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The role of water and vegetation in the distribution of solar energy and local climate: a review

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Abstract The role of plants in global climate change discussions is usually considered only in terms of the albedo and sinks/sources of CO_2 and other greenhouse gases. The main aim of this review article is to summarize the entire impact of vegetation on the climate change. It describes quantitatively the energy balance of vegetated surfaces and the effect of vegetation on the hydrological cycle. The distribution of solar energy in the landscape is dealt with in thermodynamic terms. The role of water and plants in the reduction of temperature gradients is emphasized. Papers dealing with the relationship between changes in the landscape cover and regional climates are reviewed, and the fundamental role of wetlands and forests in water cycling is outlined. Positive examples of restoration of dry landscapes, based on rainwater retention and the recovery of permanent vegetation, are described. It is recommended that the direct role of water and vegetation in cooling, reducing temperature and air pressure gradients should be included into all future recommendations for policymakers made by scientists.

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Introduction

Climate change is a widely discussed topic both in the scientific community and general public. The Intergovernmental Panel of the Climate Change (IPCC) is the international scientific body tasked to estimate the risk of climate change caused by the human activities. According to the Fifth Assessment Report (AR5) of the IPCC, the average surface temperature has risen by roughly 0.85°C between 1880–2012 (IPCC 2013: 194). It has been assumed that the main cause of this global warming is the increasing concentration of carbon dioxide in the air (IPCC 2007, 2013). Radiative forcing connected with increased concentrations of the greenhouse gases (GHG) in the atmosphere has been estimated to cause the surface energy input increase of 1–3 $W{\cdot}m^{-2}$ compared with the year 1750. In the next decade, radiative forcing is expected to rise by 0.2 $W{\cdot}m^{-2}$ (IPCC 2007).

One-third of the Earth's land surface is covered by vegetation, but its role in the surface energy budget and consequently climate change is usually overlooked. In recent years, attention has focussed on a possible effect of rapid warming on biodiversity and vegetation. Most studies consider plant stands as a passive subject of external climate

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change caused by changes of GHG in the atmosphere. The role of vegetation in the climate is often reduced to the albedo and a sink/source of CO_2 and other GHG. For example, the effect of the climate change on vegetation was studied by Overpeck et al. (1990), Cramer and Leemans (1993), Bachelet et al. (2001), Theurillat and Guisan (2001), Bakkenes et al. (2002), Lenihan et al. (2003), Burkett et al. (2005) and Ali (2013).

Civilizations have drained wetlands and cut forests for centuries in order to increase the agricultural area for growth of populations. Most crop plants do not tolerate flooding of their roots, therefore flooding must be prevented by further drainage. The effect of land cover changes (namely drainage of wetlands) on temperatures and the water cycle will be discussed on the basis of reviewed literature. We aim to summarize the main trends in assessing the impact of vegetation upon climate change. We describe quantitatively the energy balance of vegetated surfaces and the effect of vegetation on the hydrological cycle. We refer to positive examples of management implications aimed at sustainable land management and an improvement of the water cycle and local climates.

The solar energy balance at the land surface·

The Sun is the primary source of energy for all natural processes of organisms and ecosystems. It radiates energy in the form of short-wave radiation and gives 180,000 TW to the Earth. The energy that falls on the Earth's surface is absorbed, reflected, transformed into long-wave radiation and emitted back to the space. The flux of solar radiation that reaches the top of the atmosphere is called the 'solar constant'. Based on the satellite measurements this value is approximately 1,367 $W \cdot m^{-2}$ (± 20 $W \cdot m^{-2}$; Miller 1981; Brutsaert 1982; Arya 2001; Gueymard 2004). The actual direct irradiance at the top of the Earth's atmosphere varies over the year from 1,321 $W \cdot m^{-2}$ in July to 1,412 $W \cdot m^{-2}$ in January due to the Earth's elliptical 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 black body with the temperature of approximately 6,000 K and emissivity equal to 1. The solar spectrum extends from 0.01 to 4.0 μm with maximum emissivity energy of solar radiation at the wavelength of 0.47 μm (Selinger and McElroy 1965). The spectrum of short-wave radiation is changed when passing the atmosphere as a result of light absorption, reflection or scattering by atmospheric gasses, particles, aerosols and clouds. Each type of molecule has its own position of absorption bands in different parts of the 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 by about 33°C if it had not been warmed up by the absorption and emission processes of these gases (Ahrens 2008).

The Earth's surface temperature is close to 288 K, thus it emits radiation energy at 10 μm in the infrared region of the electromagnetic spectrum. This long-wave radiation near the surface can be divided in two parts: outgoing radiation, which is released from the ground surface and the vegetation, and incoming radiation emitted by the atmosphere (Arya 2001). An estimation of incoming long-wave radiation requires comprehensive knowledge about the air temperature, the air humidity and the properties of the emitting substance, such as gases or aerosols (Brunt 1932; Swinbank 1963; Brutsaert 1975; Idso 1981; Prata 1996; Niemela et al. 2001; Perez-Garcia 2004). Outgoing long-wave radiation is strongly related to the surface temperature and the surface emissivity (Arya 2001).

Ozone mainly absorbs radiation in the ultraviolet part of the electromagnetic spectrum; 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. Though the role of carbon dioxide is important, water vapour is actually the most dominant greenhouse gas, accounting for about 95 % of Earth's greenhouse effect. There is a great difference between the concentration of carbon dioxide (380 ppm), methane (1.76 ppm) and water vapour (several thousands up to 30,000 ppm; Sharma 1994; Sondergard 2009; Harrison and Hester 2014). However, water does not accumulate in the atmosphere; it has a high turnover rate. The average residence time (average amount of time that a particle spends in the atmosphere) of water vapour is a few days, while that of CO_2 is years (Michaels 1998). The water cycle (evaporation–condensation) is closely linked with binding and release of solar energy and changes of volume, and is strongly affected by land cover (Pokorný et al. 2016).

