**UNIVERSITY OF SOUTH BOHEMIA IN ČESKÉ BUDĚJOVICE**

**FACULTY OF SCIENCE**

# Effect of different types of ecosystems on their meteorological conditions and energy balance components

Ph.D. Thesis

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České Budějovice 2014

This thesis should be cited as:

Huryna H., 2014: Effect of different types of ecosystems on their meteorological conditions and energy balance components. Ph.D. Thesis Series, No. 16. University of South Bohemia, Faculty of Science, České Budějovice, Czech Republic, 171 pp.

**Annotation**

Terrestrial ecosystems play a significant role in exchange of water, heat and energy between the land surface and the atmosphere thereby affecting on a regional climate. The partitioning energy into latent, sensible and ground heat fluxes in various landscape types occur differently. Plants play a vital role in the energy partitioning via evapotranspiration. Due to its active regulation of water release, the vegetation influences the temperature and wetness therefore providing a cooling effect of environment. The studies presented in this thesis are based on original data and focused on effect of land cover on local climate conditions, such as air temperature, moisture and energy distribution in a flat mosaic landscapes.

**Declaration [in Czech]**

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This thesis originated at the Faculty of Science, University of South Bohemia, supporting doctoral studies in the biophysics study programme

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**Financial report**

The research presented in this thesis was supported by Ministry of Education, Youth and Sports of the Czech Republic (NPV 2B06023), the Ministry of Agriculture (section of environmental subsidies 2011, 2012) and the Grant Agency of the University of South Bohemia (152/2010/Z)

## ACKNOWLEDGEMENTS

I am deeply grateful to my supervisor Doc. RNDr Jan Pokorný for his continual support, invaluable advice and significant contribution throughout PhD process. His guidance helped me in all the time of research.

I gratefully thank Petra Hesslerová for her collaboration and constructive advice relating to this research. I would also like to express my gratitude to Alžbeta Rejšková-Procházková who generously read the text and made many helpful suggestions and corrections. I would like to thank Jakub Brom for his practical advice in statistical part of the project. I especially grateful to all of my colleagues in ENKI o.p.s, namely to Vladimír Jirka who was in charge of meteorological monitoring, and to all people in the castle for their administrative support and creating a pleasant working atmosphere.

I want to thank all my friends and colleagues from Nové Hrady, Třeboň and České Budějovice who shared their experience and friendship. I owe a very important debt to my friend and roommate Vasilina Zayats for her patient, kindness, great help and support. Special thanks to Alexey Bondar, Ekaterina Sviridova, Volha Shmidt, Andrei Lapitski, Daria Malakhova and Oxana Degtjarik. They make my life a source of fun.

Finally, I would like to express my love and appreciation to many people in Belarus especially my parents, grandparents and sister for their continuous supporting and encouragement during my whole life.

**List of publications and author's contribution**

Huryna H, Pokorný J (2010) Comparison of reflected solar radiation, air temperature and relative air humidity in different ecosystems: from fishponds and wet meadows to concrete surface. In Vymazal J (ed.): Water and nutrient management in natural and constructed wetlands. Springer, The Netherlands, 308–326

*Hanna Huryna carried out collecting the dataset, statistical analysis of all meteorological data and participated in writing and revising the manuscript.*

Pokorný J, Brom J, Čermák J, Hesslerová P, Huryna H, Nadezhdina N, Rejšková A. (2010) Solar energy dissipation and temperature control by water and plants. *International Journal of Water 5*(4), 311-337

*Hanna Huryna analysed meteorological data, evaluated energy fluxes and meteorological parameters which are presented in Figures 2 – 8, participated in writing and revising the manuscript.*

Huryna H, Pokorný J, Brom J (2014) The importance of wetlands in the energy balance of an agricultural landscape. *Wetlands Ecology and Management 22*(4), 363 – 381, doi: 10.1007/s11273-013-9334-2 (IF = 1.218)

*Hanna Huryna performed statistical analysis of the data, collected literature, participated in writing and revising the manuscript. according to reviewers' demands*

Huryna H, Hesslerová P, Pokorný J, Jirka V, Lhotský R (2014) Distribution of solar energy in agriculture landscape – comparison between wet meadow and crops. In Vymazal J (ed.): The role of natural and constructed wetlands in nutrient cycling and retention on the landscape. Springer, The Netherlands, 103 – 122, doi: 10.1007/978-3-319-08177-9_8

*Hanna Huryna analysed and processed the meteorological data, evaluated energy fluxes, collected literature and collaborated in writing the manuscript.*

**Publications not included in this thesis**

Pokorný J, Huryna H, Harper D. Climate regulation: case studies of wetlands and air quality. *in preparation for Wetlands Encyclopedia (Springer)*

Cudlín O, Huryna H, Pokorný J. Relationship between biodiversity and ecosystem function performance in differently disturbed territories. *in preparation*

Pokorný J, Brom J, Hesslerová P, Huryna H, Jirka V, Lhotský R (2011) Zhodnocení transformace sluneční energie a retence ţivin vybraného agroekosystému s důrazem na ochranu a kvalitu vod odtékajících ze zemědělské krajiny. *Report for the Ministry of Agriculture of the Czech Republic*

Pokorný J, Brom J, Hesslerová P, Huryna H, Jirka V, Lhotský R (2012) Provedení monitoringu a vyhodnocení celosezónní distribuce slunečního záření v typických porostech zemědělských plodin (řepka, kukuřice, travní porost), monitoring a zhodnocení kvality a kvantity odtékající vody, monitoring intenzifikačních vstupů a zhodnocení agrotechnických aktivit. *Report for the Ministry of Agriculture of the Czech Republic*

Pokorný J, Hesslerová P, Jirka J, Huryna H (2014). Termovizni snimky ukazuji jak vegetace pusobi na mistni klima a na obeh vody. Nove trendy v cistirenstvi. ENVI-PUR, CzWA, Soběslav. *in press*

## LIST OF ABBREVIATIONS AND NOTATIONS

| | | |
|---|---|---|
| $A_d$ | net loss energy due to the horizontal advection | $W\ m^{-2}$ |
| a | empiric constant | |
| b | empiric constant | |
| C | concentration of matter | $mol\ m^{-3}$ |
| c | empiric constant | |
| $c_p$ | specific heat capacity of air | $J\ kg^{-1}\ K^{-1}$ |
| D | diffusivity | $m^2\ s^{-1}$ |
| $d$ | zero plane displacement height | m |
| E | evapotranspiration rate | $mm\ d^{-1}$ |
| $e_a$ | actual vapour pressure of the air | kPa |
| $e_s$ | actual vapour pressure of the air near evaporating surface | kPa |
| $e_w$ | saturated vapour pressure at the evaporating water surface | kPa |
| $e_{wa}$ | saturated vapour pressure at the air | kPa |
| $e_{ws}$ | saturated vapour pressure at the surface | kPa |
| EF | evaporative fraction | |
| ET | actual evapotranspiration | $mm\ d^{-1}$ |
| G | ground heat flux | $W\ m^{-2}$ |
| $g$ | gravitational acceleration | $m\ s^{-2}$ |
| H | sensible heat flux | $W\ m^{-2}$ |
| $h$ | vegetation height | m |
| J | latent and sensible heat stored by vegetation | $W\ m^{-2}$ |
| j | flux of matter | $mol\ m^{-2}\ s^{-1}$ |
| $K_H$ | transfer coefficient for specific heat | $W\ m^{-2}\ K^{-1}$ |
| $K_v$ | exchange coefficient for water vapour | $W\ m^{-2}\ K^{-1}$ |
| $\kappa$ | thermal conductivity | $W\ m^{-1}\ K^{-1}$ |
| $\kappa_c$ | von-Kármán constant | |
| LE | latent heat flux | $W\ m^{-2}$ |
| $LE_p$ | potential evaporation | $W\ m^{-2}$ |
| $L_e$ | latent heat of evaporation | $J\ g^{-1}$ |
| $L_O$ | Monin-Obukhov length | m |
| $L_{\downarrow}$ | downward longwave radiation | $W\ m^{-2}$ |
| $L_{\downarrow c}$ | downward longwave radiation from clouds | $W\ m^{-2}$ |
| $L_{\uparrow}$ | upward longwave radiation | $W\ m^{-2}$ |
| M | net energy absorbed by metabolism | $W\ m^{-2}$ |
| m | empiric constant | |
| n | cloud fraction | |
| P | atmospheric pressure | kPa |
| $Q_a$ | thermal conductivity of the soil with movement of | $W\ m^{-1}\ K^{-1}$ |

| | | |
|---|---|---|
| | water and air through soil layers | |
| $Q_c$ | thermal conductivity of the soil without fluid movement | $W\ m^{-1}\ K^{-1}$ |
| $RH$ | relative air humidity | % |
| $R_n$ | net radiation | $W\ m^{-2}$ |
| $r_a$ | aerodynamic resistance | $s\ m^{-1}$ |
| $r_c$ | bulk surface resistance | $s\ m^{-1}$ |
| $S_{\downarrow}$ | incoming shortwave radiation | $W\ m^{-2}$ |
| $S_{\uparrow}$ | outgoing shortwave radiation | $W\ m^{-2}$ |
| T | absolute temperature | K |
| $T_*$ | scaling parameter in the boundary layer | °C |
| $T_a$ | ambient temperature | °C |
| $T_{atm}$ | air temperature | °C |
| $T_c$ | temperature of soil | °C |
| $T_s$ | temperature of evaporating surface | °C |
| U | wind speed | $m\ s^{-1}$ |
| $u_*$ | wind friction velocity | $m\ s^{-1}$ |
| VPD | water vapour pressure deficit | kPa |
| $z$ | distance, height above (below) earth´s surface | m |
| $z_{0h}$ | roughness for heat and vapour transfer | m |
| $z_{0m}$ | roughness length for momentum transfer (m) | m |
| | | |
| $\alpha$ | albedo | |
| $\beta$ | Bowen ratio | |
| $\gamma$ | psychrometric constant | |
| $\varepsilon$ | emissivity | |
| $\zeta$ | Monin-Obukhov stability parameter | |
| $\lambda_{max}$ | wavelength | µm |
| $\lambda$ | latent heat of water vaporization | $MJ\ Kg^{-1}$ |
| $\rho$ | air density | $kg\ m^{-3}$ |
| $\sigma$ | Stefan-Boltzmann constant | $W\ m^{-2}\ K^{-4}$ |
| $\Psi_h(\varsigma)$ | stability coefficients for heat | |
| $\Psi_m(\varsigma)$ | stability coefficients respectively for momentum transfer | |
| $\Omega$ | decoupling factor (coefficient) | |
| $\Delta$ | ratio between the saturation water vapour pressure gradient and the temperature gradient | $kPa\ °C^{-1}$ |
| | | |
| BREB | Bowen ratio energy balance | |
| IPCC | International Panel of the Climate Change | |
| TBBR | Třebon Basin Biosphere Reserve | |

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# 1. Introduction

The climate change is a widely discussed topic in the scientific community. The International Panel of the Climate Change (IPCC) is a scientific intergovernmental body tasked to estimate the risk of climate change caused by human activities. According to the Fifth Assessment Report (AR5) of IPCC, average surface temperature has risen of roughly 0.85 $^{o}$C between 1880 – 2012 (*IPCC, 2013: 194*). Apart from the temperature, there are many other indicators, such as ice and snow cover melting, sea level increase, ocean currents intensity and direction change, ocean acidity increase and surface energy budget change that support evidence of the Earth warming. It has been assumed that the main cause of the global warming is the increasing concentration of carbon dioxide in the atmosphere (*IPCC, 2007; IPCC, 2013*). The radiative forcing connected with increased concentrations of greenhouse gases in the atmosphere from 1750 to 2005 has been estimated to cause surface energy input increase of 1 – 3 W $m^{-2}$. In the next decade radiative forcing is expected to rise by 0.2 W $m^{-2}$ (*IPCC, 2007).*

The role of vegetation in the surface energy budget is usually underestimated. *IPCC 2007* and *IPCC 2013* do not engage adequately with the complex interplay of water, sunlight and vegetation. The role of plants and water in climate is not mentioned in the Summary for Policymakers of the AR5 *(IPCC, 2013).* However, plant-cover structure influences the distribution of heat in the ecosystems and can provide mitigating effects on climate on regional level *(Ryszkowski and Kedziora, 1987; 1995; Pokorný et al., 2010; Schwartz, 2013)*. Thereby, the knowledge about soil – vegetation – atmosphere interaction is a significant factor to determine the mitigating effect on local climate and land-cover change.

