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Biotic pump of atmospheric moisture as driver of the hydrological cycle on land

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Abstract

In this paper the basic geophysical and ecological principles are jointly analyzed that allow the landmasses of Earth to remain moistened sufficiently for terrestrial life to be possible. 1. Under gravity, land inevitably loses water to the ocean. To keep land moistened, the gravitational water runoff must be continuously compensated by the atmospheric ocean-to-land moisture transport. Using data for five terrestrial transects of the International Geosphere Biosphere Program we show that the mean distance to which the passive geophysical air fluxes can transport moisture over non-forested areas, does not exceed several hundred kilometers; precipitation decreases exponentially with distance from the ocean. 2. In contrast, precipitation over extensive natural forests does not depend on the distance from the ocean along several thousand kilometers, as illustrated for the Amazon and Yenisey river basins and Equatorial Africa. This points to the existence of an active biotic pump transporting atmospheric moisture inland from the ocean. 3. Physical principles of the biotic moisture pump are investigated based on the previously unstudied properties of atmospheric water vapor, which can be either in or out of hydrostatic equilibrium depending on the lapse rate of air temperature. A novel physical principle is formulated according to which the low-level air moves from areas with weak evaporation to areas with more intensive evaporation. Due to the high leaf area index, natural forests maintain high transpiration fluxes, which support the ascending air motion over the forest and "suck in" moist air from the ocean, which is the essence of the biotic pump of atmospheric moisture. In the result, the gravitational runoff water losses from the optimally moistened forest soil can be fully compensated by the biotically enhanced precipitation at any distance from the ocean. 4. It is discussed how a continent-scale biotic water pump mechanism could be produced by natural selection acting on individual trees. 5. Replacement of the natural forest cover by a low leaf index vegetation leads to an up to tenfold reduction in mean continental precipitation and runoff, in contrast to the previously available estimates made without accounting for the biotic moisture pump. The analyzed body of evidence

testifies that the long-term stability of an intense terrestrial water cycle is unachievable without the recovery of natural, self-sustaining forests on continent-wide areas.

1 Is it a trivial problem, to keep land moistened?

Liquid water is an indispensable prerequisite for all life on Earth. While in the ocean the problem of water supply to living organisms is solved, the landmasses are elevated above the sea level. Under gravity, all liquid water accumulated in soil and underground reservoirs inevitably flows down to the ocean in the direction of the maximum slope of continental surfaces. Water accumulated in lakes, bogs and mountain glaciers feeding rivers also leaves to the ocean. So, to accumulate and maintain optimal moisture stores on land, it is necessary to compensate the gravitational runoff of water from land to the ocean by a reverse, ocean-to-land, moisture flow.

When soil is sufficiently wet, productivity of plants and ecological community as a whole is maximized. With natural selection coming into play, higher productivity is associated with higher competitive capacity. Thus, evolution of terrestrial life forms should culminate in a state when all land is occupied by ecological communities functioning at a maximum possible power limited only by the incoming solar radiation. In such a state local stores of soil and underground moisture, ensuring maximum productivity of terrestrial ecological communities, should be equally large everywhere on land irrespective of the local distance to the ocean. Being determined by the local moisture store, local loss of water to river runoff per unit ground surface area should be distance-independent as well. It follows that in the stationary state the amount of locally precipitating moisture, which is brought from the ocean to compensate local losses to runoff, should be evenly distributed over the land surface.

In the meantime, passive geophysical air fluxes transporting ocean-evaporated moisture to the continents weaken exponentially as they propagate inland. The empirically established characteristic scale length on which such fluxes are damped out is of the order of several hundred kilometers, i.e. much less than the linear dimensions of the continents. Geophysical atmospheric ocean-to-land moisture fluxes cannot therefore compensate local losses of moisture to river runoff that, on forested territories, are equally high far from the ocean as well as close to it. This means that no purely geophysical explanation can be given to the observed existence of highly productive forest ecosystems on continent-scale areas of the order of tens of millions square kilometers, like those of the Amazonia, Equatorial Africa or Siberia.

To ensure functioning of such ecosystems, an active mechanism (pump) is necessary to transport moisture inland from the ocean at a rate dictated by the needs of ecological community. Such a mechanism originated on land in the course of biological evolution and took the form of forest – a contiguous surface cover consisting of tall plants (trees) closely interacting with all other organisms of the ecological community. Forests are responsible both for the initial accumulation of water on continents in the geological past and for the stable maintenance of the accumulated water stores in the subsequent periods of life existence on land. In this paper we analyze the geophysical and ecological principles of the biotic water pump transporting moisture to the continents from the ocean. It is shown that only intact contiguous cover of natural forests having extensive borders with large water bodies (sea, ocean) is able to keep land moistened up to an optimal for life level everywhere on land, no matter how far from the ocean.

The paper is structured as follows. In Sect. 2 the exponential weakening of precipitation with distance from the ocean is demonstrated for non-forested territories using the data for five terrestrial transects of the International Geosphere Biosphere Program (Sect. 2.1); it is shown that no such weakening occurs in natural forests, which points to the existence of the biotic pump of atmospheric moisture (Sect. 2.2); how the water cycle on land is impaired when this pump is broken due to deforestation is estimated in Sect. 2.3. In Sect. 3 the physical principles of the biotic pump functioning are investigated. The non-equilibrium vertical distribution of atmospheric water vapor associated with the observed vertical lapse rate of air temperature (Sect. 3.1) produces an upward directed force, termed evaporative force, which causes the ascending motion of air masses (Sect. 3.2), as well as the horizontal air motions from areas with low evaporation to areas with high evaporation. This physical principle explains the existence of deserts, monsoons and trade winds; it also underlies functioning of the biotic moisture pump in natural forests. Due to the high leaf area index, natural forests maintain powerful transpiration exceeding evaporation from the oceanic surface. The transpiration flux supports ascending fluxes of air and “sucks in” moist air from the ocean. In the result, forest precipitation increases up to a level when the runoff losses from optimally moistened soil are fully compensated at any distance from the ocean (Sect. 3.3). Mechanisms of efficient retention of soil moisture in natural forests are considered in Sect. 3.4. In Sect. 4 it is discussed how the continent-scale biotic pump of atmospheric moisture could be produced by natural selection acting on individual trees. In Sect. 5, based on the obtained results, it is concluded that the long-term stability of a terrestrial water cycle compatible with human existence is unachievable without recovery of natural, self-sustaining forests on continent-wide areas.

2 Passive physical and active biotic mechanisms of ocean-to-land moisture transport

2.1 Physical moisture fluxes in the absence of biotic control

Let F be the horizontal moisture flux equal to the amount of atmospheric moisture passing inland across a unit horizontal length perpendicular to the stream line per unit time, dimension kg $H_2O\,m^{-1}\,s^{-1}$. With air masses propagating inland to a distance x from the ocean (x is measured along the stream line), their moisture content decreases at the expense of the precipitated water locally lost to runoff. Thus, change of F per unit covered distance is equal to local runoff. Due to the homogeneity of the atmosphere, the probability that water vapor molecules join the runoff, does not depend on the distance traveled by these molecules in the atmosphere. It follows that the change dF

of the flux of atmospheric moisture over distance dx is proportional to the flux itself:

$$R(x) \equiv \frac{dF(x)}{dx} = -\frac{1}{l}F(x) \quad \text{or} \quad F(x) = F(0)\exp\{-\frac{x}{l}\}, \tag{1}$$

where $R(x)$ is the local loss of water to runoff per unit surface area, kg H_2O m^{-2} s^{-1}, l is the mean distance traveled by an H_2O molecule from a given site to the site where it went to runoff, $F(0)$ is the value of flux F in the initial point x=0. Parameter l reflects the intensity of precipitation formation processes (moisture upwelling, condensation and precipitation) (Savenije, 1995); the more rapid they are, the shorter the distance l.

As far as a certain amount, E, of the precipitated water evaporates from the surface and returns to the atmosphere, precipitation P is always higher than runoff R:

$$P = E + R \equiv kR, \quad k \geq 1. \tag{2}$$

For example, global runoff constitutes 35% of terrestrial precipitation (Dai and Trenberth, 2002), which gives a global mean $k \approx 3$. Eq. (1) can then be re-written as

$$P(x) = P(0)\exp\{-\frac{x}{l}\} \text{ or } \ln P(x) = \ln P(0) - \frac{x}{l}, \tag{3}$$

where $P(0)=kR(0)$ is precipitation in the initial point x=0.

Linear scale l can be determined from the observed decline of precipitation P with distance x on those territories where the biotic control of water cycle is weak or absent altogether. Such areas are represented by deserts, i.e. territories completely deprived of vegetation cover, as well as by low-productive ecosystems with low leaf area index, open canopies and/or short vegetation cover (semideserts, steppes, savannas, grasslands). We collected data on five extensive terrestrial regions satisfying this criterion, i.e. not covered by natural close-canopy forests, Fig. 1. These regions represent the non-forested parts of five terrestrial transects proposed by the International Geosphere Biosphere Program (IGBP) for studying the effects of precipitation gradients under global change (Canadell et al., 2002), Table 1. Within each region distance x

was counted in the inland direction approximately perpendicular to the regional isohyets (Savenije, 1995), Fig. 1.

Based on the available meteorological data, the dependence of precipitation P on distance x was investigated in each region, Fig. 2a. In all regions this dependence accurately conforms to the exponential law Eq. (3), which is manifested in the high values of the squared correlation coefficients (0.90–0.99), Table 1. This indicates that the possible dependence on x of coefficient k, Eq. (2), which we do not analyze, is weak compared to the main exponential dependence of precipitation P on distance, which is taken into account in Eq. (3). Estimated from parameter b of the linear regression $\ln P = a + bx$ as $l = -1/b$, see Eq. (3), scale length l takes the values of several hundred kilometers, from 220 km in Argentina at 31° S to 870 km in the North America, except for region 4a in Argentina at 45° S, where $l = 93$ km, Table 1. Such a rapid decrease of precipitation has to do with the influence of the high Andean mountain range impeding the movement of westerly air masses coming to the region from the Pacific Ocean (Austin and Sala, 2002). On the island of Hawaii, high (>4 km a.s.l.) mountains also create a large gradient of precipitation which can change more than tenfold over 100 km (Austin and Vitousek, 2000). In West Africa, the value of $l = 400$ km obtained for the areas with $P \leq 1200$ mm year^{-1}, Table 1, compares well with the results of Savenije (1995), who found $l \approx 970$ km for areas with $P \geq 800$ mm year^{-1} and $l \sim 300$ km for the more arid zones of this region.

Total amount of precipitation Π (kg H$_2$O year^{-1}) over the entire path $L \gg l$ traveled by air masses to the inner parts of the continent, $0 \leq x \leq L$, in an area of width D (for river basins D can be approximated by the smoothed length of the coastal line) is, due to Eq. (3), equal to

$$\Pi = D \int_0^L P(x)dx \approx P(0)l D. \quad (4)$$

As is clear from Eq. (4), l represents a characteristic linear scale equal to the width of the band of land adjacent the coast which would be moistened by the incoming oceanic air masses if the precipitated moisture were uniformly distributed over x with a density $P(0)$. For land areas with ordinary orography (regions 1–3, 4b, 5) the mean value of l is about 600 km, Table 1, i.e. it is significantly smaller than the characteristic horizontal dimensions of the continents. Thus, the passive geophysical transport of moisture to land would only be able to ensure normal life functioning in a narrow band near the ocean of a width not exceeding several hundred kilometers; the much more extensive inner parts of the continents would have invariably remained arid. Already at this stage of our consideration we come to the conclusion that in order to explain the observed existence of the extensive well-moistened continental areas several thousand kilometers in length (the Amazon river basin, Equatorial Africa, Siberia), where natural forests are still functioning (Bryant et al., 1997), it is necessary to involve a different, active mechanism of ocean-to-land moisture transport.

2.2 Biotic pump of atmospheric moisture

Let us now consider the spatial distribution of precipitation on extensive territories covered by natural forests. As far as soil moisture content ultimately dictates life conditions for all species in the ecological community, functioning of the community should be aimed at keeping soil moisture at a stationary level optimal for life. Maintenance of high soil moisture content W (units kg H_2O m^{-2}=mm H_2O) enables the ecological community to achieve high power of functioning even when the precipitation regime is fluctuating. For example, transpiration of natural forests in the Amazon river basin, where soil moisture content is high throughout the year (Hodnett et al., 1996), is limited by the incoming solar energy only. It increases during the dry season when the clear sky conditions predominate (da Rocha et al., 2004; Werth and Avissar, 2004). Dry periods during the vegetative season in natural forests of higher latitudes neither bring about a decrease of transpiration (Goulden et al., 1997; Tchebakova et al., 2002). In contrast, transpiration of open ecosystems like savannas, grasslands or shrublands incapable of maintaining high soil moisture content year round, drops radically during the dry season (Hutley et al., 2001; Kurc and Small, 2004).

Change of soil moisture content with time, dW/dt, is linked to precipitation P, evaporation E and runoff R via the law of matter conservation, $dW/dt=P-E-R$. High stationary ($dW/dt=0$) soil moisture content is inseparably associated with a significant runoff, i.e. loss of water by the ecosystem. In areas where neither the surface slope, nor soil moisture content depend on distance x from the ocean, $W(x)=W(0)$, loss of ecosystem water to runoff is spatially uniform as well, $R(x)=R(0)$. It follows that in the stationary case precipitation P, which maintains stationary soil moisture content and compensates the runoff, cannot decrease with distance from the ocean either. The conditions $W(x)=W(0)$ and $R(x)=R(0)$ are incompatible with an exponential decline of $P(x)$, Eq. (3).