The amount of energy received, reflected and emitted from the Earth's surface is defined as 'net radiation'. 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) as:

$$R_n = S_{\downarrow} - S_{\uparrow} + L_{\downarrow} - L_{\uparrow} = S_{\downarrow}(1-a) + L_{\downarrow} - L_{\uparrow}, \qquad (1)$$

where: R_n is the net radiation, $S_{\uparrow}$ and $S_{\downarrow}$ are the incoming and the outgoing short-wave radiation fluxes, respectively, $L_{\uparrow}$ and $L_{\downarrow}$ are the downward and the upward long-wave radiation fluxes measured on the surface, respectively, and α is surface reflectivity.

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 Law of Energy Conservation, as:

$$R_n = LE + H + G + J + M + A_d, \qquad (2)$$

where: LE is the latent heat flux, a product of the latent heat of vaporization of water (L) and the rate of evapotranspiration from vegetation or 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. 1; Kravcik et al. 2008). M (the amount of energy used in plant metabolism) as well as J (the amount of heat stored by vegetation) and A_d (advection and freezing of water) are very small and can usually be neglected (Thom 1975; Monteith and Unsworth 1990).

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 energy flows away from the surface and evapotranspiration occurs. The downward latent heat flux at night indicates condensation at the surface as frost or dew.

Sensible heat flux is driven by the 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 third heat flow in an ecosystem is ground heat flux. The total heat flux depends on the temperature gradient and the thermal conductivity, which in turn depends on the dry density, the mineral and texture composition, the temperature, the water content and time (De Vries 1963; Wierenga et al. 1969). In general, heat stored by soil is a small part of energy balance and represents 5–10 % of net radiation for dense vegetation canopy (Tanner and Pelton 1960; Baldocchi et al. 1985).

The ratio of sensible to latent heat is called the Bowen ratio. A great portion of available energy at the surface is transformed into latent heat if the Bowen ratio is less than one. This is usually observed at wet surfaces, vegetation and open water during a day. A major part of available energy becomes sensible heat flux if the ratio is equal to or greater than one. This is typical for sealed dry surfaces.

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

Albedo of different surfaces strongly varies according to the surfaces' spectral features. It is then influenced by the 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's 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). The albedo of plants usually ranges between 0.1 and 0.27 (Doorenbos and Pruitt 1977; Oke 1978; Brutsaert 1982; Jensen et al. 1990; Meyer et al. 1999). Approximate values of the albedo for different types of vegetation cover are given in Table 1. Doorenbos and Pruitt (1977) suggested that the albedo is equal to 0.25 for most fields; it tends to decrease,

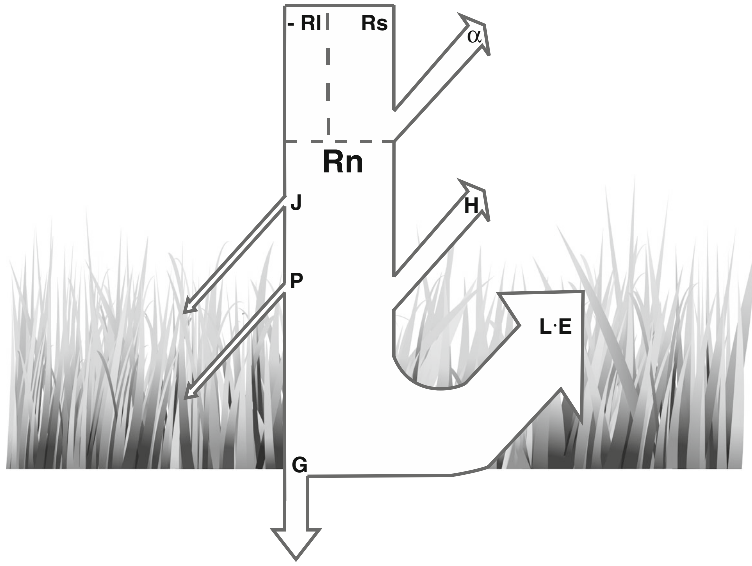

Fig. 1 Dissipation of solar energy in a plant stand. Rs – short-wave radiation, Rl – long-wave radiation, α – albedo, Rn – net radiation, H – sensible heat, L·E – latent heat of evapotranspiration, G – ground diffusion, J – accumulation of heat in biomass, P – energy consumption by photosynthesis.

however, with vegetation height and wetness of leaves, and increase with rising leaf area index (Cuf et al. 1995). The albedo of crops also usually changes during the vegetation period as well as with changing solar elevation angle and wavelength (Coulson and Reynolds 1971; Pivec 1992; Houghton et al. 2001).

Plants as dissipative structures

Life processes on the Earth are driven by solar energy. The planet receives short-wave photons from the Sun and disposes low-ordered long-wave radiation to space. The maintenance of order in living systems against dissipation is supported by solar energy input. According to Makarieva et al. (2016), the maintenance of order in life is decentralized and strongly depends on interactions between numerous and independent living systems at different levels of complexity. Solar energy is dissipated during conversion into thermal energy. The dissipation process is carried out through different channels.

Living organisms, and plants in particular, are most efficient users of solar radiation (Ripl 1995; Schneider and Sagan 2005). The energy of the Sun is captured by

Table 1 Approximate values of albedo for some cover types.

