BEHAVIOUR OF CARBON DIOXIDE AND WATER VAPOUR FLUX DENSITIES FROM A DISTURBED RAISED PEAT BOG

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ABSTRACT

Measurements of carbon dioxide and water vapour flux densities were carried out for a disturbed raised peat bog in the north of the Netherlands during an 18 month continuous experiment. Tussock grass (sp. *Molinea caerulae*) mainly dominated the vegetation of the bog area. The maximum leaf area index (LAI) of the vegetation reached a numerical value of about 1.7 in mid-August.

When the LAI is large enough (LAI > 0.2), a mean net uptake of carbon dioxide is observed with a clear daily pattern. The total evapotranspiration consists of a soil, an open water, and a plant transpiration part. When the LAI is large enough (LAI > 0.2), plant transpiration dominates the total evapotranspiration. The mean daily transpiration pattern, however, is not similar to the carbon dioxide uptake pattern.

During the summer months, the daytime carbon dioxide uptake shows a single early morning maximum value followed by a decline in uptake during the rest of the day. The evapotranspiration, however, follows more or less the incoming short-wave radiation pattern. Effects of the vapour pressure deficit are suggested as a possible cause of this discrepancy. Copyright © 2002 Royal Meteorological Society.

KEY WORDS: evapotranspiration; carbon dioxide; climate variables; peat bog; vapour pressure deficit

1. INTRODUCTION

Wetlands are biogeochemically active because of their high productivity and redox gradients. In particular, wetlands are major natural sources of carbon dioxide (CO_2), methane, and sulphur compounds and can have high rates of denitrification and nitrogen fixation (Sahagian and Melack, 1998). Surface cover and hydrology have been found to affect significantly the flux densities of carbon dioxide, water vapour, and other trace gases such as methane (Lafleur, 1990; Shurpali *et al.*, 1995; Campbell and Williamson, 1997).

Various studies have pointed out that evaporation from non-vascular *Sphagnum* wetlands was reduced remarkably when the water table dropped (Ingram, 1983). With the presence of vascular plants, evaporation was not only found to be related to the soil water availability but also to plant factors (Ingram, 1983; Kim and Verma, 1996). Changing hydrological conditions and aeration, possibly due to higher temperatures, have changed many regions in the Arctic tundra from sinks to sources of CO_2 (Oechel *et al.*, 1993).

With peat in the Northern Hemisphere's wetlands containing about one-third of the world's carbon pool (Gorham, 1991), a good understanding of the mass and energy cycles under various climatic, hydrological, and ecological conditions is of major importance to predict the effect of possible future climate changes and anthropogenic factors.

The objective of this study is to gain insight into the annual behaviour of the surface flux densities of water vapour and CO_2 from a disturbed true raised peat bog. To achieve this, a field experiment was executed during

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a one-and-a-half year period in a disturbed raised peat bog, in the north of the Netherlands. In contrast to most raised peat bogs in the Northern Hemisphere dominated by *Sphagnum*, peat bogs in the Netherlands are dominated by tussock grass and heather vegetation, e.g. sp. *Molinea caerulea*, *Calluna vulgaris*, and *Erica tetralix* (Schouwenaars and Vink, 1990). Until 1980, Dutch peat bogs were used for fuel production, and peat was cut in an extensive part of the Netherlands. However, for different reasons, some areas were never largely affected, although drainage had a large impact on all peat bog ecosystems and resulted in the typical tussock grassland vegetation (Schouwenaars and Vink, 1990).

The research was done within the framework of the Surface Layer Integration Measurement and Modelling (SLIMM) project (Vugts *et al.*, 1994). The eddy covariance technique, used in the experiment, allowed the direct measurement of mass and energy transfer at a height above the surface, resulting in a spatially integrated flux density. Here, the measured flux densities were linked to the environmental conditions, with a focus on temperature and vapour pressure deficit.

2. SITE DESCRIPTION

The experiment was done in a disturbed raised peat bog in the north of the Netherlands (lat. 53°00′30″N, long. 6°23′52″W, altitude +11 m) during a one-and-a-half year period in 1994 and 1995. The dominating plant species in this tussock-grassland was *M. caerulea* (>75% of the vegetation), but the heather species *C. vulgaris* and *E. tetralix* could also be found. In the sections in which the soil was permanently saturated, *Sphagnum* spp. (<1% of the vegetation) were also present (Altenburg *et al.*, 1993). The area taken by the individual tussocks can vary from 20 cm^2 to more than 1 m^2. The average height of the tussocks is approximately 0.4 m. A dense layer (0.1 m) of dead organic material from the previous growing seasons covered the tussocks and the hollows in-between. In these hollows, spots of open water (<5% of the surface) could be observed, sheltered by the surrounding vegetation. Underneath the tussock surface was a saturated dark-brown layer of peat. The depth of this layer varied from 0 to 2.5 m. At the measuring site, this highly porous layer was about 0.3 m deep, and at greater depth a grey sandy loam was present. Throughout the year, the water table varied, depending on the season, from 0 to 0.2 m below the tussock−soil interface, but the soil remained saturated.

The growing season of *M. caerulea* lasted from May until October, with a maximum leaf area index (LAI) of about 1.7 during mid-August. During this period, green shoots grew up to a height *h* of 1.2 m above the tussock. From November until the beginning of May, no green shoots were present. What was left above the tussock was the brownish surface of organic material and some brown shoots standing upright throughout the winter period. For 1995, it was assumed that the LAI followed the same trend, except for a different timing of the onset and end of the growing season and maximum LAI.

3. INSTRUMENTATION

A lattice tower was instrumented with an eddy-covariance system installed at a height of 8 m. This system included: a 3-D sonic anemometer (3-D Solent Res. Gill Instruments Ltd., model A1012R2), a fine wire thermocouple (home made), and an inlet tube (polyethylene, 4 mm internal diameter) leading to an infrared CO_2 and H_2O gas analyser (LI-COR Inc., Lincoln, NE, model LI-6262). The 3-D sonic anemometer and the sampling tube were placed together, 0.05 m apart.