### 1.1 Aims of the research

The objective of this thesis was to investigate the influence of the land cover structure on a local climate, i.e. on moisture, air temperature and energy balance of the terrestrial ecosystems. Based on the data from 12 continuously measuring meteostations operated by ENKI, o.p.s. The aims of this research were:

- 1) to estimate the temperature and air humidity differences between different land cover types

- 2) to evaluate the role of vegetation in solar energy balance and energy distribution of some typical ecosystems of the Czech Republic
- 3) to determine the evaporative fraction of the studied land cover types
- 4) to discuss the role of plant cover types on the local hydrological cycle

## 1.2 Radiation distribution on the Earth's surface

Energy exchanging processes in ecosystems comply with the first and the second laws of thermodynamics. The first law of thermodynamics states that energy can be converted from one form to another but can never be created or destroyed (conservation law). In the energy transforming processes a certain part of energy is dissipated as heat. The irreversible dissipation of energy is expressed by entropy. According to the second law of thermodynamics, the system tends to change itself spontaneously into a less organized form while entropy of the system grows (entropy law). This means that temperature in a system tends to reach an equilibrium. However, the living systems are able to apply solar energy to stay away from the thermodynamic equilibrium. They use the energy potential for their self-organization *(Schrodinger, 1944; Gorshkov et al., 2004; Schneider and Sagan, 2005; Lineweaver and Egan, 2008; Bunn, 2009; Klauber, 2009; Skene, 2013).*

The relationship between the temperature and the emitted energy by the Sun in space obeys the radiation laws, the most important being the Planck's radiation law. This law determines the amount of radiation emitted by a black body as a function of temperature and wavelength. Integrating Planck's distribution gives the Stefan-Boltzmann law. According to the Stefan-Boltzmann law, the total energy density of blackbody radiation is proportional to the forth power of the absolute thermodynamic temperature of the surface:

$$E = \varepsilon\sigma T^4, (\mathrm{W\ m^{-2}}) \tag{1}$$

where: T is surface temperature in K, σ is the Stefan-Boltzmann constant ($5.67e^{-8}$, $\mathrm{W\ m^{-2}\ K^{-4}}$), ε is the emissivity. Emissivity is the ability of a surface to absorb or emit energy in the form of electromagnetic radiation. The highest possible emissivity is 1 which is the emissivity of a blackbody; the lowest is intrinsic to a perfect mirror (white body) and equals 0. Emissivity of all other bodies (grey bodies) fluctuates between 0 and 1 depending on the types of material and temperature of the surface *(Michalski et al., 2001).*

Wien's displacement law describes a relation between wavelength and temperature at maximum radiation intensity.

$$\lambda_{max} = \frac{2898}{T}, (\mu m) \quad (2)$$

### 1.2.1 Shortwave radiation

Surface energy transfer is a driving force of water movement in the Earth – atmosphere system. The Sun radiates energy in the form of shortwave radiation. The Earth gets 180 000 TW energy from the Sun. The energy that falls on the Earth's surface is absorbed, reflected, transformed into longwave radiation and emitted back to the space. The flux of solar radiation reaching the top of atmosphere is called the solar constant. Based on the satellite measurements this value is approximately 1367 W $m^{-2}$ (± 20 W $m^{-2}$) *(Miller, 1981; Brutsaert, 1982; Arya, 2001; Gueymard, 2004).* The actual direct irradiance at the top of the Earth's atmosphere varies slightly over the year from 1321 W $m^{-2}$ in July to 1412 W $m^{-2}$ in January due to the Earth's elliptic orbit and to the variation in the distance between the Earth to the Sun *(Duffie and Beckman, 1991; Arya, 2001).*

The spectrum of the Sun's emission is very similar to that of a blackbody with the temperature of approximately 6000 K and ε = 1. The solar spectrum extends from 0.01 to 4.0 μm and includes ultraviolet radiation (0.01 – 0.39 μm), visible light (0.40 – 0.76 μm) and near infrared radiation (0.76 – 4.0 μm). The maximum emissivity energy of solar radiation is at the wavelength of 0.47 μm that falls in the blue-green portion of the spectrum (Fig.1) *(Selinger and McElroy, 1965).*

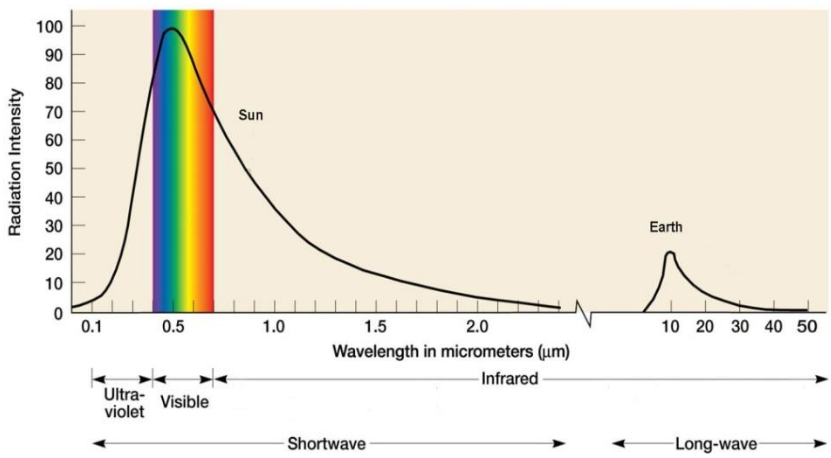


**Figure 1.** Comparison of the intensity of the shortwave (solar) and longwave radiation (terrestrial) radiation (according to *McKnight and Hess, 2007)*

The spectrum of the shortwave radiation is changed when passing the atmosphere as a result of light absorption, reflection or scattering of the atmospheric gasses, particles, aerosols and clouds. Each type of a molecule has its own position of absorption bands in different parts of electromagnetic spectrum. Gases such as water vapour, carbon dioxide, methane, ozone or carbon monoxide participate in the absorption and emission processes of radiation in the atmosphere. These gases make up less than 1 % of the volume of the atmosphere. But the average temperature of the Earth would decrease of about 33 $^{o}$C if it hadn't been warmed up by the absorption and emission processes of these gases (*Ahrens, 2008)*.

It is mainly ozone that absorbs radiation in the ultraviolet part of the electromagnetic spectrum. The carbon dioxide efficiently absorbs the energy in the mid- and far- infrared regions (13-17.5 µm). Water vapour has two most important absorption areas at 5.5 – 7.0 µm and above 27 µm (Fig. 2). Water vapour represents about 95 % of greenhouses gases *(Michaels, 1998).* Therefore, it is the water vapour that is the dominant greenhouse gas in the atmosphere although it remains in the atmosphere only for a short time: from several hours to several days.

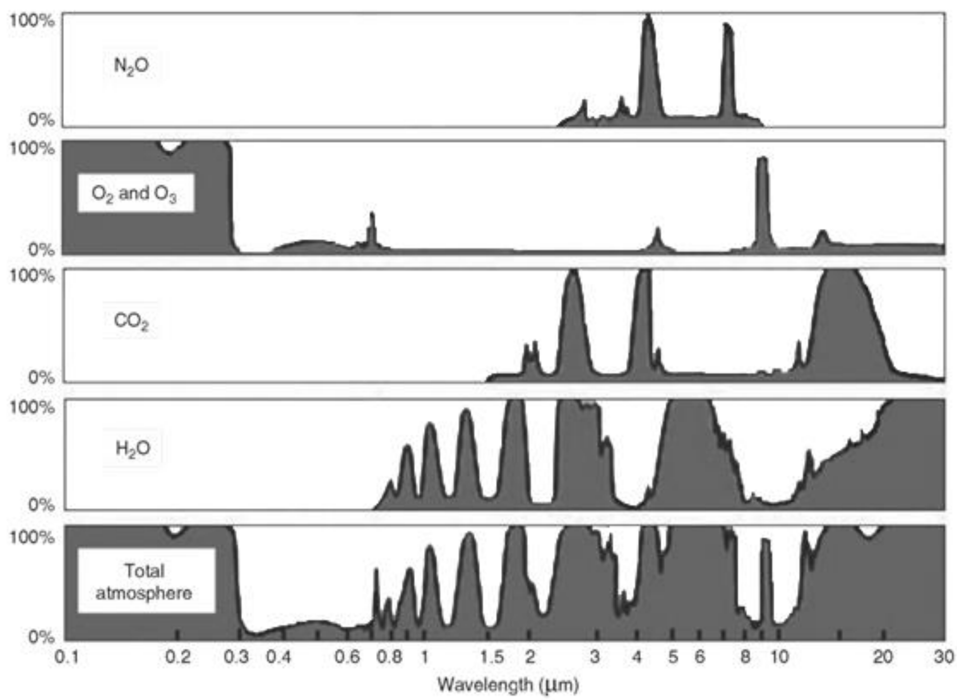


**Figure 2.** Absorption spectra of radioactively active gases in the lower atmosphere as a function of wavelength (according to *Shuttleworth, 2012)*

### 1.2.2 Longwave radiation

The Earth surface temperature is close to 288 K, therefore it emits radiation energy at 10 μm in the infrared region of the electromagnetic spectrum whereas approximately 99 % of the solar radiation occurs in the spectral interval 0.15 – 4.0 μm (Fig. 1).

The longwave radiation near the surface can be divided in two parts: the outgoing radiation ($L_{\uparrow}$) which is released from the ground surface and the vegetation and the incoming radiation ($L_{\downarrow}$) emitted by the atmosphere *(Arya, 2001).* An estimation of the incoming longwave radiation requires comprehensive knowledge about the air temperature, the air humidity and the properties of emitting substance, such as gases or aerosols *(Brunt, 1932; Swinbank, 1963; Brutsaert, 1975; Idso, 1981; Prata, 1996; Niemela et al., 2001; Perez-Garcia, 2004).* For the measurement of the longwave radiation pyrgeometers are used (*e.g. Walden et al., 1998; Philipona and Ohmura, 2001; Marty et al., 2003*). However the direct measurements of $L_{\downarrow}$ are not common in the climatological research due to difficulties and high costs connected with the accurate longwave radiation measurements *(Pluss and Mazzoni, 1994; Kjaersgaard et al., 2007; Marthews et al., 2012).* Thus, this variable is often modelled on the bases of other meteorological variables and parameterizations describing emission and absorption processes in the atmosphere *(Sedlar and Hock 2009; Kruk et al., 2010).*

Many theoretical and empirical techniques have been developed for determine $L_{\downarrow}$ with different formulae of effective atmospheric emissivity. All models consider the atmosphere a grey body, so $L_{\downarrow}$ is defined by the bulk emissivity $\varepsilon_{atm}$ and the air temperature $T_{atm}$ according to the Stefan-Boltzmann law (Eq. 1). However, it is difficult to obtain effective temperature and bulk emissivity for a vertical column of the atmosphere *(Dilley and O'Brien, 1998; Crawford and Duchon, 1999).* Generally, $L_{\downarrow}$ is evaluated from temperature $T_a$ and water vapour pressure e, measured on the surface (meteorological) screen level (~ 1.3 – 2 m above the ground) *(Monteith and Unsworth, 1990; Crawford and Duchon, 1999).* Two meters height is considered as a standard in our research.

Various methods suggest different processes of estimation the emissivity from the air temperature and water vapour pressure or vapour pressure measured in the meteorological screen *(Swinbank, 1963; Idso and Jackson, 1969; Unsworth and Montheith, 1975; Konzelmann et al., 1994).* However, this assumption is relevant under clear sky conditions only. For estimation of longwave radiation on a cloudy day it is necessary to consider emissivity of clouds *(Sugita and Brutsaert 1993; Aladosarboledas et al., 1995; Crawford and Duchon, 1999).* Thereby, the assessment of atmospheric emissivity has to be divided according to the weather into two groups: "clear sky" models for the cloudless weather and "cloudy sky" models. The comparative analyses of equations for air emissivity estimation used by different researches for cloudless conditions are presented in Table 1.

**Table 1.** Comparative air emissivity calculation approaches for computing longwave radiation under clear-sky conditions

| Publication | Formula | Note |
|---|---|---|
| *Angstrom (1918)* | $\varepsilon = a - b(10^{-ce})$ | a, b and c are empiric constants which is ranged from 0.75 to 0.82, from 0.15 to 0.33 and from 0.09 to 0.22, respectively. The values depending on the location under study |
| *Brunt (1932)* | $\varepsilon = a + b\sqrt{e}$ | a and b are empiric constant, depending on local conditions |
| *Swinbank (1963)* | $\varepsilon = 9.36 * 10^{-6} T_a^2$ | Based on the data from Aspendale, Kerang (both in Australia), the Indian Ocean and Benson (UK) |
| *Idso and Jackson (1969)* | $\varepsilon = 1 - 0.261 exp\,(-7.77 \cdot 10^{-4}(273 - T_a)^2)$ | Based on the data from Point Barrow, Phoenix (both in USA), Aspendale, Kerang (both in Australia) and the Indian Ocean |
| *Brutsaert (1975)* | $\varepsilon = 1.24 \left(\frac{e}{T_a}\right)^{\frac{1}{7}}$ | Based on the data from other researches |
| *Idso (1981)* | $\varepsilon = 0.7 + 5.95 * 10^{-7} e_a exp(\frac{1500}{T_a})$ | Based on the data from Phoenix (USA) |
| *Montheith and Unsworth (1990)* | $\varepsilon = \frac{1}{\sigma {T_c}^4}(-119 + 1.06\sigma T_a^4)$ | Based on the data from English Midlands (UK) |
| *Konzelmann et al. (1994)* | $\varepsilon = 0.23 + 0.484(\frac{e}{T_a})^{\frac{1}{8}}$ | Based on the data from Greenlands (Denmark) |

A new method for determination of the net longwave radiation was suggested by *Jirka et al. (2009).* This method is based on radiation measurement with a NET radiometer and calculation based on the Stefan-Boltzmann law.