Type of vegetation cover	Albedo	References
Boreal and temperate forest	0.12–0.15	Betts and Ball 1997; Restrepo and Arain 2005; Wang 2005
Concrete surface	0.17–0.37	Takebayashi and Moriyama 2007; Huryna and Pokorný 2010
Corn	0.20–0.24	Breuer et al. 2003
Fishpond	0.10–0.11	Ahrens 2008; Huryna and Pokorný 2010
Grassland	0.16 - 0.23	Moore 1976; Oke 1978; Sagan et al. 1979
Meadow	0.15–0.25	Fitzgerald 1974; Huryna and Pokorný 2010
Pasture	0.16–0.22	Campra et al. 2008; Huryna and Pokorný 2010
Shrubs, woodland	0.25–0.29	Breuer et al. 2003
Tropical forest	0.13–0.146	Pinker et al. 1980
Wetland	0.10–0.17	Burba et al. 1999; Sumner et al. 2011
Wheat	0.18–0.23	Fritschen 1967

plants via photosynthesis process and transmitted from one biological form to another along the food chain. This multitude of biochemical reactions taking place on the living Earth, and not on any lifeless planet, allows us to compare plants to complex genetically programmed processors (Ripl 1995). Plants route the dissipation of solar energy into non-random channels and thus stabilize local environments and climates in a way that is favourable for them. When vegetation cover is destroyed, this machinery ceases to provide its services for environmental stabilization. The self-organizing process on a micro-scale of the water cycle of trees was described by Tributsch et al. (2005). During the transpiration process, negative pressure develops within the xylem whereby the water is stretched into a tensile state. The energy from Sun helps cope with the tensile state of the water, so the latent heat of vaporization is slightly higher compared to evaporation from a free surface.

Impact of vegetation on local climate

The influence of vegetation on climate change includes both a direct and indirect effect. The indirect effect is the result of a change of climate through emission/sequestration of greenhouse gases. The effect of plants on the distribution of incoming solar energy – i.e. effects on reflection, evapotranspiration, sensible heat, ground heat flux and photosynthesis is considered as a direct effect, as these processes control the daily dynamics of temperature on a particular place (Huryna et al. 2014).

The hydrological cycle is a significant process in all ecosystems and involves the continuous circulation of water through the Earth's atmosphere, land surface and oceans. It is driven by solar radiation and gravitational force. Horton (1931) expressed the hydrological cycle as a closed system with four major processes: precipitation, surface runoff, evaporation and infiltration. Water's presence in different parts of the hydrological cycle widely varies. For instance, 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). Water has a unique feature – it exists in three aggregate states: solid, liquid and vapour. The phase transition from liquid to vapour is associated with changes of volume (18 ml of liquid forms 22,400 ml of vapour) and consumption or release of energy (2.45 $MJ \cdot kg^{-1}$ at 20°C), which is a cooling or heating environment. Having high heat capacity, water mediates the exchange of energy, which equalizes the temperature differences in time (day and night) and space (between different spaces; Eiseltová et al. 2012).

The circulation of water between the oceans and the continents is called the global water cycle. About 1,400 mm (84 %) of water evaporates annually from the ocean surface per unit area, while evaporation from the land is only 485 mm (16 %; Kuchment 2004). Over the ocean, the total annual of evaporation exceeds precipitation. Surplus evaporated water is transferred to the continents. On average, approximately 40 % of precipitation on terrestrial surface comes from ocean evaporation and 60 % comes from continental evapotranspiration (Schlesinger and Bernhardt 2013). The total annual precipitation and global evaporation are approximately equal on land. A part of precipitation infiltrates into ground water, the other part returns to seas and oceans as runoff. The remaining part is re-evaporated from vegetation and open surfaces and falls back on land. According to Brutsaert (1982) and Oki and Kanae (2006), nearly 60–70 % of annual global precipitation is re-evaporated through evapotranspiration.

The vegetation and water are thus inextricably linked through the impacts on energy and hydrological cycles (Nobre et al. 1991; Hutjes et al. 1998; Arora 2002). The circulation of water between the land and the atmosphere generates the terrestrial water cycle. The presence of water 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 systems (Milly 1997), and the leaf area index (Peel et al. 2001). Evapotranspiration is a combination of two simultaneous processes: free-water evaporation and plant transpiration from the land surface to the air. It 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). Evapotranspiration varies regionally and seasonally according to the growing season, climate, available radiation, land cover, soil moisture, land-use change. The process of evapotranspiration has been studied thoroughly since the middle of the 20th century (Budyko 1974; Ryszkowski and Kedziora 1987; Monteith and Unsworth 1990; Ryszkowski and Kedziora 1995; Schneider and Sagan 2005; Pokorný et al. 2010), and its quantitative aspects and interrelations with environmental factors and the quality of plant stands are well documented. Few studies, however, deal with the cooling effect of functional vegetation (e.g. Burba et al. 1999; Herbst and Kappen 1999; Hojdová et al. 2005; Brom and Pokorný 2009; Rejsková et al. 2010).

Evapotranspiration from forests and wetlands

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). Evapotranspiration from forest ecosystems is a complex process and depends on the tree species, trees growth and height, soil conditions, geographical location, and regional climate (Swank et al. 1988; Čermák and Kučera 1990; Roberts and Rosier 1994; Calder et al. 2003; Čermák et al. 2004; Dawson et al. 2007). Forest evapotranspiration includes soil evaporation, transpiration from leaves and interception of water by leaves, branches and trunks during rainfall.