For all wind directions the fetch was more than 1 km and between 180 and 270° the terrain was slightly sloped. The closed-path infrared gas analyser (IRGA) was placed in a box at the base of the mast. A pump was used to draw the air from the inlet to the IRGA at a rate of 10^{-4} m^3 s^{-1}. The pump was placed in such a way as to cause the pressure inside the sampling chambers to be slightly below atmospheric pressure, but minimizing the chance of condensation of water inside the tube. The IRGA ran in absolute mode and was re-calibrated every week. During this period the analyser's gain typically showed a 2 to 5% drift. The soda lime and magnesium perchlorate, used to keep the reference cell free of CO_2 and H_2O respectively, was changed every week. Every week, dry nitrogen from a cylinder was used as a zero calibration gas for CO_2 and

water vapour, while dry air gas (400 ppm CO_2) was used as a second reference. An aspirated psychrometer was used as a second water vapour calibration.

A net radiometer (Middleton & Company Pty. Ltd, model CN1-R), two pyrgeometers (Kipp & Zonen, model CG 1), and two pyranometers (Kipp & Zonen, CM11) were used to measure the net radiation R_n, the incoming long- R_l and short-wave R_s radiation, and the outgoing long- and short-wave radiation respectively. A second, 6 m high tower was instrumented with aspirated psychrometers (home made) and sensitive cup anemometers (home made, stalling speed 0.15 m s^{-1}) at 2, 4 and 6 m height above the surface. Soil temperature profiles (0.025, 0.2, 0.5 and 1 m) and the soil heat flux density G (TNO-Delft, model WS 31-Cp) at 0.05 m depth were measured in the actual peat soil underneath a tussock, but not in the tussock itself. To obtain the surface heat flux, corrections were made for differences in heat conductivity between the soil and heat plate (Morgensen, 1970; Watts *et al.*, 1990; Van Loon *et al.*, 1998) and heat capacity storage above the plate (Fritschen and Gay, 1979). The heat capacity of the soil, needed for the latter correction, was measured by using a non-stationary needle probe technique as described by Van Loon *et al.* (1989).

The LAI was measured by the leaf tracing technique, as described by Kvet and Marshall (1971).

4. DATA ACQUISITION

The meteorological instruments were sampled at 0.25 Hz. At 30 min intervals, data were averaged and stored in data loggers for later processing. The sonic anemometer was sampled at 20.8 Hz. Analog signals from the IRGA and thermocouple, connected to the anemometer's interface unit, were sampled at 5 Hz.

The raw data of the eddy covariance system were stored on a PC and processed later, using a first-order recursive digital filter with a time constant of 200 s (McMillen, 1988). Here, a moving average was subtracted from every sample to obtain the fluctuating value of all the measured components. A software program (Van den Hurk, 1996) performed the necessary corrections, including: coordinate rotation (McMillen, 1986), Webb corrections (Webb *et al.*, 1980), and frequency-response corrections (Moore, 1986), required for calculation of the half-hour averaged flux densities. The power and co-spectra of the eddy-covariance measurements (not shown) and the ability to close the surface energy balance were analysed to test the quality of the data. As an example, the available energy $R_n - G$ versus the sum of the sensible heat H and latent heat λE flux density is shown in Figure 1 for ten consecutive days in August (1995). Here, the data were excellently fitted with a linear regression line described by $R_n - G = 0.96(H + \lambda E)$ (W m^{-2}) with a coefficient of determination $R^2 = 0.94$ ($N = 470$).

The course of LAI was measured during the growing season of 1994. Here, three patches in the vicinity of the towers were selected with characteristic features, varying from extremely wet, relatively dry, or with many pools of shallow open water. These patches were sampled every 3 weeks. The course of the monthly average has been plotted in Figure 2. Following the growing stages given by Kim and Verma (1990) for a grassland ecosystem, these stages are the early growth stage, peak growth stage, early senescence stage, and senescence stage, and these have also been indicated in Figure 2.

5. RESULTS

During the one-and-a-half year period, the daily mean air temperature ($z = 2$ m) ranged between -6 and 25 °C. The air temperatures for both summers were above the (30 year) climatic normal, with a maximum deviation in July 1994 of 4 °C. The 0.025 m soil temperatures were below 20 °C, with a maximum of 19 °C in August 1994. In contrast to the air temperature, the daily amplitude of the soil temperature was typically less than 5 °C. This was believed to be due to the insulating effect of the dense layer of organic material on top of and in-between the tussocks. The monthly averaged 2 m air temperature T_a, and the 0.025 m soil temperature T_s, are shown in Figure 3, where the solid circles represent the average temperature and the open circles the maximum and minimum temperatures measured during that month.