In the presence of clouds, the flux of the longwave radiation to the surface increases. When sky is cloudy, the fluxes are subject the empirical adjustment which depend mostly on the total cloud cover *(Niemela et al., 2001). Monteith (1973)* presented an equation for estimating the longwave radiation under cloudy conditions based on clear sky estimates and their empirical corrections related to the cloud fraction:

$$L_{\downarrow c} = L_{\downarrow}(1 + mc^{n}) \quad (3)$$

where: $L_{\downarrow c}$ is the downward longwave radiation from cloudy or partly cloudy sky (W $m^{-2}$), $L_{\downarrow}$ is the clear sky radiation, $m$ is the empiric constant and $n$ is the cloud fraction.

Empiric constant mostly depends on the elevation and on the type of the present clouds. Different values of $m$ can be found for example in *Sellers (1965), Maykut and Church (1973), Jacobs (1978), Sugita and Brutsaert (1993), Konzelmann et al. (1994), Pluss and Omhura (1997), Crawford and Duchon (1999), Kjaersgaard et al. (2007), Sedlar and Hock (2009).* The cloud cover fraction is mainly estimated by visual inspection, dividing the sky into ten sectors and counting up the number of sectors covered. Cloud cover fraction ranges from 0 to 1 *(Crawford and Duchon, 1999).*

Calculation of outgoing longwave radiation using Stefan-Boltzmann equation requires knowledge of surface temperature and surface emissivity. For most natural surfaces the emissivity ranges from 0.9 to 1.0 and for green crop it is 0.98 *(Oke, 1987; Brutsaert, 1982).*

## 1.3 Energy balance of vegetative surfaces

The amount of energy received, reflected and emitted from the Earth surface is defined as “net radiation“. For computing net radiation, linear (*Kaminsky and Dubayah, 1997)* or multiple *(Irmak et al., 2003)* regression models can be used. These models are simple and require only a few independent variables.

Common methods evaluate net radiation by estimating the solar radiation balance *(Jensen et al., 1990; Allen et al., 1998; Arya, 2001; Kjaersgaard et al., 2007.):*

$$R_n = S_{\downarrow} - S_{\uparrow} + L_{\downarrow} - L_{\uparrow} = S_{\downarrow}(1 - \alpha) + L_{\downarrow} - L_{\uparrow} \quad (4)$$

where: $R_n$ is the net radiation, $S_{\uparrow}$ and $S_{\downarrow}$ are the incoming and the outgoing shortwave radiation fluxes, respectively, and $L_{\uparrow}$ and $L_{\downarrow}$ are the downward and the upward longwave radiation fluxes measured on the surface, respectively. Longwave radiation is strongly related to the surface temperature and the surface emissivity. All variables are expressed in W $m^{-2}$.

The direct field measurement of net radiation is difficult to realise *(Twine et al., 2000).* Net radiometers used in the meteorological stations to measure local net radiation are sensitive and thus demand permanent calibration *(Alados et al., 2003).* The net radiation measurement error can vary from as much as 20 % *(Halldin and Lindroth, 1992)* to 6 % *(Twine et al., 2000).* For large areas satellites can be used for net radiation measurements *(Diak et al., 2000).*

At the surface the net radiation is balanced by turbulent fluxes into the atmosphere, conduction into the ground and accumulation into biomass according to the energy conservation law as

$$R_n = LE + H + G + J + M + A_d \tag{5}$$

where: LE is the latent heat flux; it is product of the latent heat of vaporization of water (L) and the rate of evapotranspiration from the vegetation or the soil (E), H expresses vertical turbulent fluxes of sensible heat flux into the atmosphere by thermal convection, G is the heat conducted into the soil, J is the latent and sensible heat stored by vegetation, M is the net energy absorbed by metabolism (photosynthesis minus respiration) and $A_d$ is the net loss energy due to the horizontal advection (Fig. 3) *(Kravčik et al., 2008).*

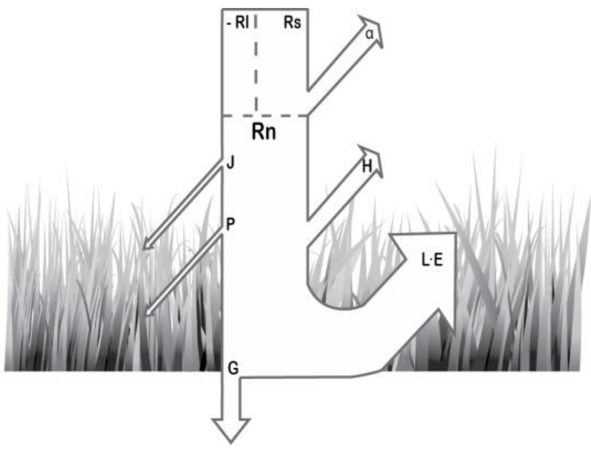


**Figure 3.** Dissipation of solar energy in a plant stand. Rs – shortwave radiation, Rl – longwave radiation, α – albedo, Rn – net radiation, H – sensible heat, L·E – latent heat of evapotranspiration, G – ground diffusion, J – accumulation of heat in biomass, P – energy consumption by photosynthesis

The amount of energy used in plant metabolism (M) as well as the amount of heat stored by vegetation (J), advection and freezing of water ($A_d$) are very small and can be usually neglected (*Thom, 1975; Monteith and Unsworth, 1990).* Therefore, the energy balance is practically simplified to the following form:

$$R_n = LE + H + G \tag{6}$$

The latent heat flux is related to evaporation of water via releasing or consuming energy during the phase-transition process. It describes the flow of energy caused by the difference in water vapour between the land surface and the atmosphere. The latent heat flux is usually dominant by wet surface, vegetation and open water during the day. The energy flows away from the surface and evapotranspiration occur. The downward latent heat flux at night indicates condensation at the surface as frost or dew. The rate of latent heat flux from the surface into the air is determined according to Fick's law as the proportional amount of the vapour pressure gradient between the evaporating surface and the environment:

$$LE = -\frac{\rho c_p}{\gamma} K_v \frac{\partial e}{\partial z} \tag{7}$$

where p is the density and $c_p$ is the specific heat of the air, $K_v$ is the exchange coefficient for water vapour and $\partial e/\partial z$ is the vertical atmospheric vapour pressure gradient and γ is the psychrometric constant.

The sensible heat flux is driven by temperature differences between the surface and the overlying air. Heat is initially transferred into the atmosphere by conduction. Then, with gradual heating air, it circulates upwardly through convection. When the surface is warmer than the overlaying air, heat will be transferred upwards into the air as a positive sensible heat transfer. If the air is warmer than the surface, heat is transferred from the air to the surface creating a negative sensible heat transfer. The rate of the sensible heat flux from the surface into the air is proportional to the temperature gradient between the land surface and its environment and it is controlled by the magnitude of the exchange coefficient that depends on the turbulent conditions above the surface:

$$H = \rho c_p K_H \frac{\partial T}{\partial z} \tag{8}$$

where $\partial T/\partial z$ is the temperature gradient, $K_H$ is the transfer coefficient for specific heat *(Rosenberg et al., 1983).*

The Bowen ratio energy balance method (BREB) is one of the most common methods used to determine the fluxes of sensible and latent heat. It is defined as *(Bowen, 1926):*

$$\beta = \frac{H}{LE} = \gamma \frac{T_s - T_a}{e_s - e_a} \tag{9}$$

where $T_s$ and $T_a$ are the temperature of an evaporating surface and the ambient temperature, respectively, $e_s$ is the actual vapour pressure of the air near the evaporating surface, $e_a$ is the actual vapour pressure of the air and $\gamma$ is a psychrometric constant approximately equal to 0.066 kPa $K^{-1}$, which can be computed as:

$$\gamma = \frac{c_p P}{0.622\lambda} \tag{10}$$

where $c_p$ is the specific heat of the air in a constant pressure, P is the atmospheric pressure and λ is the latent heat of water vaporization (2.45 MJ $kg^{-1}$). If Bowen ratio is less than one, a greater fraction of the available energy at the surface is passed to the atmosphere as the latent heat flux than as the sensible heat flux.

There are two main energy transfers that occur in soil: conduction and convection. The ground heat flux through a porous medium includes the heat transport through all the soil components, i.e. water, air, mineral and organic matter.

Conduction is a flow of heat through solid or liquid materials as a consequence of vibration and collision of molecules and free electrons. The thermal conduction takes place when bodies of various temperatures are in physical contact *(Chudnovskii, 1976; Maxwell 2001; Atkins and de Paula, 2006).*

Another process that moves heat through the soil is the convective movement of air driven by the total gas pressure gradient. It results in the movement of the entire air masses from a zone of higher pressure to a zone of lower pressure *(Scott, 2000).* The total heat flux of the soil a] with ($Q_a$) and b] without ($Q_c$) movement of water and air through soil layers is expressed as:

$$G = Q_c + Q_a = -\kappa\left(\frac{\partial \mathrm{T}}{\partial \mathrm{z}}\right) \tag{11}$$

where: ${}^{\partial \mathrm{T}}/_{\partial \mathrm{z}}$ is the vertical temperature gradient in the soil (K $m^{-1}$), and κ is the thermal conductivity of the soil (W $m^{-1}$ $K^{-1}$) that integrate all thermal conductivities of the various soil elements and processes.

The soil's thermal conductivity depends on the dry density, the saturation, the mineral and texture composition, the temperature, the water content and time *(de Vries, 1963; Wierenga et al., 1969).* Experimentally estimated values of thermal conductivity were published by *Kersten (1949), de Vries (1963), Kimball et al. (1976a, 1976b), Asrar and Kanemasu (1983), Montheith and Unsworth (1990), Stathers and Spittlehouse (1990), Gregory (1991), Sikora and Kossowski (1993), Peters-Lidard et al. (1998), Ochsner et al. (2001).* Some values of thermal conductivity for different surfaces are given in Table 2.

**Table 2.** Thermal conductivity of different types of soil, (W $m^{-1}$ $K^{-1}$)

| | $k_{min}$ | $k_{max}$ | References |
|---|---|---|---|
| Asphalt | 1.7 | 2.1 | *Anandakumar, 1999; Mrawira and Luca, 2006* |
| Clay soil, dry | 0.25 | 0.52 | *Monteith and Unsworth, 1990; Stathers and Spittlehouse, 1990; Gregory, 1991* |
| Clay soil, moist | 0.69 | 0.87 | *Gregory, 1991* |
| Clay soil, wet | 1.04 | 1.58 | *Monteith and Unsworth, 1990; Stathers and Spittlehouse, 1990* |
| Granite | 1.69 | 3.12 | *Touloukian, 1970; Cho et al., 2009* |
| Gravel | 0.9 | 1.25 | *Gregory, 1991* |
| Loam soil | 0.15 | 0.79 | *Van Wijk, 1963* |
| Peat soil, dry | 0.17 | 0.29 | *Monteith and Unsworth, 1990, Gregory, 1991* |
| Peat soil, wet | 0.50 | 2.20 | *Monteith and Unsworth, 1990, Gregory, 1991* |
| Sandy soil, dry | 0.30 | 0.69 | *Ghuman and Lal, 1985; Stathers and Spittlehouse, 1990; Gregory, 1991* |
| Sandy soil, moist | 0.87 | 1.04 | *Gregory, 1991, Ghuman and Lal, 1985* |
| Sandy soil, wet | 1.90 | 2.42 | *Monteith and Unsworth, 1990; Stathers and Spittlehouse, 1990; Gregory, 1991* |
| Sandstone | 1.60 | 2.08 | *Gregory, 1991* |

Minimum and maximum values of thermal conductivity are reached when the soil is dry or saturated with water, respectively.

There are several ways of estimating the ground heat flux from the soil temperature and soil moisture data and the heat flux measurements. The widely used four methods are: the flux plate, calorimetric, gradient and combination methods. The most common method used for estimating the soil heat flux is the soil heat flux plate. Heat flux plates are typically small, rigid, wafer-shaped sensors that are inserted into the soil horizontally in the reference depths *(Sauer, 2002).* The plate is made of materials of known thermal conductivity, and a thermopile uses to measure the temperature gradient between its lower and upper surfaces. The faces are often covered with thin metal plates in order to protect them and to ensure good thermal contact with the soil. The heat flux plate should be small, so that it does not disrupt the soil profile and to ensure low heat capacity for quick response to any temperature changes in time. For field estimation of average soil heat flux in small depths several heat plates may be buried horizontally in the soil in different depths from 1 to 50 mm.

## 1.4 Albedo of vegetation and land use changes

Albedo is defined as a ratio of reflected radiation from the Earth's surface to total incident solar radiation. Consequently, surface albedo determines the actual amount of solar energy available to transfer turbulent heat fluxes and moisture. As it affects the energy partitioning and energy balance on the Earth surface it closely links land cover alteration with the climate change *(Mintz, 1984; Betts, 2000; Bonan, 2008; IPCC, 2013*).

Albedo of different surfaces strongly varies according to the surface's spectral features. It is then influenced by meteorological conditions such as cloudiness as well as by the spectral composition of incident solar radiation or its angle of incidence *(Paltridge and Platt, 1976; Dong et al., 1992; Yin, 1998).* The average albedo of the Earth is commonly estimated to be 0.31. As vegetation covers large parts of the Earth surface its contribution to this overall estimation is substantial. It can play a very important role on the local to regional energy balance. Vegetation is a great absorber of the solar radiation in the visible part of the spectrum while it strongly reflects most of the solar radiation in the near infra-red spectrum *(Dorman and Sellers, 1989; Liang et al., 2005).* Albedo of plants usually ranges between 0.1 and 0.27 *(Doorenbos and Pruitt, 1977; Oke, 1978; Brutsaert, 1982; Jensen et al., 1990; Jones, 1992; Meyer et al., 1999).* Approximate values of albedo for different types of vegetation cover are given in Table 3.