Evapotranspiration is determined by annual runoff and precipitation. There is a difference in the water budget between broadleaved and conifer trees; conifers forests tend to evaporate more water due to high interception during the whole year, especially in the winter period. Swank and Douglass (1974) measured the change of annual runoff and evapotranspiration between broadleaf and coniferous woods. The annual reduction of runoff was about 20 % with transformation of broadleaf to coniferous forests. The results show lower annual evapotranspiration for broadleaf forest, suggesting that this type of tree is more useful for increasing runoff. According to Calder et al. (2003), annual evaporation rates from conifer trees may exceed those from broadleaf trees by 15–20 %. Bosch and Hewlett (1982) showed that an annual average reduction of runoff about 40 mm for every 10 % of catchment area covered with coniferous and eucalypt trees, compared to brush or grassland. For deciduous forests, this would associate with an average reduction of approximately 25 mm per year. Roberts and Rosier (1994) documented that seasonal transpiration for ash forest in southern Britain was 372 mm which slightly exceeded that from beech forests – 355 mm. Variability of forest transpiration also significantly depends on tree root distribution. Čermák and Nadezhdina (2000) compared transpiration in pure and mixed coniferous and broadleaf forests under sufficient and limited water supply conditions. The seasonal course of transpiration was almost equal to potential evapotranspiration in monospecific and mixed forests under sufficient soil water supply. Transpiration rate decreased significantly in trees with shallow rooting system under dry conditions. However, trees with deep root systems were less sensitive to drought and transpired water over a much longer period during the growing season. Čermák and Prax (2001, 2003, 2007) and Čermák et al. (2004) monitored transpiration of the dominant tree species (*Quercus robur*, *Fraxinus excelsior* and *Tilia cordata*) as a part of the water balance in the floodplain of the Dyje river over a period of three decades. Transpiration was monitored under various climatic conditions. After flooding in 1973, transpiration was 400 mm during 6 months with daily maxima of large individual trees of 400 litres and maximum transpiration rate 40 litres per tree per hour. In the period of unlimited water supply from soil, the actual transpiration was linearly dependent on potential evapotranspiration reaching ca 80 % of potential ET. Seventy percent of the transpired water was supplied from soil and rest from local precipitation. Later in the 1970s and the 1980s, transpiration decreased due to regulation of the river bed (straightening and deepening), trees took only 30 % of water from soil and their seasonal transpiration decreased to half over a decade. The underground water level decreased by over 2 m. The root system of trees adapted to this soil water shortage and new roots developed near the soil surface. Later on in the 1990s, a regime of flooding was introduced which resulted in an increase of transpiration, but long lasting flooding had an adverse effect due to root hypoxia. Soil moisture remained rather high during the whole monitoring period. However, trees suffered from temporal drought due to the low hydraulic conductivity of heavy soils, so that, even during moderate soil drought, the trees with unfavourable root/shoot ratio suffered and consequently died. The authors concluded that the original flood regime up to the 1970s kept the floodplain forest in good conditions and supplied both underground and flood water for high evapotranspiration; but drainage caused marked decrease of ET and the subsequent flooding regime did not result in full recovery of ET to levels before 1970s. Many studies have suggested that most of the precipitation in tropical forest originates from regional transpiration (Eltahir and Bras 1994; Costa and Foley 1997; Levia and Frost 2003). Bruijnzeel (1990) reported an annual average evapotranspiration for tropical rainforests that ranged from 1,310 to 1,500 mm, average annual transpiration was 1,045 mm with a range from 885 to 1,285 mm, and the average interception was 13 % of the rainfall. In general, annual evapotranspiration can be between 200 to 480 mm in temperate and boreal forests and from 1,200 to 1,600 mm in tropical rainforests (Blanken et al. 2001; Noguchi et al. 2004; Kume et al. 2011; Nakai et al. 2013; Suryatmojo et al. 2013; Wu et al. 2013).

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 was observed to be 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 exceptional values of evapotranspiration 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). Evapotranspiration rates in the Czech Republic from wetland dominated by *Phragmites australis* was reported to be between 6.9–11.4 mm·d^{-1} (Květ 1973). Rejsková et al. (2010) indicated that evaporation from a temperate wetland dominated by *Phalaris arundinacea* reached values of 5.3–5.9 mm·d^{-1} on hot sunny days. The daily evaporation measured in a littoral stand of *Phragmites communis*. was 5.6 and 6.9 mm·d^{-1} for two cloudless July days and 5.5 mm·d^{-1} for a hot day in August (Šmíd 1975). Huryna et al. (2014) evaluated seasonal data of evapotranspiration for crops and wet meadows and found higher rates in meadow in comparison with arable lands. In July the evapotranspiration rates ranged from 5.2 to 7.1 mm·d^{-1} for winter barley field and wet meadows, respectively. Evapotranspiration from small constructed wetlands located in a dry landscape continuously supplied by wastewater was higher than potential ET due to advection of dry warm air from the surroundings and ranged from 16.4 to 27.4 L·m^{-2} for *Salix cinerea*, from 4.3 to 21.5 L·m^{-2} for *Populus tremula* and from 8.8 to 16.0 L·m^{-2} for *Prunus padus* (Kučerová et al. 2001). Čermák J et al. (1983) measured higher transpiration on willows (expressed per m^2 of ground) than potential ET and pointed out the role of the spherical shapes of crowns in absorbing solar radiation. The annual sums of wetland evapotranspiration ranged between 1,100 and 1,600 mm·yr^{-1} (Raisin 1999; Wiessner et al. 1999; Lafleur et al. 2005).

