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Conservation of water cycle on land via restoration of natural closed-canopy forests: implications for regional landscape planning

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Abstract Investigating the role of forests for maintenance of the water cycle on land is critically important in the current situation of rapid global elimination of the natural vegetation cover. In this paper we contribute to the on-going discussion of the issue with two aspects. (1) Theoretical consideration of the water cycle on land reveals the importance of correct identification of independent and dependent terms in the water budget with respect to changing vegetation cover for understanding possible scenarios of water cycle change under anthropogenic impact. An important controlling influence of the vegetation cover is imposed on the outgoing fluxes of atmospheric moisture A^- from land to the ocean, which is maximized in deserts and minimized in forested areas, while the dependencies for runoff and precipitation are the reverse. (2) Physical mechanisms allowing for efficient water retention and minimization of A^- in forest ecosystems are investigated. Atmospheric water vapor is in aerostatic equilibrium when the temperature lapse rate is less than $G = 1.9 \text{ K km}^{-1}$ and out of aerostatic equilibrium when $G > 1.9 \text{ K km}^{-1}$. In the former case there are no vertical upward fluxes of the evaporated water. It is shown that the temperature profiles developed under the closed canopies of natural forests keep water vapor in aerostatic equilibrium preventing soil moisture loss to A^- , in contrast to the situation in open ecosystems like grasslands. The analyzed evidence allows one to conclude that an intensive water cycle on land can be restored after recovery of natural, self-sustained closed canopy ecosystems on continent-wide areas.

Keywords Forest canopy · Lapse rate · Temperature · Humidity · Water cycle · Water vapour · Hydrostatic equilibrium

Introduction

Two thirds of the Earth's surface is covered by the oceans. Some part of water evaporated from the oceanic surface is brought to land by atmospheric air fluxes, undergoes condensation on land and returns to the ocean in the form of runoff. Some parts of land are dry; on the others one observes an intensive water cycle. The extant natural forests covering about $33 \times 10^6 \text{ km}^2$ (Bryant et al. 1997) generally occupy well-moistened areas, while the drier parts of land are covered by more open ecosystems like savannas, grasslands and shrublands.

In the modern consideration of the terrestrial water cycle it is assumed that the gross flux of atmospheric moisture, A^+ , which is brought to land from the ocean, is largely independent of the vegetation cover on land, while the major control on water cycle imposed by vegetation consists in enhancing evapotranspiration (e.g. da Rocha et al. 2004). River runoff R is equal to the net influx of moisture from ocean to land (moisture convergence), $R = A^+ - A^-$ (Marengo 2005), where A^- is the gross atmospheric outflux of moisture from land to the ocean.

The main ecological question of terrestrial hydrology is the following. If one assumes that A^- is determined by the geophysical land properties like slope and distance from the ocean, then the amount of runoff, on which modern humanity is critically dependent, should not depend on the presence/absence or properties of local terrestrial vegetation. In such a case it would be possible to change the natural vegetation cover in accordance to human agricultural and industrial needs, i.e. cut down natural forests replacing them for arable lands and pastures or industrial tree stands, all this without threatening the water cycle of the currently well-moistened land parts. Similarly, the dryness areas covered by savannas, grasslands, prairies and steppes would remain

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arid irrespective of the degree of their cultivation by man. Forests, with their high demand for moisture, cannot presumably grow on such territories in any case. Despite there are many data and analyzes indicating the complexity of vegetation change effects on the various climatic parameters (e.g. Pielke 2001), modern practical forest-use policies resulting in global deforestation are grounded in the above simple views (e.g. Kaimowitz 2005; Hayward 2005).

Alternatively, the value of atmospheric moisture outflux A^- could significantly depend on the properties of vegetation cover. For example, recent studies have demonstrated that natural forests efficiently facilitate rainout from the atmospheric air masses via biotic control of biogenic cloud condensation nuclei (Andreae et al. 2004; Koren et al. 2004). The forest-induced decrease of moisture content of the outflowing air masses results in the corresponding increase of river runoff R up to the maximum value, when it becomes equal to A^+ . Turning all moisture, which is brought from the ocean, into runoff natural forests can elevate the soil water content to a high optimal level. As far as transpiration of water by plants is tightly coupled to biotic productivity, then when soil moisture is high, transpiration can be also increased to its maximum possible value dictated by the available solar radiation. With high transpiration E and an efficient mechanism of rainfall extraction, forests are able to arrange an intensive water cycle when precipitation P significantly exceeds runoff R , both being high, $P > R$. Such a situation is observed in all natural forest massives like, e.g. tropical forests of the Amazon basin (Marengo 2005) or boreal forests of Central Siberia (Zimmerman et al. 2000). Biotic water cycle on such territories can be completely autonomous, i.e. independent of the amount of atmospheric moisture brought from the ocean, $A^+ \approx R$. In this picture, large-scale deforestation will lead to a reduction of river runoff R (due to the broken control of A^-) and a very significant reduction of both P and E , i.e. decrease of biological productivity. Vice versa, facilitation of gradual natural forest recovery will lead to restoration of an intensive water cycle in currently arid zones.