**Table 3.** Approximate values of albedo for some vegetation surface cover types

| Type of vegetation cover | Albedo | References |
|---|---|---|
| Grassland | 0.16 - 0.23 | *Moore, 1976; Oke, 1978; Sagan et al., 1979* |
| Tropical forest | 0.13 – 0.146 | *Pinker et al., 1980* |
| Boreal and temperate forest | 0.12 – 0.15 | *Betts and Ball, 1997; Restrepo and Arain, 2005; Wang, 2005* |
| Wheat | 0.18 – 0.23 | *Fritchen, 1967* |
| Corn | 0.20 – 0.24 | *Breuer et al., 2003* |
| Meadow | 0.15 – 0.25 | *Fitzgerald, 1974* |
| Pasture | 0.16 – 0.22 | *Campra et al., 2008* |
| Shrubs, woodland | 0.25 – 0.29 | *Breuer et al., 2003* |
| Wetland | 0.11 – 0.17 | *Burba et al., 1999* |

Back in the 1977 *Doorenbos and Pruitt (1977)* suggested that for most fields albedo is equal to 0.25. However, albedo tends to decrease with vegetation height and wetness of leaves and increase with rising leaf area index *(Cuf et al., 1995).* Besides, albedo of crops usually changes during vegetation period *(Houghton et al., 2001).*

Many experiments have been carried out to quantify the effects of land cover change on climate. *Charney (1975)* was the first who determined the connection between the growing aridity in the semi-arid regions and the change in the albedo. He hypothesized that overgrazing could lead to desertification. *Bounoua et al. (2002)* studied the effect of land cover modification on the regional and global climate. Also other studies evaluated the influence of surface albedo on climate *(Carson and Sangster, 1981; Laval and Picon, 1986; Betts, 2000; Govindasamy et al., 2001). Govindasamy et al. (2001)* demonstrated that the global cooling of about 0.25 K from between 1000 to 1900 AD might have been caused by vegetation changes.

## 1.5 Evapotranspiration process

Evapotranspiration is the important part of the hydrological cycle and surface energy balance (*Novák; 2012*). It alters regionally and seasonally according to the growing season, climate, available radiation, land cover, soil moisture, land use change and etc. Evapotranspiration is a combination of two simultaneous processes: free-water evaporation and plant transpiration from the land surface to the air. For estimating the evapotranspiration it is necessary to distinguish between the different categories of evapotranspiration. Potential evapotranspiration refers to the rate of evaporation that would occur if sufficient water supply was available *(Rosenberg et al, 1983).* This rate depends not only on the amount of radiation and heat; it is affected also by the wind speed and the air humidity (*Stephenson, 1990*). The actual evapotranspiration is the real amount of water that is evaporated.

The most important factors affecting evapotranspiration process are (1) the availability of the solar energy and the degree of advection; (2) the vapour pressure deficit between the land and the air; (3) the soil structure; and (4) the biological factors: species composition, vegetation structure, root depth, stomatal density and aperture etc.

For the purpose of this thesis, the following terms are defined:

- Evaporation is a movement of free water from natural sources, such as water bodies, ground surfaces and canopy interception to the atmosphere *(Brutsaert, 1982; Arya 2001).*

- Transpiration is the evaporation of water within the leaves of plants and its subsequent release in the gaseous state (vapour) through stomatal cavities into the atmosphere *(Allen et al., 1998)*.

### 1.5.1. Physical properties of evaporation

The process of evaporation is depicted by the kinetic theory. According to this theory, the evaporation consists of three processes: (1) evaporation of water molecules from a surface; (2) absorption of water molecules by the evaporating surface; and (3) distribution of the water molecules in the surrounding area. In liquids, the distances between molecules are small and the attraction forces between the neighbouring molecules are strong. Molecules are in constant motion and thus possess kinetic energy. The rates of molecular movement of water molecules differ. On the liquid surface the particles move faster than those in deeper layers because they are attracted only from the interior of the liquid. Those molecules that reach the evaporating surface at high speeds overcome the attraction of other water molecules and convert into vapour. At higher temperature the rate of evaporation increases and more intensive diffusion of molecules in the surrounding area occur. In addition, with increasing temperature the distance between molecules rise, which leads to a reduction of attraction forces and causes growth of evaporation. The average kinetic energy of the liquid is reduced – i.e. the temperature of the liquid decreases. This is a molecular explanation of evaporation and its cooling effect *(Coletta, 2008).*

The number of molecules evaporating from water is connected to the available energy *(Wallace and Hobbs, 2006).*

Air normally contains water vapour. It gets to the air mainly due to evaporation. Since the evaporated molecules move near the surface of the liquid, some of them collide and return to the liquid state. This occurs by increasing the vapour pressures and the process is called condensation.

The transfer of water from the surface layer into the atmosphere complies with the laws of diffusion (Fick's law) and by the vertical turbulent air mass exchange. According to the first Fick's law of diffusion, the rate of diffusion in a matter is proportional to the concentration gradient of that matter. In differential form the Fick's law is given by:

$$j = -D\frac{\partial c}{\partial z} \qquad (12)$$

where j is the flux of matter in the direction z, C is a concentration of matter with units number per meter cubed displacement in the direction z. D is the diffusion coefficient. The negative sign signifies that the flux takes place from high to low concentration *(Atkins and Paula, 2006).*

The modern theory of evaporation was formulated by Dalton. Dalton's theory describes more specifically the principle of diffusion for water evaporation. According to this theory, the evaporation is proportional to the mass of water vapour pressure differential between the surface and the overlying air:

$$E = f(U)(e_{ws} - e_{wa}) \tag{13}$$

where E is a rate of evaporation (mm $day^{-1}$); $e_{ws}$ is a saturated vapour pressure at the water surface and $e_{wa}$ is a vapour pressure in the air, f(U) is a function of wind speed.

Dalton equation included the dependence of evaporation rates on the air humidity and on the intensity of turbulent movements in the atmosphere, but it does not contain any explicit reference to the other major factor: the energy required to provide the necessary latent heat of vaporisation *(Rosenberg et. al., 1983).* In the twenties century *Penman (1948)* and *Monteith (1965)* developed their evaporation equation based on available energy analysis and turbulent flux:

$$\lambda E = \frac{\Delta(R_n - G) + \rho c_p VPD \frac{1}{r_a}}{\Delta + \gamma(1 + \frac{r_c}{r_a})} \tag{14}$$

where $E$ is evapotranspiration (mm), $\lambda$ is latent heat of vaporisation (MJ $Kg^{-1}$), $\Delta$ is the slope of the vapour pressure curve (kPa $^{o}C^{-1}$), $\gamma$ is psychrometric constant (kPa $^{o}C^{-1}$), Rn is net radiation (W $m^{-2}$), G is ground heat flux (W $m^{-2}$), $\rho$ is atmospheric density (kg $m^{-3}$), $c_p$ is specific heat of moist air (J $kg^{-1}K^{-1}$), $VPD$ is vapour pressure deficit (kPa), $r_c$ is canopy resistance (s $m^{-1}$), and $r_a$ is aerodynamic resistance (s $m^{-1}$).

### 1.5.2 Physical properties of transpiration

Plants release water through the process of transpiration. More than 99 % of water is lost through transpiration in exchange for $CO_2$ molecules used in the process of photosynthesis *(Esau, 1977).* Only a small part of water absorbed by roots is used in plants' metabolism. Water flow is also a mean of nutrient transport from the soil to the plant. In addition, transpiration of water cools leaves and serves thus as a temperature regulating process *(Gates, 1966; Raschke, 1975).*

In spite of some beneficial effects, transpiration is considered a "necessary or an unavoidable evil" of photosynthesis *(Arnon, 1992; Kramer and Boyer, 1995; Allen et al., 1998; Pokorný et al., 2006; Cudlín et al., 2013)* since the regulation of gas exchange for photosynthesis and transpiration occurs at the same time.

Plants take water by the rooting system. Mass flow of liquid water from the roots through plants to the atmosphere is driven by the decrease in hydrostatic water pressure, thus against the gravitational force. Water is transferred along the radius of xylem towards the leaves. Then it is evaporated through the cell walls of mesophyll, cuticle and mainly through stomata into the atmosphere. A small part of water taken by plants is transferred via phloem into cells where photosynthesis occurs.

Stomata are the vents limiting diffusion. They are regulated both by external environmental conditions (irradiance, humidity, temperature) and internal physiological factors (carbon dioxide concentration within the leaf water status) *(Monteith, 1995; Bunce, 1996).* Stomata optimize the balance between absorption of carbon dioxide and loss of water from plants *(Monteith and Unsworth, 1990; Jones, 1992; Ward and Robinson, 2000).*

Number of stomata per unit area of epidermis depends on the plant species and site characteristic. According to *Schulze et al. (2002),* the number of stomata varies from 30 to more than 500 (per square millimetre of the surface). The size of stomata alters between 77x42 μm and 25x18 μm.

The rate of transpiration is directly related to the degree of stomatal opening and to the evaporative demand of the atmosphere surrounding the leaf. Factors controlling transpiration are net radiation, ambient concentration of $CO_2$, leaf temperature, leaf water content, water vapour pressure deficit in the air, aerodynamics of the boundary layer above the canopy and stomata *(Losch and Tenhunen, 1981).* The impact effect of these factors depends on vegetation structure, surface roughness and types of plant (e.g. trees, shrubs, grass). The soil characteristics (soil water supply, soil temperature) can also affect stomatal opening, and thus the rate of transpiration *(Schulze et al., 2002).*

For instance, seasonal transpiration loss from agricultural ecosystems may vary between 100 mm and 400 mm *(Hall et al., 1996; Nulsen and Baxter, 2004).* Trees generally evaporate more water than other types of vegetation. Annually forest transpiration may fluctuate between 400 and 800 mm *(Harding et al., 1992; Calder et al., 2003).*

### 1.6 Vegetation and hydrological cycle

The hydrological cycle is a significant process in all ecosystems. It involves the continuous circulation of water through Earth's atmosphere, land surface and oceans. *Horton* (1931) expressed the hydrological cycle as a closed system with the four major processes: precipitation, surface deposition of precipitation, evaporation and atmospheric moisture in transportation and storage. Water presence in different parts of a hydrological cycle widely varies. For example, the evaporated water remains in the atmosphere for a short time of about 9 – 11 days, whereas groundwater may stay in an aquifer for thousands of years (*Winter et al., 1998*). Theoretically, the hydrological cycle can be divided into three levels: global (water exchange between oceans and continents through runoff and atmospheric circulation); local or terrestrial (water exchange between the land and the atmosphere) and micro hydrological cycle (circulation of water between top soil layers and near surface layers of the atmosphere within plant communities) (*Kedziora and Olejnik, 2002*).

The vegetation and water are inextricably linked through the impacts on energy and hydrological cycle (*Nobre et al., 1991; Hutjes et al., 1998; Arora, 2002).* The water presence is an important factor for the distribution of terrestrial ecosystems, whilst vegetation structure influences evapotranspiration and runoff formation (*Gerten et al., 2004*). *Falkenmark and Rockstrom (2004)* introduced the concept of “green and blue water flows”. Runoff and groundwater flows are referred to as “blue water flow”, whereas “green water flow” is denoted as evapotranspiration. Plants impact the “blue water flow” through meteorological and biological factors such as albedo (*Trimble et al., 1987; Eckhardt et al., 2003*), temperature and humidity (*Swank and Douglass, 1974*), stomatal conductance (*Field et al., 1995*), transpiration (e.g. *Wang et al., 1996*; *Koster and Milly, 1997),* root system (*Milly, 1997)* and leaf area index (*Peel et al., 2001*). Evapotranspiration represents not only the largest contribution to the hydrological cycle but it is also essential for understanding atmospheric circulation and modelling terrestrial ecosystem production (*Willmott et al., 1985*; *Nemani et al., 2003*; *Heijmans et al., 2004; Schmidt, 2010*). According to *Brutsaert (1986)* and *Oki and Kanae (2006),* nearly 60 – 70 % of annual global precipitation is re-evaporated through evapotranspiration.

The existence of vegetation on the Earth is significant as it provides a huge cooling effect. Without vegetation the local water cycle cannot be provided

(*Fraedrich et al., 1999*). A reduction in precipitation or/and irregular pattern of precipitation is caused by destruction of vegetation cover and consequent soil erosion and desertification (*Pokorný et al., 2010b*). A number of recent studies have focused on the problem of deforestation in sub-humid, semi-arid and arid regions (*Dickinson and Henderson-Sellers, 1988; Shukla et al., 1990*; *Nobre et al., 1991; Xue and Shukla, 1993; Eltahir and Bras, 1994*; *Dirmeyer and Shukla, 1995; Dirmeyer and Shukla, 1996; Pickup, 1998*). In particularly, *Xue and Shukla* (1993) described the effect of desertification on Sahel drought. They found that change in vegetation cover associated with land surface modification leads to anomalies in rainfall. Rainfall was decreased and the rainy season was delayed by half a month. Moreover, the axis of maximum rainfall had been changed. According to *McGuffie et al. (1995); Chagnon and Bras (2005),* desertification in Amazonian basin results in a local reduction in evaporation and decrease in precipitation over the region.