During the summer months, extensive periods of relatively hot weather with high daily maximum air temperatures recorded at 2 m (>25 °C) were followed by shorter periods with showers and thunderstorms.
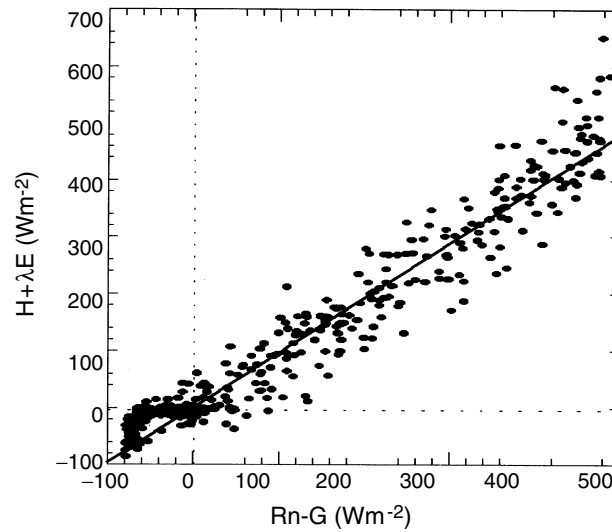
Figure 1. The energy balance closure. The half-hourly averaged available energy ($R_n - G$) is plotted against the sum of the partitioning components ($H + \lambda E$) for ten consecutive days in August 1995. The linear regression line $R_n - G = 0.96(H + \lambda E)$ could be fitted to the data with $R^2 = 0.94$ ($N = 470$)
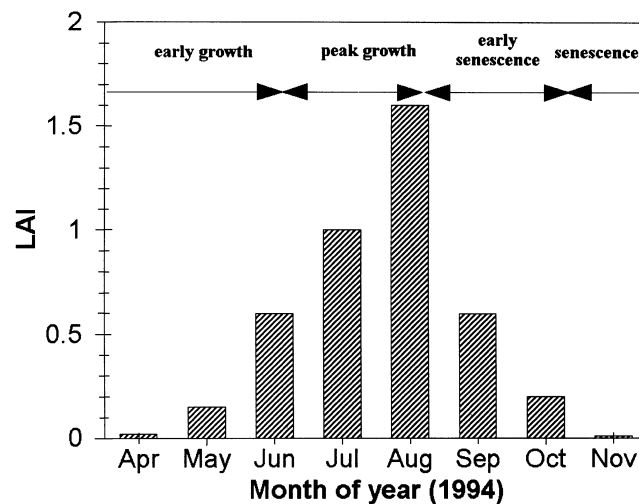

Figure 2. Annual courses of the LAI from the three patches around the experimental site during the growing season of 1994. Also indicated are the growing stages of the dominating vegetation, according to Kim and Verma (1990)

The amount of precipitation was, on average, 25% less than the long-term climatic normal during this period, except for September 1994 and 1995, which were about 50% wetter. The water table of the area was kept constant as much as possible in order to get back the original vegetation. However, it was observed that, during this particular dry summer period, the water table at the observation site dropped about 0.2 m during the period between mid-July until mid-August. During autumn and winter, precipitation was more regularly distributed and higher than on average.

Monthly averaged diurnal courses of the CO_2 and H_2O flux densities can give insight into climatic- and plant-related factors on the exchange of CO_2 and H_2O between the surface and the atmosphere. Here, half-hourly flux densities for a particular month and at a particular time are averaged to obtain the ensemble averaged diurnal course for the month. Such a representation of measurements will smooth the extremes due to random errors in the measurements and extreme weather events.
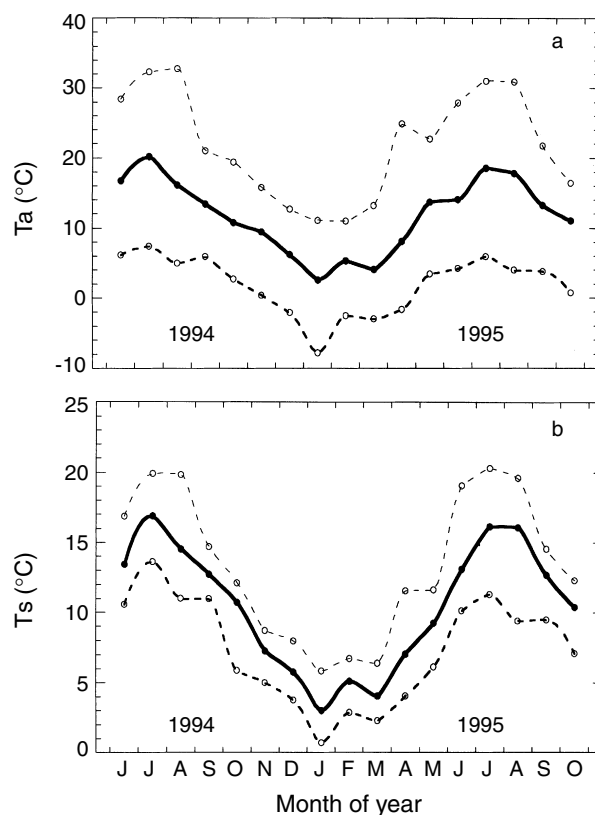
Figure 3. (a) Monthly averaged 2 m air temperature T_a during 1994 and 1995. (b) Monthly averaged 2.5 cm soil temperature T_s during 1994 and 1995. Also indicated are the averaged maximum and minimum temperatures as measured during a particular month

Figures 4 and 5 depict these monthly averaged CO_2 flux density F_c, and H_2O flux density, λE, respectively, during different stages of the vegetation, from early spring until late summer irrespective of the measurement year. A negative flux density is directed towards the surface, and a positive flux density is directed away from the surface. Following other recent literature, we use the terms 'peak' or 'maximum' F_c for the largest negative CO_2 flux density (Baldocchi, 1994; Neumann *et al.*, 1994; Verhoef *et al.*, 1996).

Figure 4 displays the monthly averaged CO_2 flux density for March to September. All data were used here, with the exception of data with apparent errors, such as data measured during brief power failures. In Figure 4, the absence of green vegetation in March and April (no growth stage) is clear. The average exchange of CO_2 during this month was apparently totally caused by respiration, resulting in a net release of CO_2 (positive F_c during the course of the day. In May (early growth stage) the LAI had increased from nearly 0 to 0.2, as can be inferred from Figure 2. Despite the presence of green shoots and the associated photosynthetic activity, there was a net release of CO_2 throughout the day and night. Soil and plant respiration, in this case, still exceeded the photosynthesis. Probably due to a higher soil temperature, respiratory losses were higher than in March.