When the soil moisture content is sufficiently high, transpiration, which makes up a major part of total evaporation from the surface, E, is dictated by solar energy. Therefore, when x is counted along the parallel, on well-moistened continental areas we have $E(x)=E(0)$, i.e. evaporation does not depend on the distance from the ocean. Coupled with constant runoff, $R(x)=R(0)$, this means that precipitation P is similarly independent of the distance from the ocean, $P(x)=E(0)+R(0)=P(0)$. When the considered area is oriented, and x counted, along the meridian, transpiration increases towards the equator following the increasing flux of solar energy. In such areas, at constant soil moisture content and runoff, precipitation must also grow towards the equator irrespective of the distance from the ocean.

We collected precipitation data for three extensive terrestrial regions spreading along 2.5 thousand kilometers in length each and representing the largest remnants of Earth's natural forest cover (Bryant et al., 1997). These are the Amazon basin, the Congo basin (its equatorial part) and the Yenisey basin, regions 6, 7 and 8 in Fig. 1. As can be seen from Fig. 2b, precipitation in the Amazon and Congo basins is independent of the distance from the coast at around 2000 mm year^{-1}. In the Yenisey river basin, which has a meridional orientation, Fig. 1, precipitation increases with distance from the ocean from about 400 mm year^{-1} at the mouth to about 800 mm year^{-1} on the upper reaches of the river, Fig. 2b.

Similar precipitation, $P(0)=790$ mm year^{-1}, is registered at 125° E in that part of the

North East China Continental Transect (NECT) (region 2), which is closest to the Pacific Ocean, 400 km from the coast. In the meantime, the upper reaches of Yenisey river are about four thousand kilometers away from the Pacific Ocean, and about six thousand kilometers away from the Atlantic Ocean; in fact, it is one of the innermost continental areas on the planet, Fig. 1. Due to the low oceanic temperature in the region of their formation, Arctic air masses that dominate the Yenisey basin (Shver, 1976) are characterized by low moisture content (less than 8 mm of precipitable water in the atmospheric column compared to 16 mm in the Pacific Ocean at NECT (Randel et al., 1996). According to Eq. (1), this moisture content should decrease even further as these air masses move inland to the south.

In other words, if the modern territory of the forest-covered Yenisey basin were, instead, a desert with precipitation of the order of 100 mm year^{-1}, this would not be surprising from the geophysical point of view (indeed, one is not surprised at the fact that the innermost part of NECT and other non-forested regions, Fig. 2a, are extremely arid). This could easily be explained by the character of atmospheric circulation and large distance from any of the oceans. In contrast, the existence of a luxurious water cycle (Yenisey is the seventh most powerful river in the world) as well as the southward increase of precipitation in this area is quite remarkable, geophysically unexpected and can only be explained by functioning of an active biotic mechanism pumping atmospheric moisture from ocean to land. Similar biotic pumps should ensure high precipitation rates throughout the natural forests of the Amazon basin and Equatorial Africa.

It can be concluded, therefore, that all the largest and most powerful river basins must have formed as an outcome of the existence of forest pumps of atmospheric moisture. Forest moisture pump ensures an ocean-to-land flux of moisture, which compensates for the runoff of water from the optimally moistened forest soil. This makes it possible for forests to develop the maximally possible transpiration fluxes that are limited by solar radiation only. Thus, precipitation over forests increases up to the maximum value possible at a given constant runoff (i.e., coefficient k in Eq. (2), the precipitation/runoff ratio, is maximized for a given R.) Forest moisture pump determines both the ultimate distance to which the atmospheric moisture penetrates on the continent from the ocean, as well as the magnitude of the incoming moisture flux per unit length of the coastal line. Dictated by the biota, both parameters are practically independent of the geophysical fluctuations of atmospheric moisture circulation. The biotic pumps of atmospheric moisture enhance precipitation on land at the expense of decreasing precipitation over the ocean. This should lead to appearance of extensive oceanic "deserts" – large areas with low precipitation (see, e.g., Fig. 1 of Adler et al., 2001).

2.3 Deforestation consequences for the water cycle on land

Let us denote as $P_f(x)=P_f(0)$ the spatially uniform distribution of precipitation over a river basin covered by natural forest (low index f stands for forest), which spreads over distance L inland and over distance D along the oceanic coast. Total precipitation Π_f on this territory is equal to $\Pi_f=P_f(0)LD$, where the product $LD=S$ estimates the area occupied by the river basin. According to Eq. (4) and the results of Sect. 2.1, total precipitation Π_d on a territory deprived of the natural forest cover (low index d stands for deforestation, desertification) is equal to $\Pi_d=P_d(0)l_dD$, where $l_d \sim 600$ km. One thus obtains

$$\frac{P_f}{P_d} = \frac{\Pi_f}{\Pi_d} = \frac{P_f(0)L}{P_d(0)l_d}. \tag{5}$$

Equation (5) shows that mean precipitation $\overline{P} \equiv \Pi/S$ in a deforested river basin of linear size L will decrease inversely proportionally to L. For example, for the Amazon river basin $L \approx 3\times10^3$ km, which means that Amazonian deforestation would have led to at least a $L/l_d \approx$ 5-fold decrease of mean precipitation in the region. This effect, i.e. a 80% reduction in precipitation, is several times larger than the available estimates that are based on global circulation models not accounting for the proposed biotic moisture pump. According to such model estimates, deforestation of the Amazon river basin would have led to $(\Pi_f-\Pi_d)/S=270\pm60$ mm year^{-1} (±1 s.e., $n=22$ models) (McGuffie and Henderson-Sellers, 2001), i.e. only 13% of the modern basin mean precipitation of 2100 mm year^{-1} (Marengo, 2004).

Reduction of the characteristic distance of the inland propagation of atmospheric moisture from L to $l_\mathrm{d} \ll L$ is not the single consequence of deforestation. Impairment of the biotic moisture pump in the course of deforestation causes the ocean-to-land flux of atmospheric moisture via the coastal zone to diminish. In the result, the amount of precipitation $P(0)$ in the coastal zone decrease from the initial high biotic value $P(0)=P_\mathrm{f}(0)$ down to $P(0)=P_\mathrm{d}(0)<P_\mathrm{f}(0)$. As will be shown in Sect. 3.3, in the case of complete elimination of the vegetation cover, precipitation in the coastal zone can be reduced practically to zero, $P_\mathrm{d}(0)=0$, see Fig. 4a. In the case when the natural forest is replaced by an open-canopy, low leaf area index ecosystem, cf. Fig. 4b, c in Sect. 3.3, the characteristic magnitude of reduction in $P(0)$ can be estimated comparing the observed values of precipitation $P(0)$ in the coastal zones of forested versus non-forested territories under similar geophysical conditions of atmospheric circulation. A good example is the comparison of the arid ecosystem in the northeast Brazil, the so-called caatinga, which receives about $P_\mathrm{d}(0)=800$ mm year^{-1} precipitation (Oyama and Nobre, 2004), with the forested coast of the Amazon river basin, where $P_\mathrm{f}(0)>2000$ mm year^{-1} (Marengo, 2004), which gives $P_\mathrm{f}(0)/P_\mathrm{d}(0)>2.5$. According to Eq. (5), the cumulative effect of deforestation can amount to more than a tenfold reduction of the mean basin precipitation. The innermost continental areas will be most affected. For example, at $x=1200$ km (in the Amazon river basin this approximately corresponds to the city of Manaus, Brazil) local precipitation will decrease by

$$\frac{P_\mathrm{f}(x)}{P_\mathrm{d}(x)} = \frac{P_\mathrm{f}(0)}{P_\mathrm{d}(0)} \frac{1}{\exp(-x/l)} = \frac{2.5}{\exp(-1200/600)} = 18.5 \tag{6}$$

times, while in Rio Branco, Brazil ($x \approx +2500$ km) precipitation will decrease by 160 times, i.e. the internal part of the continent will turn to a desert. Total river runoff from the basin to the ocean, which is equal to $\int_0^L R(x)dx = \Pi/k$, see Eq. (2), will undergo the same or even more drastic changes as the total precipitation, Eq. (5), due to decrease of coefficient k, Eq. (2).

Ratio $\Pi_f/\Pi_d>10$, Eq. (5), characterizes the power of the biotic pump of atmospheric moisture: the biotically induced ocean-to-land moisture flux does not decrease exponentially with distance from the ocean and it is more than an order of magnitude larger than in the biotically non-controlled, purely geophysical state. At any distance from the ocean and under any fluctuations of external geophysical conditions this moisture flux prevents forest soil from drying. As follows from the above ratio, geophysical fluctuations of the precipitation regime can be no more than 10% as powerful as the biotic pump. Relative fluctuations of river runoff are dictated by fluctuations in the work of the biotic pump, i.e. they do not exceed 10% of its power either. This prevents floods in the forested river basin. Summing up, the undisturbed natural forests create an autonomous cycle of water on land, which is decoupled from whatever abiotic environmental fluctuations. We will now consider the physical and biological principles along which this unique biotic mechanism functions.

3 Physical foundations of the biotic pump of atmospheric moisture

3.1 Non-equilibrium vertical distribution of atmospheric water vapor

In this section we describe a physical effect, which, as we show, plays an important role in the meteorological processes on Earth, but so far remains practically undiscussed in the meteorological literature.

Atmospheric air is in hydrostatic equilibrium (everywhere in Sect. 3 under air we understand dry air). Hydrostatic equilibrium means that air pressure p at a given height z in the atmosphere is balanced by the weight of atmospheric column above z. Change dp of air pressure over vertical distance dz is equal to the weight of air in the atmospheric column of thickness dz (Landau and Lifschitz, 1987):

$$-\frac{dp}{dz} = MNg. \tag{7}$$

Here N (mol m^{-3}) is molar air density at height z, M=29 g mol^{-1} is air molar mass, which is practically independent of z, g=9.8 m s^{-2} is the acceleration of gravity. Atmospheric air is close to ideal gas and conforms to the equation of state, $p=NRT$, where T is absolute air temperature at height z, R=8.3 J K^{-1} mol^{-1} is the universal gas constant. Therefore, Eq. (7) and its solution can be written in the following well-known form (Landau and Lifschitz, 1987):

$$\frac{dp}{dz} = -\frac{p}{h}, \quad p(z) = p_s \exp\{-\int_0^z \frac{dz}{h}\}, \quad h \equiv \frac{RT}{Mg}, \tag{8}$$

where p_s is air pressure at the Earth's surface. In the good approximation of z-independent air molar mass M and temperature T taken to be equal to the mean global surface temperature, $T=T_s$=288 K (15°C), we have $h=h_s$=8.4 km and $p(z)=p_s \exp(-z/h_s)$.

Hydrostatic equilibrium of atmospheric water vapor with molar mass M_w=18 g mol^{-1} is described by Eq. (8) with h replaced by $h_w \equiv RT/M_w g$, $h_{ws} \equiv RT_s/M_w g$=13.5 km.

Immediately above the wet soil or open water surface, water vapor is in the state of saturation. The dependence of partial pressure p_{H_2O} of saturated water vapor on air temperature T is governed by the well-known Clapeyron-Clausius law (Landau et al., 1965; Raval and Ramanathan, 1989):

$$p_{H_2O} = p_{H_2Os} \exp\{\frac{T_{H_2O}}{T_s} - \frac{T_{H_2O}}{T}\}, \; T_{H_2O} \equiv \frac{Q_{H_2O}}{R} \approx 5300\,\mathrm{K}, \tag{9}$$

where low index s refers to corresponding values at the Earth's surface, $Q_{H_2O} \approx 44$ kJ mol^{-1} is the molar latent heat of evaporization. Taking the derivative of the partial pressure of saturated water vapor $p_{H_2O}(z)$ over height z we obtain from Eq. (9):

$$\frac{dp_{H_2O}}{dz} = -\frac{p_{H_2O}}{h_{H_2O}}, \quad p_{H_2O} = p_{H_2Os} \exp\{-\int_0^z \frac{dz}{h_{H_2O}}\}, \; h_{H_2O} \equiv \frac{T^2}{\{-\frac{dT}{dz}\} T_{H_2O}}. \tag{10}$$

Equations (10) have the same form as the hydrostatic equilibrium Eqs. (8) for atmospheric air, but with a different height parameter h_{H_2O}, which, in the case of water vapor, depends on the atmospheric lapse rate Γ of air temperature, $\Gamma \equiv -dT/dz$. The second equation in Eqs. (10) is another mathematical form of the Clapeyron-Clausius law, Eq. (9).

Written for water vapor with h_w instead of h, Eq. (8) formally coincides with Eq. (10) if the equality $h_{H_2O}=h_w$ is fulfilled. This equality can be solved for the atmospheric temperature lapse rate Γ. The obtained solution $\Gamma=\Gamma_{H_2O}$ corresponds to the case when water vapor is saturated in the entire atmospheric column being at the same time in hydrostatic equilibrium, when at any height z its partial pressure is equal to the weight of water vapor in the atmospheric column above z. Equating the scale heights $h_w \equiv RT/M_w g$, Eq. (8) and h_{H_2O}, Eq. (10), we arrive at the following value of Γ_{H_2O}:

$$-\frac{dT}{dz} = \Gamma_{H_2O} = \frac{T_s}{H} = 1.2\,\mathrm{K\,km^{-1}},\ H \equiv \frac{RT_{H_2O}}{M_w g} = 250\,\mathrm{km}. \tag{11}$$

Note that due to the large value of $H \gg h_w$, $h_w/H \approx 0.05 \ll 1$, one can put $\exp(-z/H) \approx 1$ for any $z \le h_w$, which we did when obtaining Eq. (11). In Eq. (11) $\Gamma_{H_2O}=1.2$ K km^{-1} is calculated for the mean global surface temperature $T_s=288\,\mathrm{K}$. Differences in the absolute surface temperatures of equatorial and polar regions change this value by no more than 10%.

