Wetlands (2016) 36:1055–1067
DOI 10.1007/s13157-016-0821-0

ORIGINAL RESEARCH

Wetland Evapotranspiration: Eddy Covariance Measurement in the Biebrza Valley, Poland

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Received: 26 January 2016 / Accepted: 7 September 2016 / Published online: 27 September 2016

Abstract This paper presents eddy covariance measurements of evapotranspiration and sensible heat flux in wetland area. The rate of evapotranspiration is characterized by a distinct seasonal pattern and assumes the highest values during summer. From April to September, the latent heat flux is characterized by a distinct diurnal pattern with the highest average values (at the noon hours) at a level of ca 200 W·m^{-2}. Moreover, in the period from June to August the values of latent heat flux far outweigh those of sensible heat flux (the Bowen ratio in the noon hours ranging between 0.4 and 0.6) and account for 50–60 % of radiation balance. The average daily totals of evapotranspiration vary from about 0.5 mm·d^{-1} during the winter months to about 3–3.5 mm·d^{-1} in summer.

Keywords Latent heat fluxes · Surface energy budget · Water balance · Wetland

Introduction

Wetlands, occupying an area of ca 5.7 × 10^6 km^2, represent a considerable percentage of land area, especially in the zone of middle latitudes (Aselman and Crutzen 1989). Due to their characteristic hydrological conditions, these areas are also an important element of the water cycle. The high levels of groundwater and dense vegetation cause wetlands to have a surface heat balance different from e.g. agricultural or forest areas, and thus they can significantly influence the weather and climatic conditions. In addition, a large biodiversity of species, numerous breeding grounds of waterfowl and marsh birds as well as high biomass production make these areas an important component of the system of nature protection. Wetlands are also an element of the global carbon cycle. Moreover, the rate of exchange of carbon dioxide and of methane between the atmosphere and wetlands is strictly dependent on soil moisture conditions (Moore et al. 1998). They determine the course of aerobic and anaerobic processes of biomass decomposition (Fraser et al. 2001; Billett et al. 2004). Therefore, due to their specificity and diversity of environmental conditions, these areas are often subject to different forms of nature conservation. Therefore, the measures to preserve the original features of such areas require a deeper knowledge of the principles of wetland environment functioning, with particular emphasis put on their water balance.

The process of evapotranspiration mainly depends on there being sufficient energy resources which can be consumed to evaporate a certain amount of water. This quantity depends primarily on the net all-wave radiation Q* and other elements of surface heat balance, approximated by:

$$Q^{*} \approx Qh + Qe + Qg \tag{1}$$

where Qh is sensible heat flux, Qe is latent heat flux and Qg is ground heat flux.

However, the high soil moisture, dense vegetation and thus a large surface area of foliage, as well as the roughness of surface result in the fact that wetlands are usually characterized by higher evaporation in relation to evaporation from an

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open water surface (e.g. Kim and Verma 1996; Herbst and Kappen 1999; Pauliukonis and Schneider 2001; Acreman et al. 2003). Evaporation in such an environment usually takes place by several parallel paths, including transpiration from vegetation, evaporation from surface water under the vegetation and evaporation from an open water surface (numerous lakes, rivers, canals characteristic of the landscape of wetlands). At the same time, the complexity of the process of evapotranspiration in wetlands can be evidenced by the fact that in many cases, however, higher evaporation from an open water surface was observed (Campbell and Williamson 1997; Burba et al. 1999) in relation to evapotranspiration from wetlands. In addition, a number of studies point to evapotranspiration as the main component of the water balance of marshland areas, and (the energy equivalent of evapotranspiration) the latent heat flux dominates in the surface heat balance (Priban and Ondok 1985; Drexler et al. 2004). Therefore, a detailed knowledge of evapotranspiration of wetlands can be crucial for the functioning of these ecosystems or for restoring the land transformed as a result of land reclamation or other human activities to its original condition. In the second half of the twentieth century, wetland areas in many countries were used for economic purposes and subjected to land amelioration. However, including such an area in the programme of nature conservation (e.g. the Birds Directive – 79/409/EEC, the Water Framework Directive – 2000/60/EC) involves restricting agricultural activities (such as grazing and mowing peat meadows), which results in the observed succession of forest and bushy vegetation onto the peat meadow areas (Grygoruk et al. 2014; Wassen et al. 2006). Therefore, the results of evapotranspiration evaluation are essential to assess the water balance of these areas in order to restore them to their original condition.

In the studies of wetland evapotranspiration, a number of methods to assess this phenomenon are applied. For example, in the direct measurement of evapotranspiration lysimeters (and more specifically, a set of several lysimeters) are used, enabling simultaneous evaluation of evaporation loss of water from individual bog formations and from an open water surface (Abtew and Obeysekera 1995; Abtew and Melesse 2013). In the direct measurements, microclimatic methods based on the Bowen ratio are also used (Burba et al. 1999; Campbell and Williamson 1997; Kellner 2001; Peacock and Hess 2004), as well as methods based on the measurements of fluctuations of groundwater levels (Gerla 1992; Lafleur et al. 2005; Grygoruk et al. 2011; Grygoruk et al. 2014; Mazur et al. 2014). Furthermore, the results of evaporation evaluation, obtained using the Penman equation, the Penman-Monteith equation or the water balance are also widely presented (Wessel and Rouse 1994; Campbell and Williamson 1997; Souch et al. 1998; Jacobs et al. 2002; Gasca-Tucker et al. 2007). In the last two decades, the eddy covariance method has been more and more commonly used. The method enables the direct measurement of the vertical flux of water vapour. An important advantage of the eddy covariance method is the fact that the obtained water vapour flux comes from a source area within a distance of 200–300 m from the sensor. At the same time, due to the high costs of measuring instruments, and also because of various technical factors affecting the quality of the data obtained, a number of studies devoted to the measurement of evapotranspiration present the results of measurement experiments lasting for several weeks or seasons. They mainly present the daily patterns of water vapour flux or the daily totals of evapotranspiration against the background of meteorological conditions (Souch et al. 1998; Kurbatova et al. 2002; Gasca-Tucker et al. 2007), as well as the results of impact assessment of parameters such as foliage surface area or surface resistance on the rate of evapotranspiration (Acreman et al. 2003; Gulden et al. 2007).

The primary aim of this study is to present the variability of daily and seasonal evaporation from the wetlands located in north-eastern Poland in the Biebrza National Park. Thanks to the simultaneous measurements of the radiation balance, the remaining elements of heat balance, as well as selected meteorological parameters, the results allow to present the following:

- a number of characteristics of the daily pattern of latent heat flux as compared to other components of the heat balance of bog surface,
- the variability of evapotranspiration in individual seasons of the year as well as the variability of the relationship of latent heat fluxes with the radiation balance values and sensible heat fluxes at various stages of development of vegetation.