In what is to follow we present arguments in favor of the second view on the forest-water issue with an emphasis on the biotic control of A^- .

Terrestrial water budget and vegetation

As already noted, in the stationary state runoff R (under runoff we shall understand all losses of liquid water from land to the ocean, of which river runoff makes the dominant part) is equal to atmospheric moisture convergence, $R = A^+ - A^-$, where A^+ is the influx of atmospheric moisture from ocean to land, and A^- is the outflux of atmospheric moisture from land to the ocean. The second terrestrial water budget equation is $P = E + R$, which says that precipitated water P is

either evaporated or transpired back to the atmosphere, E , or runs off to the ocean, R .

Studying the effects of vegetation cover on the water cycle is difficult due to the complex interdependence of all the budget terms (Rodríguez-Iturbe 2000; Pielke 2001; Rodríguez-Iturbe and Porporato 2004). Traditionally, precipitation in ecology has been considered as an abiotic constraint on ecosystem processes (Austin and Sala 2002). Since modern city-based civilization with spatially concentrated industrial and human settlements is heavily dependent on river runoff, much research has been done to understand how changes in vegetation cover could affect runoff. In these studies precipitation P was commonly chosen as an independent variable. That is, water yields from forested and unforested catchments were compared at equal values of precipitation. Based on the data from 250 water catchments worldwide, Zhang et al. (2001) established empirical curves relating evapotranspiration E from catchments with different types of vegetation cover to precipitation P . The overall finding was that evapotranspiration from forests E_f is higher than evapotranspiration E_u from unforested areas like grasslands or shrubs, the difference reaching over 500 mm year⁻¹ for $P > 2,000$ mm year⁻¹ (Zhang et al. 2001). As far as $P = E + R$, the relationship $E_f > E_u$ at constant P means that $R_f < R_u$, i.e. water runoff from forested areas is lower.

However, this conclusion, interpreted in support of the attitude that deforestation poses no threat to the runoff component of the water cycle or might even enhance it, is in strong disagreement with the important global-scale observations. All Earth's most powerful rivers are either running through, or originating in, natural forested areas (Dai and Trenberth 2002), while territories with scarce vegetation like savannas, steppes, semi-deserts are generally characterized by weak, ephemeral or non-existent river systems. This suggests that setting precipitation P as an independent variable in the water cycle problem can be misleading.

Another approach to assessing vegetation cover impact on the water cycle makes use of the notion of precipitation (moisture) recycling (Savenije 1995; Eltahir and Braas 1996; Cowling 2004). Precipitation recycling $0 \leq \rho \leq 1$ on a given area is equal to the ratio of locally originating precipitation to total precipitation, which includes moisture brought to the area from elsewhere. If one assumes that all locally evaporated water precipitates in the considered area, which is true for large regions, then $\rho \approx E/P$. (This is an overestimate, because some part of locally evaporated water can be brought away via the atmosphere as A^- , so a more accurate estimate of recycling is $\rho = (E - A^-)/P$ (Eltahir and Braas 1996).)

As far as precipitation recycling is positively related to evapotranspiration E , large values of ρ should be characteristic of forested areas, where E is high. However, as far as $P = E + R$, high precipitation recycling $\rho \approx E/P = E/(E + R)$ can be also reached at the expense of a very small runoff $R \rightarrow 0$. For example, for

the arid Sahel zone (low runoff) precipitation recycling can be as high as 90% (Savenije 1995), while for the Amazon river basin with huge runoff it is only around 60% (Eltahir and Braas 1996; Marengo 2005), creating a misleading impression that it is in Sahel, but not in the Amazon, that the local water cycle is controlled by evapotranspiration from the local vegetation cover and soil. This consideration illustrates that, similarly to precipitation P , runoff R cannot be fixed as an independent variable when studying the vegetation impact on water cycle.

Finally, the seemingly transparent idea that increased evapotranspiration E leads to increased precipitation P , as far as $P = E + R$, does also have its caveat. Increased evapotranspiration means returning more water to the atmosphere, which can leave the considered area with the outflowing air masses. Thus, if the increased in E is accompanied by an equivalent increase in A^- , then precipitation $P = E + A^+ - A^-$ will remain unaffected, irrespective of whatever high E might be.

We propose that a consistent approach for studying the complex effects of vegetation cover on water cycle should be based on consideration of the vegetation-induced changes in A^- (if one assumes that A^+ is unaffected by vegetation, see, however, Gorshkov and Makarieva (2006)). Two extreme cases can be envisaged. In what might be called an ideal desert, there is a certain inflow A^+ of atmospheric moisture from the ocean. However, there is no precipitation. First, dust inherent to surfaces uncovered by vegetation strongly suppresses precipitation (e.g. Rosenfeld et al. 2001). Second, absence of evaporation from the dry surface leads to increased heating of the surface, which also prevents condensation of water vapor. Thus, in the ideal desert (d) atmospheric moisture travels above the surface without interacting with it and is taken away by the outflowing air masses in exactly the same quantity as it was brought there from the ocean, $A_d^- = A^+$. We thus have for the desert (d) $R_d = A^+ - A_d^- = 0$ (no runoff) and $P_d = E_d = 0$ (no precipitation, no evaporation), Fig. 1a.