The obtained value of $\Gamma_{H_2O}=1.2\,\mathrm{K\ km^{-1}}$, Eq. (11), is a fundamental parameter dictating the character of atmospheric processes.

When $\Gamma<\Gamma_{H_2O}$, water vapor in the entire atmosphere is in hydrostatic equilibrium, but it is saturated at the surface only, i.e. $p_w(z)<p_{H_2O}(T(z))$ for $z>0$ and $p_w(z)=p_{H_2O}(T_s)$ for $z=0$, where p_w is partial pressure of water vapor at height z (and p_{H_2O}, as before, is the saturated pressure of water vapor at $T(z)$). Relative humidity p_w/p_{H_2O} decreases with height. As far as in the state of hydrostatic equilibrium air pressure p and water vapor pressure p_w at a given height z are compensated by the weight of air and water vapor in the atmospheric column above z, in this state macroscopic fluxes of air and water vapor in the atmosphere are absent. Solar radiation absorbed by the Earth's surface makes water evaporate from the oceanic and soil surface, but the evaporated water undergoes condensation immediately at a microscopic distance above the surface, which is of the order of one free path length of water vapor molecules. Energy used for evaporation is ultimately released in the form of thermal radiation of the Earth's surface, with no input of latent heat into the atmosphere.

At $\Gamma > \Gamma_{H_2O}$ the situation is quite different. In this case water vapor is saturated at all heights, $p_w(z) = p_{H_2O}(T(z))$, but it is out of hydrostatic equilibrium. Due to the steep decline of air temperature with height, the atmospheric column above any height z cannot bear a sufficient amount of water for its weight to compensate water vapor partial pressure at height z. The excessive moisture condenses and precipitates. Partial pressure of the saturated water vapor near the surface exceeds the weight of water vapor in the entire atmospheric column. The lapse rate of water vapor partial pressure, $-dp_w/dz$, becomes larger than the weight of a unit volume of water vapor, cf. Eq. (7). There appears an uncompensated force acting on atmospheric air and water vapor. Under this force, upward fluxes of air and water vapor originate that are accompanied by the vertical transport of latent heat. With water vapor continuously leaving the surface layer for the upper atmosphere, saturation of water vapor near the surface is maintained by continuous evaporation. Notably, if evaporation discontinues, atmospheric water vapor ultimately comes to the state of hydrostatic equilibrium with relative humidity at the surface much less than unity.

Violation of hydrostatic equilibrium is manifested as a strong compression of the vertical distribution of water vapor as compared to the distribution of atmospheric air, Eq. (8). At the observed mean atmospheric value of $\Gamma = \Gamma_{ob} = 6.5\,\mathrm{K\ km^{-1}}$, see Appendix A, we obtain from Eqs. (8), (10) and (11):

$$\frac{h}{h_{H_2O}} = \frac{\Gamma_{ob}}{\Gamma_{H_2O}} \frac{M_w}{M} \frac{T_s}{T} \equiv \beta \equiv \beta_s \frac{T_s}{T}, \quad \beta_s = 3.5. \tag{12}$$

The compression coefficient β grows weakly with z due to the z-dependent drop of temperature T corresponding to $\Gamma_{ob} = 6.5\,\mathrm{K\ km^{-1}}$. At $z = h_{H_2O}$, which defines the characteristic vertical scale of water vapor distribution, β increases by 5% compared to its value at the surface β_s. Ignoring this change and putting β constant at $\beta=\beta_s$ we obtain from Eqs. (8), (10), and (12):

$$\frac{p_{H_2O}(z)}{p_{H_2Os}} = \exp\{-\int_0^z \frac{dz}{h_{H_2O}}\} = \exp\{-\int_0^z \beta\frac{dz}{h}\} \approx \{\frac{p(z)}{p_s}\}^{\beta}. \tag{13}$$

Equation (13) 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. Its scale height h_{H_2Os} is calculated as $h_{H_2Os}=h_s/\beta_s=2.4$ km. This theoretical calculation agrees with the observed scale heights ≈2 km of the vertical profiles of atmospheric water vapor (Goody and Yung, 1989; Weaver and Ramanathan, 1995).

Let us now emphasize the difference between the physical picture that we have just described and the traditional consideration of atmospheric motions. The latter resides on the notion of convective instability of the atmosphere associated with the adiabatic lapse rate Γ_a (dry or wet). If an air parcel is occasionally heated more than the surrounding air, it acquires a positive buoyancy and, under the Archimedes force, can start an upward motion in the atmosphere. In such a case its temperature decreases with height z in accordance with Γ_a. If the environmental lapse rate is steeper than Γ_a, $\Gamma>\Gamma_a$, the rising parcel will always remain warmer and lighter that the surrounding air, thus infinitely continuing its ascent. Similar reasoning accompanies the picture of a descending air parcel initially cooled to a temperature lower than that of the surrounding air. On such grounds, it is impossible to determine either the degree of non-uniformity of surface heating responsible for the origin of convection, or the direction or velocity of the resulting movement of air masses. After averaging over a horizontal scale exceeding the characteristic height h of the atmosphere, mean Archimedes force turns to zero. This means that the total air volume above an area greatly exceeding h^2 cannot be caused to move anywhere by the Archimedes force.

According to the physical laws that we have discussed, upward fluxes of air and water vapor always arise when the environmental lapse rate exceeds $\Gamma_{H_2O}=1.2\,\mathrm{K\,km^{-1}}$.

This critical value is significantly lower than either dry or wet adiabatic lapse rates Γ_a (9.8 and $\approx$6 K km^{-1}, respectively). The physical cause of these fluxes is not the non-uniformity of atmospheric and surface heating, but the fact that the partial pressure of water vapor is not compensated by its weight in the atmospheric column, Fig. 3a. The resulting force is invariably upward-directed, Fig. 3b. It equally acts on air volumes with positive and negative buoyancy, in agreement with the observation that atmospheric air updrafts exhibit a wide range of both positive and negative buoyancies (Folkins, 2006). Quantitative consideration of this force, which creates upwelling air and water vapor fluxes and supports clouds over large areas of the Earth's surface, makes it possible to estimate characteristic velocities of the vertical and horizontal motions in the atmosphere, which is done in the next sections.

3.2 Vertical fluxes of atmospheric moisture and air

The Euler equation for the stationary vertical motion of air masses under the force generated by the non-equilibrium pressure gradient of water vapor can be written as follows (Landau and Lifschitz, 1987):

$$\frac{1}{2}\rho\frac{dw^2}{dz} = \frac{dp_{\mathrm{H_2O}}}{dz} - \frac{p_{\mathrm{H_2O}}}{h_w} = (\beta - \beta_0)\rho\gamma g \equiv f_{ev}. \qquad (14)$$

Here $p_{\mathrm{H_2O}}$ is the partial pressure of saturated water vapor at height z under conditions of the observed atmospheric lapse rate Γ_{ob}=6.5 K km^{-1}, i.e. $p_{\mathrm{H_2O}}$ is given by Eq. (13) at $\beta \approx \beta_s$=3.5, $\beta_0 \equiv M_w/M$=0.62, w is the vertical velocity of air masses at height z; $\rho \equiv NM = p/gh$ is air density, $\gamma \equiv p_{\mathrm{H_2O}}/p$ is water vapor mixing ratio; $h = RT/(Mg)$, $h_w = RT/(M_w g)$, p is air pressure at height z given by Eq. (8).

The right-hand part of the first equality in Eq. (14) represents force f_{ev} acting on a unit air volume. It is equal to the difference between the upward directed pressure gradient force $f_\uparrow(z)$ associated with partial pressure of the vertically compressed saturated water vapor, $f_\uparrow(z) \equiv -p_{\mathrm{H_2O}}(z)/dz$, and the downward directed weight $f_\downarrow$ of a unit volume of the saturated water vapor, $f_\downarrow \equiv p_{H_2O}(z)/h_w$: $f_{ev}=f_\uparrow - f_\downarrow$, Fig. 3b. Due to the vertical compression of water vapor as compared to the state of hydrostatic equilibrium, at any height z the pressure of moist air, $p+p_{H_2O}$, becomes larger than the weight of the atmospheric column above z, so force f_{ev} arises. Acted upon by this upward directed force at any height z, volumes of moist air start to rise in order to compensate the insufficient weight of the atmospheric column above z.

As can be seen from Eq. (14), force f_{ev} is proportional to the local concentration $N_{H_2O}=p_{H_2O}/RT$ of the saturated water vapor. As far as the ascending water vapor molecules undergo condensation, the stationary existence of force f_{ev} is only possible in the presence of continuous evaporation from the surface, which would compensate for the condensation and maintain high concentration of water vapor. It is therefore natural to term force f_{ev}, Eq. (14), as the evaporative force.

This force is hundreds of times larger than the force associated with the horizontal barometric gradient in the atmosphere, Fig. 3b. The barometric gradient force accelerates air masses up to certain velocities when the relevant Coriolis force caused by Earth's rotation comes into play. It grows proportionally to velocity and is perpendicular to the velocity vector. These two forces along with the centrifugal force of local rotational movements, the turbulent friction force describing the decay of large air eddies into smaller ones, and the law of momentum conservation together explain the observed atmospheric circulation patterns like gradient winds in cyclones and anticyclones, cyclostrophic winds in typhoons and tornadoes, as well as geostrophic winds in the upper atmosphere where turbulent friction is negligible. However, the origin, magnitude and spatial distribution of the horizontal barometric gradient – the primary cause of atmospheric circulation – has not so far received a satisfactory explanation (Lorenz, 1967). Below we show that the observed values of the barometric gradient are determined by the magnitude of the evaporative force. The various patterns of atmospheric circulation correspond to particular spatial and temporal changes in the fluxes of evaporation, so the evaporative force drives the global atmospheric circulation.

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In agreement with Dalton's law, partial pressures of different gases in a mixture independently come in or out of the equilibrium. The non-equilibrium state of atmospheric water vapor cannot bring about a compensating deviation from the equilibrium of the other air gases, to zero the evaporative force. (In this hypothetical case the vertical distribution of air would be "overstretched" along the vertical, in contrast to the distribution of water vapor which is vertically compressed as compared to the state of hydrostatic equilibrium.) A non-equilibrium vertical distribution of air concentration would initiate downward diffusional fluxes of air molecules working to restore the equilibrium. As soon as air molecules undergo a downward diffusional displacement, the weight of the upper atmospheric column diminishes rendering partial pressure of the water vapor uncompensated, and the evaporative force reappears. (A similar effect when a fluid-moving force arises in the course of diffusion of fluid mixtures with initially non-equilibrium concentrations is inherent to the phenomenon of osmosis.) Thus, the only possible stationary state in the case of non-equilibrium vertical distribution of water vapor, Eq. (13), is the dynamic state when parcels of moist air move under the action of the evaporative force. According to the law of matter conservation, movement trajectories should be closed for air molecules and partially closed (taking into account condensation and precipitation of moisture) for the water vapor. The particular shape of these trajectories will be dictated by the boundary conditions.

In the stationary case the ascending flux of moist air removing wN_{H_2O} mol water vapor from unit area per unit time, where w is vertical velocity, should be compensated by the incoming flux K bringing moisture to the considered area, $K=wN_{H_2O}$. If K is fixed by the flux of evaporation from the Earth's surface, $K=E$, then any increase of vertical velocity w will lead to a decrease in the concentration of the water vapor N_{H_2O}, and, as a consequence, decrease of the evaporative force, Eq. (14). As soon as the evaporative force diminishes, vertical velocity w also diminishes due to dissipation of the energy of moving air masses. Thus, the stable stationary value of the vertical velocity w is equal to $w=E/N_{H_2O}$. For the global mean value of $\overline{E}\approx10^3$ kg H_2O m^{-2} year^{-1} $\approx55\times10^3$ mol m^{-2} year^{-1} and saturated water vapor concentration at the surface $N_{H_2Os}=0.7$ mol m^{-3}

at the global mean surface temperature T_s=288 K we obtain

$$w_{\overline{E}} = \overline{E} / N_{H_2O} = 2.5\,\mathrm{mm\,s^{-1}}. \qquad (15)$$

If K is determined by evaporation maintained by horizontal heat fluxes or directly by the horizontal influx of moisture from the neighboring areas, its value can exceed the local mean evaporation E. In such a case vertical velocity of moist air movement under the action of the evaporative force can reach its maximum value $w_{\max}$, when the evaporative force accelerates air parcels along the entire atmospheric column. Using the expression for air pressure p given by Eq. (8), the approximate equality of Eq. (13), and recalling that $\rho = p/gh$, one can estimate $w_{\max}$ from Eq. (14) as

$$w_{\max} = \sqrt{2(\beta-\beta_0)g\gamma_s \int_0^\infty dz\, \exp\{-\int_0^z (\beta-1)\frac{dz'}{h}\}} \approx \sqrt{2\gamma_s g h_s} \approx 50\,\mathrm{m\,s^{-1}}. \qquad (16)$$

where $\gamma_s \equiv p_{H_2Os}/p_s \approx 2\times10^2$, h_s=8.4 km, g=9.8 m s^{-2}.

The obtained theoretical estimate Eq. (16) is in good agreement with the maximum updraft velocities observed in typhoons and tornadoes (e.g. Smith, 1997). The value of $w_{\max}$, Eq. (16), exceeds the global mean upward velocity $w_{\overline{E}}$, Eq. (15), which is stationarily maintained by the global mean evaporation $\overline{E}$ at the expense of solar energy absorbed by the Earth's surface, by 2×10^4 times. Such velocities can be attained if only there is a horizontal influx of water vapor and heat into the considered local area where the air masses ascend, from an adjacent area which is 2×10^4 times larger, i.e. from a distance one hundred of times larger than the linear size of the tornado.