Effects of land cover changes on climate

The land surface plays a crucial role in regulating energy fluxes and water cycle in the land-biosphere-atmosphere system and controls many processes in the climate. Landscape change can dramatically modify local and regional climate conditions (Pielke et al. 1998; Snyder 2010). Human activities modify the land cover, which results in destruction of vegetation cover, soil moisture content and change in surface structure such as albedo, aerodynamic and surface conductance, roughness of vegetation, as well as change in the surface temperature (Masson et al. 2003; Pielke et al. 2011; Eiseltová et al. 2012; Hesslerová et al. 2013). Temperature plays a key role in ecosystem functioning and can be considered as an indicator of land cover change. Transformation of sustainable vegetated ecosystems well supplied with water into arable or urban land alters the partition of available energy between sensible and latent heat fluxes and the partition of precipitation between evapotranspiration, soil water and runoff. The plants and soil moisture of natural ecosystems help in regulating surface temperature due to evapotranspiration process by releasing water vapour and dissipating ambient heat. That contributes to a reduction of temperature and surface cooling. By contrast, the lack of vegetation and drainage are associated with the release of a relatively large amount of heat. As a result the local surface temperature sharply increases (Fig. 2). A decrease of evapotranspiration of 3 mm from 1 km^2 per day results in an increase of sensible heat flux of 2.1 GWh (Pokorný and Rejsková 2008).

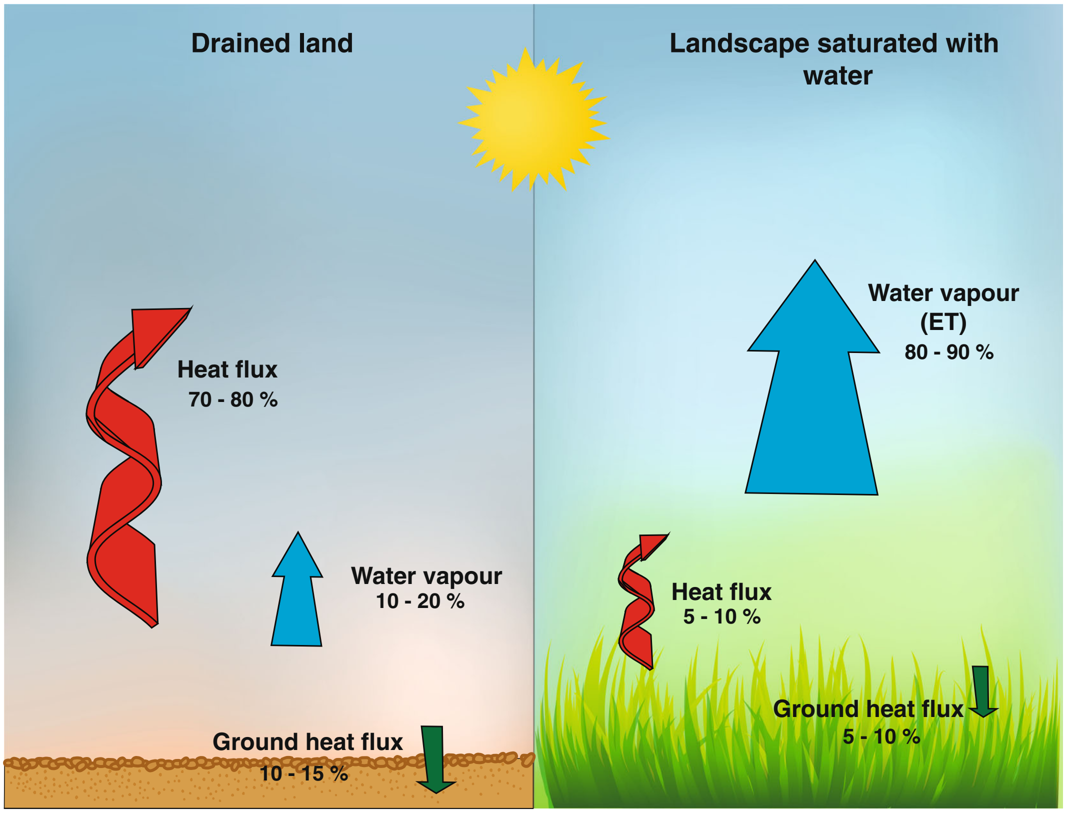

Fig. 2 Solar energy distribution in a drained landscape and in a landscape well supplied with water.

Remote sensing provides data for landscape functioning assessment (Hesslerová 2009; Eiseltová et al. 2012). Hesslerová (2009) focused on monitoring surface temperature in three locations (forest, bare ground, non-forest vegetation) during the vegetation period. The analysis demonstrated that areas with natural vegetation provided a more balanced temperature fluctuation through the growing season, reduced average temperature and increased relative humidity, while non-forest and bare-ground localities were characterized by the lowest dissipation ability and high temperature gradient. Procházka et al. (2011) examined the impact of landscape cover structure on temperature and humidity with different state of the vegetation cover. Direct (field measuring) and indirect (analysis of satellite data) methods were applied. They indicated significant temperature increase at surfaces without well-developed vegetation cover during sunny days. Remote sensing analysis showed the highest temperatures and the lowest moistures on localities covered with minimum vegetation. 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 correlated with the intensity of incoming solar radiation and water availability. In early morning, the difference in surface temperature between the localities was insignificant. The difference in temperatures increased with increasing solar radiation. During the high solar radiance differences of about 20°C were reached between forest and asphalt. The obvious role of green vegetation was shown from comparison of a wet meadow and a harvested meadow where difference at noon was approximately 13°C (Fig. 3).