Owing to the multi-seasonal measurements in the period 2013–15, it was possible to determine the daily and monthly evapotranspiration totals and their annual pattern. In addition, during the measurement period the average weather conditions in individual seasons were characterized by a fairly wide diversity. Therefore, the results of the study also present the long-term variability of evapotranspiration in wetland areas both under the conditions of a relatively humid growing season and of a drought.

The study results are also crucial from the point of view of works aimed at protecting the wetland areas of the Biebrza National Park. The knowledge of evapotranspiration is essential to assess the water balance of these areas in order to restore them to their original condition. In addition, despite the ever wider research into this issue using state-of-the-art measurement methods, there still are only a handful of studies on the complex characteristics of evapotranspiration in the wetlands of Poland (Kleniewska et al. 2009; Chojnicki et al. 2007; Urbaniak 2006; Fortuniak et al. 2013).

Methods

Study Site

The Biebrza National Park (Polish: *Biebrzański Park Narodowy*) (BNP) is the largest national park in Poland (59 thousand ha) and occupies the biggest marshland areas in the valley of the Biebrza river in north-eastern Poland (ca 255 km^2). The Biebrza basin is characterized by a typical valley landscape with high levels of groundwater and strong paludification. Hydrogenic, mostly peat soils dominate, and in the areas subject to amelioration these are primarily peat-and-muck soils. A smaller percentage of the area of the national park consists of mineral soils, such as rusty soils, luvisols, brown soils and spodic soils (Banaszuk and Banaszuk 2004). In the wetlands, rush and sedge communities dominate, as well as wet and variably wet meadows. The transition zones between peat soils and mineral soils are overgrown with forests and thickets (Malinowska 2004).

The measurement station is located in the neighbourhood of the village of Kopytkowo (53°35′20″N, 22°53′31″E), within the so-called Central Basin, covering the area of the largest marshlands and peatlands (Fig. 1) which are named the Red Marsh. In the immediate vicinity of the measurement site (~ 500 m), there occur rush communities with the dominating reed sweet-grass (*Glycerietum maximae*) and the common reed (*Phragmitetum communis*), as well as moss and sedge communities represented by the *Carici canescentis-Agrostietum caninae*. These communities widely occur in the whole Biebrza valley (Matuszkiewicz 2004). During the measurements, in the immediate vicinity of the weather station, the average height of vegetation in the cold half of the year and in spring was approximately 0.4 m, whereas in summer and autumn the average height was approximately 1 m.

Eddy Covariance Measurement

The measurements of the turbulent flux of water vapour were performed in the period between June 2013 and September 30, 2015. In the eddy covariance method applied, the turbulent flux of water vapour E is calculated as follows:

$$E = \rho \cdot \overline{w' \cdot q'} \qquad (2)$$

where: ρ – air density, w' – fluctuations in vertical wind velocity, q' – specific density fluctuations, the flux from the Earth's surface to the atmosphere being assumed as positive. The overbar and primes indicate the mean for the sampling interval (30-min frequency) and deviations from the mean, respectively. Then standard data-verification steps were used, including: elimination of artificial electrical impulses, rotations of coordinate systems, temperature and humidity corrections, the WPL correction (Webb et al. 1980), as well as corrections for spectral losses (Fortuniak 2010). The next step taken in the preparation of the data was to verify the requirement of series stationarity (Foken 2008; Fortuniak 2010). This issue was solved using the following three tests: the test proposed by Foken (Foken 2008), by Mahrt (Mahrt 1998), and by Dutaur as modified by Affre (Dutaur et al. 1999; Affre et al. 2000). In addition, the data were verified for precipitation and conditions of the presence of frost on instruments and the data from such situations were also omitted in further analysis. The criteria so adopted for the verification of data quality, as well as errors resulting from interference of measuring instruments cause gaps in the measurement series. With the aforementioned criteria adopted and measurements made in rainy or frosty weather eliminated, the percentage of good data was approximately 80 %.

To prevent underestimation of the actual evapotranspiration, data gap filled before determining the daily, monthly or annual totals. Gaps of several hours were filled using the average daily pattern of We values calculated over a period of 14 days (Falge et al. 2001; Foken 2008). On the other hand, in assessing the daily pattern of evapotranspiration, series with unfilled measurement gaps were used. Such approach allows to determine the basic statistics of variability only on the basis of data obtained from the eddy covariance measurements.

The eddy covariance method requires measurements made with a very short time step (10-20 Hz). The set of measuring instruments includes: a sonic anemometer (RM Young 81,000, USA) enabling the measurement of three components of wind velocity, and a gas analyzer (LI 7500, LI-COR, Inc., Lincoln, USA) which allows the identification of fluctuations of water vapour and carbon dioxide content. The instruments were installed on a mast at a height of 3.7 m above ground level. Data were recorded at a frequency of 10 Hz. The components of radiation balance were measured using a net radiometer CNR1 (Kipp&Zonen, the Netherlands) consisting of two pyrgeometers (CG3) for measuring the atmospheric infrared radiation, and two pyranometers (CM3), allowing the simultaneous measurements of sunlight received and reflected by the Earth's surface.

The measurements of temperature and humidity were performed at a height of 2.2 m and 0.5 m using HMP 60 sensors (Vaisala, Finland). In addition, the measurement station was equipped with instruments for the measurements of heat flux in the soil (HFP01SC, Campbell Scientific, Inc. USA), soil temperature (T107, Campbell Scientific, Inc. USA), soil moisture and the amount of precipitation.

Weather Conditions

The Biebrza basin, located in north-eastern Poland, is characterized by different meteorological conditions as compared to central and western Poland. In addition, due to a great diversity of its soil cover and plant communities, the area also