### 1.6.1 Forest evapotranspiration

Evapotranspiration represents a fundamental component of water circulation especially in wetlands and forest ecosystems (*Campbell and Williamson, 1997*; *Makarieva and Gorshkov 2007; Makarieva et al., 2013*).

Forest evapotranspiration is a complex process and depends on tree species, trees growth and height, soil condition, geographical location and regional climate (*Swank et al., 1988*; *Roberts and Rosier, 1994; Calder et al., 2003; Dawson et al., 2007*).

There is a large difference in the water loss among various forest types. *Swank and Douglas (1974)* measured the change of annual reduction of water yield after transformation of the broad-leaved wood to the coniferous. They indicated the steam flow was reduced of about 20 % after transformation of broad-leaved forest into a coniferous forest. According to *Calder* et al. (2003), annual rainfall lost by interception from conifer trees was 15 - 20 % higher compared to broad-leaved forest. *Bosch and Hewlett (1982)* showed that an annual average reduction of water yield of about 40 mm for every 10 % of catchment area covered with coniferous and eucalypt trees, compared to brush or grassland. For deciduous forest this would associate with an average reduction of approximately 25 mm per year. Many studies are focused on tropical forest relations. They suggest that most of precipitation in Amazonian region originates from regional transpiration (*Eltahir and Bras, 1994*; *Costa and Foley, 1997; Levia and Frost, 2003).* Annual evapotranspiration in a tropical rainforest can range from 1200 to 1600 mm (*Noguchi et al. 2004; Kume et al., 2011; Suryatmojo et al., 2013*). According to *Suryatmojo et al. (2013),* land use change, canopy cover reduction and surface disturbance can increase the runoff and reduce evapotranspiration. Reduction of canopy cover in a cultivated forest reduced the annual evapotranspiration by about 45 % while the annual runoff from natural forest increased approximately by 33 %.

*Makarieva and Gorshkov (2007; 2010)* introduced a hypothesis describing the forests as active attractors of moist air. Analysing the interrelation between rate of rainfall for forested and deforested regions across various continents and their distance from the sea level the authors concluded that annual rainfall decreases in deforested parts of continents while in areas covered by natural forests the rainfall can increase even over a distance of several thousand kilometres. Authors also presented a concept of a “biotic pump” – condensation of water vapour on forested areas results in a drop of air pressure and “horizontal sucking“ of wet air from oceans or other donor places.

### 1.6.2 Wetland evapotranspiration

Water evaporation from wetland plants represents one of the mechanisms for removal of pollutants from drainage water. Wetland transpiration is connected to mineral absorption and can be used as an indicator of purification capability of plants (*Perttu and Kowalik, 1997; Aronsson and Perttu, 2001; Randerson, 2006*). Wetland plants also participate in the removal of ammonia and nitrogen from the water and for the release of methane and carbon dioxide from the soil (*Duggan, 2005*; *Randerson, 2006*). Moreover, wetlands are considered to have a higher evapotranspiration than many other ecosystems due to the vegetation cover density, high wetness and water saturated or inundated soils (*Acreman et al., 2003*).

A number of recent studies have focused on the range of evapotranspiration rates for various wetlands types and on the importance of evapotranspiration during hot periods. For instance, meadow dominated by *Typha latifolia* and *Scirpus californicus* evaporated 3 – 4 mm $d^{-1}$ in summer in California, USA (*Goulden et al., 2007).* The evapotranspiration rate in reed beds were observed between 0.5 and 5.5 mm $d^{-1}$ in Kent, UK; and between 0.1 and 5.8 mm $d^{-1}$ in the Liaohe Delta, Northern China (*Peacock and Hess, 2004*; *Zhou and Zhou, 2009*). *Acreman et al. (2003)* reported evapotranspiration rates from reed bed exceeded that of wet grassland by 14 % over a five-month period. *Herbst and Kappen (1999)* indicated the unordinary values of evapotranspiration recorded up to 20 mm $d^{-1}$ for reed beds in northern

Germany. Evapotranspiration rates from wetlands planted with *Phragmites australis* exceeded 10 mm $d^{-1}$ in sub-tropical Australia (*Headley et al., 2012*). For the Czech Republic, evapotranspiration rates from wetland dominated by *Phragmites australis* was reported to be between 6.9 – 11.4 mm $d^{-1}$ (Květ, 1973). *Rejšková et al. (2010)* indicated that evaporative losses from a temperate wetland dominated by *Phalaris arundinacea* reached values of 5.3 – 5.9 mm $d^{-1}$ on hot sunny days. The annual sums of wetland evaporative losses can be ranged between 1100 and 1600 mm $yr^{-1}$ (*Raisin, 1999*; *Wiessner et al., 1999; Lafleur et al., 2005*).

### 1.6.3. Agricultural evapotranspiration

Agriculture is by far the largest water-use consumer, accounting for about 70 % of water used worldwide (*Billib et al., 2009*). Water movement in agriculture can be classified into three categories: 1) agriculture and the aquatic systems (runoff change); 2) agriculture and the soil (groundwater); and 3) agriculture and the atmosphere (evapotranspiration) (*Gordon et al., 2008*).

Evapotranspiration is a hot topic in agricultural management, such as arable water distribution, monitoring of crop growth, drought detection and assessment (*Allen et al., 1998*). Previous studies reported a range of evapotranspiration rates and its environmental control for different agricultural lands (e.g. *Ryszkowski and Kedziora, 1987; Baldocchi, 1994; Jara et al., 1998; Inman-Bamber and McGlinchey, 2003; Watanabe et al., 2004; Burba and Verma, 2005; Eulentstein et al., 2005; Merta et al., 2006; Li et al., 2009; Attarod et al., 2009*).

Salinity is the most widespread problem in agriculture and can impact both photosynthesis and evapotranspiration of crops (*Shalhevet, 1994; Mojid et al., 2012*). Salts in groundwater reduce evapotranspiration by making soil water less available for extraction by root of plants (*Heidarpour et al., 2009*). Irrigation is necessary when plants cannot satisfy all their water needs through natural precipitation but it is the major supplier of salts into the soil. Additionally, salts may be introduced through application of fertilizers (*Grattan and Grieve, 1999*). High salt concentration in groundwater reduces the ability of crops and plants to take up water and leads to lower yields. It is common in arid and semi-arid areas particularly in countries of Asia and Africa where evapotranspiration exceeds annual precipitation (*Pitman and Lauchli, 2002*).

Evapotranspiration is a very effective way of plant cooling. The plant cooling effect mainly depends on vegetation types and water availability. *Hesslerová et al.*

*(2013)* analysed the surface temperature of seven localities in a temperate landscape – harvested meadow, wet meadow, alder stand, mixed forest, bare field, open surface water and asphalt surface. They found that surface temperature and consequently cooling of plants correlates with the intensity of the incoming solar radiation. In early morning the difference in surface temperature between the localities was insignificant. The distinction in temperature increased with the increasing solar radiation. During the high solar radiance the distinct differences reached of about 20 $^{o}$C between forest and asphalt localities. The obvious role of green vegetation was shown from comparison of a wet meadow and a harvested meadow where difference at noon was approximately of 13 $^{o}$C.

### 1.6.4 Changes in natural vegetation patterns

Change of pattern of vegetation from sustainable ecosystems like forests and wetlands into unsustainable like agricultural lands may dramatically modify local climate conditions (*Shukla et al., 1990; Tinker et al., 1996; Ryszkowski and Kedziora, 2008; Hesslerová and Pokorný, 2010).* Replacement of forest ecosystems by arable lands reduces evapotranspiration and precipitation and increases the annual runoff (e.g. *Bosch and Hewlett, 1982; Bonan et al., 2002*).

The problem of deforestation from this point of view was discussed by *Bosch and Hewlett (1982), Shukla et al. (1990), Norbe et al. (1991), Dunn and Mackay (1995), Sud et al. (1996), Zhang et al. (2001), Gordon et al. (2003), Rost et al. (2008).* According to *Gordon et al*. (2003), reduction of forest cover decreased global vapour flows by 4 %. The most sensitive regions that felt the reduction are Southeast Asia, South-Western and Central Africa, the areas around the edge of the Amazon basin, northern South America, and the temperate forest regions of the eastern United States and Central and Western Europe. *Rost et al. (2008)* investigated the effect of human land management on terrestrial water cycle. They found that the effect of land cover change from forest to agriculture led to reduction in interception loss by 26 %, while river discharge increased by 7 % as compared to the wildwood. Furthermore, their data indicated that deforestation decreased transpiration by 11 % due to the shorter growing seasons and change in rooting depth of plants.

*Costa and Foley (1997)* pointed out that annual average evapotranspiration decreases by ~ 12 % as a consequence of replacement forests, woodlands and savannas by grasslands in the Amazonian basin. The conversion of tropical forest into a pasture increases surface temperature of about 2.5 $^{o}$C and reduces evapotranspiration of approximately 30 %. If the Amazonian region was totally converted from a forest to grassland, the continent would permanently show drier climate (*Shukla et al., 1990*).

The hydrologic effects of land use changes, dams, and irrigation led to decrease of evapotranspiration by over the past 300 years in North America and Asia. The simulation research showed that the expansion of arable lands for the period of 1700 – 1992 has resulted in increases in annual runoff by 2.5 % and 6 % for North America and Asia, respectively (*Haddeland et al., 2007).* Expansion of agricultural land will result in reduction of total evapotranspiration by 2.5 % and in increase of river discharge by 3.9 % in 2046 – 2055 (*Rost et al., 2008*).

Changes in vegetation cover can alter precipitation *(Feddema et al., 2005; Makarieva and Gorshkov, 2007).* In semi-arid and arid regions reduction of vegetation leads to a decrease of precipitation and increase of temperature (*Schlezinger et al., 1990; Shukla et al., 1990*). *Andrich and Imberger (2013)* focused on the effects of land-use change on rainfall in southwest Western Australia. The authors considered two hypotheses regarding the inland rainfall decline in southwest Western Australia: (1) the decline of rainfall had been caused by global meteorological conditions affected by ‘natural variation’ and anthropogenic greenhouse gas emissions affecting ocean temperatures and (2) the decline of rainfall caused by land clearing. Based on the available data, the authors declared that the first hypothesis was not confirmed as Northern coastal rainfall had been stable from 1984 to 2010 whereas the inland precipitation had decreased. The results suggested that the dominant factor affecting rainfall decline in this region was the land-use change. The results indicated that deforestation in the coastal area causing native vegetation reduction from 60 % to 30 % correlated with a 15 % to 18 % decline in annual winter rainfall relative to rainfall at the coast. It was also found that land-use change reduced stream flow by around 300 GL year$^{-1}$.

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# 2. Method and study area description

## 2.1 Location and climate

The research was conducted in a lowland within the Třeboň Basin Biosphere Reserve (TBBR) in the southern part of Bohemia region ($49^{o}$ $05^{'}$ N, $14^{o}$ $46^{'}$ E), near the Czech-Austrian border (Fig. 4). The Třeboň Basin is a flat bottom or slightly wavy area (410 – 470 a.s.l.) with domination mainly clay and sandy soil. The TBBR (700 $km^2$ in total) has about 500 fishponds total area of which reaches 7 500 ha which were constructed mostly in the $16^{th}$ century. The high diversity of habitats and species of this region was recognized in the UNESCO Man and Biosphere Programme in 1977 and declared as a Protected Landscape Area in 1979 (*Květ et al., 2002*).

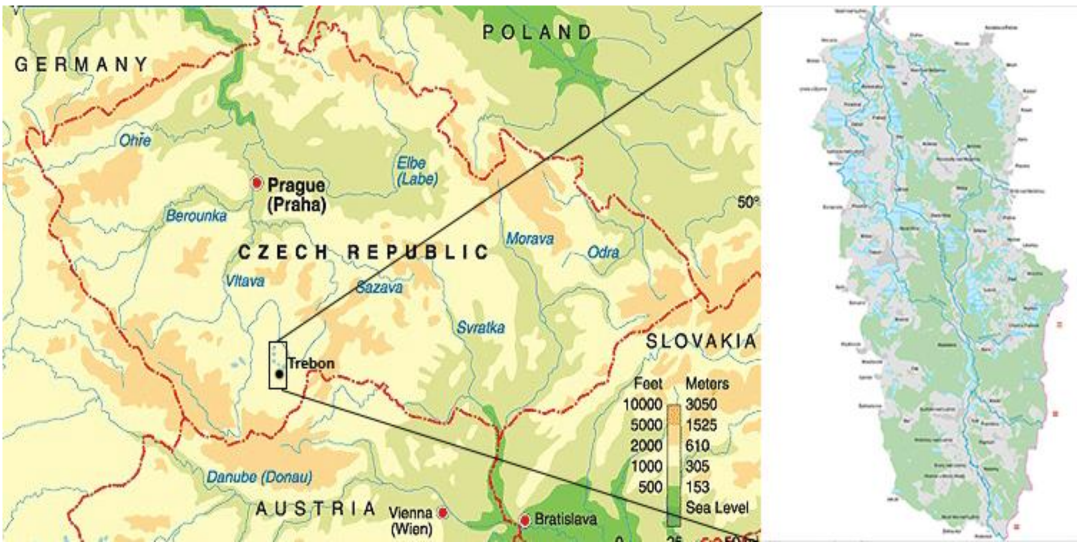


**Figure 4.** The geographic position of research locality Třeboň Basin Biosphere Reserve within the Czech Republic.

The Třeboň Basin belongs to the moderately warm and mildly moist region of the Central European temperate zone. The mean annual temperature is about of 7.8 $^{o}$C, with January the coldest month at – 2.8 $^{o}$C and July the warmest at 18 $^{o}$C. The average temperature during the vegetation period (April – September) is up to 14 $^{o}$C. The mean annual precipitation is about 600 – 650 mm, with the most rains falling in July during summer storms. The average yearly sunshine is about 1750 hours per year, with a maximum of 240 hours in June (*Květ et al., 2002*).