In what might be called an ideal forest, there is the same incoming flux A^+ of atmospheric moisture from the ocean. However, the forest efficiently extracts rainfall from the flowing air masses by means of dust removal, release of biogenic condensation nuclei (Andreae

et al. 2004; Koren et al. 2004), transpiration-induced decrease of canopy temperature (Bruijnzeel 2001) and presumably other mechanisms, which still wait to be studied. This minimizes the amount A^- of atmospheric moisture leaving the forested area. In the extreme case of $A_f^- \ll A_f^+$, we have $R_f = A^+$, which means that practically all atmospheric moisture brought from the ocean is returned to the ocean as runoff, Fig. 1b. Soil moistening allows for increase of transpiration up to the maximum possible value E_{\max} dictated by the available solar radiation. In the ideal forest we thus have $P_f = E_{\max} + A^+$ and $R_f = A^+$, compare to desert $P_d = 0$ and $R_d = 0$.

From this consideration it is clear that the water cycle on land is fully determined by vegetation cover even if the incoming flux of atmospheric moisture A^+ from the ocean is not affected by vegetation change. The major hydrological function of forest is not only the increased flux E of evaporated water, but also the decreased flux A^- of the out-flowing atmospheric moisture, which is often overlooked. As far as A^- decreases from ideal desert to ideal forest, $A_d^- \gg A_f^-$, this shows that transition to forest can only increase runoff $R = A^+ - A^-$, Fig. 1, but not decrease it as it could be incorrectly inferred from the analysis with precipitation P as an independent parameter (Zhang et al. 2001).

If the outflux of moisture A^- is expressed in terms of evaporation E as $A^- = \alpha E$, then in the desert where $E \ll E_{\max}$ and $A^- \approx A^+$, the value of α is very large. For a typical extreme desert with $E \sim 50 \text{ mm year}^{-1}$ (Nicholson 2000) and tropical A^+ of the order of $10^3 \text{ mm year}^{-1}$, α would be of the order of several dozens (in the ideal desert α is infinite). With increasing forest cover $E \rightarrow E_{\max}$, while A^- diminishes. This means that coefficient α should drop rapidly. In the ideal forest it is zero, as far as $A^- = 0$. In real forests α is much less than unity. For example, for Amazonian forests $\rho \equiv (E - A^-)/P \approx 0.3$, while $E/P \approx 0.6$ (Eltahir and Braas 1996), which means that $E = 0.5 A^-$ and $\alpha = 0.5$. For comparison, for the Mississippi river basin with significantly scarcer vegetation cover as compared to Amazon, $\rho \equiv (E - A^-)/P \approx 0.2$ and $E/P \approx 0.8$ (Eltahir and Braas 1996), which gives $\alpha = 0.75$. That is, via the atmosphere the Mississippi river basin loses a significantly larger portion of locally evaporated water than does the Amazon basin.

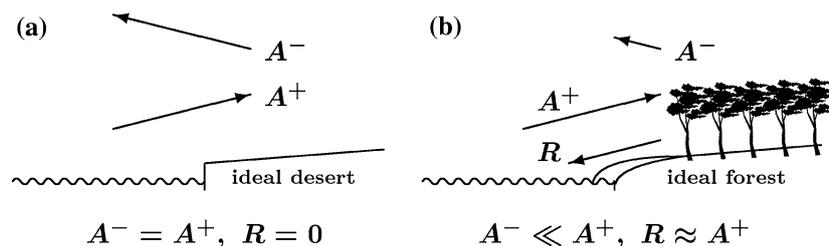


Fig. 1 Schematic representation of the influence of the vegetation cover on the water cycle budget. A^+ and A^- represent the inflow and outflow of atmospheric water vapor to and from the

ecosystem, respectively; $R = A^+ - A^-$ is the runoff. See text for further details

These examples show that forests should be capable not only of water extraction from the flowing air masses, but also of efficient retention of water in the local stores, e.g. soil water. The mechanisms of water extraction have recently received considerable attention (e.g. Andreae et al. 2004; Koren et al. 2004). In the next section we dwell on the physical principles that can be used by forest biota to prevent uncontrolled water losses to the atmosphere.