In the following three summer months (early growth, peak growth, and early senescence stages), F_c exhibited a distinct diurnal course. Night-time efflux of CO_2 varied between 0.05 and 0.1 mg CO_2 m^{-2} s^{-1} (1.1–2.3 µmol CO_2 m^{-2} s^{-1}). In June, the CO_2 flux density peaked at 11:00 with a maximum of −0.18 mg CO_2 m^{-2} s^{-1} (−4.1 µmol CO_2 m^{-2} s^{-1}). July, on the other hand, showed a maximum CO_2 flux density around 9:00 in the morning of about −0.22 mg CO_2 m^{-2} s^{-1} (−5.0 µmol CO_2 m^{-2} s^{-1}). After this maximum, as the day progressed, the rates of net CO_2 uptake diminished with time. Similar behaviour was also reported for vegetation suffering from water stress in combination with an increase in air temperature and vapour pressure
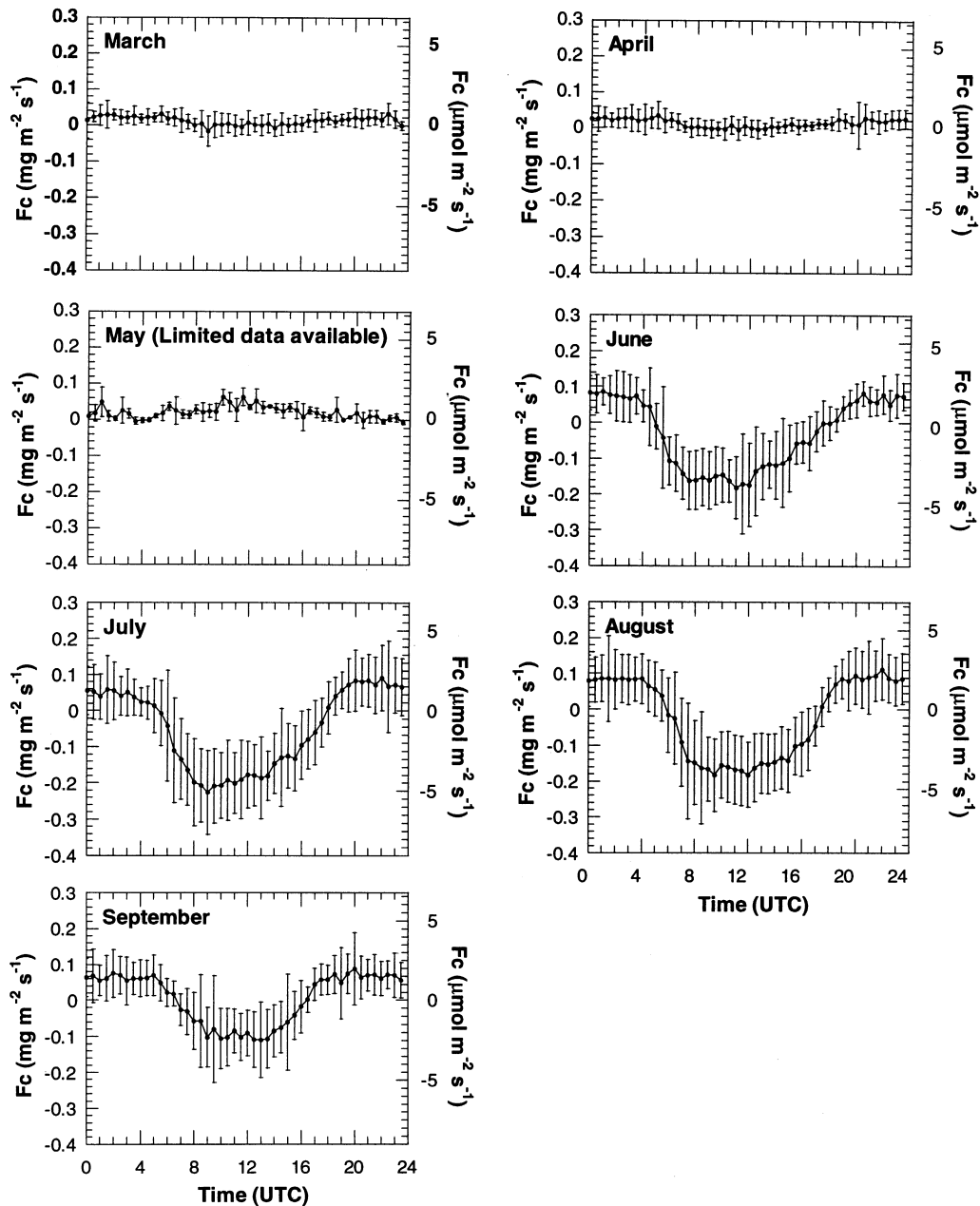
Figure 4. Diurnal patterns of monthly averaged half-hourly CO_2 exchange rates from March 1995 to September 1995. Bars indicate the standard deviation of a particular half-hourly mean value. UTC = local time − 2 h

deficit (Greco and Baldocchi, 1996; Neumann *et al.*, 1994; Shurpali *et al.*, 1995; Verhoef *et al.*, 1996). In our study area, however, *M. caerulea* was never observed to suffer from water stress.