## 2.2 Meteorological observation data

To determine the effect of vegetation on local climate, complete series of data during the vegetation seasons (2008 - 2012) when evapotranspiration was high were used for the analysis. Data were collected from several localities: an artificial concrete land, a wet meadow, an open water body, a pasture and widespread crops in the region: a winter barley field, a cornfield, a wheat and a rapeseed. The detailed description of all localities is given in Chapters 3 – 6.

The dataset, used for the estimation of energy fluxes, evapotranspiration and micrometeorological conditions, was obtained from automatic meteorological stations M4016 placed at the study locations (Fig. 5). The details of measurements, equipment and sensors' accuracy are summarised in Table 4.

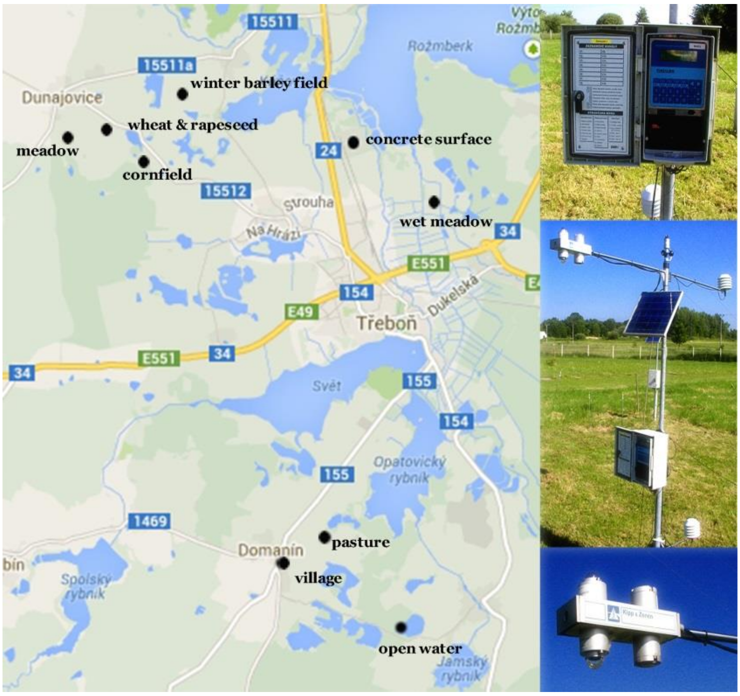


**Figure 5.** The satellite image of the experimental ecosystems and the example of meteorological station used for continuous monitoring on individual sites.

The dataset of air temperature, relative air humidity, global radiation, soil temperature and atmospheric pressure were measured every 1 s and logged automatically as 10-min average intervals. Wind speed and wind direction were recorded automatically at 10-min intervals, with an integration time of 30 s.

**Table 4.** Details of permanently deployed instruments at the localities, measuring the height, the units and the instrumental brand.

| Meteorological variable | Height (m) | Units | Instrumental brand |
|---|---|---|---|
| Air temperature | 0.3; 2 | °C | T/RH probes, accuracy +/- 0.3 °C |
| Relative air humidity | 0.3; 2 | % | T/RH probes, accuracy +/- 2 % |
| Temperature at the soil surface | 0; -0.1; -0.2 | °C | Pt 100, accuracy +/- 0.15 °C |
| Shortwave (global) radiation | 2 | W $m^{-2}$ | CM3 pyranometers, Kipp & Zonen, the Netherlands, spectral range from 310 to 2800 nm, accuracy +/- 5 % |
| Wind speed and wind direction | 2 | m $s^{-1}$ | W12 TM Praha, Czech Republic |
| Atmospheric pressure | 2 | mbars | PTB 100 A Vaisala sensor, Finland, accuracy +/- 0.3 mbars |
| Volumetric content of liquid water in the soil | -0.05 | % | Wirrib, AMET, Czech Republic, accuracy +/- 0.01 $m^3.m^{-3}$ |
| Incoming longwave radiation | 2 | W $m^{-2}$ | CNR1 Net radiometer, Kipp & Zonen, the Netherlands, spectral range 5 to 50 µm, accuracy +/- 10 % |

The equipped unit includes data logger, telemetric station with build-in GSM module, programmable control automat and multiple flows. Electrical signals have been converted by data logger from equipment to scientific units; perform calculation and data outputs for later analysis. Transmission of measured and archived data from automatic weather stations was occurred by GPRS transfers in GSM net and store data to the internet server. The described meteorological stations were purchased and operated by ENKI (*Jirka et al., 2012*).

## 2.3 Methods

Different techniques have been derived for estimating actual evapotranspiration (*Drexler et al., 2004; Merta et al., 2006; Pauwels and Samson, 2006; Verstraeten et al., 2008*). All these methods can be separated into two fundamental groups: direct/water budget methods (e.g. pan evaporation, weighing lysimeter and soil moisture depletion) and indirect/water vapour transport (e.g. eddy covariance, Bowen ratio energy budget) (*Held et al., 1990; Bausch and Bernard, 1992; Yrisarry and Naveso, 2000; Unlu et al., 2010).* The water budget methods are based on direct measurement of water loss from a container. The principle of water vapour transport methods is based on measuring the flow of water vapour into the air using meteorological equipment. In more cases, indirect methods are recommended to use for estimation of evapotranspiration but direct methods can be used when more accurate measurements of evapotranspiration are necessary *(Gulliver et al., 2010).* Nevertheless, all methods have strengths and weaknesses *(Allen et al., 2011).* The summary of advantage and disadvantage of more popular evapotranspiration methods is presented in Table 5.

**Table 5.** Summary of strengths and weaknesses of evapotranspiration measurement methods (according to *Johnson and Odin, 1978*; *Lal, 1991; Rana and Katerji, 2000; Drexler et al., 2004; Verstraeten et al., 2008; Allen et al., 2011*)

| | Measurement techniques | Advantage | Disadvantage |
|---|---|---|---|
| Water budget methods | Pan evaporation | Simple easy and cheap to construct / easy to use and clean | Can give erroneous data, water can heat up, complex wind speed effects associated with lip. Water in pan stores and releases water differently than plants |
| | Weighing lysimeter | Possibility of mechanical calibration and total automation | Expensive/ stationary/ difficult to maintain or to reconstruct original soil profile characteristics / difficult to maintain exact field conditions for vegetation and soil water / difficult to reproduce rooting characteristics / difficult to detect the presence of low levels of water stress / difficult to accurately measure ET from non-potential vegetation. |
| | Soil water balance | Inexpensive/ soil moisture simple to be evaluated with gravimetric method | Large spatial variability / difficult to be applied when drainage and capillary rising are important / difficult to measure soil moisture in cracked soils/ inaccurate / changes of water amount in plant tissue are not measured |
| Water vapour transport methods | Bowen ratio balance method | Low cost in comparison with eddy covariance / the aerodynamic data isn't required / simple measurement/ suitable for tall crops | Requires medium to large fetch /accuracy depends on the representativeness and accuracy of $R_n$ and G / measurements of temperature and water vapour pressure deficit must be unbiased |
| | Eddy covariance (EC) | Automated / non-destructive, continuous direct sampling of the turbulent boundary layer / | Requires substantial fetch/ requires consistently horizontal flowlines / problems in eddy formation / instrumentation is relatively fragile and expensive / difficult software for data acquisition |

In this research the Bowen ratio balance method (BREB) was provided for assessment of energy fluxes, evaporative fraction and etc. This method is widely used to determine the sensible and the latent heat fluxes over vegetation, forests, arable lands, bare soil and water bodies (*Bremer and Ham, 1999*; *Burba et al., 1999*; *Todd et al., 2000; Scott et al., 2003; Lee et al., 2004; Peacock and Hess, 2004; de Teixeira et al., 2007; Perez et al., 2008; Savage et al., 2009*).

The distinctive feature of the BREB method is its simplicity: enabling to compute the BREB by measuring of air temperature and water vapour pressure at two different levels above the canopy within the adjusted boundary layer. To estimate the balance between sensible and latent heat fluxes the Bowen ratio was expressed in the form (*Bowen, 1926*) -

$$\beta = \frac{H}{LE} = \gamma \frac{T_s - T_a}{e_s - e_a} \tag{14}$$

by approximating the fluxes by the temperature and humidity gradient. In equation, $\gamma$ is the psychrometric constant (kPa $K^{-1}$), $T_s$ - $T_a$ is the temperature differences between the air temperature at 2 m above surface and air temperature of canopy surface (0.3 m), $e_s$ - $e_a$ – is the difference of water vapour pressure (kPa) in these levels.

The BREB method is based on the energy balance equation, which for uniform surfaces can be simplified to -

$$R_n = LE + H + G \tag{15}$$

where: $R_n$ is the net radiation, H is the sensible heat flux, LE is the latent heat flux and G is the ground heat flux. All units of this equation are expressed in $Wm^{-2}$.

The following sign conversion has been used through the thesis: the positive net radiation is directed to the surface (gain energy) and the negative net radiation is gone away from the surface (loss of energy). Positive latent and sensible heat fluxes indicated that energy tends away from the earth surface to the atmosphere and vice versa. Whilst, positive ground heat showed energy transportation from the earth surface into the ground. The individual components of the equation 15 are presented in more detail below.

The net radiation was computed from shortwave and longwave energy balance by using the following formula -

$$R_n = R_{s\downarrow} - R_{s\uparrow} + R_{L\downarrow} - R_{L\uparrow} \tag{16}$$

where:

longwave radiation emitted by surface ($R_{L\uparrow}$, W $m^{-2}$) was computed using by the equation (*Brutsaert, 1982*) –

$$R_{L\uparrow} = \varepsilon\sigma(T_s + 273.16)^4 \tag{17}$$

where: ε is emissivity, $T_s$ – surface temperature and σ – Stefan-Boltzmann constant (W $m^{-2}$ $K^{-4}$). Emissivity was set 0.98 for all stations. Different methods for computing longwave radiation have been described in Chapter 1.1.2.

The incoming longwave radiation ($R_{L\downarrow}$) was calculated using the same expression, but the temperature of air ($T_a$) is used.

Albedo was obtained as a ratio between reflected ($R_{s\uparrow}$) and incoming ($R_{s\downarrow}$) shortwave radiation –

$$\alpha = \frac{R_{s\uparrow}}{R_{s\downarrow}} \tag{18}$$

Ground heat flux was given by Fourier's law of heat conduction using the vertical method (*Oke, 1987; Montheith and Unsworth, 1990*) –

$$G = k\frac{T_c - T_{0.2}}{z_c - z_{0.2}} \tag{19}$$

where: *k* is the thermal conductivity of soil (W $m^{-1}$ $K^{-1}$), $T_c$ and $T_{0.2}$ are the soil temperatures in the depth $z_c$ and $z_{0.2}$.

Thermal conductivity depends on soil moisture content, mineral composition, dry density of soil, particle types, size distribution and temperature of the soil (*Wierenga et al., 1969*). In more detail the thermal conductivity is written in Chapter 1.2.

Ground heat flux of the concrete surface was computed using the empirical equation due to the difficulty in exact positioning and sealing of thermometers in concrete (*Brutsaert, 1982*) –

$$G = cR_n \tag{20}$$

where: Rn is the net radiation, c is the constant value (for bare soil = 0.3).

The essential part of BREB is enabling to calculate the temperature and vapour vertical differences and subsequently Bowen ratio itself. But prior this step, the relative humidity should be converted into vapour pressure. The water vapour pressure *e* (kPa) was computed for the 2 m above surface ($T_a$, $^oC$) and canopy level ($e_a$ and $e_c$, respectively) by the formula–

$$e = \frac{RH \cdot e_w}{100} \tag{21}$$

where: *RH* – is the relative air humidity, $e_w$– is the saturation pressure of saturated water (kPa) in the air and at the canopy level, respectively.

The values of saturation pressure were obtained using the modify empirical Magnus-Teten´s equation (*Buck, 1996*) -

$$e_w = 0.61121exp\left(\left(18.678 - \frac{T}{234.5}\right)\left(\frac{T}{257.14+T}\right)\right) \tag{22}$$

where: *T* is temperature in 2 m above surface ($T_a$) and at canopy level ($T_c$), respectively.