A physical mechanism of efficient water retention in closed-canopy forests

Vertical distribution of atmospheric water vapor

Atmospheric air is in the so-called aerostatic equilibrium, when air pressure p at a given height z in the atmosphere is balanced by the weight of atmospheric column above z , while change dp of air pressure over vertical distance dz is equal to the weight of air in the atmospheric column of thickness dz :

$$-\frac{dp}{dz} = MNg. \quad (1)$$

Here N is molar air density at height z , $M = 29 \text{ g mol}^{-1}$ is air molar mass, $g = 9.8 \text{ m s}^{-2}$ is the acceleration of gravity. Atmospheric air is close to ideal gas and conforms to the equation of state

$$p = NRT, \quad (2)$$

where $R = 8.3 \text{ J K}^{-1} \text{ mol}^{-1}$ is the universal gas constant, T is absolute air temperature. From Eqs. 1 and 2 we obtain the following equations for air pressure p :

$$\frac{dp}{dz} = -\frac{p}{h}, \quad p(z) = p_s \exp \left\{ -\int_0^z \frac{dz}{h} \right\}, \quad (3)$$

$$h \equiv \frac{RT}{Mg}, \quad h_s \equiv \frac{RT_s}{Mg} = 8.4 \text{ km},$$

where p_s is air pressure at the Earth's surface, h_s is calculated for mean global surface temperature $T_s = 288 \text{ K}$. If one neglects the 10% change of temperature T with height in the atmosphere and considers $h \approx h_s$ constant in Eq. 3, then the solution for $p(z)$ (Eq. 3) corresponds to air pressure exponentially dropping with height z , $p(z) = \exp(-z/h)$.

Water vapor is a minor constituent of atmospheric air, accounting for about 1% of air pressure. Immediately above water surface or wet soil water vapor is saturated. The dependence of partial pressure p_w of water vapor on air temperature T is governed by the Clapeyron-Clausius law (Raval and Ramanathan 1989; Wentz and Schabel 2000):

$$p_w = p_{ws} \exp \left\{ \frac{T_w}{T_s} - \frac{T_w}{T} \right\}, \quad T_w \equiv \frac{Q_w}{R} \approx 5,300 \text{ K} \quad (4)$$

where low index s refers to corresponding values at the Earth's surface, $Q_w \approx 44 \text{ kJ mol}^{-1}$ is the molar latent heat of evaporation.

If air temperature did not change with height, atmospheric water vapor would remain in aerostatic equilibrium as all the other gases in the dry air. Partial pressure of water vapor would follow the same dependence (Eq. 3) of decrease with height as air pressure does. In such a case saturated concentration of water vapor would be present at the Earth's surface only. With increasing height relative humidity would drop, so that no condensation of water vapor in the atmosphere were possible.

When air temperature T decreases with height z , water vapor at all heights approaches the state of saturation. Taking derivatives over z of water vapor partial pressure $p_w(z)$ we obtain from Eq. 4:

$$\frac{dp_w}{dz} = -\frac{p_w}{h_w}, \quad h_w \equiv \frac{T^2}{\left(-\frac{dT}{dz}\right)T_w}, \quad (5)$$

$$p_w = p_w \exp \left\{ -\int_0^z \frac{dz}{h_w} \right\}. \quad (6)$$

Equation 5 and its solution 6 (that is but a different mathematical expression of Eq. 4) have the same form as the aerostatic equilibrium equation 3 for atmospheric air, but with a different value of height h_w , which, in the case of water vapor, depends on the air temperature lapse rate $G \equiv -dT/dz$. To ensure that water vapor is in aerostatic equilibrium in the entire atmosphere, so that expressions 5 and 6 coincided with 3, the equality $h = h_w$ must hold. This equality constrains the value of atmospheric lapse rate $G = G_w$ and the rate of air temperature decrease with height (Gorshkov et al. 2002; Makarieva et al. 2003, 2004):

$$h_w = h \text{ or } \frac{dT}{dz} = -\frac{T}{H}, \quad H \equiv \frac{RT_w}{Mg} = 155 \text{ km}, \quad (7)$$

$$T = T_s e^{-z/H}, \quad G_w = -\frac{dT}{dz} = \frac{T_s}{H} e^{-z/H}.$$

As far as H (Eq. 7) is very large, the scaling exponent in the last expression in Eq. 7 can be put equal to unity, which means that the lapse rate G_w is independent of z throughout the atmospheric column and coincides with its value at the surface:

$$G_w = \frac{T_s}{H} = 1.9 \text{ K km}^{-1}. \quad (8)$$

The absolute value of G_w (Eq. 8) corresponds to mean global surface temperature $T_s = 288 \text{ K}$ (Mitchell 1989). As can be easily checked, differences in the absolute surface temperatures between the equatorial and polar regions change the obtained value (Eq. 8) by no more than 10% and do not affect any subsequent conclusions.

It follows from Eqs. 5–8 that when air temperature lapse rate $G \equiv -dT/dz$ is equal to G_w , $G = G_w$, atmospheric water vapor at all heights is saturated and it is in aerostatic equilibrium. At smaller lapse rates, $G < G_w$, water vapor is in aerostatic equilibrium, but remains saturated only near the surface. At $G \leq G_w$ the behavior of water vapor and its vertical distribution in the atmosphere coincide with those of all other, non-condensable, air constituents.