Movement of air masses under the action of the evaporative force follow closed trajectories, which have to include areas of ascending, descending and horizontal motion. The vertical pressure difference Δp_z associated with the evaporative force is equal to $\Delta p_z \approx p_{H_2O} \approx 2\times10^2 p_s$, where $p_s = 10^5$ Pa is atmospheric pressure at the Earth's surface. The value of Δp_z, a few per cent of standard atmospheric pressure, should give the scale of atmospheric pressure changes at the sea level in cyclones and anticyclones; this agrees well with observations. Given that the scale length r of the areas with consistent barometric gradient does not exceed several thousand kilometers, at $r \approx 2 \times 10^3$ km the horizontal barometric gradient is estimated as $\partial p / \partial x \sim \Delta p_z / r \sim 1$ Pa km^{-1}. Again, this theoretical estimate agrees well with characteristic magnitude of horizontal pressure gradients observed on Earth.

Taking into account that movement of air masses under the action of the evaporative force with a mean vertical velocity $w_{\overline{E}}$, Eq. (15), is the cause of the turbulent mixing of the atmosphere, it is also possible to obtain a theoretical estimate of the turbulent diffusion coefficient for the terrestrial atmosphere. If there is a gradient of some variable C in the atmosphere, flux F_c of this variable is proportional to the gradient $\partial C / \partial z$ and is described by the equation of turbulent (eddy) diffusion, $F_c = -\nu \partial C / \partial z$, where the eddy (turbulent) diffusion coefficient ν (m^2 s^{-1}) (kinematic viscosity) does not depend on the nature of the variable or its magnitude. This relationship is formally similar to the equation of molecular diffusion. However, molecular diffusion fluxes are caused by concentration gradients alone, with no forces acting on the fluid. Molecular diffusion coefficient is unambiguously determined by the physical properties of state, in particular, by the mean velocity and free path length of air molecules. In contrast, turbulent fluxes are caused by air eddies that are maintained by certain physical forces acting on macroscopic air volumes and making them move; thus, the eddy diffusion coefficient depends on air velocity. Therefore, using the empirically established eddy diffusion coefficient for the determination of the characteristic air velocity via the scale length of the considered problem, a common feature of many theoretical studies of global circulation, e.g., (Fang and Tung, 1999), represents a circular approach. It sheds no light on the physical nature and magnitude of the primary forces responsible for air motions.

In the meantime, the theoretically obtained air velocity $w_{\overline{E}}$, Eq. (15), and the scale height $h_{H_2O} \sim 2$ km of the vertical distribution of atmospheric water vapor make it possible to estimate the global mean atmospheric eddy diffusion coefficient (which, for atmospheric air, coincides with the coefficient of turbulent kinematic viscosity) as $\nu \sim w_{\overline{E}} h_{H_2O} \sim 5$ m^2 s^{-1}. This estimate agrees in the order of magnitude with the phenomenological value of $\nu \sim 4$ m^2 s^{-1} used in global circulation studies (Fang and Tung, 1999). This lends further support to the statement that the observed turbulent processes in the atmosphere and atmospheric circulation are ultimately conditioned by the process of evaporation of water vapor from the Earth's surface at the observed Γ_{ob}=6.5 K km^{-1} and are generated by the evaporative force, Eq. (14). Finally, we note that the evaporative force equally acts on all gases irrespective of their molar mass, so when the ascending air parcels expand, the relative amount of the various dry air components does not change. This explains the observed constancy of air molar mass M (i.e. constant mixing ratios of the dry air constituents) over height z.

3.3 Horizontal fluxes of atmospheric moisture and air

Based on the physical grounds discussed in the previous two sections it is possible to formulate the following physical principle of atmospheric motion. If evaporation fluxes in two adjacent areas are different, there appear horizontal fluxes of moist air from the area with weaker evaporation to the area where evaporation is more intensive. The resulting directed moisture flow will enhance precipitation in the area with strong evaporation and diminish precipitation in the area with weak evaporation. In particular, it is possible that moisture will be brought by air masses from dry to wet areas, i.e. against the moisture gradient. This is equivalent to the existence of a moisture pump supported by the energy spent on evaporation. This pattern provides clues for several important phenomena.

First, it explains the existence of deserts bordering, like Sahara, with the ocean. In deserts where soil moisture stores are negligible, evaporation from the ground surface is practically absent. Atmospheric water vapor is in the state of hydrostatic equilibrium, Eq. (8), so the evaporation force, Eq. (14), in desert is equal to zero. In contrast, evaporation from the oceanic surface is always substantial. The upward-directed evaporative force is always greater over the ocean than in the desert. It makes oceanic air rise and effectively "sucks in" the desert air to the ocean, where it replaces the rising oceanic air masses at the oceanic surface, Fig. 4a. The backward ocean-to-desert air flux occurs in the upper atmosphere, which is depauperate in water vapor. This moisture-poor air flux represents the single source of humidity in the desert. Its moisture content determines the stationary relative humidity in the desert. To sum up, due to the absence of surface evaporation, deserts appear to be locked for oceanic moisture year round, Fig. 4a.

In less arid zones like savannas, steppes, irrigated lands, some non-zero evaporation from the surface is present throughout the year. Land surface temperature undergoes greater annual changes than does the surface temperature of the thermally inertial ocean. In winter, the ocean can be warmer than land. In such a case partial pressure p_{H_2O} of water vapor in the atmospheric column over the ocean is higher than on land. The evaporative force is greater over the ocean as well. In the result, a horizontal land-to-ocean flux of air and moisture originates, which corresponds to the well-known phenomenon of winter monsoon (dry season), Fig. 4b.

In contrast, as the warm season sets in, land surface heats up more quickly than does the ocean and, despite the preceding dry winter season, the evaporation flux from the land surface can exceed the evaporation flux over the colder ocean. There appears an air flux transporting oceanic moisture to land known as summer monsoon (rainy season), Fig. 4c. In the beginning, only the wettest part of land, the coastal zone, can achieve an evaporation flux in excess of the oceanic one. This initiates the fluxes of moist air, precipitation and further enhances terrestrial evaporation. As the evaporation flux grows, so do the fluxes of moist air from the ocean. They gradually spread inland to the drier parts of the continent and weaken exponentially with distance from the ocean, Fig. 2a. Notably, an indispensable condition for summer monsoon is the considerable store of water on land, which sustains appreciable evaporation year round. In deserts, in spite of even greater seasonal differences between land and ocean surface temperatures, the evaporation on land is practically absent, so no ocean-to-land fluxes of moisture can originate in any season.

Although the vegetation of savannas does support some non-zero ground stores of moisture and fluxes of transpiration, the absence of a contiguous cover of tall trees with high leaf area index prevents such ecosystems from increasing transpiration up to a level when the appearing flux of atmospheric moisture from the ocean would compensate runoff from the optimally moistened soil. The biotic pump of atmospheric moisture does not work in such scarcely vegetated ecosystems; precipitation weakens exponentially with distance from the ocean, Fig. 2a.

The phenomenon of trade winds (Hadley circulation) can also be explained on these grounds. As far as in the stationary case solar radiation is the source of energy for evaporation, the increase of the solar flux towards the equator should be accompanied by a corresponding increase of evaporation flux E, evaporative force f_{ev}, Eq. (14), and vertical velocity w, Eq. (15). The intensive ascent of moist air on the equator has to be compensated by horizontal air fluxes originating at higher latitudes and moving towards the equator, where they ascend and travel back in the upper atmosphere, Fig. 4d. Subsidence of dry air masses at the non-equatorial border of Hadley cells leads diminishes water vapor concentration in these areas, producing an additional, unrelated to the geography of solar radiation, decrease of the evaporative force. This creates favorable conditions for Ferrel circulation, i.e. movement of air masses from subtropics to higher latitudes.

Finally, natural forest ecosystems can ensure the necessary ocean-to-land flux of atmospheric moisture in any direction. Due to the high leaf area index, which is equal to the total area of all leaves of the plant divided by the plant projection area on the ground surface, the cumulative evaporating surface of the forest can be much higher than the open water surface of the same area. Forest transpiration can be several times higher than the evaporation flux in the ocean, approaching the maximum possible value limited by solar radiation. Maximum evaporation, corresponding to the global mean solar flux absorbed by the Earth's surface $I=150\,\mathrm{W\,m^{-2}}$, is about $I/(\rho_{H_2Os}Q)\approx 2\ \mathrm{m\,year^{-1}}$, where $\rho_{H_2Os}=10^{-2}\ \mathrm{kg\,m^{-3}}$ is the water vapor density at the surface at $T_s=288\,\mathrm{K}$. In the meantime, global evaporation from the oceanic surface is substantially lower, about $1.2\,\mathrm{m\,year^{-1}}$ (L'vovitch, 1979). Intensive ascending fluxes of moist air generated by forest transpiration induce the compensating low-level horizontal influx of moisture-laden air from the ocean. When the incoming air fluxes ascend, the oceanic moisture condenses and precipitates over the forest. Unburdened of moisture, dry air returns to the ocean from land in the upper atmosphere.

As far as in the natural forest with high leaf area index transpiration, limited by solar radiation only, can exceed evaporation from the ocean all year round, the corresponding horizontal influx of oceanic moisture into the forest can persist throughout the year as well, Fig. 4e. Here lies the difference between the undisturbed natural forest and open ecosystems with low leaf area index. In the latter ecosystems the ocean-to-land flux can only originate when the land surface temperature is much higher than the oceanic surface temperature, Fig. 4b, c.

Total force causing air above the natural forest canopy to ascend is equal to the sum of local evaporative forces generated by the transpiration of individual trees. On the other hand, this cumulative force acts to pump the atmospheric moisture inland from the ocean via the coastline. Therefore, the total flux $F(0)$ of moisture from the ocean to the river basin, which compensates the total runoff, is proportional to the number of trees in the forest and, consequently, to the area of the forest-covered river basin. According to the law of matter conservation (continuity equation), the horizontal flux of moisture via the vertical cross-section of the atmospheric column along the coastline, $F(0){=}W_a uDh$, where D is the coastline length (basin width), h is the height of the atmospheric column, u is horizontal air velocity and W_a (kg H_2O m^{-2}) is the moisture content in the atmospheric column, should be equal to the ascending flux of moisture across the horizontal cross-section of the atmospheric column above the entire river basin of area DL, which is equal to $W_a w_f DL$, where w_f is the vertical velocity of air motion. From this we obtain $w_f{=}uh/L$.

The magnitudes of the velocities w_f and u are determined by the condition that the power developed by the evaporative force in the atmospheric column above the forest canopy is equal to the power of dissipation of the horizontal air fluxes near the Earth's surface. The power A_{ev} of the evaporative force f_{ev} over the forest canopy is equal to $A_{ev} = f_{ev} w_f h_{H_2O} DL \sim \rho(\beta-\beta_0)\gamma_s g w_f h_{H_2O} DL$, see Eq. (14). The power of dissipation A_{fr} associated with the friction force f_{fr} is equal to $A_{fr}{=}f_{fr} u z_s DL$. Friction force is equal to $f_{fr}=\rho\nu\partial^2 u/\partial z^2=\rho\nu u/z_s^2$ (Lorenz, 1967), where z_s is the height of the surface layer where velocity u changes rapidly with height due to substantial friction. Taking into account that $\nu\sim w_{\mathrm{f}}h_{\mathrm{H_2O}}$, we have $f_{fr}=\rho w_{\mathrm{f}}h_{\mathrm{H_2O}}u/z_s^2$. Equating powers A_{ev} andA_{fr} we obtain $(\beta-\beta_0)\gamma_s g\sim u^2/z_s$. Taking $z_s\sim$ 25 m and $\gamma_s\sim 2\times10^{-2}$, we obtain an estimate of the horizontal velocity $u\sim 4\,\mathrm{m\,s^{-1}}$, which coincides in its order of magnitude with the observed global mean wind speed (Gustavson, 1979).

Horizontal velocity u sufficient for the compensation of the river runoff R from the optimally moistened soil in the river basin of area DL covered by natural forest is calculated from the relationship $RDL=W_a uD$. For example, for the tropical Amazon we have $R\sim10^3$ kg $\mathrm{H_2O}$ $\mathrm{m^{-2}}$ $\mathrm{year^{-1}}$ (Marengo, 2004), $W_a\sim50$ kg $\mathrm{H_2O}$ $\mathrm{m^{-2}}$ (Randel et al., 1996) and $L\sim3000$ km, Fig. 2b, so $u=RL/W_a\sim1.6\,\mathrm{m\,s^{-1}}$ and $w_{\mathrm{f}}=Rh/W_a\sim 5.6\,\mathrm{mm\,s^{-1}}$ at $h=8.4$ km, see Eq. (8). It follows from the relationship $w_{\mathrm{f}}>w_{\overline{E}}\approx2.5\,\mathrm{mm\,s^{-1}}$, where $w_{\overline{E}}$, Eq. (15), corresponds to the global mean evaporation from the Earth's surface, that only undisturbed natural forests with closed canopies and high leaf area index can maintain optimal soil moisture content at any distance from the ocean at the expense of pumping atmospheric moisture from the ocean. This is because only such ecosystems are able to ensure transpiration fluxes exceeding the fluxes of evaporation from the open water surface of the ocean.