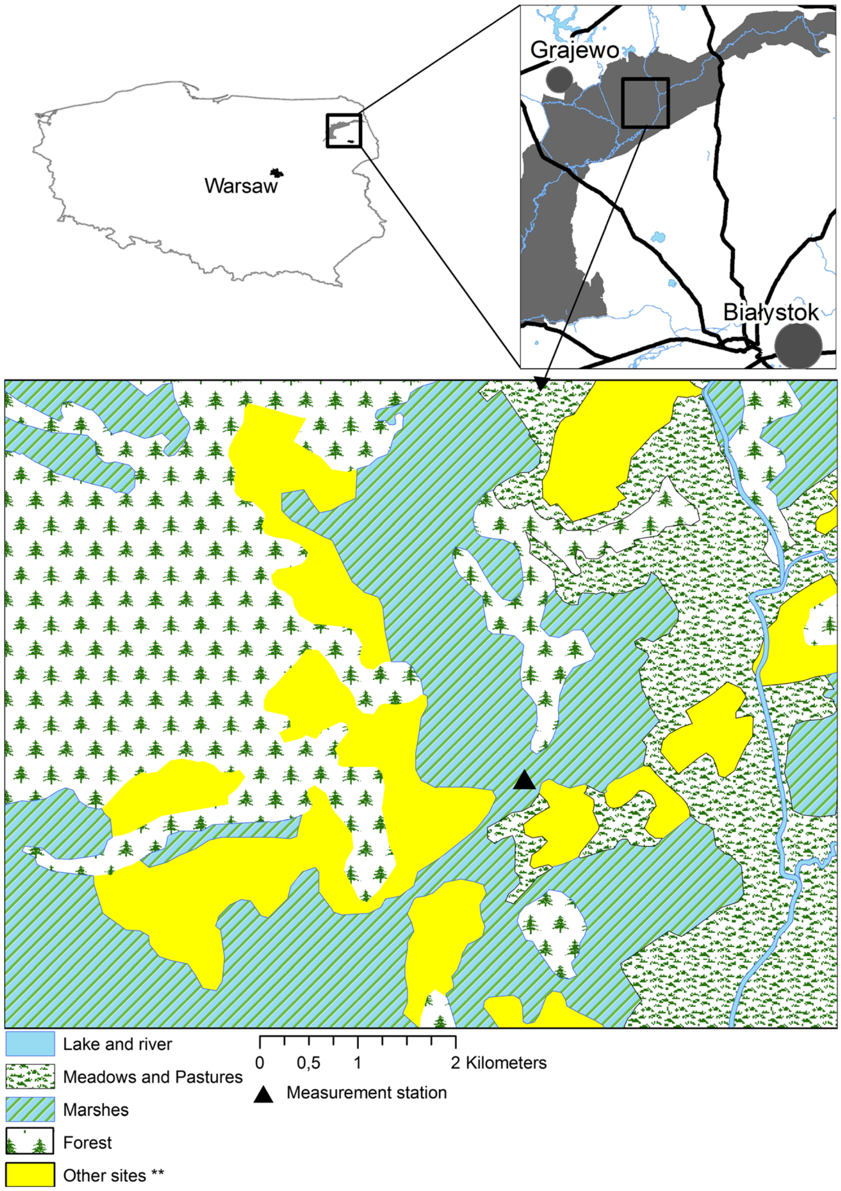


Fig. 1 Location map showing the Biebrza National Park with spatial distribution of different vegetation types near the measurement station (based on CORINE Land Cover data available from European Environment Agency (EEA 2006)). ** Other sites – mixed group included agricultural fields and areas covered by farm buildings

presents a rich mosaic of microclimates (Kossakowska-Cezak et al. 1991), which requires thorough research, especially under conditions of progressing changes in soil moisture associated with the activities of nature conservation and restoration of the Biebrza marshes to their original features.

In the period under analysis, the thermal conditions were quite typical. In the winter period (both in 2014 and 2015), generally, the mean daily air temperatures oscillated between −5 °C and 5 °C. There were also periods of several days of very frosty air advection when the mean daily air temperature dropped below −20 °C (Fig. 2a). In spring, there was a gradual increase in air temperature and, for example, days with an average daily temperature above 20 °C were recorded at the measurement station in May 2014 (Fig. 2a). During the studied three summer seasons, the months of June, July and the first half of August were the period of the year with the highest air temperatures. There were periods of several days when the average daily air temperature was about 20 °C, while the maximum temperature exceeded 30 °C. From the second half of August, notably in 2013 and 2014, a gradual decrease in air temperatures began. In September, the average daily air temperature fell below 10 °C; moreover, there already occurred the first cases of ground frost.

The humidity conditions significantly affect the rate of evaporation. In the annual pattern of vapour pressure deficit (Fig. 2b), a marked seasonality is visible with the highest values occurring in summer. During winter, in turn, the vapour pressure deficit is low (less than 5 hPa), which indicates

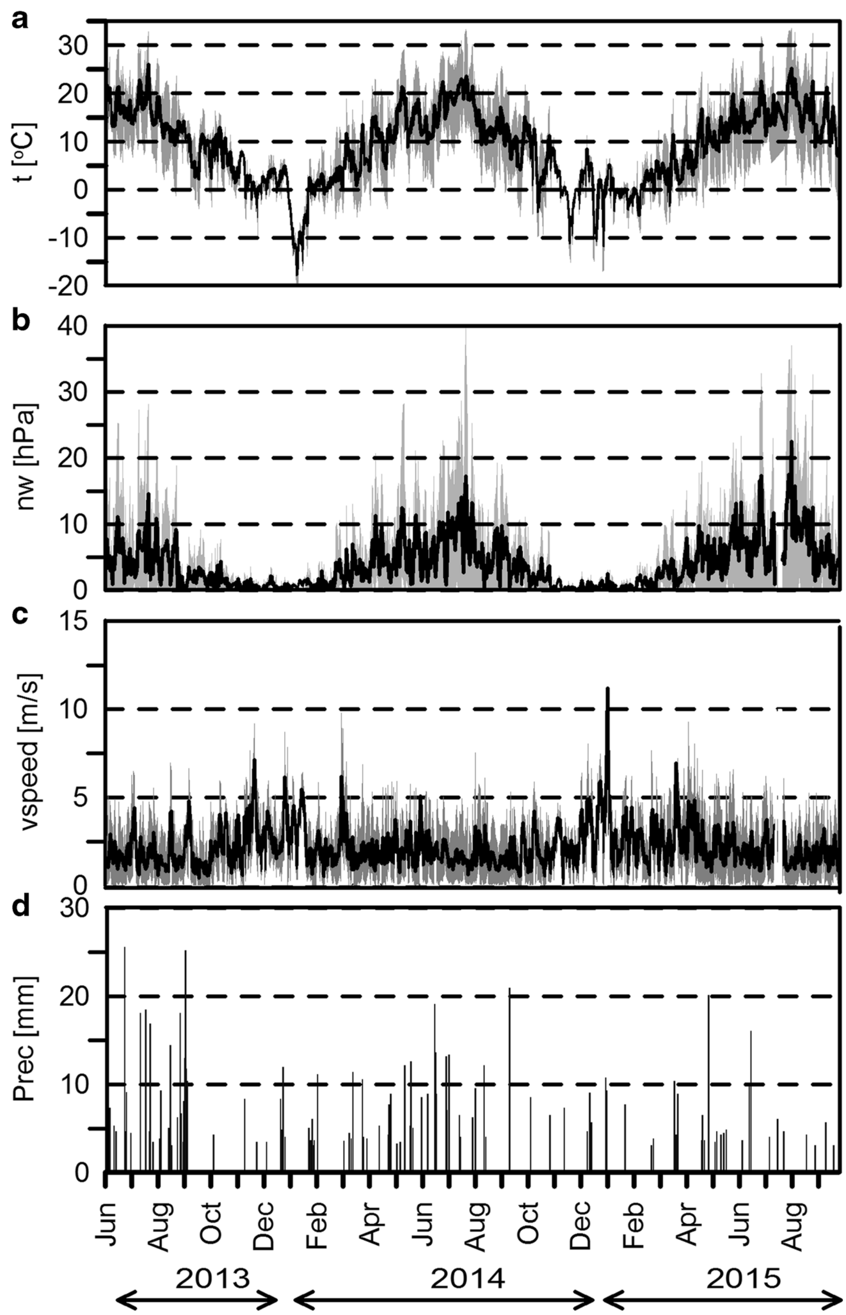


Fig. 2 Daily characteristics of selected weather conditions during study period: **a** air temperature, **b** vapour pressure deficit (VPD), **c** wind speed, **d** daily precipitation total. Solid line represents daily average, shaded area indicates daily minimum and maximum values

conditions similar to those of saturation of the air with water vapour. In the spring and summer periods, the average daily values of vapour pressure deficit grow. In particular, August 2015 was characterized by extremely low humidity, when several days with very high values of vapour pressure deficit occurred.

