji, N., Mastrorilli, M. and El Moujabber, M., Evapotranspiration and canopy resistance of grass in a mediterranean region. *Theor. Appl. Climatol.*, 50, 61-71, 1994.

Stephan, K. and Mayinger, F., *Thermodynamik* Band 1, Springer Verlag, p. 215, 1974.

Tiktak, A. and Bouten, W., Soil water dynamics and long-term water balances of a Douglas fir stand in the Netherlands, *J. of Hydrol.*, 156, 265-283, 1994.

Wright, J.L., Crop coefficientsfor estimates of daily crop evapotranspiration. *Irrig. Scheduling for water and energy conversation in the 80's. Proc. Irrig. Scheduling Conf. Chicago, Il.*, ASAE, St. Joseph, MI, 1981.

Wright, J.L., New evapotransiration crop coefficients. *J. Irrig. Drain. Div.*, *Am. Soc. Civ. Eng.*, 108,(2), 57-74, 1982.