Water vapour pressure deficit (VPD, kPa) was computed for 2 m above surface level by the formula –

$$VPD = e_w - e_a \tag{23}$$

From equations 14 and 15 the latent heat and the sensible heat fluxes were calculated by (*Bowen, 1926*):

$$LE = \frac{R_n - G}{1+\beta} \tag{24}$$

$$H = \frac{\beta(R_n - G)}{1+\beta} \tag{25}$$

BREB method may infrequently produce large errors due to its theoretical limitations. The errors by using BREB has been studied by *Ohmura, 1982; Nie et al., 1992; Malek and Bingham, 1993; OrtegaFarias et al., 1996; Unland et al., 1996; Prueger et al., 1997; Foken, 2008).* The vagueness of the method mostly occurs when the Bowen ratio (β) approaches to -1 since the values in eq. 24 and 25 approach to infinity and these equations lose their physical significance (*Ohmura, 1982*; *Prueger et al., 1997; Perez et al., 1999; Perez et al., 2008; Xing et al., 2008).* For this case, the Bowen ratio values within interval $-1.3 < \beta < -0.75$ has to be rejected (*Tanner et al., 1987; OrtegaFarias et al., 1996; Unland et al., 1996; Savage et al., 2009*). The next erroneous situation occurs when fluxes change their sign and depend on the limitation of accuracy instruments. This condition may occur during sunrise and/or sunset (*Prueger et al., 1997*), under cloudy conditions (*Pruitt et al., 1987)* or with precipitation *(Ohmura, 1982). Unland et al. (1996)* also found that the latent and the sensible heat fluxes are almost equal but has opposite direction when the relation $|1+\beta| < 0.3$ occurs. This is indicating that the BREB quantify the fluxes from the surface. To determine the correct sign, *Ohmura (1982)* indicated that reliable data should satisfy the following criteria:

$$\text{if } R_n - G > 0 \text{ then } \partial T + {}^{\partial e}/_{\gamma} < 0 \tag{26}$$

$$\text{if } R_n - G < 0 \text{ then } \partial T + {}^{\partial e}/_{\gamma} > 0 \qquad (27)$$

The data should be removed if the inequalities are not fulfilled. *Gavin and Agnew (2004)* also excluded values for periods when β > 10 and ∂e is zero or very small. Another requirement for using BREB method is that the wind should move over a sufficient distance from vegetation or terrain before it reaches detectors (*Campbell, 1977*). Therefore, it is recommended to increase the vertical distance between the sensors as much as possible *(Foken, 2008).* All these recommendations have been implemented.

Evapotranspiration rate (ET, g $m^{-2}$ $s^{-1}$) was computed from latent heat flux in the following way -

$$ET = \frac{LE}{L_e} \qquad (28)$$

where: $L_e$ is the latent heat of evaporation (J $g^{-1}$). Daily sums of ET were expressed in mm.

The ratios of the energy fluxes to the net radiation were used for analysis of the study sites behaviour. Furthermore, evaporative fraction (EF, rel.) was used for analysis of available energy amount consumed for evaporation of water (*Lhomme and Elguero, 1999; Suleiman and Crago, 2004; Gentine et al., 2007).* EF was computed using equation -

$$EF = \frac{LE}{R_n - G} \qquad (29)$$

To determine parameterisation of fluxes between the land surface and the atmosphere the decoupling coefficient, surface and aerodynamic resistance have been evaluated.

Dimensionless decoupling coefficient (Ω) was used for analysis of the coupling between vegetation and the atmosphere. It was computed using equation (*Jarvis and McNaughton, 1985*) -

$$\Omega = \frac{LE}{LE_p} = \frac{\Delta + \gamma}{\gamma + 1\left(\frac{r_c}{r_a}\right)} \qquad (30)$$

where: $LE_p$ is flux of potential evaporation (W $m^{-2}$), Δ is the ratio between the saturation water vapour pressure gradient and the temperature gradient (kPa °$C^{-1}$), $r_c$ is bulk surface resistance (s $m^{-1}$) and $r_a$ is aerodynamic resistance (s $m^{-1}$).

According to *Jarvis and McNaughton (1985),* the decoupling coefficient (factor) describes how closely the saturation deficit at the canopy (or leaf) surface is linked to that of the air outside the canopy boundary layer. It describes sensitivity of evaporation to stomatal or surface conductance (*Jones, 1992*). The decoupling coefficient is a dimensionless factor that assumes values range from 0 (transpiration is mainly controlled by canopy conductance and vapour pressure deficit) to 1 (transpiration is mainly controlled by available energy).

The transfer of water vapour and heat between canopy and air is determined by aerodynamic resistance, which depends on wind speed and vegetation structure. Aerodynamic resistance was computed using the Thom equation (*Thom, 1975*) –

$$r_a = \frac{\left[ln\left(\frac{z-d}{z_{0m}}\right)-\psi_m(\varsigma)\right]\left[ln\left(\frac{z-d}{z_{0h}}\right)-\psi_h(\varsigma)\right]}{Uk_v{}^2} \quad (31)$$

where: $z$ is the height where measurement where taken to describe the physical conditions of the air (m), $d$ is the displacement height (m), $z_{0m}$ and $z_{0h}$ are aerodynamic roughness parameters respectively for the momentum and for heat transfer (m), $\Psi_m(\varsigma)$ and $\Psi_h(\varsigma)$ are stability coefficients respectively for momentum and for heat transfer (unitless), where $\varsigma$ is the Monin-Obukhov stability parameter, U is wind speed (m $s^{-1}$) and $\kappa_v$ is the von-Kármán constant.

The displacement height was computed as (*Allen et al., 1998*) –

$$d = \frac{2}{3}h \quad (32)$$

where: $h$ is vegetation height. $z_{0m}$ and $z_{0h}$ were computed in simple way according to *Allen et al. (1998)* –

$$z_{0m} = 0.123h \quad (33)$$

$$z_{0h} = 0.1z_{0m} \quad (34)$$

Stability conditions are characterized by the Monin-Obukhov stability parameter as –

$$\varsigma = \frac{z}{L_O} \quad (35)$$

where: $L_o$ is the Monin-Obukhov length (m) computed by means of the following equation (*Kalma, 1989*) –

$$L_o = \frac{u_*^3 pc_p(T_a+273.16)}{kgH} = \frac{u_*^2(T_a+273.16)}{kgT_*} \quad (36)$$

where: $u_*$ is wind friction velocity (m $s^{-1}$), $\rho$ is air density (kg $m^{-3}$), $c_p$ is specific heat at constant pressure (J $kg^{-1}$ $K^{-1}$), $g$ is acceleration due to gravity (m $s^{-2}$) and $T_*$ is a scaling parameter in the boundary layer, analogous to the friction velocity (°C). The wind friction velocity was computed by equation (*Kalma, 1989*) –

$$u_* = \frac{kU}{ln\frac{z-d}{z_{0m}} - \Psi_m(\varsigma)} \quad (37)$$

and $T_*$ using formula (*Kalma, 1989*) –

$$T_* = \frac{k(T_a - T_s)}{ln\frac{z-d}{z_{0h}} - \Psi_h(\varsigma)} \quad (38)$$

Stability parameters $\Psi_m(\varsigma)$ and $\Psi_h(\varsigma)$ were computed for stable ($\varsigma \geq 0$) and unstable or near neutral stable atmosphere conditions ($\varsigma < 0$) (*Foken, 2008*). Stability parameters for unstable and near neutral atmospheric condition were computed in the following way (*Liu et al., 2007*)

$$\Psi_m(\varsigma) = 2ln\left(\frac{1+x}{2}\right) + ln\left(\frac{1+x^2}{2}\right) - 2\arctan(x) + \frac{\pi}{2} \quad (39)$$

$$\Psi_h(\varsigma) = 2ln\left(\frac{1+x^2}{2}\right) \quad (40)$$

where:

$$x = (1 - 16\varsigma)^{0.25} \quad (41)$$

The stability parameters for stable atmospheric condition ($\varsigma \geq 0$) were computed according to *Beljaars and Holstag (1991)* –

$$\Psi_m(\varsigma) = -\left[a\varsigma + b\left(\varsigma - \frac{c}{d}\right)\exp(-d\varsigma) + \frac{bc}{d}\right] \quad (42)$$

$$\Psi_h(\varsigma) = -\left[\left(1 + \frac{2a}{3}\varsigma\right)^{1.5} + b\left(\varsigma - \frac{c}{d}\right)\exp(-d\varsigma) + \left(\frac{bc}{d} - 1\right)\right] \quad (43)$$

where: $a$ = 1, $b$ = 0.667, $c$ = 5, $d$ = 0.35.

Bulk surface resistance was calculated according to the Penman–Monteith equation (*Jackson et al., 1981; Wallace, 1995*); we used the following modification -

$$r_c = r_a \frac{\frac{\gamma r_a(R_n - G)}{pc_p} - (T_c - T_a)(\Delta + \gamma) - VPD}{\gamma\left[(T_c - T_a) - \frac{r_a(R_n - G)}{(pc_p)}\right]} \quad (44)$$

All stability parameters, the Monin-Obukhov length, the friction velocity and the scaling parameter $T_*$ were computed by means of the iterative procedure proposed by *Itier (1980*, cited in *Kalma, 1989*). For more details see *Kalma (1989)* and *Liu et al. (2007).*

## 2.4 Processing data

Several statistical routines were used in the research. The dataset was converted and imputed into IBM SPSS software to compute energy fluxes and meteorological variables.

For providing the quality of the data, careful selection was performed to identify and reject erroneous data. Some conditions were made to the raw data. At first, the incorrect dataset was removed and replaced through linear interpolation. Next, the missing data were replaced by linear interpolation of data from the nearest values. And finally, the row dataset was smoothed once by applying a 1:2:1 smoothing algorithm to the all meteorological values.

The data were arranged into the following three groups, according to the amount of total incoming daily shortwave solar irradiance: overcast (0 – 3,000 Wh $m^{-2}$); cloudy (3,000 – 6,000 Wh $m^{-2}$) and clear (over 6,000 Wh $m^{-2}$).

The significance of the seasonal changes in measured and calculation parameters was analysed by stepwise linear regression. As stated above, all statistical analysis was performed using IBM SPSS 18.0 program for Windows (SPSS Inc., Illinois, USA) Linear regression analysis was used to assess existing relationships between the meteorological parameters, energy fluxes and at time at different ecosystems. A confidence interval of 95 % was used and $p$ values $< 0.05$ were considered significant.

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# 3. Comparison of reflected solar radiation, air temperature and relative air humidity in different ecosystems: from fishponds and wet meadows to concrete surface

This chapter is based on Paper I.:
Huryna H, Pokorný J (2010) Comparison of reflected solar radiation, air temperature and relative air humidity in different ecosystems: from fishponds and wet meadows to concrete surface. In Vymazal J (ed): Water and nutrient management in natural and constructed wetlands. Springer, The Netherlands, 308–326

**Abstract**

Incoming and reflected shortwave solar radiation, temperature and relative humidity of air at 0.3 and 2 m were measured continuously during the vegetation season (from April 1 to September 30, 2008) at six different sites: fishpond, wet meadows, pasture, barley field, village and concrete surface. All six sites are located in the basin of Třeboň Biosphere Region, Czech Republic. For data evaluation the 183 days of the vegetation season were divided into the three classes according to the amount of incoming solar energy they received. The albedo of water was two times lower than the albedo of the field and three times lower than the albedo of the concrete surface, which also had the highest average temperature. The lowest average temperature was measured at the wet meadows. The lowest average daily amplitude in temperature (difference between daily maximum and minimum) was measured at the fishpond. The highest difference (3.58 $^{o}$C) in average temperature at 0.3 m among sites was found for wet meadows and concrete surface on clear days and 2.49 $^{o}$C for all days of the vegetation season. Daily time courses of relative air humidity on sunny days show the ability of vegetation to buffer air humidity extremes. We conclude that changes of land cover results in changes of average temperature.

**Keywords** Albedo · Local climate · Solar energy income · Wetlands

# 4. Solar energy dissipation and temperature control by water and plants

This chapter is based on Paper II.:
Pokorný J, Brom J, Čermák J, Hesslerová P, Huryna H, Nadezhdina N, Rejšková A. (2010) Solar energy dissipation and temperature control by water and plants. *International Journal of Water 5*(4), 311-337

## Abstract

Ecosystems use solar energy for self-organisation and cool themselves by exporting entropy to the atmosphere as heat. These energy transformations are achieved through evapotranspiration, with plants as 'heat valves'. In this study, the dissipative process is demonstrated at sites in the Czech Republic and Belgium, using landscape temperature data from thermovision and satellite images. While global warming is commonly attributed to atmospheric CO2, the research shows water vapour has a concentration two orders of magnitude higher than other greenhouse gases. It is critical that landscape management protects the hydrological cycle with its capacity for dissipation of incoming solar energy.