Aerostatic equilibrium of atmospheric water vapor at $G \leq G_w = 1.9 \text{ K km}^{-1}$ implies that there is no atmospheric precipitation (because relative humidity is lower than unity at all $z \geq 0$) and no upward fluxes of water evaporated from the planetary surface to the upper atmosphere. Solar radiation absorbed by the Earth's surface results in evaporation of water from the ocean and soil; this water is immediately condensated at microscopic distances from the surface, while the corresponding latent heat is released near the surface in the form of sensible heat and thermal radiation.

Mean tropospheric lapse rate is $G = 6.5 \text{ K km}^{-1}$ and is practically uniform over the globe. This lapse rate exceeds $G_w = 1.9 \text{ K km}^{-1}$ by 3.5 times. This means that at the observed global mean lapse rate of 6.5 K km^{-1} aerostatic equilibrium of atmospheric water vapor is impossible, because partial pressure of water vapor at the Earth's surface significantly exceeds the weight of the atmospheric column of water vapor. Due to this uncompensated pressure, there appear upward fluxes of water vapor, while the resulting deviation from saturation at the surface is replenished by continuous evaporation from the surface.

Absence of aerostatic equilibrium results in a strong compression of the vertical distribution of water vapor as compared to atmospheric air (Eq. 3). As follows from Eq. 7,

$$\frac{h}{h_w} = \frac{G}{G_w} \equiv \beta, \quad \beta = 3.5 \text{ at } G = 6.5 \text{ K km}^{-1} \quad (9)$$

Compression coefficient β does not depend on z due to the observed approximate constancy of G_w and G . Combining Eqs. 3 and 6 we obtain

$$\frac{p_w}{p_{ws}} = \left(\frac{p}{p_s}\right)^\beta = \exp\left\{-\beta \int_0^z \frac{dz}{h}\right\}. \quad (10)$$

Relationship (Eq. 10) shows that the vertical distribution of water vapor in the troposphere is compressed 3.5-fold as compared to the vertical distribution of atmospheric air. Mean height h_w characterizing vertical distribution of water vapor is $h_w = h/\beta = 2.4 \text{ km}$, which agrees well with observations (Weaver and Ramanathan 1995).

Thus, we conclude that depending on the ambient temperature gradient, atmospheric water vapor can be either in aerostatic equilibrium (at $G \leq 1.9 \text{ K km}^{-1}$), when there are no upward fluxes of water; or out of aerostatic equilibrium (at $G > 1.9 \text{ K km}^{-1}$), when, due

to compression of the vertical distribution of water vapor, there are continuous upward fluxes of water evaporated from the surface. In the next section we discuss how these physical phenomena work in forest ecosystems to prevent extra water losses to the atmosphere.

Retention of water by closed canopies

In natural forest ecosystems with well-developed closed canopies air temperature during the daytime increases in the upward direction, i.e. it is higher in the canopy than at the ground surface (Shuttleworth 1989; Kruijt et al. 2000; Szarzynski and Anhof 2001), Fig. 2a. This is caused by the fact that in closed-canopy ecosystems solar radiation is absorbed predominantly in the canopy thus heating it. When canopy temperature T_c exceeds ground temperature T_g , air temperature lapse rate becomes negative, $G = -dT/dz \approx (T_c - T_g)/z_c < 0$, where z_c is the canopy height reaching several tens of meters in natural forests, Fig. 2a. As far as $G < 0 < G_w = 1.9 \text{ K km}^{-1}$, in this case, according to the results of the previous section, water vapor under canopy remains in aerostatic equilibrium, and the upward fluxes of water vapor from beneath the canopy are absent. Water vapor partial pressure p_w conforms then to Eq. 3 and remains practically constant under the canopy with $z_c \ll h$, see Eq. 3, $p_w(z) \approx p_w(0) \equiv p_{sw}$. Relative humidity $\text{RH}(z)$, which is equal to 100% immediately above the surface of the wet soil, decreases with height as

$$\text{RH}(z) = 1/\exp\{[T_w/T_s] - [T_w/T(z)]\}, \quad (11)$$

cf. Eqs. 3 and 4 and see Fig. 2b. Thus, the daytime under-canopy temperature inversion ensured by closed canopies keeps saturated water vapor above ground surface in aerostatic equilibrium, thus preventing biotically uncontrolled losses of soil water to the atmosphere. This mechanism explains why ground surface of closed-canopy forests always remains wet, which is manifested as low susceptibility of closed canopies to fires (Cochrane et al. 1999; Nepstad et al. 2004).

In higher latitudes, where the solar angle is lower than in the tropics and solar beams at midday are slanting rather than perpendicular to the surface, the daytime temperature inversion within the canopy can arise at a lesser degree of canopy closure than in the tropics, as far as the lower solar angle diminishes the difference in the solar radiation obtained by canopy and inter-canopy patches (e.g. Breshears et al. 1998).