An important meteorological factor conditioning the rate of evapotranspiration is wind speed. During the measurement period, rather low wind speeds prevailed (Fig. 2c). In the warm season of the year, the daily average wind velocity oscillated around 2 m s^{-1}, and the maximum wind speed rarely exceeded 5 m s^{-1}. Higher wind speeds occurred during the winter months when several day periods with the daily mean wind speed exceeding 5 m s^{-1} were recorded.

Precipitation is a major component of water balance conditioning the level of surface waters, as well as soil moisture. Seasonal precipitation during the summer of 2013 and the spring and summer of 2014 at the study site was similar to the long-term average values (Ustrnul et al. 2013; Ustrnul et al. 2014). High precipitation (110–140 % of the norm from the years 1971–2000) in north-eastern Poland occurred in September 2013 and during the winter and spring seasons of 2015 (Ustrnul et al. 2015a). In contrast, the summer of 2015 was characterized by very low precipitation at a level of 30–

70 % of the average seasonal total of the period 1971–2000 (Ustrnul et al. 2015b). In the pattern of daily precipitation totals at the measurement station, longer periods without precipitation occurred mainly in the autumn season. The spring and summer periods were characterized by a greater number of days with precipitation; in addition, there also was precipitation with the daily total exceeding 10 mm. Against this background, the summer of 2015 stands out because in August and September there were only a few days with rainfall, and its daily totals did not exceed 5 mm (Fig. 2d).

Results and Discussion

Mean Daily Pattern of Surface Energy Balance Components

Between April and September, the net all-wave radiation at the measurement station is characterized by a clear daily pattern with the highest values at around midday. The average value of Q* in the afternoon from May to August reach a value of 400 $W{\cdot}m^{-2}$, the Q* at the same time being characterized by a high variability due to more dynamic weather conditions in summer (Fig. 3a). In the night hours, in turn, the surface of the marsh loses heat through radiation. After sunset, the average radiation balance values in the summer months assume negative values at a level of −10 to −40 $W{\cdot}m^{-2}$.

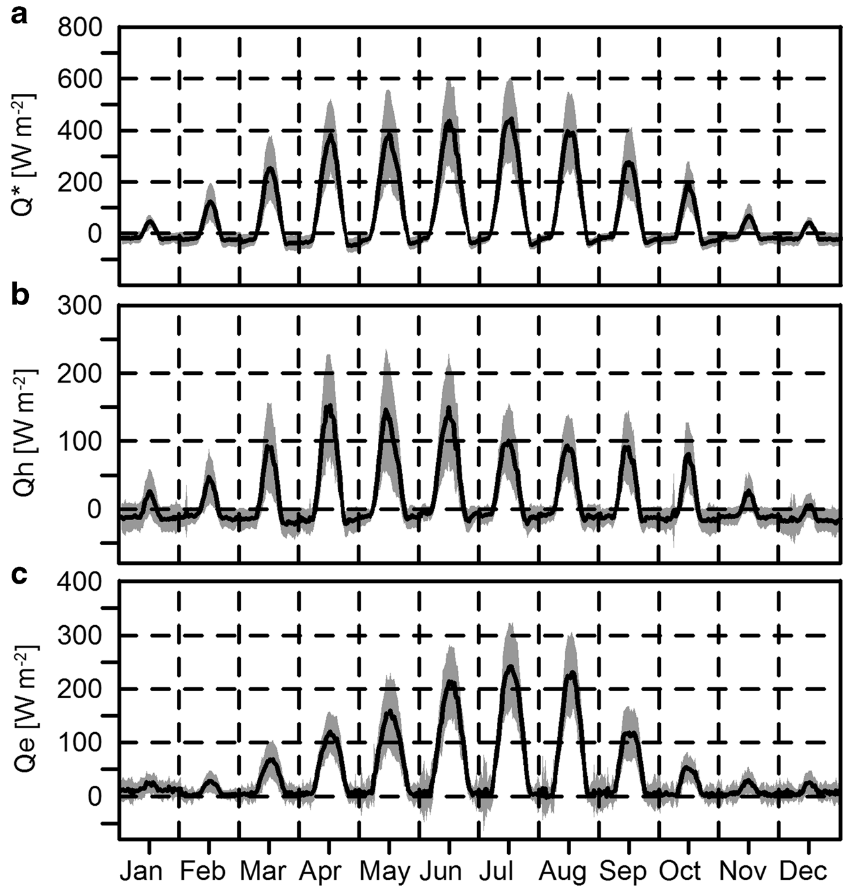
In the warm half of the year, an increase in solar radiation and higher air temperatures lead to the development of turbulent sensible heat fluxes Qh and latent heat fluxes Qe. From April to September, with an increase in Q*, the pattern of the Qh and Qe values show a distinct diurnal cycle (Fig. 3 b, c); at the same time, in the spring season, i.e. in the months of March, April and May an increase in the values of sensible heat flux is clearly visible. Around noon, the average values of Qh vary from about 100 $W{\cdot}m^{-2}$ in March to about 150 $W{\cdot}m^{-2}$ in May (Fig. 3b) and account for 30–40 % of the net all-wave radiation (Fig. 4a). The average ratio of sensible heat flux and radiation balance (Qh/Q*) in the noon hours assume values of 0.29 to 0.36 (Table 1). The latent heat flux Qe is lower than Qh and in the noon hours the average values of Qe vary from ca 50 $W{\cdot}m^{-2}$ in March to 100 $W{\cdot}m^{-2}$ in April (Fig. 3c) which is approximately 30 % of the Q* (Fig. 5a). A sensible heat flux exceeding the values of Qe in spring was also detected in the multi-season measurements in the wetlands in north-eastern China (Zhou and Zhou 2009). The recorded mean values of sensible heat flux in the noon hours reach a level of 300 $W{\cdot}m^{-2}$, while the values of Qe range between 80 and 100 $W{\cdot}m^{-2}$. A similar trend is also characteristic of evapotranspiration from the surface of marshland areas in the Central United States (Lenters et al. 2011) where, for example, in April the average Bowen ratio is 3.3.