By the end of August (start of early senescence stage), the course of CO_2 uptake was rather constant during daytime, with a maximum of −0.18 mg CO_2 m^{-2} s^{-1} (−4.1 μmol CO_2 m^{-2} s^{-1}). The period of net uptake of CO_2 was about 2 h shorter than during June and July. With the start of autumn, green vegetation was still present but started to diminish rapidly. September (senescence stage), therefore, showed a diurnal course with significantly lower maximum daytime values for λE and F_c. The period of net uptake of CO_2 was about 2 h
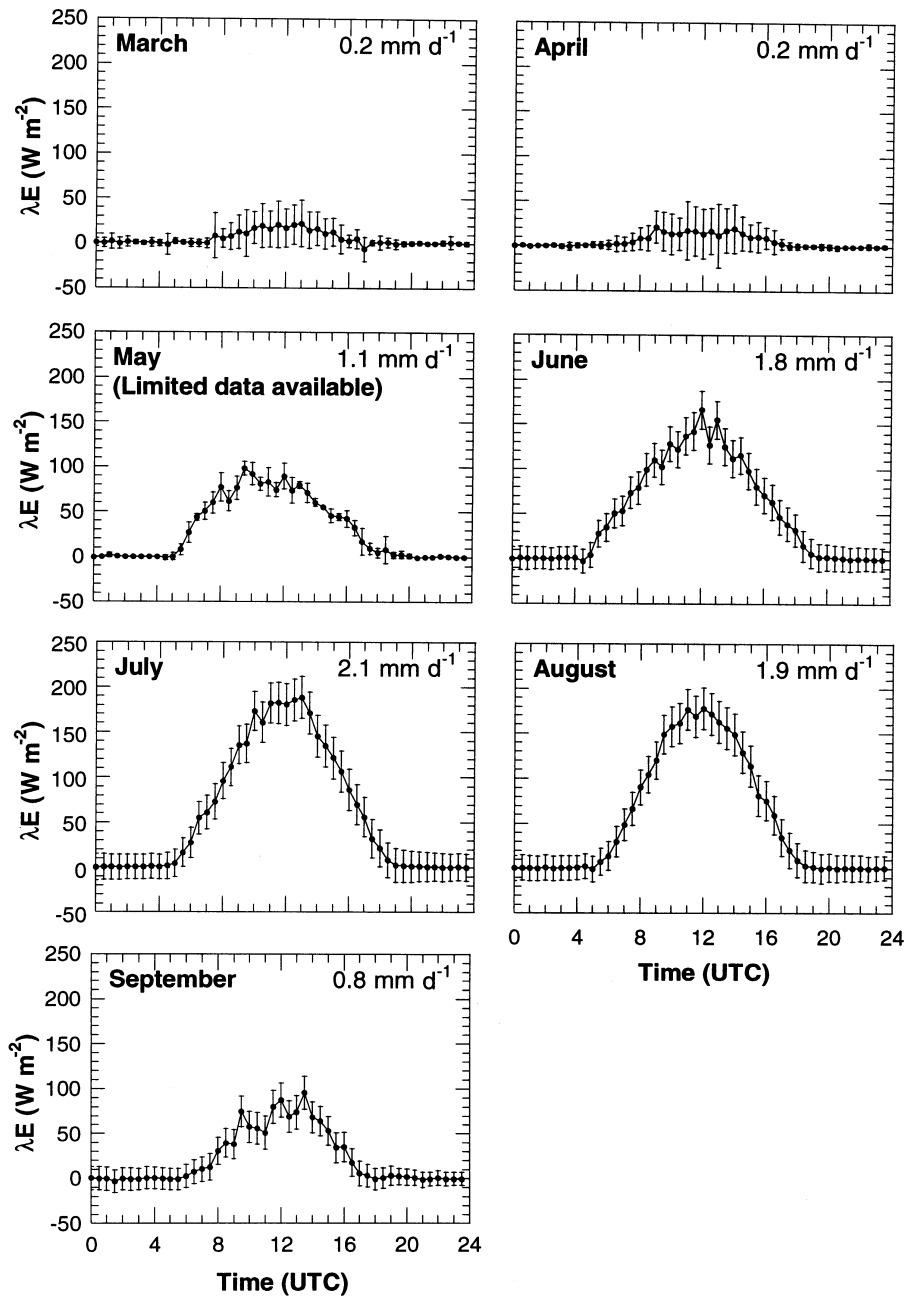
Figure 5. Diurnal patterns of monthly averaged half-hourly latent heat exchange rates from March 1995 to September 1995. Bars indicate the standard deviation of a particular half-hourly mean value. UTC = local time − 2 h

shorter than in August. Despite the higher soil temperature T_s, and higher LAI during July and August, there was no significant increase in night-time CO_2 efflux in comparison with June and September. This might be caused by the relatively large contribution of calm nights (wind speed at 2 m height <4 m s^{-1}) when the exchange of CO_2 respiration is hindered by the near-surface stable stratification of the atmosphere.

The results during autumn and winter are not shown here, but they demonstrated diurnal patterns similar to March as presented in Figure 4. For F_c this meant a net release of CO_2 during night and day, depending

on the soil temperature. Earlier, Nieveen *et al.* (1998) found a soil temperature (2.5 cm) dependence for soil respiration, expressed in a Q_{10} of 4.8 for this area. This is in the range of values found by others for wetland ecosystems (Chapman and Thurlow, 1996).

Figure 5 shows the monthly averaged diurnal course of the latent heat flux density λE. Also indicated is the value of the monthly averaged daily evaporation rate (0.2 mm day^{-1}). The monthly averaged λE in March (no growth yet) showed a maximum of about 20 W m^{-2} during daytime hours and a mean daily evaporation of 0.2 mm day^{-1}. This flux density consisted of soil and open water evaporation. With a mean daily evaporation of 0.2 mm day^{-1}, April showed a pattern similar to March. In the absence of green vegetation, soil and open water evaporation were presumably again the main components of λE. In May (early growth stage), the evaporation demonstrated a clear diurnal course, where λE started to increase from 0 W m^{-2} at 6:30 to a maximum of 95 W m^{-2} at midday. Around 18:00, λE dropped to 0 W m^{-2} and remained constant during night-time. Soil and open water evaporation could still be the major contributors to the total λE. Transpiration was probably only a minor component, due to the small LAI.

During June, July, and August (early growth, peak growth, and early senescence stages), λE peaked around noon at maximum rates between 170 and 190 W m^{-2} in synchrony with the incoming global radiation. Monthly averaged daily evaporation rates, as shown in Figure 5, ranged between 1.8 and 2.1 mm day^{-1}; this is lower than found in other wetland areas that are mostly dominated by *Sphagnum* species (Koerselman and Beltman, 1988; Lafleur, 1990; Price, 1991; Kim and Verma, 1996). Observed evaporation rates compared well with measurements from an *Empodisma minus*-dominated raised peat bog in northern New Zealand that contained an extremely dense canopy of dead vegetation interwoven with living stems (Campbell and Williamson, 1997). Campbell and Williamson (1997) reported a mean evaporation rate of 1.54 mm day^{-1} for a dry canopy with a maximum daily mean of 2.13 mm day^{-1}. As the water table lowered, the insulating layer became thicker and the puddles present became more sheltered by vegetation. Transpiration was probably the major contributor to the total evaporation.

