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SHORT COMMUNICATIONS

The Forest Biotic Pump of River Basins

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Moist soil is necessary for life on the land. Soil moisture determines the formation of river runoff, which is correlated with it. Given an optimal moisture content of soil and constant slope of the ground, the river runoff density (per unit area) is also constant throughout the river basin. In a steady state, the river runoff to the ocean must be compensated by the opposite flow of moisture from the ocean to the land in all parts of the river basin, irrespective of their distances from the ocean.

Geophysical flows of moisture from the ocean to the land, such as monsoons, trade winds, atmospheric fronts, etc., exponentially decay as they spread into the land to distances of several hundred kilometers, which is many times shorter than the size of continents and the lengths of the largest river basins (Figs. 1, 2a). Only river basins covered with undisturbed natural forests spread over unlimited distances into continents (Figs. 1, 2b).

Biotic control of moisture flow from the ocean to the land, precipitation, and evaporation over a continuous forest cover is based on the physical law expressed by the Clapeyron–Clausius equation and its consequences, which have not been taken into consideration thus far. The Clapeyron–Clausius equation describes an increase in the partial pressure of saturated water with an increase in temperature vapor as a geometric series: the pressure of saturated water vapor doubles as the temperature increases by ten degrees centigrade. According to satellite survey data, the moisture content of Earth's atmosphere averaged over temporal fluctuations precisely fits this fundamental physical law on the regional scale (Wentz and Schabel, 2000). In contrast to other (noncondensable) atmospheric gases, water vapor in the atmosphere is in aerostatic equilibrium (when the gas pressure is compensated by its weight in the air column) only if the negative vertical gradient of air temperature (G) does not exceed the critical value $G = G_{H_2O} = 1.2$ K/km (see Appendix). At $G < G_{H_2O}$, water vapor reaches saturation only over moist ground, there are no flows of evaporated water or latent heat from the hydrosphere and moist soil to the atmosphere, clouds are not formed, and precipitation is absent.

The observed negative vertical gradient of air temperature in the atmosphere ($G_{ob} = 6.5$ K/km) is 5.4 times higher than the critical value $G_{H_2O} = 1.2$ K/km. Water vapor at all heights is not in aerostatic equilibrium: the pressure of atmospheric water vapor is several times higher than its weight. At the observed value of $G = G_{ob} = 6.5$ K/km, the negative gradient of the water vapor pressure concentration minus the weight of the unit volume of water vapor at each height creates an upward force resulting in upward flow of humid air and maintaining the cloud cover at fixed heights (see Appendix). The condensation and precipitation of moisture accompanying the ascent of the air are compensated by water evaporation from the ground.

A difference in evaporation between two neighboring regions leads to differences in the forces and rates of the upward air flows. This results in horizontal flows of humid air in lower atmospheric layers from regions with smaller evaporation to regions with greater evaporation (see Appendix). The air that has come to the region with a greater evaporation ascends, and the moisture contained in it is condensed and precipitates over this region. In higher atmospheric layers, the air with now decreased humidity goes back to regions with smaller evaporation. This corresponds to a geophysical

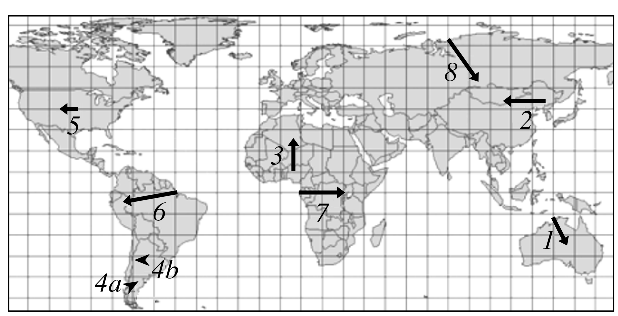

Fig. 1. Inland geophysical regions where the dependence of annual mean precipitation on the distance from the ocean has been studied.