Hurtt et al. (2006) suggested that 42–68 % of global land surfaces have been modified by land-use practices (transformation to cultivation lands and pastures and wood harvesting) since 1700. Agriculture is by far the largest water consumer, accounting for about 70 % of water used worldwide (Billib et al. 2009). Water movement in agriculture can be classified into three

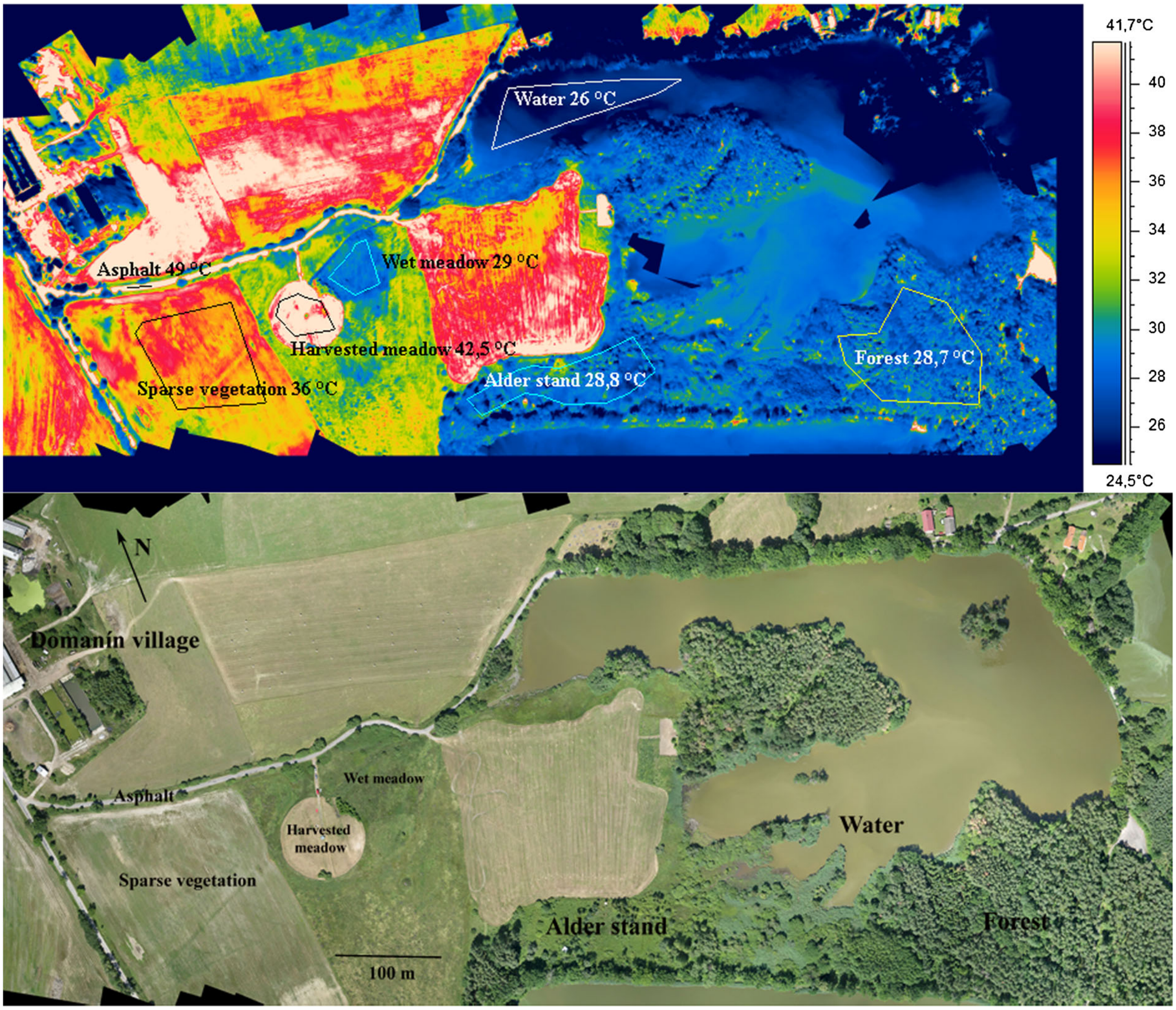

Fig. 3 Surface temperatures of different land cover of the cultural landscape in the Třeboň Biosphere Reserve on a sunny summer day at 1pm (July 9th 2010). Marked differences in surface temperature on relative small area (ca 90 ha) are evident on the IR image (Hesslerová et al. 2013).

categories: (1) agriculture and aquatic systems (runoff change); (2) agriculture and 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; Eulenstein et al. 2005; Merta et al. 2006; Li et al. 2009; Attarod et al. 2009). Evapotranspiration and even water retention are viewed narrowly as 'lost water' in agriculture, while their stabilizing effects on climate are frequently not appreciated.

A reduction in precipitation and/or an irregular pattern of precipitation is caused by destruction of the vegetation cover and consequent soil erosion (Feddema et al. 2005; Makarieva and Gorshkov 2007; Pokorný et al. 2010). Deforestation leads to warmer climate conditions in tropical regions due to decreasing of evapotranspiration (Bounoua et al. 2002). According to McGuffie et al. (1995) and Chagnon and Bras (2005), deforestation in the Amazonian basin results in a local reduction in evaporation and decrease in precipitation over the region. Deforestation has been pointed out to reduce evaporation and enhance surface temperature by 3–5 K (Dickinson and Henderson-Sellers 1988) and a reduction of net surface radiation in the tropics (Eltahir and Bras 1994). Shukla et al. (1990) and Nobre et al. (1991) mentioned a significant increase in the mean surface temperature, with decline in the annual evapotranspiration and precipitation after the replacement of tropical forest by pasture over Amazonia. There was also an increase in the length of the dry season in the southern half of the Amazon basin.