From September (senescence stage) onwards, the monthly averaged λE showed a rapid decline. The mean daily evaporation rate dropped about 60% to 0.78 mm day^{-1} in comparison with the previous months. In the following months, λE continued to diminish to values similar to, or smaller than, those measured in March. Again, soil evaporation was probably the only contributor to the total evaporation.

The diurnal course of the CO_2 flux density in July showed a clear mid-morning depression and continuation of this trend towards the evening. This suggests a limitation of photosynthetic activity, due to stomatal closure, high leaf tissue temperatures, or increased respiratory losses induced by high leaf tissue temperatures (Tenhunen *et al.*, 1984; Schulze 1986). Often, midday stomatal closure is associated with soil water stress or a high atmospheric vapour pressure deficit D (Lösch and Tenhunen, 1981; Jacobs, 1994; Valentini *et al.*, 1996).

Soil water stress as an explanation for stomatal closure is not considered here. Verry (1988) demonstrated that a drop in the water table of about 0.2 to 0.3 m had little effect on the total evaporation from an open *Sphagnum* bog with vascular plant cover. A similar result was found earlier by Romanov (1968), who explained an abrupt decline in evaporation by the fact that the water table had dropped below 0.4 m in a raised peat bog with dwarf shrubs. During the summer months, the contribution of vascular plants with an extended root system was relatively high in the Fochteloöer area. The water table dropped to only 0.2 m at most below the actual soil surface underneath the tussocks. This small drop is unlikely to have caused a decrease in evapotranspiration and an associated increase in stomatal closure. This is in marked contrast with reports from *Sphagnum*-dominated peat bogs, where a slight drop in water table resulted in a substantial decrease in the total evaporation (Ingram, 1983; Kim and Verma, 1996).

What remains as a possible explanation for stomatal closure is the vapour pressure deficit D. In spite of high D, which stimulates transpiration, the stomatal closure results in decreased transpiration (Lösch and Tenhunen, 1981; Tenhunen *et al.*, 1987; Grantz, 1990). In our case, however, a decrease in the total evapotranspiration, synchronized with F_c, was not observed. To investigate this further, we studied the behaviour of F_c and λE on three more or less contrasting days.

To demonstrate the relatively small contribution of soil evaporation to the total evapotranspiration, day 123 (3 May 1995) is analysed. On this day, air temperatures at 2 m rose to a maximum of 23 °C, with a maximum
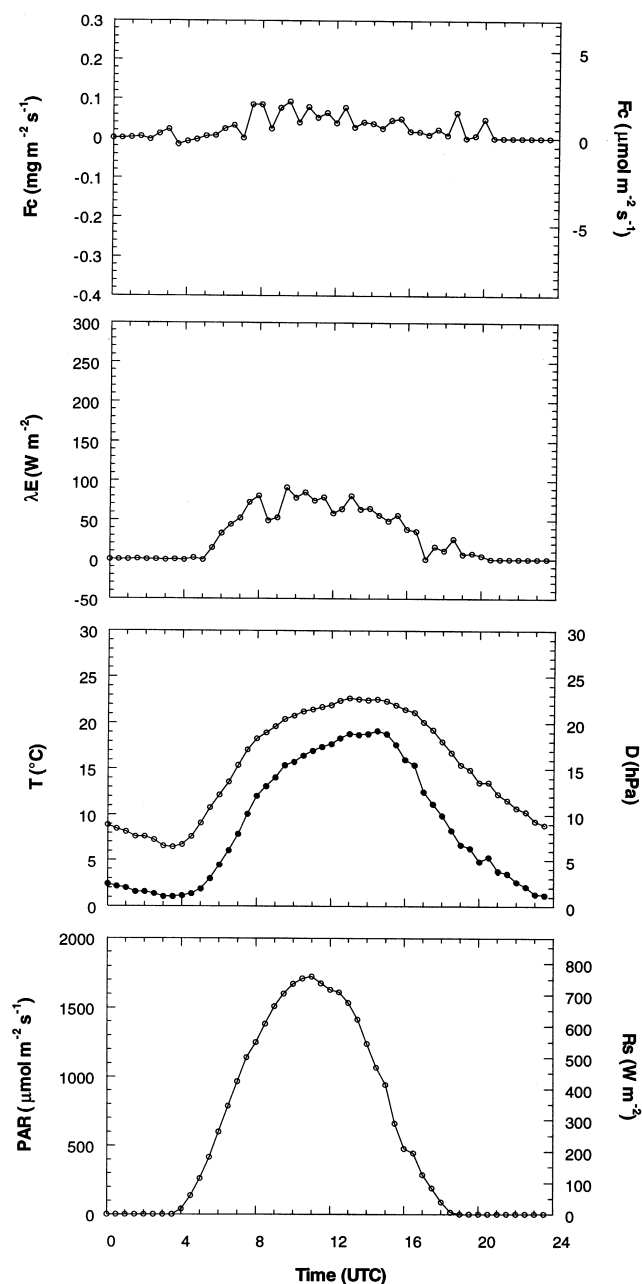
Figure 6. Diurnal patterns of canopy CO_2 and latent heat exchange rates, 2 m air temperature, vapour pressure deficit (solid circles) and PAR for day 123. UTC = local time $-$ 2 h

vapour pressure deficit of 19 hPa. Green shoots started to show on top of the tussock. Throughout the day and night there was a net release of CO_2, as shown in Figure 6 (daily net exchange: 2.2 g CO_2 m^{-2} day^{-1}). Nieveen et al. (1998) found a soil respiration R_d dependence on soil temperature of $R_d = 0.012 \exp(0.16T_s)$ for this area. This would lead to an approximate respiratory flux density of 0.08 mg CO_2 m^{-2} s^{-1} at a midday soil temperature at depth 0.025 m of 12 °C, which is similar to the values measured. This confirms our earlier assessment that the photosynthetic activity at this stage of the growing season contributes little to the CO_2 flux density. At the same time, latent heat loss showed a midday maximum of 90 W m^{-2} and an evaporation

rate of 1.1 mm day^{-1}. With little green vegetation, this water originated solely from the soil, which, in this part of the year, was waterlogged and showed numerous shallow puddles. As the growing season progressed, the water table lowered and the puddles disappeared, or were sheltered by the new shoots, possibly resulting in a smaller contribution of soil evaporation to the total latent heat flux density later in the season.