In May, the sensible and latent heat fluxes achieve similar values. In the noon hours, the average values of Qh and Qe reach a level of 150 $W{\cdot}m^{-2}$. In July and August, the high value of solar radiation in conjunction with fully developed vegetation and high precipitation lead to high values of the latent heat flux. The highest average values of Qe in the noon hours reach a level of 250 $W{\cdot}m^{-2}$ (Fig. 3c), while the values of the sensible heat flux at the same time amount to 100 $W{\cdot}m^{-2}$ (Fig. 3b). Despite the different climatic conditions (warm temperate climate conditions with the monsoon variant), quite similar values of turbulent latent heat flux are characteristic of evapotranspiration from the wetland area overgrown mainly with reeds in north-eastern China. For example, in the period of strong growth of vegetation (May and June), as well as in summer, the average values of Qe in the noon hours vary between 180 and 200 $W{\cdot}m^{-2}$ (Zhou and Zhou 2009). The average values of Qe at a level of about 200 $W{\cdot}m^{-2}$ are also characteristic of the marshland areas of central Sweden (Kellner 2001); the average value of Qe for the noon hours at a level of 300 $W{\cdot}m^{-2}$ was found, in turn, in the case of evapotranspiration from the surface of wetlands in the subtropical zone (Jacobs et al. 2002; Wu and Shukla 2014) or coastal marshes (Souch et al. 1998).

In summer, the values of Qe obtained in the daytime account for an average of 50–60 % of the radiation balance (Fig. 5b), while the ratio of Qh/Q* has a value of 0.2 to 0.3 (Fig. 4b). In the diurnal pattern of the Qe/Q*, a clear u-shaped pattern is characteristic for the daylight hours (Fig. 5b). After sunrise in the summer months, the ratio Qe/Q* assumes values close to 1 (solar radiation energy received is consumed to evaporate dew), and next these values fall to about 0.5 in the noon hours to reach values equal to or greater than 1 in the evening, which means that the energy accumulated in the soil and the downward sensible heat flux can be used for evaporation (Fig. 5b). At night, due to the fact that the measured values of the turbulent latent heat flux are low, the result of the ratio Qe/Q* assumes values rather randomly (without a distinct diurnal pattern), ranging between −1 and 1.

September is another month in which the average pattern of Qh and Qe achieve similar values (around 100 $W{\cdot}m^{-2}$). In October, like in early spring, Qh exceeds Qe. In this season, the values of Qe obtained in the daytime account for an average of 35–50 % of the radiation balance (Fig. 5c), while the ratio of Qh/Q* has a value of 0.2 to 0.4 (Fig.4c). In winter, when the Q* is low, the turbulent sensible and latent heat fluxes thereby assume low values, and moreover, without a clear diurnal cycle. Furthermore, in the daily patterns of the relation of Qh and Qe to Q* it is difficult to identify a clear diurnal cycle (Fig. 4d and Fig. 5d). The low measured values, close to the accuracy of measuring instruments, in combination with the conditions of weak turbulence mean that the results obtained during this period may be subject to a considerable error.

Fig. 3 Mean daily pattern of surface energy balance components (*solid line*) and its mean variance described as ± standard deviation (*shaded area*)



The presented comparison of the average daily patterns of the sensible and latent heat fluxes in the warm half of the year shows that in spring the increase of solar radiation leads in the first place to the development of the sensible heat flux. It is only in summer, thanks to well-developed flora, evaporation heat losses are greater than those consumed for the sensible heat flux. This is also confirmed by the average pattern of the value of the Bowen ratio (Fig. 6). In the three summer seasons of study, the quotient Qh/Qe assumed values smaller than 1. The autumn period, namely the end of September and early

Fig. 4 Mean daily patterns of Qh and Q* ratios

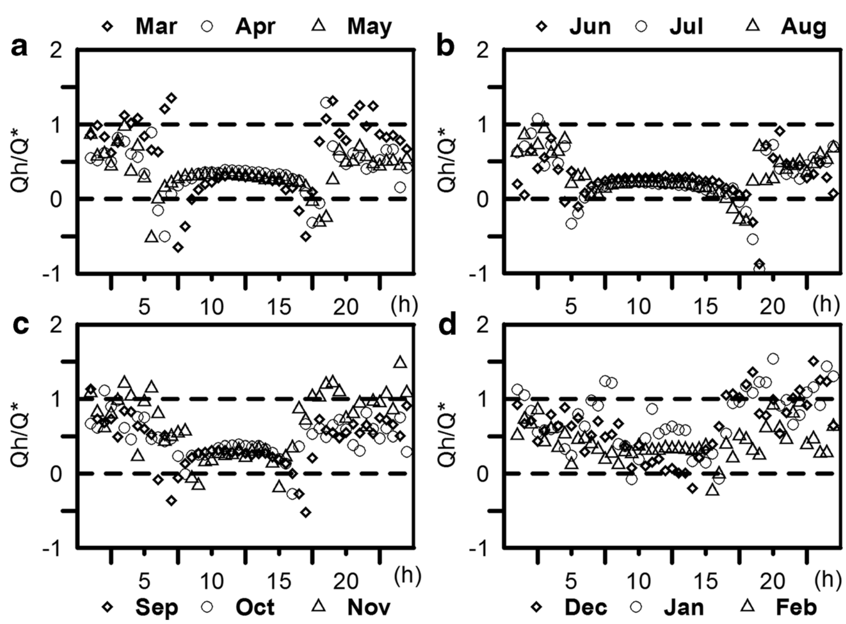


Table 1 Average daytime (from 10.00 a.m. to 2.00 p.m.) flux ratio

Month	Qe/Q*	Qh/Q*	Qh/Qe
Jan	0.69	0.48	1.14
Feb	0.28	0.33	1.59
Mar	0.30	0.29	1.28
Apr	0.35	0.36	1.18
May	0.48	0.29	0.78
Jun	0.53	0.28	0.59
Jul	0.59	0.18	0.35
Aug	0.59	0.19	0.35
Sep	0.51	0.27	0.62
Oct	0.38	0.34	1.47
Nov	0.54	0.21	0.78
Dec	0.58	0.09	0.16

October, showed a clear increase in the value of the Bowen ratio (> 1). Changes in the vegetative cycle lead to reduced transpiration of water vapour through the stomata of plants. At the same time, due to the accumulation of thermal energy in the soil after the summer period, and at relatively high values of solar radiation, the sensible heat flux assumes high values which exceed evaporation heat losses. The high values of the ratio Qh/Qe (> 1) are characteristic of the winter period, the visible high variability of the observed values being likely caused by the fact that the obtained values of Qh and Qe are small, within the accuracy of the measurement method (Fig. 6).