Bradshaw (2012) showed that over 40 % of forest territory has been lost in Australia since the first European settlement. This massive deforestation has led to an increase in average temperature in eastern parts of the continent, with decreased rainfall in the south-eastern and increases in the northwest regions (Nicholls and Lavery 1992; Nicholls et al. 2006). Andrich and Imberger (2013) focused on the effects of land-use change on rainfall in southwest Western Australia. They 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 has been caused by land clearing. They concluded 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 land-use change. It was also found that land-use change reduced stream flow by around 300 GL·year^{-1}.

Other studies have focused on the problem of deforestation and desertification in semi-arid and arid regions. Xue and Shukla (1993) described the effect of desertification on Sahel drought. They found that change in vegetation cover associated with land surface modification has lead to anomalies in rainfall. Rainfall has decreased, and the rainy season was delayed by half a month. Moreover, the axis of maximum rainfall had changed. Zheng and Eltahir (1997) noted that deforestation along the southern coast of West Africa may result in the complete destruction of monsoon circulation, resulting in a significant rainfall reduction. Changes in annual and seasonal rainfall have been analysed in different climatic zones of Ethiopia over 1965–2002 (Seleshi and Zanke 2004). The results indicated a significant decline of summer rainfall in eastern, southern and south-western parts of the country.

Land use change is not only associated with an increase of rain water runoff and surface temperature enhancement. Land transformation leads to substantial loss of nutrients from the soil and reduction of soil organic material content (Gupta and Germida, 1988; Post and Mann 1990; Davidson and Ackerman 1993; Jurgensen et al. 1997). Davidson and Ackerman (1993) reported of about 30 % carbon decline after cultivation. The rapid soil carbon loss occurred within the first few years. Vitorello et al. (1989) concluded that the reduction of organic carbon reached by about 50 % after twelve years of cultivation in Brazil. The rapid decline in soil organic content after land transformation was indicated by Burke et al. (1989) in the USA, Ellert and Gregorovich (1996) in Canada, and Sanchez and Logan (1992) in the tropics.

Management implications

Observational evidence indicates that the annual mean surface temperature of the Earth has increased by about 0.85°C since 1880, with rapid warming in the recent past decades (about 0.72°C after 1951; IPCC 2013). Global warming is generally explained by emission of greenhouse gases. The 'Summary for policymakers' in the IPCC (2013) report focused on the correlation between average surface temperature and concentration of specific greenhouse gases. According to the IPCC Report (2013), water vapour is a feedback agent rather than a forcing agent of climate change and has a negligible impact on the global climate. Its impact on climate change is not calculated, since its concentration depends mainly on air temperature and varies widely. Moreover, its residence time is only several days, compared with years for carbon dioxide and methane. The IPCC Report (2013; Chapter 8: Radiative forcing) asserts that amount of water vapour in atmosphere is controlled by temperature and temperature is controlled by GHG like CO_2 or CH_4.

However, despite the overwhelming focus on carbon in current IPCC efforts, it is recognized that atmospheric water – the main greenhouse substance owing to its concentration 1–3 orders of magnitude higher than that of other GHGs – is at the same time the biggest source of uncertainty in future climate projections (Bony et al. 2015). For instance, there is persistent ambiguity regarding the role of the planetary cloud cover−whether it reduces or enhances climate sensitivity to external forcing. However, the formation of clouds is profoundly affected by the land cover via biotically mediated synthesis of condensation nuclei (Pohlker et al. 2012). Second, modern circulation models do not properly reproduce the continental water cycle (Hagemann et al. 2011). Again, according to rapidly accumulating knowledge, vegetation cover plays a major role in sustaining the regional water cycle on land (Makarieva and Gorshkov 2007; Makarieva et al. 2010, 2013; Sheil 2014; Lawrence and Vandecar 2015). Furthermore, theoretical studies suggest that evaporation can serve as an efficient climate stabilizing mechanism not only on a local, but also on a global scale (Bates 1999). Recent evidence suggests that the overwhelming part of evaporation from land is mediated by life (Jasechko et al. 2013). Finally, as we discuss here, vegetation has a direct role in distribution of solar energy, reducing temperature gradients and damping extremes of air temperature. Thus, while the IPCC (2013) report has apparently undervalued the importance of plant cover, the climatic role of vegetation on the Earth is significant and should be urgently made a focus of an intense, cross-disciplinary and well-coordinated global research effort. Changes in terrestrial ecosystems like drainage of wetlands and deforestation reduce precipitation and evapotranspiration, enhance runoff and modify surface temperature by shifting the energy balance from latent heat (cooling through evapotranspiration) to sensible heat loss (turbulent flux of hot air). This leads to the destruction (open) of both matter and water cycles.

Retention of water is often considered in a negative meaning, such as water 'loss' (Cudlin et al. 2013; Huryna et al. 2014). The process of transpiration is sometimes even regarded as an unavoidable evil, in the sense that water is sacrificed for the sake of enabling intake of CO_2 for photosynthesis. However, the interaction of water and plants dampens the temperature maxima and in the process of water evaporation the cooling effect is important. Vegetation well supplied with water is able to dampen the vertical exchange of sensible energy and enhance the latent heat flux (evapotranspiration) between the surface and the atmosphere, while the absence of vegetation intensifies the flux of sensible heat. The important difference in water evaporation is observed between wetlands and croplands. Croplands evaporate water but mainly at intermediate growth stages when accumulation of plant biomass occurs. Water vapour from crop plants rises faster than from wetlands which have dense vegetation and therefore lower temperatures at the ground. Conversion of natural to agricultural fields changes land surface characteristics, which lead to redistribution of surface energy components (Esau and Lyons 2002). More than 51 % (45.9×10^6 ha) of the total area of wetland has been replaced by cropland in the USA since Presettlement (Mitsch and Hernandez 2013). About of 400 $W{\cdot}m^{-2}$ has thus been shifted from latent to sensible heat flux for days with the highest solar irradiance (Huryna et al. 2014). Thereby, we can assume that more than 175,000 GW of energy has been converted into sensible heat in the territory of the USA, which strongly affects dynamic processes in the atmosphere.