The outlined principles of horizontal air motions, Fig. 4a–e, allow one to make several generalizations. First, if the coastal zone of width $l\sim600$ km is deforested, the flow of oceanic moisture to the inner part of the continent is switched off, thus dooming the inner continental forest to perish. On the other hand, a narrow band of forest along the coast cannot develop a power high enough for pumping atmospheric moisture from the ocean in amounts sufficient for keeping the entire continent moistened and for supporting powerful river systems. Moreover, if the inner part of the continent is turned to an extensive desert with negligible evaporation, the originating horizontal land-to-ocean air fluxes, cf. Fig. 4a, may become more powerful than the small ocean-to-land air fluxes maintained by the narrow forest band near the coast. In such a case the forest will aridify and die out despite its immediate closeness to the open water surface of the ocean or inner sea. Third, as far as the direction and velocity of horizontal air motions is dictated by the difference in evaporative forces between the considered areas, it should apparently be easier for the biota to pump atmospheric moisture from a cold ocean with low evaporation, i.e. from an ocean situated in higher latitudes than the forested river basin itself. This explains the stable existence of the forested basins of the Great Siberian Rivers.

In the meantime, pumping of atmospheric moisture to land from a warm ocean, which is realized in the tropical river basins, is a more complex problem. If the flux of evaporation from the warm oceanic surface exceeds the transpiration flux from the forest canopy, there appears a land-to-ocean flux of atmospheric moisture, which, together with the river runoff, will act to deplete forest moisture stores. To avoid this, it is necessary for the forest to maintain high transpiration at all times. Even if during the most unfavorable season of the warmest oceanic temperatures this flux will nevertheless prevent the moisture of the atmospheric column above the forest from being blown away to the ocean. Thus, we come to a non-trivial conclusion that the more moisture is evaporated from the land surface, the less moisture is lost by land. High transpiration rates observed in the Amazon forests during the dry season (da Rocha et al., 2004) can well serve this goal.

In the next section we consider the physical principles of efficient moisture retention under the closed canopy, which prevent uncontrollable losses of atmospheric moisture and allow the trees to maintain high transpiration power independent of rainfall fluctuations.

3.4 Water preservation by closed canopies

In natural forest ecosystems with well-developed closed canopies the daytime air temperature 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 Anhuf, 2001), because the incoming solar radiation is predominantly absorbed in the canopy. When canopy temperature T_c exceeds ground temperature T_g, the under-canopy lapse rate becomes negative, $\Gamma = -dT/dz \approx (T_c - T_g)/z_c < 0$, where z_c is the canopy height reaching several tens of meters in natural forests. As far as $\Gamma < 0 < \Gamma_{H_2O} = 1.2\,\mathrm{K\,km^{-1}}$, in this case, according to the results of the previous sections, water vapor under the canopy remains in hydrostatic equilibrium. Evaporation from soil and the upward fluxes of water vapor from beneath the canopy are absent. Water vapor partial pressure p_w conforms to Eq. (8) (with p and h changed for p_w and h_w, respectively) and remains practically constant under the canopy with $z_c \ll h_w$, $p_w(z) \approx p_w(0) \equiv p_{ws}$. Relative humidity $RH(z)$, which is equal to 100% immediately above the wet soil surface, decreases with height as $RH(z) = 1/\exp\{[T_w/T_s] - [T_w/T(z)]\}$, cf. Eqs. (8) and (9) (Szarzynski and Anhuf, 2001).

The daytime hydrostatic equilibrium of the saturated water vapor above the ground surface under the closed canopy prevents biotically uncontrolled losses of soil water to the upper atmosphere. This mechanism explains why the ground surface of undisturbed closed-canopy forests always remains wet, which is manifested as low fire susceptibility of undisturbed natural forests with closed canopies (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).

At nighttime the soil surface under the closed canopy is warmer than the canopy due to the rapid radiative cooling of the latter (Shuttleworth et al., 1985; Szarzynski and Anhuf, 2001). The observed temperature lapse rate Γ_{ob} is positive and can exceed the mean global value of $6.5\,\mathrm{K\,km^{-1}}$ by dozens of times (Szarzynski and Anhuf, 2001). This leads to a very high value of the compression coefficient β, Eq. (12), for water vapor. Water vapor is saturated in the entire atmospheric column under the canopy. For example, in the tropical forests of Venezuela the nighttime lapse rate under the canopy is about Γ_{ob}=70 K km^{-1}$\gg$$\Gamma_{H_2O}$ (Szarzynski and Anhuf, 2001), which corresponds to β=36, see Eq. (12). Relative humidity under the canopy is then equal to 100% at all heights. In spite of the huge evaporative force, Eq. (14) that arises at large values of β, the ascending fluxes of air under the closed canopy of natural forests remain small due to the high aerodynamic resistance of the trees.

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. Air temperature increases with height up to several hundred meters (Karlsson, 2000; Acevedo et al., 2004). Temperature inversions result in the condensation of water vapor in the lower cooler layer near the surface (or canopy) often accompanied by formation of fog. As far as at $\Gamma<0<\Gamma_{H_2O}$ there are no ascending fluxes of water vapor, fog moisture remains near the ground surface z=0. However, with increasing solar heating during the daytime and Γ growing up to $\Gamma>\Gamma_{H_2O}$, there appear upward water vapor fluxes. On open areas and areas covered by low vegetation, fog moisture is then taken away from the ground layer 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 the moisture is prevented from leaking to the upper atmosphere by the daytime temperature inversion.

This analysis illustrates that both large canopy height and high degree of canopy closure inherent to undisturbed natural forests are important for efficient soil water retention.

During the day air temperature above the forest canopy rapidly decreases with height, $\Gamma\gg\Gamma_{H_2O}$ (Szarzynski and Anhuf, 2001). Transpiration of water vapor from the leaf stomata lead to formation of an ascending flux of water vapor and air, which supports clouds. When stomata are closed and transpiration ceases, the ascent of air masses discontinues as well. The evaporative force f_{ev}, Eq. (14), vanishes; liquid atmospheric water, no longer supported, precipitates under gravity. Precipitation fluxes can be additionally regulated by the biota via directed change of various biogenic condensation nuclei. Observations indicate that stomata of most plants close during the midday (Bond and Kavanagh, 1999; Goulden et al., 2004). This allows one to expect that precipitation on land should predominantly occur in the second part of the day. In the ocean, where there is no diurnal biological cycle of water vapor ascent, the diurnal cycle of precipitation is not expected to be pronounced either. These predictions are in agreement with the available data on the diurnal cycle of rainfall in the tropics (Nesbitt et al., 2003).

4 Biological principles of the biotic pump of atmospheric moisture

Based on the analysis of the spatial distribution of precipitation over forested and non-forested areas, in Sect. 2 we demonstrated that the biotic pump of atmospheric moisture must exist, as follows from the law of matter conservation and the small length $l{\sim}600$ km of the physical dampening of precipitation fluxes on land. In Sect. 3 the physical principles of the biotic pump were described. Regarding the ecological and biological grounds of its functioning, the inherently complex applications of the laws of physics by natural forest ecosystems have been refined and polished during the hundred million years of evolution, so it will hardly be ever possible to study them in detail. However, it is worthy to check that the existence of the biotic moisture pump does not contradict the known biological principles of life organization.

Moisture enters a forest-covered river basin via the linear coastline, while it is spent on the two-dimensional area of the basin. Area-specific loss of water to runoff is independent of distance x from the ocean, $R(x){=}R(0)$, due to the constant moisture content $W(x){=}W(0)$ of the uniformly moistened forest soil, Sect. 2. Thus, the atmospheric flux of moisture via the coastline (dimension kg H_2O m^{-1} s^{-1}), which compensates river runoff in a river basin of length L (L is counted inland from the coast), $F(0){=}\int_0^L R(x)dx$, should grow linearly with basin length L, $F(0){=}R(0)L$. The biotic moisture pump works to ensure this effect. Without biotic pump the value of $F(0)$ would be independent of L, cf. Eq. (1), $R(x)$ would exponentially decrease with growing distance x from the ocean, while the mean basin runoff would decrease inversely proportionally to L, Sect. 2.

The problem of meeting the demand of matter or energy of an n-dimensional area (n=2 for the river basin) by a flux of matter or energy via an ($n-1$)-dimensional area (one-dimensional coastline of the river basin) is a fundamental biological problem repeatedly faced by life in the course of evolution. Living organisms consume energy via the two-dimensional body surface and spend it within the three-dimensional body volume. In very much the same manner as local runoff and precipitation in natural forests are maintained independent of the distance from the ocean, living cells are supplied by an energy flux that is on average independent of the size of the organism.

Analysis of the metabolic power of living organisms from bacteria to the largest mammals revealed that, independent of organismal body size, mean energy consumption of living tissues constitutes 1–10 W kg^{-1} (Makarieva et al., 2005a,b,c). The smallest organisms like unicells can satisfy their energetic needs at the expense of passive diffusion fluxes of matter, if their linear body size L is much less than the scale length l of exponential weakening of the diffusion matter fluxes. For example, bacteria have to transport the food obtained via cell surface over less than one micron (bacterial cell length), while the distance between elephant's trunk (the food-gathering organ) and its brain or heart is about ten million times longer. Passive diffusion fluxes cannot meet the energetic demands of large organisms with $L \gg l$. Thus, large organisms had to invent active pumps (e.g., lungs, heart), which pump matter and energy into the organism and distribute them within it. Similarly, passive geophysical fluxes of moisture can ensure sufficient soil moistening at small distances $L \ll l$ from the ocean, while to keep large territories with $L \gg l$ biologically productive, the biotic pump of atmospheric moisture is necessary.

Biological pumps like lungs or heart, which supply living cells of large organisms with energy and matter, are complex mechanisms. They have evolved in the course of natural selection of individual organisms possessing most efficient pumps in the population. The biotic pump of atmospheric moisture, grounded in plant physiology and including sensing of the environmental parameters and reaction to their change by a corresponding change in the physiological state of the plant (e.g., stomata opening/closure), is also a complex, highly-ordered biotic mechanism. However, in contrast to an individual organism of a given body size, natural forest ecosystem of linear size L is not an object of natural selection, as it consists of a large number of competing individuals like, e.g., trees. A fundamental theoretical question is therefore how the biotic pump of atmospheric moisture or any other biotic mechanisms regulating regional or global environmental parameters could have been produced by natural selection which acts on individual organisms.

The discussion around this question has already a several decades' history in the biological literature, e.g., (Doolittle, 1981; Baerlocher, 1990), and is largely organized along two opposing lines of reasoning. The first attitude (shared, among others, by many supporters of the Gaia hypothesis) is that the state when ecosystem as a whole regulates regional or global environment for its benefit appears is the most probable macroscopic physical property of the ecosystem, like Maxwell's or Boltzmann's distributions characterizing gas as a whole rather than particular molecules. No natural selection is needed for such a state to become established.

The opponents of this attitude implicitly recognize that the biotic control of any environmental conditions is a very complex process having nothing to do with the physical processes of thermodynamics (either linear, or non-linear). As such, it could have only evolved as the product of natural selection, like any other complex function of living organisms. However, continue the opponents, one individual performing some job on environmental regulation makes the regional environment slightly better for all individuals, including those who do nothing. Hence, individuals capable of environmental control do not have a selective advantage over their conspecifics. Moreover, doing something for the global environment, the regulators spend their metabolic energy and can therefore lose to the non-regulators who spend all their energy on competitive interaction. Thus, since natural selection cannot favor regulatory capabilities on the individual level, the biotic control of regional or global environment could not have appeared in the course of biological evolution and, hence, it does not exist.

Most recently, this logical opposition could be followed in the discussion of A. Kleidon (first attitude) and V. Arora (second attitude) (Kleidon, 2004, 2005; Arora, 2005). The contradiction between the observational evidence lending support to the existence of biotic mechanisms of environmental control, including the biotic pump of atmospheric moisture, and the apparent impossibility of finding a logically coherent physical and biological explanation of this evidence, prevents the wide scientific recognition of the decisive role the natural ecosystems play in maintaining the environment in a state suitable for life, including the biotic control of water cycle on land.

This contradiction is solved in the biotic regulation concept (Gorshkov, 1995; Gorshkov et al., 2000). First, it was shown that the degree of orderliness of ecological and biological systems is about twenty orders of magnitude higher than the degree of orderliness of any of the so-called physically self-organized open non-linear systems (Gorshkov and Makarieva, 2001). Natural selection of competing biological objects (individuals) is the only way by which the biological information accumulates in the course of evolution. Therefore, any highly organized life property, including the biotic pump of atmospheric moisture, can only originate as a product of natural selection.

Second, a fundamental parameter of the biotic sensitivity ε_b was introduced (Gorshkov, 1984, 1995; Gorshkov et al., 2000, 2004), which, if finite, makes it possible for various biotic mechanisms of global environmental control (including the biotic pump of atmospheric moisture) to originate in the course of natural selection acting on individual organisms (trees). Under tree below we shall understand an individual tree together with all the other organisms of the ecological community that are rigidly correlated with this tree (e.g., soil bacteria and fungi). Each tree works to maintain optimal soil moisture conditions on the area it occupies. If the occasional relative change of soil moisture content under the tree is less than ε, the tree does not react to it. According to our estimates (Gorshkov et al., 2000, 2004), biotic sensitivity with respect to changes of major environmental parameters is of the order of $\varepsilon \sim 10^{-2}-10^{-3}$. Trees' ability to remain highly sensitive to environmental conditions is genetically programmed via individual selection of trees with $\varepsilon < 10^{-2}$. Those individuals who have lost this ability and have a poor sensitivity $\varepsilon > 10^{-2}$, suffer from large fluctuations of local soil moisture content, which adversely affect their biological performance. Such trees lose competitiveness and are forced out from the population by normal trees with $\varepsilon < 10^{-2}$.