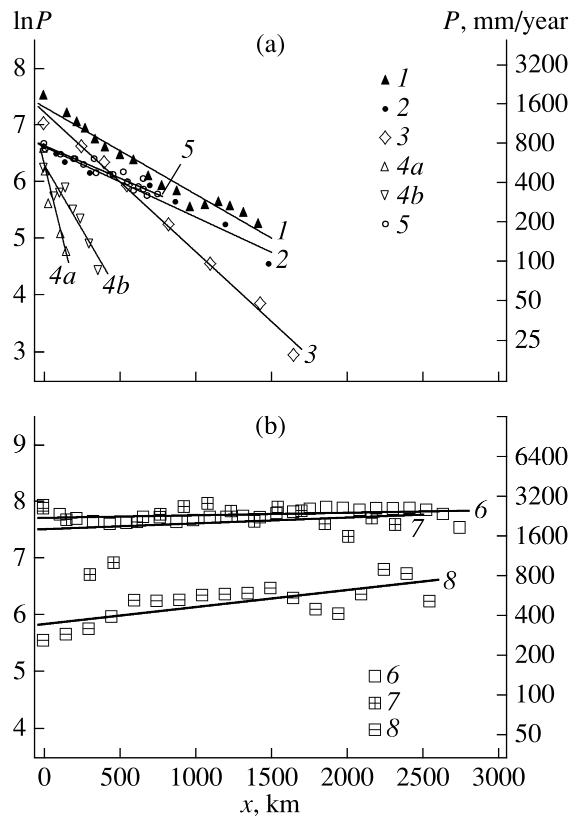

Fig. 2. Dependence of the precipitation (P, mm/year) on the distance (x, km) from the source of atmospheric moisture (the ocean coast) in regions (a) lacking a forest cover and (b) covered with natural forests. The locations of the regions are shown in Fig. 1. Meteorological data from the Carbon Cycle Model Linkage-CCMLP global database of monthly precipitation referred to a geographical grid 0.5 × 0.5 degrees in size have been used; the period of time is 1950–1995 (Gorshkov and Makar'eva, 2006). Designations: *1*, northern Australia; *2*, northeastern China, 42°N; *3*, western Africa; *4a*, Argentina, 45°S; *4b*, Argentina, 31°S; *5*, North America, 40°N; *6*, Amazon River; *7*, Congo River; *8*, Yenisei River.

pump carrying moisture from regions with small evaporation to regions with greater evaporation even if the moisture content of the former is lower than that of the latter; i.e., it can pump moisture against the gradient of moisture concentration. The flow of solar energy supplies energy to this pump; it is spent on evaporation and maintenance of the observed air temperature gradient (G_{ob}).

In a continuous canopy of tall trees in undisturbed forests, the area of the evaporating surface of leaves is tens of times larger than the area of the canopy projection onto the ground. Therefore, provided that heat is supplied and the temperature is constant, the evaporation from the leaf surface (forest evapotranspiration) may be considerably higher than the evaporation from an open ocean area of the same size as the area covered with the forest. This difference causes a flow of humid air, controlled by trees, from the ocean to the forest river basin, which compensates for the river runoff at any distance away from the ocean throughout the year, irrespective of the difference in temperature between the land and the ocean. This maintains a constant soil moisture, thereby preventing floods and forest fires. This constitutes the biotic pump carrying moisture from the ocean to natural forests. The constant drive of the biotic pump also prevents the formation of hurricanes and whirlwinds over undisturbed forests, including the wooded coastal frontier (Fig. 3a).

In a desert, there is no water evaporation; therefore, ground layers of air are driven to the ocean, which makes the desert permanently locked for humid air from the ocean surface (Fig. 3b). This is also true for winter monsoons in savannas, when the evaporation over the ocean is higher than that over a colder savanna, which results in a flow of near-ground air from the land to the ocean and, hence, a dry season in the savanna (Fig. 3c). However, in summer, when the savanna is warmer than the ocean, the water evaporation in the savanna is higher than in the ocean. This results in a flow of humid water from the ocean to the land in lower atmospheric layers and a rainy season in the savanna (Fig. 3d). The savanna vegetation cannot have an excess of the evaporation from the land over that from the ocean and optimal moisture content of the soil throughout the year. Therefore, dry and rainy seasons alternate in savannas, and summer monsoons carrying moisture from the ocean decay at short distances away from the ocean (Fig. 2a). The biotic pump does not work in savannas, and their moistening is determined by geophysical processes.