The high variability of the Bowen ratio (Qh/Qe) values in winter is also confirmed by the average daily patterns for the winter months with highest values in the middle of the day and smaller or negative values around sunrise and sunset (Fig. 7a). However, in the spring, summer as well as autumn seasons a distinct average daily cycle is visible, especially for the daylight hours. For the months of March and April, in the morning hours the mean values of the Bowen ratio range between 0 and 1 to assume values greater than 1 in the afternoon (Fig. 7b). In May and in the summer months (Fig. 7c), the mean values of the Bowen ratio in the afternoon are in the range 0.35–0.6 (Table 1). In addition, from May to August a marked asymmetry is observed in the daily profile, where the values of afternoon and evening Qh/Qe quotient are lower than those of the morning. It is probably heat stored in the ground which may be an additional source of energy intensifying evapotranspiration at this time of day. However, due to difficulties in measuring Qg in an environment of lush vegetation and high moisture content, these results are not presented in this paper.

Evapotranspiration

The conversion of the latent heat flux values from Wm^{-2} to $mm \cdot h^{-1}$ allows to determine water losses associated with evaporation and transpiration. In the winter months, the mean hourly evapotranspiration totals are not big and the highest mean hourly evapotranspiration totals slightly exceed 0.05 mm/h, while the maximum values oscillate around 0.2 mm/h. From March to September, evapotranspiration it is characterized by a marked diurnal cycle (Fig. 8). In the spring months, the mean hourly totals of evapotranspiration at noon change from 0.1 $mm \cdot h^{-1}$ to 0.3 $mm \cdot h^{-1}$ in June. Equally high mean hourly totals of evapotranspiration are typical of the summer months. The highest hourly totals of evapotranspiration observed so far (using the adopted criteria for

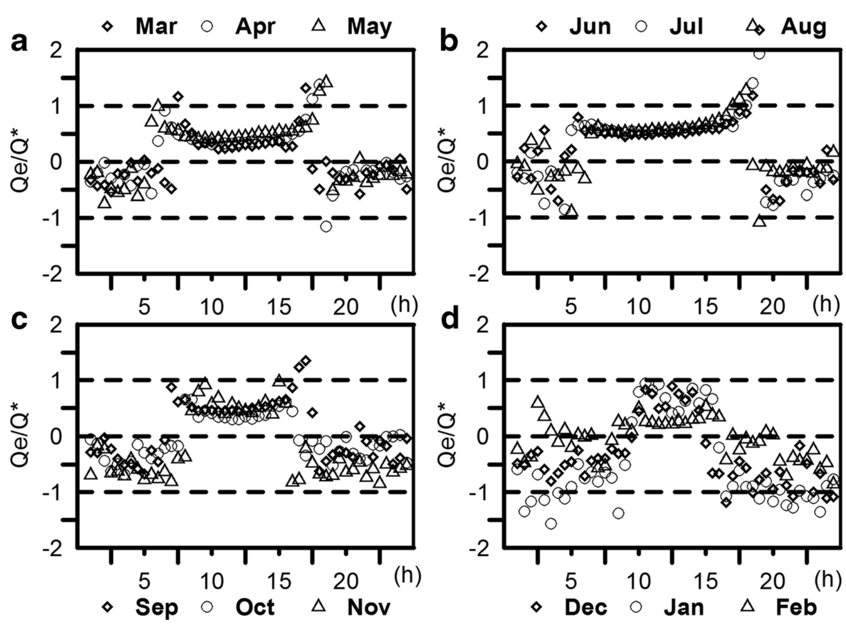


Fig. 5 Mean daily patterns of Qe and Q* ratios

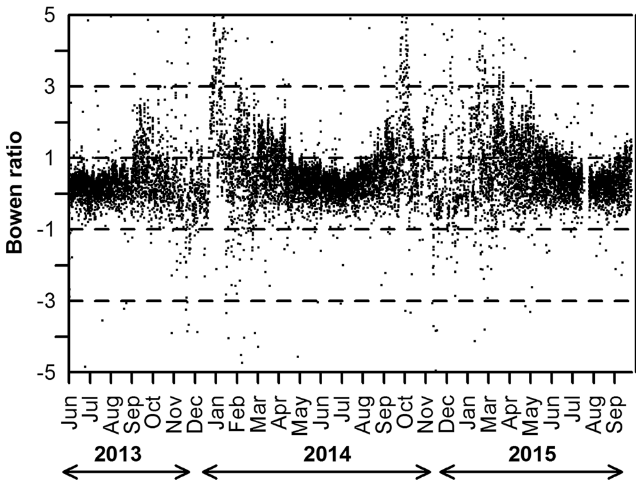


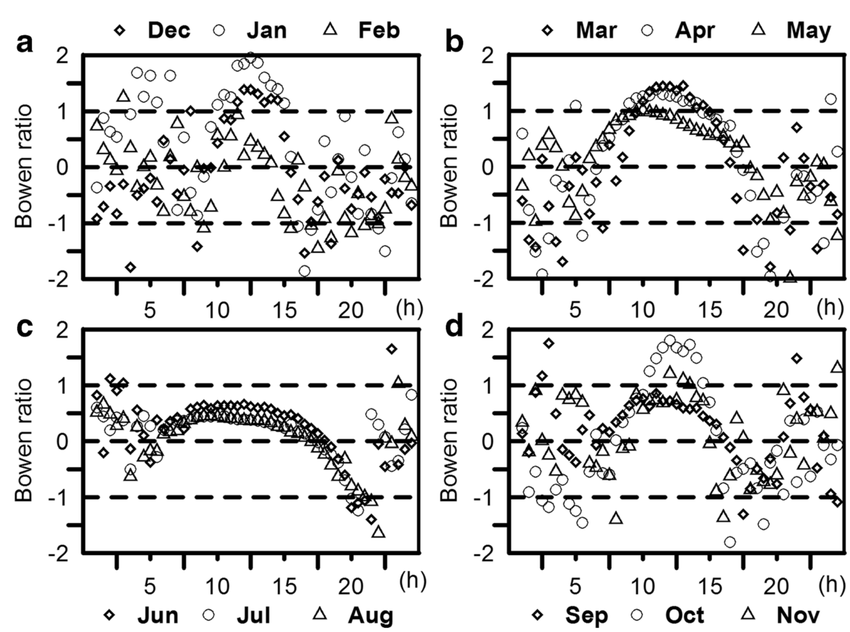
Fig. 6 Pattern of Bowen ratio values in daytime hours (Kd > 5 W·m^{-2})

data verification) hovered around 0.5 mm·h^{-1} (Fig. 8). Similar values of hourly evapotranspiration totals in the noon hours are characteristic of the wetlands constituting the west coast of Lake Huron, Michigan, USA (Mazur et al. 2014). Based on the water table fluctuation method, the presented daily patterns of evapotranspiration recorded during the measurement sessions of several days in the summer season show a clear pattern with maximum values of 0.5 to 0.8 mm·h^{-1}.