Under open canopies, as well as in open ecosystems like grasslands and savannahs, the daytime temperature inversion does not form and the air temperature decreases rapidly with height at $G \gg G_w$, Fig. 2c. In this case, as discussed above, all water evaporated from soil is carried away from the canopy by the upwelling fluxes of water vapor caused by the uncompensated pressure of saturated water vapor. The decrease of air temperature with height is also observed above closed canopies, Fig. 2a.

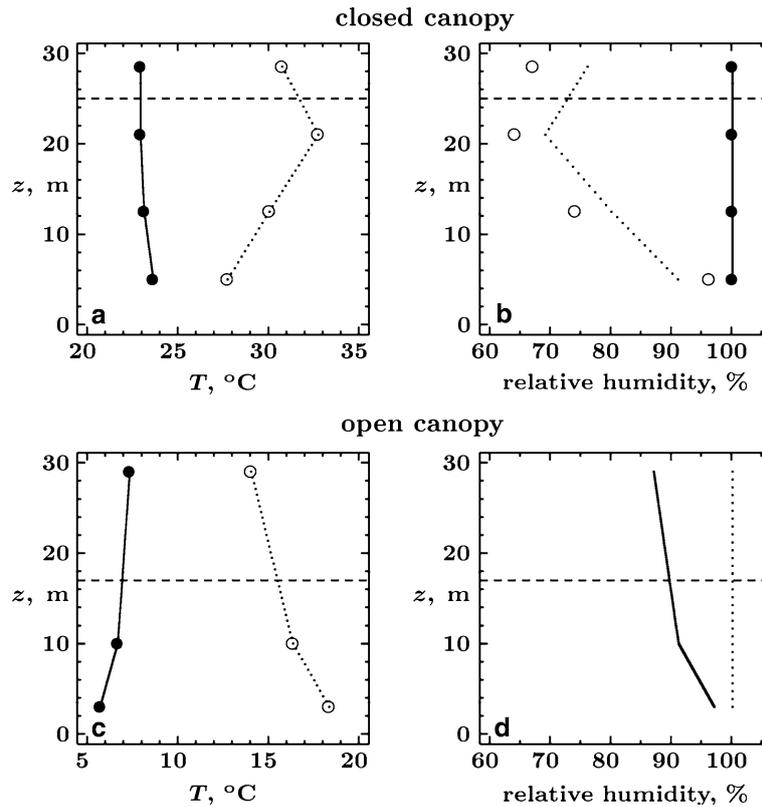


Fig. 2 Diel cycle of vertical profiles of under-canopy air temperature and relative humidity in the closed-canopy versus open-canopy forests. z (m) is height above the ground surface; *dashed lines* denote canopy height. *Closed and open circles* correspond to measurements taken at a particular height during the nighttime (approx. 5 a.m.) and daytime (approx. 5 p.m.), respectively. *Dotted and solid lines* in **b** and **d** are theoretical curves calculated for the corresponding temperature lapse rates, see Eq. 11, and assuming

100% relative humidity at the ground surface. The closed-canopy data are for the primary forest in Venezuela (Szarzynski and Anhof 2001); the open-canopy data are for the subalpine spruce-fir forest in Wyoming, USA (Zeller and Nikolov 2000). Note that if the relative humidity at the ground surface in the open-canopy forest is less than 100%, it will increase with height during the daytime and decrease with height during the nighttime. The proposed schemes can be directly tested with experimental data

At nighttime soil under the closed canopy is warmer than the canopy due to the rapid radiative cooling of the latter (Shuttleworth et al. 1985; Szarzynski and Anhof 2001). The temperature lapse rate $G \equiv -dT/dz$ is positive and can exceed the observed mean global value $G = 6.5 \text{ K km}^{-1}$ by dozens of times, Fig. 2a. This leads to a very high value of the compression coefficient β for water vapor, which thus becomes saturated in the entire atmospheric column under the canopy. For example, in the tropical forests of Venezuela nighttime lapse rate under the canopy is about $G = 70 \text{ K km}^{-1} \gg G_w$, Fig. 2a. Relative humidity under the canopy is unity at all heights, Fig. 2b.

Above the closed canopy, as well as on open areas like pastures and within open canopies (e.g. Mahrt et al. 2000), nighttime temperature inversions are common, caused by the rapid radiative cooling of the ground surface or canopy, Fig. 2c. Air temperature increases with height up to several hundred meters (Karlsson 2000; Acevedo et al. 2004). These inversions result in condensation of water vapor in the lower cold layer near the surface (or canopy) with formation of fog. On open areas with low vegetation cover, due to the temperature

inversion with $G < 0 < G_w$ and absence of rising fluxes of water vapor, fog moisture does not move upward and remains near the ground surface $z = 0$ where it was formed. However, with increasing solar heating during the daytime and appearance of rising water vapor fluxes, on open areas fog moisture is brought from the ground up to the upper atmospheric layers and ultimately leaves the ecosystem. By contrast, fog formed at night above the closed canopy at $z = z_c$ gravitates to the ground layer $z = 0$ under the canopy, where during the daytime it is prevented from leaking to the external atmosphere by daytime temperature inversion, Fig. 2a.