If the local soil moisture content occasionally deviates from its optimum value by a relative amount exceeding biotic sensitivity ε, the tree does notice this change and initiates compensating environmental processes in order to increase or decrease the local moisture content restoring the optimal conditions. Environmental impact of the tree can take a variety of forms, including change of transpiration fluxes via stomata opening/closure, regulation of the vertical temperature gradient below the canopy, facilitation of precipitation by biogenic aerosols, and other unknown processes, of which some are possibly unknowable due to the high complexity of the object under study. On average, functioning of the tree results in pumping moisture out of the atmosphere and increase of the local precipitation, which compensates for the local losses of soil water to runoff. If the environmental changes performed by the tree lead to optimization of the local soil moisture content to the accuracy of the biotic sensitivity ε, the tree and its associated biota acquire competitive advantage over neighboring trees incapable of performing the needed environmental changes. These deficient neighbors lose in competition with, and are replaced by, normal trees capable of environmental control. In the result, all trees remaining in the forest are genetically programmed to act in one and the same direction, thus forming a regional biotic pump of atmospheric moisture, which compensates for the river runoff. At the same time, individual trees continue to compete with each other, and forest as a whole does not have a “physiology” of a superorganism.

5 Conclusions

In this paper we introduced and described the biotic pump of atmospheric moisture. It makes use of the fundamental physical principle that horizontal fluxes of air and water vapor are directed from areas with weaker evaporation to areas with stronger evaporation. Natural forest ecosystems, with their high leaf area index and high transpiration exceeding evaporation from open water surface, are capable of pumping atmospheric moisture from the ocean in amounts sufficient for the maintenance of optimal soil moisture stores, compensating the river runoff and ensuring maximum ecosystem productivity.

The biotic moisture pump, as well as the mechanisms of efficient soil moisture preservation described in Sect. 3.4, work in undisturbed natural forests only. Natural forest represents a contiguous cover of tall trees that are rigidly associated with other biological species of the ecological community and genetically programmed to function in the particular geographic region. The vegetation cover of grasslands, shrublands, savannas, steppes, prairies, artificially thinned exploited forests, plantations, pastures or arable lands is unable to switch on the biotic moisture pump and maintain soil moisture content in a state optimal for life. Water cycle on such territories is critically dependent on the distance from the ocean; it is determined by random fluctuations and seasonal changes of rainfall brought from the ocean. Such territories are prone to droughts, floods and fires. We emphasize that the contemporary wide spread of scarcely vegetated ecosystems, in particular, African savannas, is rigidly correlated with anthropogenic activities during the last several thousand years (see, e.g., Tutin et al., 1997). Savannas represent a succesional state of the ecosystem returning to its undisturbed forest state; this transition spontaneously occurs as soon as the artificial disturbances like fire and overgrazing are stopped (van de Koppel and Prins, 1988). Since savannas and other open ecosystems suffer from rapid soil erosion (Lal, 1990), their prolonged existence is only possible due to continuous cycling with the forest state, when the mineral and organic content of soil is restored. Therefore, the growing anthropogenic pressure on savannas, which prevents their periodic transitions to forests, gradually turns savannas to deserts. The same is true for steppes and prairies of the temperate zone.

Only primary aboriginal forests are able to ensure the long-term stability of the biotic moisture pump functioning, as far as the genetic properties of aboriginal forests are correlated with the geophysical properties of the region they occupy. Artificially planted exotic vegetation with geographically irrelevant genetic programs cannot persist on an alien territory for a long time; their temporal prosperity is followed by environmental degradation and ecological collapse. On the other hand, secondary aboriginal forests that are in the process of self-recovery from anthropogenic or natural disturbances like fires, cutting or windfall, are not capable of efficiently running the biotic pump either. In such forests all mechanisms of environmental regulation, including the biotic moisture pump, are under repair and cannot yet function efficiently.

For the biotic moisture pump to work properly, it is also important that the natural forest cover have an immediate border with the ocean or, at least, the distance to the ocean be much less than the scale length l, Eq. (1), of the exponential weakening of the geophysical ocean-to-land moisture fluxes. The two largest tropical river basins of Amazon and Congo are covered by rainforests spreading inland from the oceanic coastline. Northern river basins of Russia, Canada and Alaska are covered by natural taiga forests, which, at the northernmost forest line, border with tundra wetlands linking them to the oceanic coast. If the natural forest cover is eliminated along the oceanic coastline on a band $l \sim 600$ km wide, the biotic moisture pump stalls. The remaining inland forests are no longer able to pump atmospheric moisture from the ocean. There is no longer surplus to runoff to rivers or to recharge groundwater. Soil water either leaves to the ocean as runoff or is blown away via the atmosphere after being transpired by the forests. The river basin ceases to exist, the forests die back after the soil dries up. The total store of fresh water on land including the water of soil, bogs, mountain glaciers and lakes, can be estimated by the lake store alone and constitutes around 1.5×10^{14} m^3 (L'vovitch, 1979), while the global river runoff is of the order of 0.37×10^{14} m^3 yr^{-1} (Dai and Trenberth, 2002). This means that all liquid moisture accumulated on land runs to the ocean in about four years. Total destruction of the biotic moisture pump due to deforestation (i.e., complete elimination of the contiguous forest cover bordering with the ocean) will in several years turn any river basin to desert.

In Australia, the continent-scale forested river basins ceased to exist about 50–100 thousand years ago, a time period approximately coinciding with the arrival of first humans. There is a host of indirect evidence suggesting that humans are responsible for the ancient deforestation of the Australian continent (see discussion in Bowman, 2002). It is clear how this could have happened. To deforest the continent, it was enough to destroy forests on a narrow band of width l along the continent's perimeter. This could be easily done by the first human settlements in the course of their household activities or due to the human-induced fires. This done, the biotic water pump of the inner undisturbed forested part of the continent was cut off from the ocean and stalled. Rapid runoff and evaporation eliminated the stores of soil moisture and the inland forests perished by themselves even in the absence of intense anthropogenic activities or fires in the inner parts of the continent. As estimated above, this forest-to desert transition should have been instantaneous on the geological time scale, so it is not surprising that practically no paleodata were left to tell more details about this ecological catastrophe. Notably, most deserts of the world border with the ocean or inner seas. As far as the coastal zone is the preferred area for human settlements, the modern extensive deserts of Earth should all be of anthropogenic origin.

Modern practice of forest cutting and exploitation, which is responsible for the unprecedented high rate of deforestation world over (Bryant et al., 1997), was born in Western Europe. Remarkably, in this part of land there are no areas separated from the ocean or inner seacoast by more than $l \sim 600$ km, the scale length of exponential weakening of the geophysical ocean-to-land moisture fluxes, Sect. 2.1. Therefore, elimination of natural European forests, which today is being finalized in Scandinavia, has not led to complete desertification of Europe. This worked to support the illusion that the forest cutting tradition can be safely imported to the other parts of the planet, despite the accumulating evidence about the disastrous consequences of such practices when applied to vast continental areas. Even in Western Europe one has been recently witnessing an increase in catastrophic droughts, fires and floods, likely facilitated by the on-going elimination of the remaining natural forests in the mountains. The decline of Alpine forests, which, via the biotic moisture pump, used to enhance precipitation in the mountains, should have led to the decline of mountain glaciers, although the latter phenomenon is almost exclusively considered in the framework of global warming and atmospheric CO_2 build-up, while the biotic pump effects are ignored.

We also note that when the forest cover is (partially) replaced by an artificial water body, like the water storage reservoirs of hydropower plants, the controlling function of forest transpiration and the biotic pump weaken causing reduction of runoff and precipitation in the corresponding river basin.

The results obtained form the basis of a possible strategy to restore human-friendly water conditions on most part of the Earth's landmasses, including modern deserts and other arid zones. As we have shown, elimination of the forest cover in world's largest river basins would have the following consequences: at least one order of magnitude's decline of the river runoff, appearance of droughts, floods and fires, partial desertification of the coastal zone and complete desertification of the inner parts of the continents, see Eqs. (5) and (6), associated economic losses would by far exceed the economic benefits of forest cutting, let alone such a scenario would make life of millions of people impossible. Therefore, it is worthy to urgently reconsider the modern forest policy everywhere in the world. First of all, it is necessary to immediately stop any attempts of destroying the extant natural forest remnants and, in particular, those bordering with the ocean or inner seas. Further on, it is necessary to initiate a world-wide company on facilitating natural gradual recovery of aboriginal forest ecosystems on territories adjacent to the remaining natural forests. Only extensive contiguous natural forests will be able to run a stable water cycle and subsequently intensify it, gradually extending the river basin at the expense of newly recovering territories.

In intact primary forests the relative area of natural disturbances like tree gaps does not exceed 10% (Coley and Barone, 1996; Szarzynski and Anhuf, 2001). Approximately the same share of land surface is currently exploited by man as arable land, pastures and industrial forest plantations. Therefore, if natural forests are restored over most parts of currently unused arid territories, such a state will be as stable as are the natural forests that have persisted through millions of years. The recovered luxurious water resources on land will be at full disposal of the ecologically literate humanity of the future.

Appendix A

Physical basis of the tropospheric lapse rate of air temperature

The decrease of air temperature with height observed in the troposphere is conditioned by the presence of atmospheric greenhouse substances; however, it is not related to the magnitude of the planetary greenhouse effect.

Greenhouse substances perform resonance absorption of thermal radiation emitted by the Earth's surface and re-emit it isotropically in all directions. Half of the absorbed thermal radiation is re-directed back to the surface. This random walk of thermal photons in the atmosphere results in an increase of the density of radiation energy near the surface and its drop with atmospheric height.

The resulting negative vertical gradient of radiative energy density can be related to the lapse rate of the so-called brightness temperature. It is defined as the temperature of a blackbody, which emits a radiation flux equal to the upward flux of thermal radiation observed at height z in the atmosphere. In the course of inelastic collisional interaction with molecules of the greenhouse substances, air molecules lose their kinetic energy (this energy is imparted to the molecules of the greenhouse substances and ultimately lost into space in the form of thermal radiation). In the result, air temperature approaches brightness temperature of thermal radiation; there appears a negative vertical gradient (lapse rate) of air temperature close to the lapse rate of radiative brightness temperature. It is due to these physical processes that the environmental temperature lapse rate arises.

Thermal radiation spectrum of the Earth's surface is close to the blackbody spectrum at Earth's surface temperature. If the greenhouse substances of the terrestrial atmosphere had but one or a few absorption lines (narrow bands) with respect to thermal radiation, then all the radiation emitted from the Earth's surface, except for those lines, would leave into space unimpeded, without interacting with the atmosphere. Brightness temperature of the entire spectrum, except for those lines, would not change with height. In such a case the surface temperature would coincide with the brightness temperature of thermal radiation emitted into space, i.e. the planetary greenhouse effect would be close to zero.

However, if the amount of greenhouse substances with those narrow absorption bands is large and the absorption lines have large absorption cross-sections, brightness temperature of thermal radiation in the corresponding spectral interval can rapidly drop with height, theoretically down to absolute zero. Collisional excitation of these absorption lines by air molecules would deplete the kinetic energy of the latter, resulting in a decrease of air temperature with height. Thus, there appears a large negative vertical gradient of air temperature, which, as described in Sect. 3, causes the observed upward fluxes of latent heat and evaporation. Condensation of the ascending water vapor with release of latent heat increases air temperature, enhances the collisional excitation of the greenhouse substances and diminishes the lapse rates of both brightness and air temperatures. In the result, the magnitude of the air temperature lapse rates is fixed at the observed value of Γ_{ob}=6.5 K km^{-1}.

Therefore, the observed value of Γ_{ob}=6.5 K km^{-1} can be a product of one narrow absorption line with large absorption cross-section present in the atmosphere. The greenhouse substance with that narrow efficient line absorbs all the released latent heat and thermal energy of the dissipating processes of atmospheric circulation, turns this energy into thermal radiation and partially re-directs to the surface. At the surface the radiative energy corresponding to the narrow absorption line is distributed over the entire thermal spectrum and leaves into space.

On the other hand, if the absorption lines of the greenhouse substances covered the whole thermal spectrum of the surface, but they exclusively interacted with thermal photons under conditions of the so-called radiative equilibrium, when the collisional excitation of these lines is negligible, in such a case the planetary greenhouse effect could reach large values (i.e. the planetary surface would be much warmer than in the absence of greenhouse substances), while the lapse rate of air temperature remained close to zero.

Therefore, the magnitudes of the planetary greenhouse effect and air temperature lapse rate Γ are not unambiguously related to each other. However, in the absence of all greenhouse substances in the atmosphere, both air temperature lapse rate, evaporative force (Sect. 3.3) and greenhouse effect are all equal to zero.