The results of measurements of hourly evapotranspiration totals with gaps in the data filled were used to determine the daily and monthly evapotranspiration totals. In the period of study, in the winter months the daily evapotranspiration totals (except in a few cases) ranged between 0.1 and 1 mm (Fig. 9). The median of daily totals varied from 0.20 mm in December to 0.48 mm in January. At the end of March, a significant increase in the daily evaporation totals is seen. The mean daily total both in 2014 and 2015 amounted to 0.8 and 0.9 mm, respectively, evaporation also being characterized by a greater variability due to more dynamic weather conditions. For example, as compared to January or February, in March the inter-quartile range of daily totals of evapotranspiration is determined by the values 0.1 and 0.6 mm.

In the subsequent months, the obtained daily evapotranspiration totals increase, to achieve the highest values in June and July when their average value ranges between 3.0 and 3.5 mm. In addition, the summer season, especially in July, is characterized by the highest daily totals with values of 4–5 mm (Fig. 9). This was particularly evident in the summers of 2013 and 2014. However, in 2015, when summer was relatively dry, the highest daily evapotranspiration totals (only in a few cases) reached a level of 4–4.4 mm. Another characteristic feature of evaporation in summer is the fact that the daily totals may assume values within a very wide range. In June and July, the lowest daily totals were 1.2 and 1.7 mm, while the absolutely highest daily totals reached values of 5.0 and 4.6 mm (Fig. 10). It is worth mentioning that the maximum evapotranspiration was observed during sunny and hot weather (incoming solar radiation reached 800–900 W m^{-2} and the maximum daily temperature exceeded 25 °C), whereas very low daily totals were associated with cloudy weather and advection cold air mass (maximum daily temperature below 15 °C). At the same time, with a large difference between the minimum and maximum values, the inter-quartile range of daily totals changed by about 1 mm. The value of the 1st quartile varied from about 2.5 mm in June and August to 3 mm in July, while the values for the 3rd quartile were, respectively, 3.5 and 4.0 mm. In the autumn season, in the

Fig. 7 Mean diurnal patterns of Bowen ratio



period under analysis the daily totals steadily decreased. The average daily totals for September, October and November amounted to 1.6 mm, 0.8 mm and 0.4 mm, respectively.

The results of the daily evapotranspiration totals in summer, obtained using the eddy covariance methods, are quite similar to the results obtained for marshland areas in Europe, Siberia and Canada. Daily totals at a level of 2–3.5 mm were obtained during a few weeks of summer measurements in the wetlands overgrown by reeds (Reed Belt) (Peacock and Hess 2004), or also in wet meadows (Gasca-Tucker et al. 2007) in southern England. Also in northern Germany (Herbst and Kappen 1999) and Canada (Lafleur et al. 2005), evapotranspiration in summer from the surfaces of marshlands where the dominant vegetation is reed is characterized by daily totals at a level of 2–4 mm·d^{-1}. Similar values of daily evapotranspiration totals are characteristic of wetlands in north-western Russia (Kurbatova et al. 2002) or western Siberia (Shimiyama et al. 2004), but in both the cases the dominant vegetation is Sphagnum moss.

In hydrological practice, in assessing the water balance of a specific area from the point of view of evaporation and evapotranspiration water losses, monthly or seasonal totals are normally used. In the annual pattern, the period from October to February is characterized by the lowest values of the monthly totals of evapotranspiration (Fig. 11c). Both in the winter season of 2013/14 and that of 2014/2015, the radiation balance and humidity conditions were similar. The marshland area radiated out more energy than it gained from solar radiation and diffuse sky radiation, i.e. the radiation balance from December to February was negative (Fig. 11a). In addition, low air temperatures (monthly averages in January and February below 0 °C) also resulted in high relative humidity. The average vapour pressure deficit in the winter seasons of 2013/14 and 2014/2015 was low and amounted to 1–2 hPa. Under such weather conditions, evaporation water losses were low, at a level of 9–15 mm per month. At the same time, it is also the period when precipitation exceeds evapotranspiration, resulting in an increase in soil moisture content. This is particularly evident in the pattern of the mean values of the volumetric water content in the soil (VWC). For instance, the

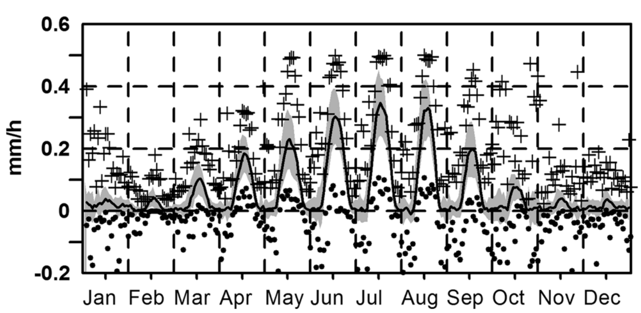


Fig. 8 Mean hourly evapotranspiration totals, shaded area – mean +/− standard deviation of hourly evapotranspiration totals, dots and cross – observed minimum and maximum values of evapotranspiration

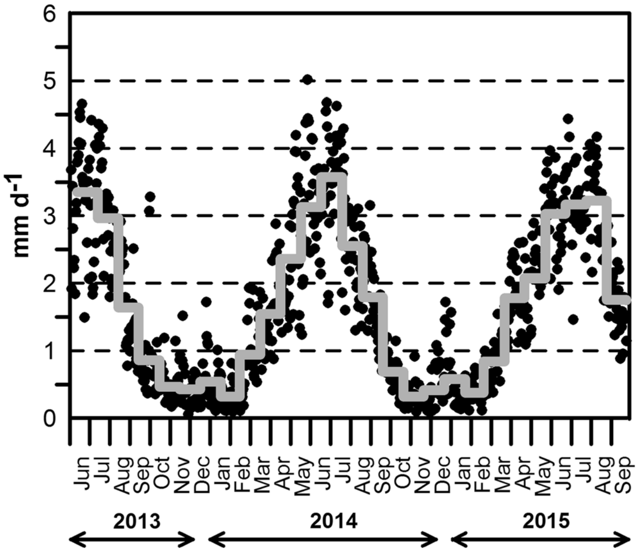


Fig. 9 Patterns of daily totals (*dots*) and mean daily total (*gray step plot*) of evapotranspiration during the study period

winter season 2014/15 in the months from November to March marked an increase of VWC from 0.7 to about 0.9. In contrast, the low values of VWC in February and March 2014 were due to problems with the measurements due to very low temperatures in January 2014 and frozen soil (Fig. 11a).