This analysis illustrates that both large canopy height z_c and high canopy closure of natural forests are important for efficient soil water retention. So far the influence of canopy closure of soil moisture storage has been studied in terms of canopy control over ground surface temperature. During the day, open patches of bare soil heat to a higher temperature than the under-canopy soil (Breshears et al. 1998; Martius et al. 2004). This means that the daytime saturated concentration of water vapor above the soil is higher on bare patches. However, as we have shown, if there are no upward

fluxes of water vapor, as is the case at $G < 1.9 \text{ K km}^{-1}$, the evaporated water can remain near the ground surface and return to soil when air temperatures drop at night. Conversely, when the lapse rate of air temperature is high, all water evaporated during the daytime rapidly leaves the ecosystem. These results suggest that it is the control of the vertical temperature profiles rather than of ground surface temperature per se that allows for efficient water retention under closed canopies.

Discussion

Water is brought to land from the ocean via the atmosphere (A^+) and leaves land both via the atmosphere (A^-) and via runoff (R), $A^+ = A^- + R$. On land, precipitated water P is divided between evapotranspiration E and runoff R , $P = E + R$. In this paper we presented a conceptual framework for analyzing the effects of vegetation cover on local water cycle based on the vegetation impact on the flux A^- of out-flowing atmospheric moisture. Until now, studies of the ecological dimension of the terrestrial water cycle have been almost invariably based on the concept that the main vegetation impact consists in controlling the evapotranspiration term E . Accordingly, the main question of ecological hydrology has been how runoff R changes with changing vegetation cover, and, hence, E , at fixed precipitation P (Watson et al. 1999; Zhang et al. 2001; Engel et al. 2002).

We have shown that conclusions which can be drawn from such a consideration (e.g. that increased evapotranspiration in forested vs unforested catchments leads to decreased runoff, $R = P - E$) can be controversial and misleading, as is the traditional ecological premise that precipitation P is an abiotic constraint on ecosystem processes. We emphasized that, for a consistent representation of the vegetation impact on water cycle to be obtained, one has to take into account the behavior of A^- over areas with different vegetation cover but the same A^+ . Tall, closed canopy forests are efficient not only in extracting water from the flowing air masses but, as shown in the previous section, they are also capable of efficient water retention under the canopy. This allows forests to minimize the outflow of atmospheric moisture A^- as compared to ecosystems with more scarce, short canopy vegetation. In the extreme case of deserts, the inflow of atmospheric water A^+ is equal to the outflow A^- ; moisture is brought in and out from the area without generating precipitation, evaporation, or runoff, Fig. 1a.

Obviously, the framework of analysis as outlined here does not capture the water cycle problem in every detail or conceptual integrity; it leaves a wide field for further speculations and needs to be refined with many more lines of experimental evidence. We have shown that the under-canopy vertical gradients of temperature and relative humidity, Fig. 2, have direct impact on the loss of soil water by the ecosystem. To our knowledge, so far

these profiles have not received a proper attention in the literature, despite the wide-spread of eddy-covariance flux measurements with use of forest towers make the data needed readily available. Measurements of closed-canopy natural temperate and boreal forests are particularly welcome.

The influence of under-canopy vegetation as well as of vegetation height on the development of temperature and humidity gradients deserve special attention. We have shown that the upward fluxes of water vapor from the ground surface arise when the temperature lapse rate exceeds 1.9 K km^{-1} and the water vapor is out of hydrostatic equilibrium. As can be seen from Fig. 2, this situation is realised during the daytime in the open-canopy forests and during the nighttime in the closed-canopy forests. The absolute magnitude of this evaporation flux increases with growing ground temperature. Thus, the taller the vegetation, the lower the ground temperature under the closed canopy and, hence, the smaller the water losses to the atmosphere. These predictions await further exploration and experimental testing.

It is intriguing that the proposed ideas have not found a proper place in modern ecohydrology despite there are many relatively well-studied examples of an ecological control of A^- . It is well known that biogenic condensation nuclei produced by natural forests stabilize local precipitation cycle (Andreae et al. 2004; Koren et al. 2004). There are studies illustrating preferential cloud formation above natural vegetation rather than agricultural fields (e.g. Lyons 2002). Another transparent mechanism is that intense evapotranspiration by trees reduces the characteristic temperature of the atmospheric boundary layer and, hence, facilitates condensation of water vapor. This effect is particularly vivid in mountain cloud forests, which extract an appreciable amount of precipitation from the surrounding clouds. Bruijnzeel (2001) reports that after a hurricane defoliated a patch of mountain cloud forests in Puerto Rico in 1989, the resulting decrease in evapotranspiration by trees caused air temperature to rise significantly, so that the cloud base moved upwards making extraction of water from clouds by plants impossible. The effect disappeared after regrowth of leaves (Bruijnzeel 2001). Finally, in the previous section we described a novel, previously unstudied, mechanism of efficient water retention under closed canopies based on control of the vertical profiles of air temperature. This mechanism prevents the evaporated soil water from leaking into A^- and thus from leaving the ecosystem.