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References

Acevedo, O. C., Moraes, O. L. L., da Silva, R., Fitzjarrald ,D. R., Sakai, R. K., Staebler, R. M., and Czikowsky, M. J.: Inferring nocturnal surface fluxes from vertical profiles of scalars in an Amazon pasture, Global Change Biology, 10, 886–894, 2004. 2650

Adler, R. F., Huffman, G. J., Bolvin, D. T., Curtis, S., and Nelkin, E. J.: Tropical rainfall distributions determined using TRMM combined with other satellite and rain gauge, J. Appl. Meteorol., 39, 2007–2023, 2001. 2631

Arora, V.: Comment on “Optimized stomatal conductance of vegetated land surfaces and its effects on simulated productivity and climate” edited by: Kleidon, A., Geophys. Res. Lett., 32, L08708, doi:10.1029/2004GL022110, 2005. 2654

Austin, A. T. and Sala, O. E.: Carbon and nitrogen dynamics across a natural precipitation gradient in Patagonia, Argentina, J. Vegetation Sci., 13, 351–360, 2002. 2627, 2669

Austin, A. T. and Vitousek, P. M.: Precipitation, decomposition and litter decomposability of Metrosideros polymorpha in native forests on Hawaii, J. Ecology, 88, 129–138, 2000. 2627

Baerlocher, F.: The Gaia hypothesis: A fruitful fallacy? Experientia, 46, 232–238, 1990. 2653

Barrett, J. E., McCulley, R. L., Lane, D. R., Burke, I. C., and Lauenroth, W. K.: Influence of climate variability on plant production and N-mineralization in Central US grasslands, J. Vegetation Sci., 13, 383–394, 2002. 2669

Bond, B. J. and Kavanagh, K. L.: Stomatal behavior of four woody species in relation to leaf-specific hydraulic conductance and threshold water potential, Tree Physiology, 19, 503–510, 1999. 2651

Bowman, D. M. J. S.: The Australian summer monsoon: a biogeographic perspective, Australian Geographic Studies, 40, 261–277, 2002. 2658

Breshears, D. D., Nyhan, J. W., Heil, C. E., and Wilcox, B. P.: Effects of woody plants on microclimate in a semiarid woodland: soil temperature and evaporation in canopy and intercanopy patches, Int. J. Plant Sci., 159, 1010–1017, 1998. 2649

Bryant, D., Nielsen, D., and Tangley, L.: The last frontier forests: ecosystems and economies on the edge, World Resources Institute, Washington, 1997. 2628, 2629, 2658, 2669

Cabido, M., Gonzalez, C., Acosta, A., and Diaz, S.: Vegetation changes along a precipitation gradient in central Argentina, Vegetatio, 109, 5–14, 1993. 2669

Canadell, J. G., Steffen, W. L., and White, P. S.: IGBP/GCTE terrestrial transects: Dynamics of terrestrial ecosystems under environmental change – Introduction, J. Vegetation Sci., 13, 297–300, 2002. 2626, 2669

Cochrane, M. A., Alencar, A., Schulze, M. D., Souza, C. M., Nepstad, D. C., Lefebvre, P., and Davidson, E. A.: Positive feedbacks in the fire dynamic of closed canopy tropical forests, Sci., 284, 1832–1835. 2649

Coley, P. D. and Barone, J. A.: Herbivory and plant defenses in tropical forests, Annual Review of Ecology and Systematics, 27, 305–335, 1996. 2659

Cook, G. D. and Heerdegen, R. G.: Spatial variation in the duration of the rainy season in monsoonal Australia, Int. J. Climatology, 21, 1723–1732, 2001. 2669

da Rocha, H. R., Goulden, M. L., Miller, S. D., Menton, M. C., Pinto, L. D. V. O., de Freitas, H. C., and Silva Figueira, A. M. E.: Seasonality of water and heat fluxes over a tropical forest in eastern Amazonia, Ecological Appl., 14(Suppl), S22–S32, 2004. 2628, 2648

Dai, A. and Trenberth, K. E.: Estimates of freshwater discharge from continents: latitudinal and seasonal variations, J. Hydrometeorol., 3, 660–687, 2002. 2626, 2657

Doolittle, W. F.: Is nature really Motherly?, CoEvolution Quaterly, 29, 58–63, 1981. 2653

Fang, M. and Tung, K. K.: Time-dependent nonlinear Hadley circulation, J. Atmos. Sci., 56, 1797–1807, 1999. 2642

Folkins, I.: Convective damping of buoyancy anomalies and its effect on lapse rates in the tropical lower troposphere. Atmos. Chem. Phys., 6, 1–12, 2006. 2638

Goody, R. M. and Yung, Y. L.: Atmospheric radiation, theoretical basis, 2nd edn, Oxford Univ.

Press, New York, 1989. 2637

Gorshkov, V. G.: Ecology of man, Leningrad Polytechnical Institute, Leningrad, 1984. 2654

Gorshkov, V. G.: Physical and biological bases of life stability, Springer, Berlin, 1995. 2654

Gorshkov, V. G., Gorshkov, V. V., and Makarieva, A. M.: Biotic regulation of the environment: Key issue of global change, Springer, London, 2000. 2654

Gorshkov, V. G. and Makarieva, A. M.: On the possibility of physical self-organization of biological and ecological systems, Doklady Biological Sci., 378, 258–261, 2001. 2654

Gorshkov, V. G., Makarieva, A. M., and Gorshkov, V. V.: Revising the fundamentals of ecological knowledge: The biota-environment interaction, Ecol. Complexity, 1, 17–36, 2004. 2654

Goulden, M. L., Daube, B. C., Fan, S.-M., Sutton, D. J., Bazzaz, A., Munger, J. W., and Wofsy, S. C.: Physiological response of a black spruce forest to weather, J. Geophys. Res., 102D, 28 987–28 996, 1997. 2628

Goulden, M. L., Miller, S. D., da Rocha, H. R., Menton, M. C., de Freitas, H. C., Silva Figueira, A. M. E., and de Sousa, C. A. D.: Diel and seasonal patterns of tropical forest CO_2 exchange, Ecol. Appl., 14(Suppl), S42–S54, 2004. 2651

Gustavson, M. R.: Limits to the wind power utilization, Sci., 204, 13–17, 1979. 2647

Hodnett, M. G., Oyama, M. D., Tomasella, J., and Marques Filho, A. de O.: Comparisons of long-term soil water storage behaviour under pasture and forest in three areas of Amazonia, in: Amazonian deforestation and climate, edited by: Gash, J. H. C., Nobre, C. A., Roberts, J. M., and Victoria, R. L., 57–77, John Wiley & Sons, Chichester, 1996. 2628

Hutley, L. B., O'Grady, A. P., and Eamus, D.: Monsoonal influences on evapotranspiration of savanna vegetation of northern Australia, Oecologia, 126, 434–443, 2001. 2628

Karlsson, I. M.: Nocturnal air temperature variations between forest and open areas, J. Appl. Meteorol., 39, 851–862, 2000. 2650

Kleidon, A.: Optimized stomatal conductance of vegetated land surfaces and its effects on simulated productivity and climate, Geophys. Res. Lett., 31, L21203, doi:10.1029/2004GL020769, 2004. 2654

Kleidon, A.: Reply to comment by V. Arora on "Optimized stomatal conductance of vegetated land surfaces and its effects on simulated productivity and climate", Geophys. Res. Lett., 32, L08709, doi:10.1029/2005GL022355, 2005. 2654

Kruijt, B., Malhi, Y., Lloyd, J., Nobre, A. D., Miranda, A. C., Pereira, M. G. P., Culf, A., and Grace, J.: Turbulence statistics above and within two Amazon rain forest canopies, Boundary-Layer Meteorol., 94, 297–331, 2000. 2648

Kurc, S. A. and Small, E.E.: Dynamics of evapotranspiration in semiarid grassland and shrubland ecosystems during the summer monsoon season, central New Mexico, Water Resour. Res., 40, W09305, doi:10.1029/2004WR003068, 2004. 2628

Lal, R.: Soil erosion and land degradation: the global risks, in: Advances in soil science, 11, Soil degradation, edited by: Lal, R. and Stewart, B. A., 129–172, Springer, New York, 1990. 2656

Landau, L. D., Akhiezer, A. I., and Lifshitz, E. M.: General Physics., Mechanics and Molecular Physics, Nauka, Moscow, 1965. 2634

Landau, L. D. and Lifshitz, E. M.: Course of Theoretical Physics, 6, Fluid Mechanics, 2nd ed., Butterworth-Heinemann, Oxford, 1987. 2633, 2634, 2638

Lorenz, E. N.: The nature and theory of the general circulation of the atmosphere, World Meteorological Organization, Geneva, 1967. 2639, 2647

L'vovitch, M. I.: World water resources and their future, American Geological Union, Washington, 1979. 2645, 2657

Mahrt, L., Lee, X., Black, A., Neumann, H., and Staebler, R. M.: Nocturnal mixing in a forest subcanopy, Agric. Forest Meteorol., 101, 67–78, 2000. 2650

Makarieva, A. M., Gorshkov, V. G., and Li, B.-L.: Biochemical universality of living matter and its metabolic implications, Functional Ecol., 19, 547–557, 2005a. 2652

Makarieva, A. M., Gorshkov, V. G., and Li, B.-L.: Revising the distributive networks models of West, Brown and Enquist (1997) and Banavar, Maritan and Rinaldo (1999): Metabolic inequity of living tissues provides clues for the observed allometric scaling rules, J. Theor. Biol., 237, 291–301, 2005b. 2652

Makarieva, A. M., Gorshkov, V. G., and Li, B.-L.: Energetics of the smallest: Do bacteria breathe at the same rate as whales?, Proceedings of the Royal Society of London B, 272, 2219–2224, 2005c. 2652

Marengo, J. A.: Interdecadal variability and trends of rainfall across the Amazon basin, Theor. Appl. Climatology, 78, 79–96, 2004. 2632, 2647

McGuffie, K. and Henderson-Sellers, A.: Forty years of numerical climate modelling, Int. J. Climatology, 21, 1067–1109, 2001. 2631

McGuire, A. D., Prentice, I. C., Ramankutty, N., Reichenau, T., Schloss, A., Tian, H., Williams, L. J., and Wittenberg, U.: Carbon balance of the terrestrial biosphere in the twentieth century: Analyses of CO_2, climate and land-use effects with four process-based ecosystem models, Global Biogeochem. Cycles, 15, 183–206, 2001. 2669

Miller, J. M., Williams, R. J., and Farquhar, G. D.: Carbon isotope discrimination by a sequence of Eucalyptus species along a subcontinental rainfall gradient in Australia, Funct. Ecol., 15, 222–232, 2001. 2669

Nepstad, D., Lefebvre, P., da Silva, U. L., Tomasella, J., Schlesinger, P., Solorzano, L., Ray, P. M. D., and Benito, J. G.: Amazon drought and its implications for forest flammability and tree growth: a basin-wide analysis, Global Change Biol., 10, 704–717, 2004. 2649

Nesbitt, S. W. and Zipser, E. J.: The diurnal cycle of rainfall and convective intensity according to three years of TRMM measurements, J. Climate, 16, 1456–1475, 2003. 2651

Ni, J. and Zhang, X. S.: Climate variability, ecological gradient and the Northeast China Transect (NECT), J. Arid Environments, 46, 313–325, 2000. 2669

Nicholson, S. E.: The nature of rainfall variability over Africa on time scales of decades to millenia, Global and Planetary Change, 26, 137–158, 2000. 2669

Oyama, M. D. and Nobre, C. A.: Climatic consequences of a large-scale desertification in Northeast Brazil: A GCM simulation study, J. Climate, 17, 3203–3213, 2004. 2632

Randel, D. L., Haar, T. H. V., Ringerud, M. A., Stephens, G. L., Greenwald, T. J., and Comb, C. L.: New global water vapor dataset, Bulletin of the American Meteorol. Society, 77, 1233–1254, 1996. 2630, 2647

Raval, A. and Ramanathan, V.: Observational determination of the greenhouse effect, Nature, 342, 758–761, 1989. 2634

Savenije, H. H. G.: New definitions for moisture recycling and the relationship with land-use change in the Sahel, J. Hydrology, 167, 57–78, 1995. 2626, 2627

Shver, C. A.: Atmospheric precipitation on the territory of the USSR, Meteoizdat, Leningrad, 1976. 2630

Shuttleworth, W. J.: Micrometeorology of temperate and tropical forest, Philosophical Transactions of the Royal Society of London B, 324, 207–228, 1989. 2648

Shuttleworth, W. J., Gash, J. H. C., Lloyd, C. R., Moore, C. J., Roberts, J., Marques-Filho, A. O., Fisch, G., Filho, V. P. S., Ribeiro, M. N. G., Molion, L. C. B., Sa, L. D. A., Nobre, J. C. A., Cabral, O. M. R., Patel, S. R., and Carvalho, J. M.: Daily variations of temperature and humidity within and above Amazonian forest, Weather, 40, 102–108, 1985. 2649

Smith, S. B.: Comments on “An interesting mesoscale storm-environment interaction observed just prior to changes in severe storm behavior”, Weather and Forecasting, 12, 368–372, 1997. 2641

Szarzynski, J. and Anhuf, D.: Micrometeorological conditions and canopy energy exchanges of a neotropical rain forest (Surumoni-Crane Project, Venezuela), Plant Ecol., 153, 231–239, 2001. 2648, 2649, 2650, 2659

Tchebakova, N. M., Kolle ,O., Zolotoukhine, D., Arneth, A., Styles, J. M., Vygodskaya, N. N., Schulze, E.-D., Shibistova, O., and Lloyd, J.: Inter-annual and seasonal variations of energy and water vapor fluxes above a Pinus sylvestris forest in the Siberian middle taiga, Tellus B, 54, 537–551, 2002. 2628

Tutin, C. E. G., White, L. J. T., and Mackanga-Missandzou, A.: The use by rain forest mammals of natural forest fragments in an equatorial African Savanna, Conservation Biol., 11, 1190–1203, 1997. 2656

Van de Koppel, J. and Prins, H. H. T.: The importance of herbivore interactions for the dynamics of African savanna woodlands: an hypothesis, J. Tropical Ecol., 14, 565–576, 1998. 2656

Weaver, C. P. and Ramanathan, V.: Deductions from a simple climate model: Factors governing surface temperature and atmospheric thermal structure, J. Geophys. Res., 100D, 11 585–11 591, 1995. 2637

Webber, S. R. and Wilmott, C. J.: South American Precipitation: 1960–1990 gridded monthly time series (version 1.02), Center for Climatic Research, Department of Geography, University of Delaware, Newark, Delaware, 1998. 2669

Werth, D. and Avissar, R.: The regional evapotranspiration of the Amazon, J. Hydrometeorol., 5, 100–109, 2004. 2628