In spring, as a result of increased turbulent water vapour flux, the monthly totals are also higher. In both the analyzed spring seasons, the weather conditions were very similar. The obtained monthly totals varied from about 25 mm in March to about 70 mm in May. The highest monthly evapotranspiration totals are characteristic of summer. In the three summer seasons studied, the monthly evapotranspiration totals from June to August exceeded 80 mm. The highest evaporation occurred in July 2014, with a monthly total of 110 mm. That month was exceptionally warm (a monthly average of 19.1 °C)(Fig. 11b), and the surface of the wetland area received, via radiation

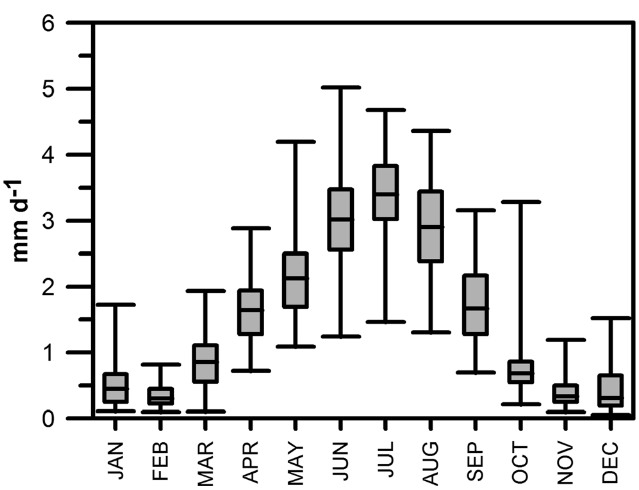


Fig. 10 Median, inter-quartile ranges, maximum and minimum daily values of evapotranspiration

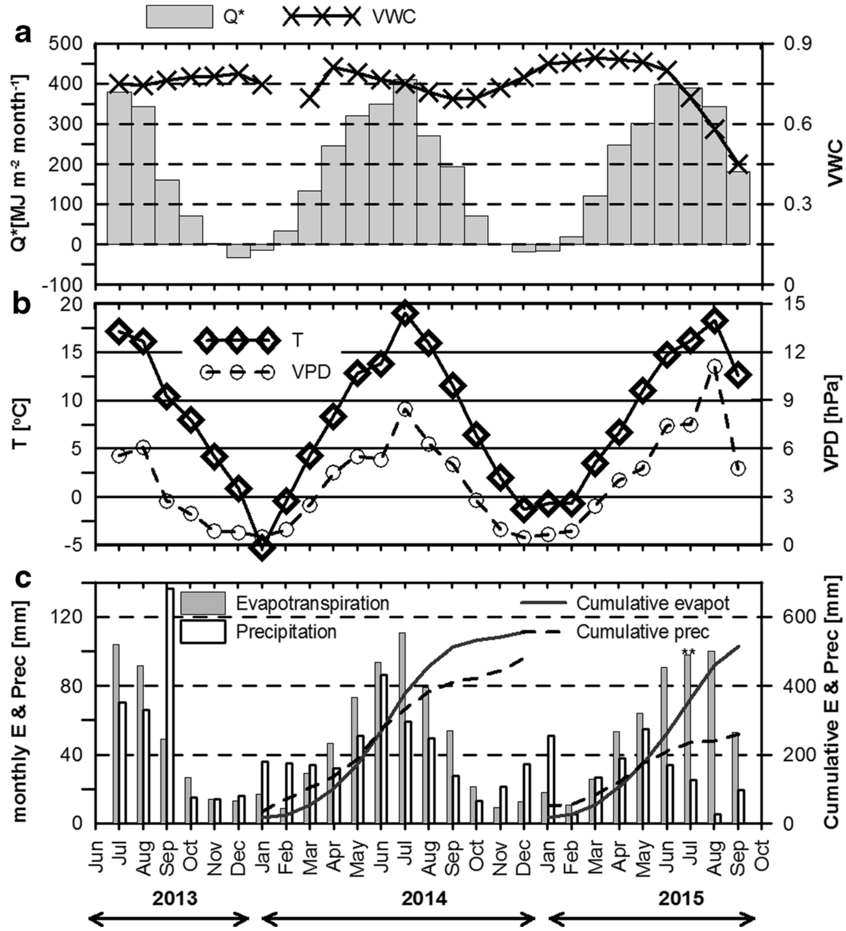


Fig. 11 Patterns of monthly evapotranspiration (E) and precipitation (P) totals during the study period with cumulative evapotranspiration and precipitation totals, mean monthly air temperature (T), vapour pressure deficit (VPD) and volumetric water content (VWC), monthly total of net radiation (Q*). ** Monthly sum of evapotranspiration taken from 20 days

exchange, a total of 410 MJ m^{-2} of energy. Another factor contributing to the high evaporation that month was high rainfall which led to high soil moisture values (VWC monthly averages in June and July amounted to ca 0.75). In the summer months of 2015, evapotranspiration was not so high, although the summer was exceptionally dry. The monthly rainfall did not exceed 40 mm and the average monthly value of vapour pressure deficit was the highest(11.2 hPa) throughout the measurement period (Fig. 11b). Under such conditions, the monthly evapotranspiration totals increased from 90 mm in June to 100 mm in August due to water resources in the soil (average monthly values of the VWC index dropped from 0.8 in May and June to 0.4 in September). Such high evaporation with very low rainfall confirms the pivotal role played by wetlands in the water balance of a specific area as well as in its surface heat balance. In the autumn months, the obtained monthly totals decreased from about 50–60 mm in September to about 10–15 mm in November.

A characteristic feature of the seasonal pattern of evapotranspiration in Poland is that in the warm half of the year it exceeds precipitation (Daniela and Lenart 1989; Musiał and Rojek 2007). A similar pattern was observed for the marshland area studied. From April to November, evapotranspiration was higher than the recorded precipitation (except September 2013, when there was exceptionally high rainfall). For example, from April to October 2014 evaporation losses amounted to more than 450 mm, while total precipitation at the measurement station during this period was 320 mm. The annual precipitation and evapotranspiration totals, in turn, amounted to 475 mm and 555 mm, respectively (Fig.11c).

Conclusion

Evapotranspiration in the marshland areas of north-eastern Poland is characterized by a distinct diurnal cycle, particularly in the growing season. The average values of latent heat in the noon hours vary from about 100 Wm^{-2} in early spring to about 250 Wm^{-2} in summer. During this season, the losses of radiation energy available for evaporation vary from about 30 % in March and April to 60 % in July and August. At the same time, with the growth of wetland vegetation during the growing season, the redistribution of radiation energy received between the sensible and latent heat fluxes changes quite significantly. The average Bowen ratio in the spring season assumes values greater than 1, while in the summer season they are in the range 0.35 to 0.7.

The obtained average daily totals in the summer months at a level of 3–3.5 mmd^{-1} are similar to the results obtained when characterizing evapotranspiration in other wetland areas in Europe, Canada and Siberia. At the same time, multi-seasonal measurements under different weather conditions in individual years confirmed the role which the marshlands play for the catchment water balance and the surface heat balance. For example, the soil moisture reserves are able to maintain evapotranspiration rates under drought conditions that are only slightly lower than under standard seasonal conditions.

Acknowledgments We would like to thank the Biebrza National Park for the cooperation and the permission to conduct the study in the Park. The study was supported by the Polish National Science Centre under projects UMO-2011/01/B/ST10/07550 and N N306717740.

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