Wetlands and forests, in comparison to their surrounding terrestrial biotopes in the same region, often show more balanced temperature patterns and increased air humidity (Priban and Ondok 1986; Brom and Pokorný 2009). By transpiring large amounts of water, wetland plants influence the temperature of the surface as well as that of the air above the stand. To evaporate 1 litre of water, 0.69 kW·h (2.5 MJ) of energy is needed. The wetland under study, which evapotranspirated about 7 $mmol{\cdot}m^{-2}{\cdot}s^{-1}$ (i.e. 126 $mg{\cdot}m^{-2}{\cdot}s^{-1}$) during a sunny afternoon, converted about 315 W of energy per square metre of its surface into latent heat flux. According to our results, the Mokré louky (Wet meadows) wetland with its area of about 4 km^2 evapotranspired about 500 kg of water every second, which is a flow rate of a small river. This invisible stream represents the latent heat flux of approximately 1,260 MW. Thus, this ecosystem regulates the surface temperature through energy and water fluxes with a power equivalent to that of a fairly large power station (Rejsková et al. 2010).

In addition, the IPCC 'Summary for policymakers' (IPCC 2014) has not specified that the presence and quality of vegetation can affect the amount and distribution of precipitation. Makarieva and Gorshkov (2007, 2010) introduced a physical theory on forest function. The theory describes forests as active attractors of moist air. By analysing the interrelation between annual rainfall in both forested and deforested regions across various continents and at varying distance from the sea coast, the authors found that annual rainfall decreased in deforested parts of continents whereas in areas covered by natural forests the rainfall could increase even over a distance of several thousand kilometres. The authors presented the concept of a 'biotic pump' – condensation of water vapour in forested areas results in a drop of air pressure and 'horizontal sucking' of wet air from oceans or other donor areas.

Renewal of the fundamental ecological functioning of landscapes is the most crucial condition for the development of a sustainable landscape. Continuous interaction of water, energy and matter are the three basic components of sustainable ecosystems. Violation of one of the components leads to the destruction of the whole system. Human activities often cause the uncoupling and opening of water and matter cycles. High surface temperature can be considered as a cause of a non-sustainable environment in areas with high matter loss. Ripl et al. (2004) and Ripl and Eiseltová (2009) investigated natural processes in unmanaged virgin forest and agricultural land in Germany. The data obtained from virgin forest showed very clearly how ecosystems can achieve minimization of water and matter losses. The daily temperature amplitude did not exceeded 9°C, and no overheating periods were observed. The runoff from virgin forest was very low with minimum loses of nutrients and minerals. High irreversible matter losses occurred from agricultural landscape. Areal matter losses were 50–100 times higher than those from virgin forest. This led to violations of the water cycle and occurrence of overheated areas.

Sustainable management practice should include fundamental changes in land use. Rehabilitation and recovery techniques help stop land degradation and restore already degraded land. Innovation in farming practices was shown in Australia, where Peter Andrews developed his method of Natural Sequence Farming. The method emulates role of natural water courses in an effort to reverse salinity, slow erosion and increase soil and water quality to enable native vegetation to restore the riparian zone (Andrews 2006; Tane 2006). Water retention practice was also implemented in India. The Tarun Bharat Sangh project pioneered by Rajendra Singh was based on the revival of traditional water reservoirs. The work is aimed at designing water harvesting structures (johads). These are simple, mud barriers built across the hill slopes to arrest the monsoon runoff. The height of the embankments varies from one johad to another, depending on the site, water flow and topography contours. A johad serves two functions: holds water for livestock and allows the liquid to percolate down through the soil. It recharges the aquifer below, as far as a kilometre away. These water harvesting structures have successfuly provided irrigation water to an estimated 140,000 ha. The revival of the system of johads decreased the ground water depth from about 100–120 m depth to 3–13 m. The area under single cropping increased from 11 to 70 % out of which the area under double cropping increased from 3 to 50 %. Forest cover, which was around 7 %, increased to 40 %. More than 5,000 johads have been built and over 2,500 old structures rejuvenated by village communities in 1,058 villages since 1985 (Gupta 2011; Hussain et al. 2014; Bhattacharya 2015).

Reintroducing vegetation cover and restoring the water-holding capacity of soil can lead to closing of both water and matter cycles and will bring much water in a system. Increased evapotranspiration leads to temperature dampening and minimized matter losses. The direct role of vegetation cannot be ignored, and changes in land-cover use must be included in both regional and global strategies to effectively mitigate climate change. Temperature-balancing measures, such as water retention and introduction of permanent vegetation, should be taken in the landscape in order to retain its sustainable functioning because global gradients drive dynamic processes in the atmosphere. Introduction of the principles of sustainable development, such as improvement of ecological stability and its coordination with human interest, are necessary. Persistence of the dogma that the greenhouse effect acts alone results in ignoring the most important functions of wetlands, namely their direct effect on the climate and water cycling through life processes. Such an approach facilitates further land drainage and deforestation.

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