Zhou, J. and Lau, K.-M.: Does a monsoon climate exist over South America?, J. Climate, 11, 1020–1040, 1998. 2669

Table 1. Precipitation on land P (mm year^{-1}) versus distance from the source of moisture x (km) as dependent on the absence/presence of natural forests

Region					Parameters of linear regression[1] $\ln P = a - bx$				
No	Name	$x=0$ °Lat, °Lon	$x=x_{max}$ °Lat, °Lon	x_{max}, km	$a \pm 1$ s.e.	$(b \pm 1$ s.e.$)\times10^3$	r^2	$P(0)\equiv e^a$, mm year^{-1}	$l \equiv 1/b$, km
	Non-forested regions[2]								
1	North Australia[3]	−11.3, 130.5	−25, 137	1400	7.37± 0.05	1.54± 0.06	0.96	1600	650
2	North East China[4]	42, 125	42, 107	1500	6.67± 0.08	1.24± 0.10	0.96	790	800
3	West Africa[5]	10, 5	25, 5	1650	7.28± 0.09	2.46± 0.09	0.99	1450	400
4a	Argentina[6], 45° S	−44.8, −71.7	−45, −69.8	150	6.36± 0.17	10.8± 2.0	0.90	580	93
4b	Argentina[6], 31° S	−31.3, −65.3	−31.7, −68.3	360	6.35± 0.12	4.57± 0.59	0.91	570	220
5	North America[7]	39.8, −96.7	41.2, −105.5	750	6.69± 0.04	1.15± 0.08	0.93	800	870
	Natural forests[8]								
6	Amason river basin[9]	0, −50	−5, −75	2800	7.76± 0.04	−0.05 ± 0.02	0.13	2300	-2×10^4
7	Congo river basin[10]	0, 9	0, 30	2300	7.56± 0.17	−0.10 ± 0.12	0.05	1900	-1×10^4
8	Yenisey river basin[11]	73.5, 80.5	50.5, 95.5	2800	6.06± 0.17	−0.01 ± 0.10	0.05	430	-1×10^4

Table 1. Continued.

Notes:

[1] Statistics are significant at the probability level $p<0.0001$ for regions 1, 2, 3 and 5; $p=0.013$ for region 4a and $p<0.001$ for region 4b; $p\geq0.05$ for regions 6, 7 and 8 (i.e., there is no exponential dependence of P on x in these regions).

[2] Regions 1, 2, 3, 4 (a and b) and 5 correspond to the non-forested parts of the North Australian Tropical Transect, North East China Transect (NECT), Savannah on the Long-Term Transect, Argentina Transect and North American Mid-Latitude Transect of the International Geosphere Biosphere Program, respectively (Canadell et al., 2002).

[3] Precipitation data for $x=0$ (Prilangimpi, Australia) taken from Cook and Heerdegen (2001), all other data taken from Fig. 2a of Miller et al. (2001) assuming 1° Lat.=110 km.

[4] Data taken for 42° N of NECT, because at this latitude NECT comes most closely to the ocean. Location of $x=0$ approximately corresponds to the border between forest and steppe zones; the dependence between P and x obtained from the location of isohyets taken from Fig. 3c of Ni and Zhang (2000) assuming 10° Lon.=825 km at 42° N.

[5] Southern border of the non-forested part of West Africa approximately coincides with the 1200 mm isohyet, hence the choice of $x=0$ at 10° N 5° E, where $P=1200\,\text{mm year}^{-1}$. The dependence between P and x was obtained from the location of isohyets taken from Fig. 2 of Nicholson (2000) assuming 1° Lat.=110 km.

[6] Data for region 4a and 4b taken from Table 1 of Austin and Sala (2002) and Figs. 1 and 2 of Cabido et al. (1993), respectively. In region 4b atmospheric moisture comes from the Pacific Ocean (Austin and Sala, 2002), in region 4a the ultimate source of moisture is the Atlantic Ocean (Zhou and Lau, 1998), hence the opposite directions of counting x in the two regions.

[7] Data taken from Table 1 of Barrett et al. (2002), x calculated assuming 1° Lat.=110 km.

[8] Precipitation values for the three regions covered by natural forests are taken from the data sets distributed by the University of New Hampshire, EOS-WEBSTER Earth Science Information Partner (ESIP) at http://eos-webster.sr.unh.edu. Regions 6 and 8 correspond to the Amazon (LBA) and Central Siberia Transect of IGBP, respectively (Canadell et al., 2002).

[9] Precipitation data taken from the gridded monthly precipitation data bank LBA-Hydronet v1.0 (Water Systems Analysis Group, Complex Systems Research Center, University of New Hampshire), time period 1960–1990, grid size 0.5×0.5 degrees (Webber and Wilmott, 1998); statistics is based on P values for 26 grid cells that are crossed by the straight line from $x=0$ to $x=x_{\max}$, Fig. 1.

[10] The transect was chosen in the center of the remaining natural forest area in the Equatorial Africa (Bryant et al., 1997). Precipitation data taken from the gridded annual precipitation data bank of the National Center for Atmospheric Research (NCAR)'s Community Climate System Model, version 3 (CCSM3), time period 1870–1999, grid size 1.4×1.4 degrees; statistics is based on P values for 16 grid cells that are crossed by the straight line from $x=0$ to $x=x_{\max}$, Fig. 1.

[11] Precipitation data taken from the gridded monthly precipitation data bank Carbon Cycle Model Linkage-CCMLP (McGuire et al., 2001), time period 1950–1995, grid size 0.5×0.5 degrees; statistics is based on P values for 20 grid cells that are crossed by the straight line from $x=0$ to $x=x_{\max}$, Fig. 1.

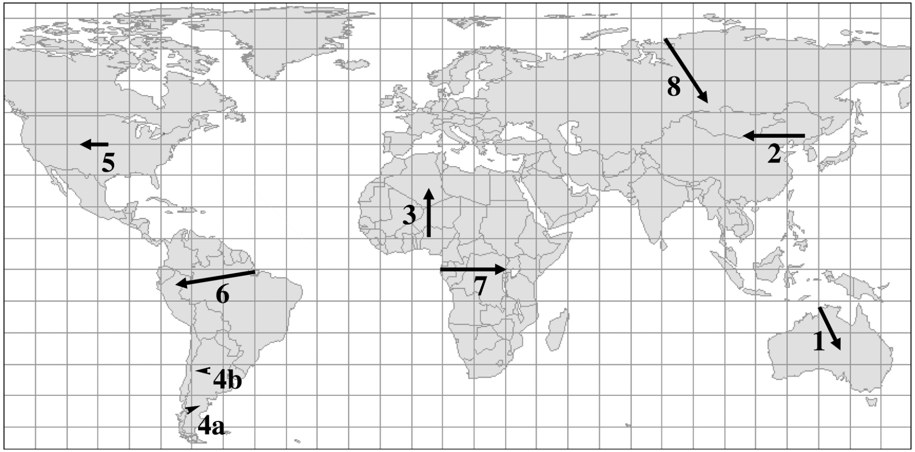


Fig. 1. Geography of the regions where the dependence of precipitation P on distance x from the source of moisture was studied. Numbers near arrows correspond to regions as listed in Table 1. Arrows start at $x=0$ and end at $x=x_{\max}$, see Table 1 for more details.

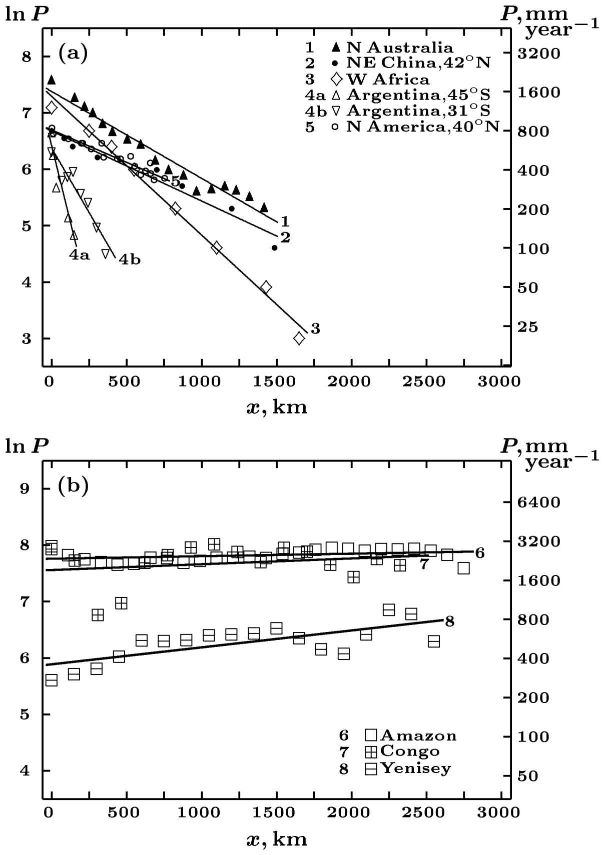


Fig. 2. Dependence of precipitation P (mm year^{-1}) on distance x (km) from the source of atmospheric moisture on non-forested territories **(a)** and on territories covered by natural forests **(b)**. Regions are numbered and named as in Table 1. See Table 1 for parameters of the linear regressions and other details.

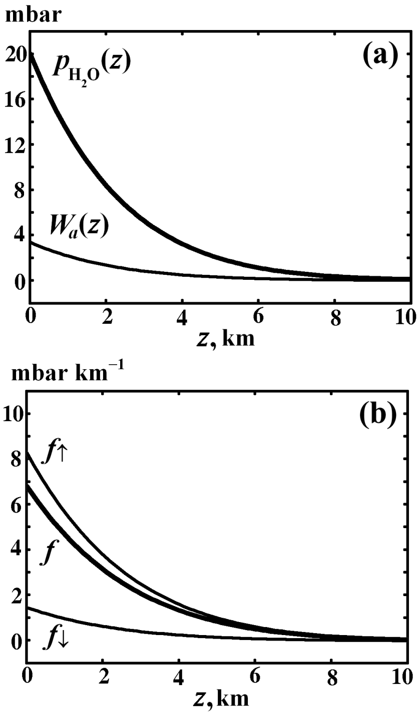
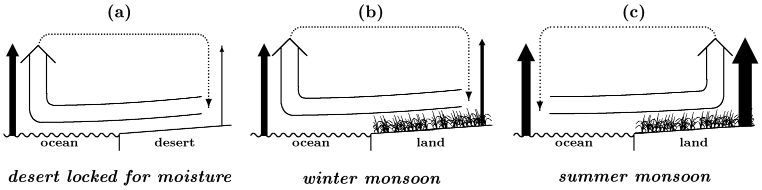
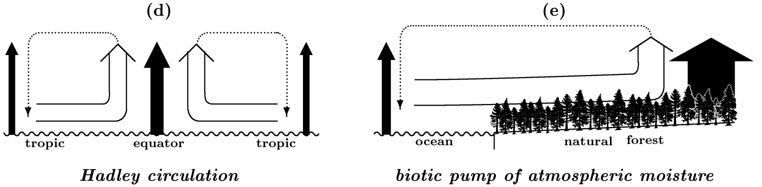


Fig. 3. Water vapor partial pressure and evaporative force in the terrestrial atmosphere. **(a)** Saturated partial pressure of water vapor $p_{H_2O}(z)$, Eq. (9), and weight of the saturated water vapor $W_a(z)\equiv\int_z^\infty \frac{p_{H_2O}(z')}{h_w(z)'}dz'$ in the atmospheric column above height z at $\Gamma_{ob}=6.5\,\mathrm{K\,km^{-1}}$. Saturated partial pressure at the surface is $p_{H_2O}(0)=20$ mbar. **(b)** The upward-directed evaporative force f_{ev}, Eq. (14), equal to the difference between the upward directed pressure gradient force $f_\uparrow(z)$ and the downward directed weight $f_\downarrow$ of a unit volume of the saturated water vapor, $f_{ev}=f_\uparrow-f_\downarrow$. The evaporative force is hundreds of times larger than the forces of the horizontal barometric pressure gradient that are of the order of 1 mbar $(100\,\mathrm{km})^{-1}=0.01$ mbar km^{-1}.



desert locked for moisture *winter monsoon* *summer monsoon*



Hadley circulation *biotic pump of atmospheric moisture*

Fig. 4. The physical principle that the low-level air moves from areas with weak evaporation to areas with more intensive evaporation provides clues for the observed patterns of atmospheric circulation. Black arrows: evaporation flux, arrow width schematically indicates the magnitude of this flux (evaporative force). Empty arrows: horizontal and ascending fluxes of moisture-laden air in the lower atmosphere. Dotted arrows: compensating horizontal and descending air fluxes in the upper atmosphere; after condensation of water vapor and precipitation they are depleted of moisture. **(a)** Deserts: evaporation on land is close to zero, so the low-level air moves from land to the ocean year round, thus ”locking” the desert for moisture. **(b)** Winter monsoon: evaporation from the warmer oceanic surface is larger than evaporation from the colder land surface; the low-level air moves from land to the ocean. **(c)** Summer monsoon: evaporation from the warmer land surface is larger than evaporation from the colder oceanic surface; the low-level air moves from ocean to land. **(d)** Hadley circulation (trade winds): evaporation is more intensive on the equator, where the solar flux is larger than in the higher latitudes; low-level air moves towards the equator year round; seasonal displacements of the convergence zone follow the displacement of the area with maximum insulation. **(e)** Biotic pump of atmospheric moisture: transpiration fluxes regulated by natural forests exceed oceanic evaporation fluxes to the degree when the arising ocean-to-land fluxes of moist air become large enough to compensate losses of water to runoff in the entire river basin year round.

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