A possible reason for the fact that the idea of a biotic control of A^- has not become popular in ecological studies might be the following. Cloud formation, rainfall extraction and other ways of controlling A^- are all processes with spatial scale by far exceeding the spatial scale of an individual tree. Ecology, as a biological science, traditionally deals with such properties of individuals that can be straightforwardly explained by natural selection. For example, increased evapotranspi-

ration is associated with increased productivity and, hence, it can impart competitive advantage to individual trees (see also Eagelson 2002). By contrast, formation of an extensive cloud cover spreading over many kilometers can be only controlled by synchronous actions of all trees growing on the corresponding area. Unless all trees work in concert, they would not “feel” the advantage of the large-scale regulation of the atmospheric moisture flow. But trees do not work in concert, they compete with each other—hence there seems to be no way for an ecological mechanism of a large-scale control of A^- to evolve. Similar arguments against large-scale biotic regulation of global environmental conditions first appeared in the literature in relation to Gaia hypothesis (e.g. Doolittle 1981; Baerlocher 1990) and persist until now (e.g. Arora 2005).

The apparent contradiction between the observational evidence lending support to biotic regulation of atmospheric moisture flows and the apparent impossibility of explaining this evidence on the basis of natural selection, can be logically resolved (Gorshkov 1984, 1995; Gorshkov et al. 2000, 2004). The solution lies in introducing the fundamental parameter of biotic sensitivity ε_b . Suppose that in the course of genetic modifications there appears a tree capable of a more efficient rainfall extraction from the atmosphere than its neighbors (e.g. via producing more condensation nuclei). Functioning of this single tree will not be able to change the precipitation regime in a large neighborhood in any significant way. However, in the long run and after the many rain events occurring during the tree’s lifespan, such tree will on average receive a slightly greater amount of precipitation, $\Pi + \Delta\Pi$, than its neighbors (Π is the mean amount of precipitation per tree in the considered area and during the considered time period). If this small relative increment of rainfall, $\varepsilon = \Delta\Pi/\Pi$, can be discerned by natural selection, $\varepsilon \geq \varepsilon_b$, then such a tree will receive competitive advantage over its neighbors. Its offspring will have better chances to spread. In the result of competitive exclusion of trees will relatively inefficient rainfall extraction, all trees in the considered area will ultimately be equally efficient in regulating the atmospheric moisture flow. The value of biotic sensitivity with respect to such environmental parameters as atmospheric CO_2 concentration was estimated to be of the order of $10^{-2} - 10^{-3}$ (Gorshkov et al. 2000). In general terms, this means that when all individuals on a large area share a global environmental parameter (e.g. atmospheric composition, precipitation regime etc.), which on a local scale differs from one individual to another by as little as $\varepsilon_b \sim 0.1-1\%$, the process of natural selection appears to be sensitive enough to see such differences. Natural selection then favors individuals performing that small change towards a more optimal value of the global environmental parameter against those not performing such a change. In the result, there appears an evolutionary gradient facilitating propagation of individual properties that make global environmental regulation possible.

Thus, the idea that natural vegetation cover controls the atmospheric moisture outflow A^- to ensure an optimal precipitation regime is free from logical controversies, it is supported by the available evidence and calls for further exploration on the ecological grounds. As illustrated by our analysis, at a given flux of atmospheric moisture inflow A^+ changing vegetation cover can result in dramatic changes of local terrestrial water cycle, from an ideal, tall, closed canopy forest with $A^- = 0$, $P = E_{\max} + A^+$ and $R = A^+$, to an ideal desert with $P = E = R = 0$ and $A^+ = A^-$, with all intermediate stages.

A likely global example of such a dramatic change is the appearance of a continent-wide desert in Australia, which used to be covered by forests 40–100 thousand years ago. Paleodata testify that the monsoonal system bringing moisture to modern Australia is very ancient (Bowman 2002); i.e. the abiotic flux A^+ of incoming moisture is unlikely to have changed in unforested vs forested Australia. However, with the arrival of first humans, who initiated large-scale deforestation on the continent with fires and logging (see discussion in Bowman 2002), the natural forest cover was destroyed, leading to disappearance of once an intensive water cycle on the Australian land.

Within the proposed framework of analysis of the forest-water problem one is led to unambiguously conclude that the long-term stability of water cycle on land is only possible on areas with extensive natural forest cover. It is important to stress that only natural indigenous forests possess the necessary genetic information for ensuring a self-sustainable local water cycle. We have shown that both large canopy height and high degree of canopy closure, especially in low latitudes, is essential for efficient control of A^- . Large-scale deforestation will inevitably result in desertification of the non-coastal continental areas. Conversely, conservation of the remaining natural forests and facilitation of their gradual natural recovery on deforested areas will lead to enhancement and restoration of regional water cycles even in the present-day zones of aridity.

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