**Keywords** ecosystems; evapotranspiration; sensible heat; albedo; radiative forcing; temperature variation; remote sensing

# 5. The importance of wetlands in the energy balance of an agricultural landscape

This chapter is based on Paper III.:
Huryna H, Pokorný J, Brom J (2014) The importance of wetlands in the energy balance of an agricultural landscape. *Wetlands Ecology and Management 22*(4), 363 – 381. doi: 10.1007/s11273-013-9334-2

**Abstract**

Energy fluxes, including net radiation, latent heat flux and sensible heat flux were determined on clear days during the vegetative period in 4 types of land cover: wet meadow, pasture, arable field, and an artificial concrete surface. The average net radiation ranged between 123 W m$^{-2}$ at the concrete surface and 164 W m$^{-2}$ at the wet meadow. The mean maximum daytime latent heat ranged between 500 W m$^{-2}$ and 600 W m$^{-2}$, which corresponds to an evapotranspiration rate of about 0.2 g m$^{-2}$ s$^{-1}$ under the prevailing conditions of the wet meadow. The results demonstrated that the wet meadow dissipated about 30 % more energy through evapotranspiration than the field or the pasture, and up to 70 % more energy than the concrete surface. The evaporative fraction indicated that more than 100 % of the energy released by the wet meadow was dissipated through evapotranspiration; this was attributed to local heat advection. Wetland evapotranspiration thus contributes significantly to the cooling of agricultural landscapes; the energy released can reach several hundred MW km$^{-2}$. Wetland evapotranspiration has a double 'air conditioning' effect through which it equalizes temperature differences: 1. Surplus solar energy is bound into water vapour as latent heat; 2. The vapour moves towards cooler portions of the atmosphere where the energy is released. The air-conditioning effect of wetlands plays an important role in mitigating local climate extremes; this ecosystem service tends to be disregarded in relation to other better known wetland functions such as nutrient retention and provision of high biodiversity.

**Keywords** Heat balance · Bowen ratio · Evapotranspiration · Evaporative fraction · Heat advection · Local climate

# 6. Distribution of solar energy in agriculture landscape – comparison between wet meadow and crops

This chapter is based on Paper IV.:
Huryna H, Hesslerová P, Pokorný J, Jirka V, Lhotský R (2014) Distribution of solar energy in agriculture landscape – comparison between wet meadow and crops. In Vymazal J (ed.): The role of natural and constructed wetlands in nutrient cycling and retention on the landscape. Springer, The Netherlands, 103 – 122, doi: 10.1007/978-3-319-08177-9_8

## Abstract

The study examines the impact of plant cover on the water and energy exchange between land and atmosphere in Třeboň Biosphere Reserve, Czech Republic. Energy fluxes, evapotranspiration and evaporative fraction were determined over typical crops of agriculture landscape and compared with fluxes in an adjacent wet meadow. The results show distinct differences in heat and water exchange between these ecosystems. Diurnal average difference in evapotranspiration rates for days with high irradiance at the wet meadow and arable crops ranged from 1.1 mm $d^{-1}$ to 3.4 mm $d^{-1}$. Furthermore, the evapotranspiration differences between C3 (rapeseed) and C4 (cornfield) was about 2.3 mm $d^{-1}$. Analysis of thermovision pictures showed that temperature variation reached of about 9 $^{o}$C between the ploughed field and meadows at the time of maximum intensity of solar radiation. Heat exchange (sensible heat flux) was greater over arable lands while water exchange (latent heat flux) was stronger over the wet meadow. The evaporative fraction displayed that more than 100 % of available energy was released by the wet meadow through evapotranspiration due to advection of dry air from surroundings. Wetlands show equal or even inverse temperature in vertical profile whereas corn and wheat show noticeable higher temperature at soil surface in comparison with plant stand surface. Therefore, we suggest that introduction of wetlands to agricultural land is one of important instrument for management of water and heat balance of the landscape.

**Keywords** Wetlands · Arable Land · Evapotranspiration · Energy Fluxes · Evaporative Fraction · Thermo-Vision Pictures

# 7. Summary and conclusion

Based on the micrometeorological observations, magnitudes and diurnal courses of energy fluxes, air temperature, relative air humidity, surface reflectance and evaporative fraction in some typical continental European ecosystems were estimated for vegetation seasons 2008, 2011 and 2012.

Temperature of the sites was connected with albedo and energy dissipation into latent heat, sensible heat and ground heat fluxes. On overcast days, when most of the incident radiation was reflected or absorbed by clouds, the air temperature at canopy height (0.3 m) on sites with different land covers did not differ. Also on cloudy days, when incident solar radiation reached 3 to 6 kWh $m^{-2}$ $d^{-1}$, the difference of air temperature at midday did not exceed 1 $^{o}$C at 0.3 m height among vegetated sites, showing no deciding role of the plant cover on temperature characteristics. On the clear days, however, when the incident radiation exceeded 6 kW h $m^{-2}$ $d^{-1}$ (maximum flux almost 1000 W $m^{-2}$), mean air temperature differences between vegetated localities measured at the canopy height varied from 4 $^{o}$C to 7 $^{o}$C. Relative air humidity variations reached of about 25 % at 0.3 m among the ecosystems. The average air temperature and relative air humidity differences at 2 m height had small deviations of about 2 $^{o}$C and 8 %, respectively, even on sunny days. High differences in air temperature and relative air humidity at the canopy height showed a strong dependence with the amount of vegetation cover at each locality. While, air temperature and air relative humidity measured at 2 m above ground did not differ significantly between the ecosystems. Wind movement apparently caused sufficient mixing of air, to prevent development of different profiles of temperature and humidity between ecosystems.

The cooling effect of plants on sunny days was studied by a thermovision camera. At the beginning of the growing season, the analysis of thermovision pictures showed only slight differences in radiometric surface temperatures of different land cover types. The most distinct differences in radiometric temperature (up to 9 $^{o}$C) were found between the ploughed field and the meadows. The impact of crop vegetation on temperature extremes was clearly demonstrated on a cornfield locality. Due to the intensive transpiration the difference between the apex of the maize leaves and the lowest part of the stem reached values of 6 $^{o}$C (26 and 32 $^{o}$C, respectively), whereas the surface temperature of bare soil surface reached up to 50 $^{o}$C.

Even though albedo of the concrete sites was of about 6 – 7 % higher than albedo of the vegetated sites (27 % and 20 – 21 %, respectively), the air temperature at the concrete surface was also higher (20 $^{o}$C and 16 – 18 $^{o}$C, respectively). Absence of water and vegetation was during clear days always linked with higher temperatures. The localities with well-functioning vegetation and sufficient water supply were always cooler than the bare soil sites. Plants modified both the reflective characteristics of the surface as well as the energy fluxes.

The net radiation of the concrete surface was much lower than that of vegetated sites mainly due to much higher reflection of solar radiation, as expressed by albedo. Net radiation fluctuated in accordance with incoming shortwave radiation. Mostly at all sites, monthly net radiation peaked at the beginning of the vegetation season and then began to decline steadily. The detailed study of solar energy distribution during vegetation periods has shown that most fluxes were linked with sensible or with the latent heat of evapotranspiration differently. The seasonal variation of energy fluxes over the study periods showed distinct differences among the localities and were clearly associated with changes in vegetation growth. For majority of the sites, the most dramatic changes in dissipation of the net radiation occurred during June and July, when latent heat flux rapidly increased (up to 50 kWh m$^{-2}$) and sensible heat flux decreased (less than 30 kWh m$^{-2}$) due to the increase in vegetation cover and active plant biomass. However, the rapeseed field showed high growth rate in spring which is linked with high evapotranspiration. The ground heat flux considered a small variation throughout the months with low magnitudes (less than 12 kWh m$^{-2}$).

The diurnal variations of energy fluxes were studied during clear days when the thermal differences between the adjacent surfaces were expected to be more pronounced. On clear days averaged net radiation peaked around 430 – 600 W m $^{-2}$. The largest difference between the sites occurred from 10:00 to 15:00, the magnitude of the average available energy at midday was of about 70 W m$^{-2}$ higher at the pasture and the field than at the concrete surface. The diurnal course of latent heat flux showed different patterns at different sites with highest values at the wet meadow (up to 630 W m$^{-2}$) and lowest at the concrete surface (less than 230 W m$^{-2}$). Conversely, at the concrete surface sensible heat was the dominant flux measured, while at all vegetated sites relatively low values of sensible heat flux were observed. At most sites the maximum latent heat flux occurred later than the peak in the sensible heat flux. This gap is connected to the peak in air temperature and vapour pressure deficit that both occur after the peak in the available energy. The highest diurnal ground heat flux represents only a low portion of available energy, on average it was only about 7 % of the net radiation (below 50 W m$^{-2}$). The maximum diurnal evapotranspiration for vegetated sites ranged between 163 and 265 mg $m^{-2}$ $s^{-1}$. The hourly energy fluxes measured on hot days at the canopy height level showed that well water-supplied vegetation used 3 times less energy for heating the surrounding air when compared with the concrete surface. The monthly and overall evapotranspiration rates from the seven localities during three growing seasons showed that arable land demonstrated less variability and lower magnitude of evapotranspiration than the wet meadow (Table 1).

**Table 1.** Monthly mean evaporative fraction (EF = LE/($R_n$-G)) and evapotranspiration (mm) of some typical ecosystems in the Czech Republic, evaluated only for clear days

| | | 2008 | | 2011 | | | 2012 | |
|---|---|---|---|---|---|---|---|---|
| | | barley field | wet meadow | meadow | cornfield | wheat | meadow II | rapeseed |
| Evaporative fraction | May | 0.69 | 0.56 | 0.52 | 0.17 | 0.79 | 0.51 | 0.81 |
| | June | 1.12 | 1.08 | 0.60 | 0.72 | 0.83 | 0.54 | 0.79 |
| | July | 0.62 | 1.49 | 0.60 | 0.72 | 0.45 | 0.66 | 0.64 |
| | August | 0.56 | 1.42 | 0.59 | 0.77 | 0.56 | 0.56 | 0.53 |
| | **Overall** | **0.75** | **1.14** | **0.58** | **0.59** | **0.66** | **0.57** | **0.69** |
| Evapotranspiration | May | 59 | 40 | 56 | 19 | 76 | 53 | 89 |
| | June | 84 | 72 | 64 | 78 | 91 | 44 | 54 |
| | July | 79 | 91 | 31 | 44 | 19 | 35 | 38 |
| | August | 31 | 69 | 19 | 13 | 13 | 51 | 48 |
| | **Overall** | **253** | **272** | **169** | **154** | **200** | **183** | **229** |

Evaporative fraction showed that during dry spells, at the wet meadow more than 100 % of available energy was transformed through evapotranspiration. The fact that latent heat of vaporization exceeded net radiation indicates that sensible heat was probably drawn from the air outside of wet meadow canopy and here it was used for evapotranspiration. Excess of solar energy was associated with local advection originating in the temperature difference between the adjacent areas occurring during sunny and hot days. The results demonstrated that the wet meadow converted about 30 % more energy than the arable lands and as high as 70 % more to the latent heat of evapotranspiration than the concrete surface. During vegetation seasons the diurnal difference in evapotranspiration rates between the wet meadow and the concrete surface reached 4 mm $d^{-1}$. The diurnal average differences in evapotranspiration rates at the wet meadow and arable crops ranged from 1.1 mm $d^{-1}$ to 3.4 mm $d^{-1}$. The comparison of the ratios of latent heat and sensible heat to net radiation showed that wet meadows and arable lands used 90 % and 65 % (average for crops) of net radiation for evapotranspiration and only 3 % and 16 % of energy for air heating (sensible heat).

Based on the results presented in this thesis it can be concluded that the plant cover structure can change the meteorological conditions through (1) partitioning solar energy into energy fluxes of evapotranspiration or air heating and (2) horizontal exchange of heat between adjacent ecosystems in advection process. Plants show the ability to mitigate climate changes on regional level because they control the surface characteristics of the landscape. Vegetation well supplied with water, like that in a wetland, is able to dampen vertical exchange of sensible energy and enhance the latent heat flux between the surface and the atmosphere, while the bare surface is the factor intensifying these fluxes. Thus, over bare soil the sensible heat fluxes dominate over the latent heat flux resulting in heating up of both air and ground surface, via the "heater" effect. Over the wet meadow, the substantial latent heat flux does not allow the heating up of the air and soil. The arable lands also evaporate water but namely at intermediate growth stages when accumulation of plant biomass occurs. However, the cornfield, wheat and barley fields were treated with herbicides that led to areas of bare soil where no weeds exist. Bare soil under crop plants has high temperature. The air heated by a soil goes upwards and eliminates water vapour. Crops lose water during sunny days, furthermore crop fields are drained. We assume that water vapour from crop plants rises faster than from wet meadow which has dense vegetation and therefore lower temperature at ground. Based on results of the dissertation can be concluded that vegetation equalizes the temperature differences in landscape thereby affecting on a microclimate and intensity of the energy exchange.  In order to neutralise undesirable climatic trends, people have to design and control the structure of agricultural landscape so that it functions like natural vegetation (forest, wet meadows) with inversion of temperature vertical profile.

Further research is needed for better understanding of the local hydrological cycle and the role the vertical structure of vegetation plays in it. There is also a need for empirical proofs of the biotic pump concept on a regional scale, which could be studied by direct monitoring of dew point duration and due point formation in various types of vegetation. Understanding and modelling of these processes and assessment of various vegetation types as possible acceptors of the air humidity acceptors is a very important task. Dew formation is linked with latent heat release, decrease of air pressure and catalytic effect of both condensation nuclei and leaf surfaces and its characteristics are surely deciding for functioning of water relations in many vegetation stands.

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Effect of different types of ecosystems on their meteorological conditions and energy balance components
Ph.D. Thesis Series, No.16. 2014, 171 pp.



Printed in the Czech Republic by Vlastimil Johanus
Edition of 15 copies

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