A high vapour pressure deficit D has been designated as a possible explanation for decreasing photosynthetic activity during the day. Day 223 (11 August 1995), as depicted in Figure 7, is an example of a day with high temperature T and high vapour pressure deficit D: maximum T reached 30 °C and maximum D reached 26 hPa in the afternoon. During the early morning, when the photosynthetically active radiation (PAR) was increasing, F_c reached a maximum of -0.17 mg CO$_2$ m^{-2} s^{-1} (-3.8 µmol CO$_2$ m^{-2} s^{-1}). However, after 9:00, F_c started to decrease while the PAR was still increasing to a maximum of about 1760 µmol m^{-2} s^{-1} at noon (net exchange: $+0.98$ g CO$_2$ m^{-2} day^{-1}). This mid-morning peak and the decline in uptake was similar to observations made in other *Sphagnum*-dominated wetland ecosystems (Neumann *et al.*, 1994; Shurpali *et al.*, 1995). During the same period, λE did not noticeably reduce and more or less followed the PAR trend, reaching a maximum of 292 W m^{-2} at 13:00 (evaporation rate: 3.1 mm day^{-1}). From the result on day 123, we conclude that transpiration is the major contributor to this daily rate.

As suggested, a possible explanation for the decrease in F_c is that high D caused the stomatal resistance of the vegetation to increase, as found by Jacobs *et al.* (1996) was in a grape vineyard in Spain. If this the case, a stronger reaction of λE to stomatal closure would possibly have been visible. In addition to a high D, however, a high T could have negatively affected the photosynthetic activity and/or have increased the soil and plant respiration, as can be seen from Figure 8.

In order to gain insight into the seasonal variation in the net carbon flux density, the net monthly fluxes have been plotted in Figure 9. The hatched bars in Figure 9 represent months with nearly complete (>80%) data coverage, and the blank bars represent months with limited (between 30 and 80%) data coverage. Months with too limited or no available data (<30%) have been left out. From these results we can infer that during three months (June, July, and August) the peat bog behaves as a sink for CO$_2$. During this period, the averaged CO$_2$ uptake by the vegetation is larger than the averaged respiratory losses of CO$_2$ by the plants and the soil. The difference in net release of CO$_2$ during both years probably reflects the effect of the inter-annual variability of the environmental conditions on the carbon balance. More details about the seasonal variations of CO$_2$ exchange can be found elsewhere (Nieveen *et al.*, 1998).

6. CONCLUSIONS

We measured the flux densities of CO$_2$ and water vapour from a *M.caerulea*-dominated raised bog over 18 months in 1994 and 1995. The growing season lasted from early May until the end of October, during which a sparse *M.caerulea* vegetation developed with a maximum LAI of about 1.7. Year round, the tussock grassland was covered with a dense layer of dead organic matter. This dense layer, along with the sparse vegetation, appeared to have a large influence on the exchange of CO$_2$ and water vapour from the area.

During the months of October until June, the LAI is zero, or it is too small to show a notably diurnal course in net CO$_2$ uptake. During this time of the year, a small production of daily CO$_2$ due to soil respiration can be observed. Nevertheless, the evaporation does show a diurnal pattern, which is mainly caused by soil and open water evaporation. However, the soil evaporation was minimized due to the dense layer of dead organic matter covering the surface.

From June until October, there is a clear pattern in the mean net daily CO$_2$ flux. During daytime there is an uptake pattern, whereas during night-time there is small release of CO$_2$ mainly due to soil respiration. During this time of the year the evapotranspiration is mainly caused by the transpiration of the vegetation. The evapotranspiration more-or-less follows the incoming short-wave radiation pattern. The daily patterns of the evapotranspiration and CO$_2$ uptake are not similar.

Diurnal courses often show a single early morning peak in CO$_2$ uptake; but, as the morning progresses, rates of CO$_2$ uptake continue to diminish with time. Simultaneously, water vapour flux densities seem unaffected and continue to maximum values around noon, in synchrony with incoming global radiation. The results of
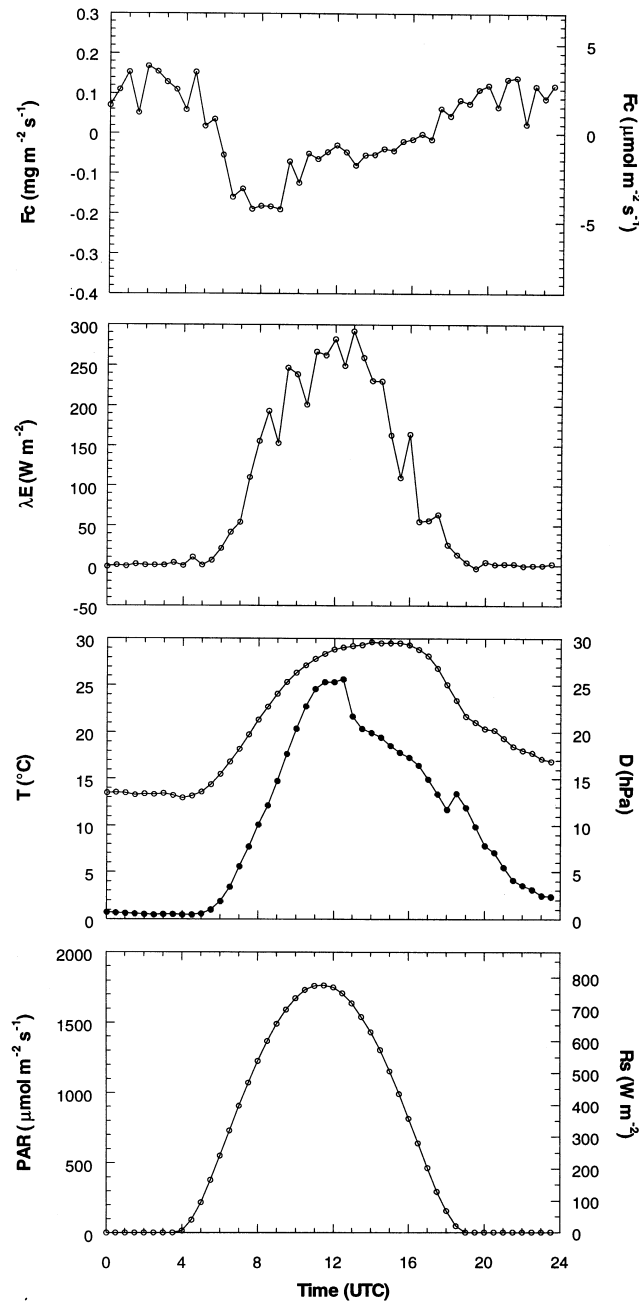
Figure 7. Diurnal patterns of canopy CO_2 and latent heat exchange rates, 2 m air temperature, vapour pressure deficit (solid circles) and PAR for day 223. UTC = local time − 2 h

this study indicate that the observed decrease in the CO_2 flux density is not likely to be caused by stomatal closure, which is often associated with soil water stress or a high atmospheric vapour pressure deficit. Soil water stress is excluded as a cause, as the dominating vegetation is rooted in a peat soil permanently close to saturation. In addition, for both high (>25 hPa) and moderate (<15 hPa) vapour pressure deficits, it is shown that the CO_2 uptake decreases during the day. High leaf tissue temperatures, or increased respiratory losses induced by high leaf tissue temperatures, most likely cause the limitation of the net CO_2.
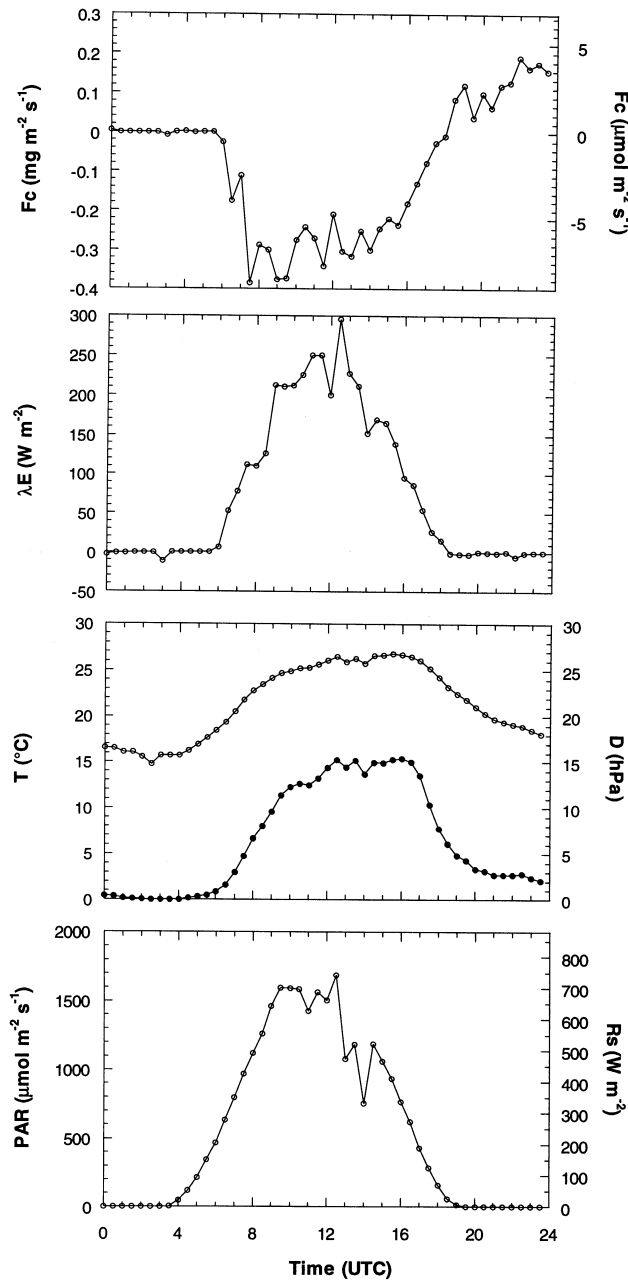
Figure 8. Diurnal patterns of canopy CO_2 and latent heat exchange rates, 2 m air temperature, vapour pressure deficit (solid circles) and PAR for day 210. UTC = local time − 2 h
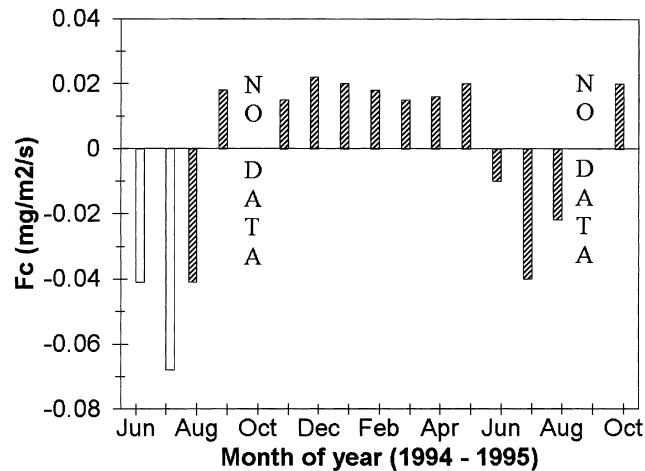
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Figure 9. Monthly variation in the mean net CO_2 flux density. Hatched bars mean nearly complete ($>80\%$) data coverage and blank bars mean limited data coverage (between 30 and 80%)

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