APPENDIX

The dependence of the air pressure (p_a) in the state of aerostatic equilibrium on the altitude (z) is described by the well-known equation (Landau and Lifshitz, 1954)

$$-\frac{dp_a}{dz} = \frac{p_a}{h_a}, \quad p_a(z) = p_{as}\exp\left\{-\int_0^z \frac{dz'}{h_a}\right\}, \qquad (1)$$
$$h_a \equiv \frac{RT}{M_a g}, \quad h_{as} \approx 8.4 \text{ km},$$

where the subscript s indicates that the given parameter was measured near the ground, T is the absolute air temperature, $\bar{T}_s$ = 288 K, R = 8.3 J/(K mol) is the gas constant, M_a = 29 g/mol is the molar mass of air, and g = 9.8 m/s^2 is the acceleration of gravity. The dependence of the partial pressure of saturated water vapor (p_{H_2O}) on the air temperature (T), which depends on z, is described by the Clapeyron–Clausius equation (Landau and Lifshits, 1954; Landau et al., 1965):

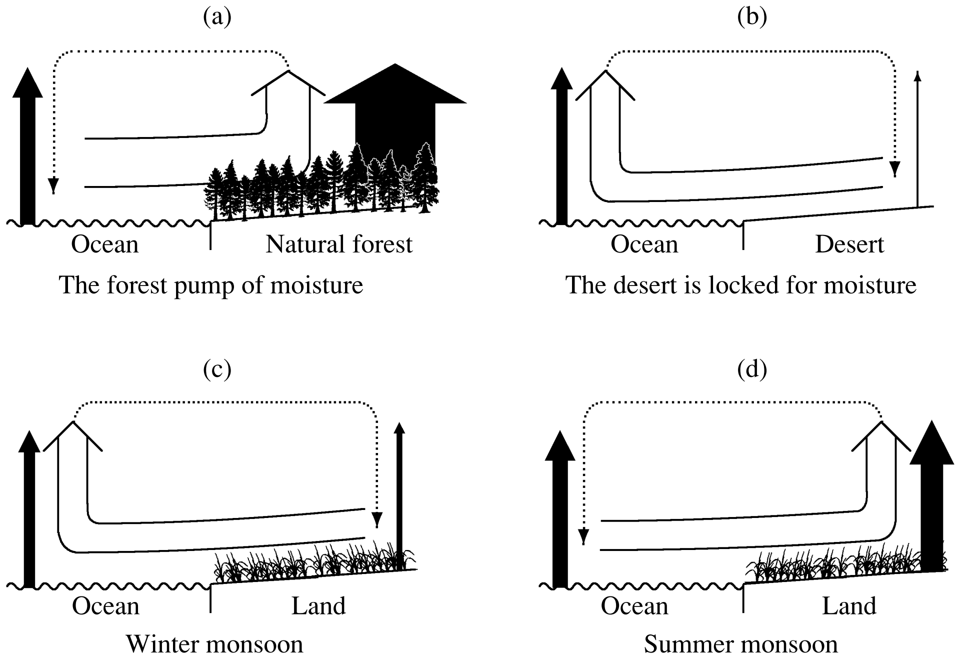

Fig. 3. The physical principle of the spread of air near Earth's surface from regions with a smaller evaporation to regions with a greater evaporation. Black arrows show evaporation flows, the flow rate being proportional to the arrow width; white arrows show horizontal and upstream flows of humid air; dotted arrows show horizontal and downstream flows of dry air.

$$p_{H_2O}(z) = p_{H_2O_s}\exp\left\{\frac{T_{H_2O}}{T_s} - \frac{T_{H_2O}}{T}\right\}$$

$$\equiv p_{H_2O_s}\exp\left\{-\int_0^z \frac{dz'}{h_{H_2O}}\right\}, \quad h_{H_2O} \equiv \frac{T^2}{\left(-\frac{dT}{dz}\right)T_{H_2O}}, \qquad (2)$$

where $T_{H_2O} \equiv Q_{H_2O}/R \approx 5300$ K, and $Q_{H_2O} = 4.4 \times 10^4$ J/mol is the molar latent heat of evaporation.

Saturated water vapor is in aerostatic equilibrium in the entire atmospheric column provided that $h_{H_2O} = h_w \equiv \frac{RT}{M_w g}$ ($M_w = 18$ g/mol is the molar mass of water), which means fixation of an air temperature gradient of

$$G \equiv \left(-\frac{dT}{dz}\right) = G_{H_2O} = \frac{T_s}{H} = 1.2 \text{ K/km},$$

$$H \equiv \frac{RT_{H_2O}}{M_w g} = 250 \text{ km},$$

where it is taken into account that $\exp(-z/H) = 1$ and $T = T_s$ at $z \le h_w$ by virtue of $h_w/H \approx 0.05 \ll 1$.

At $G < G_{H_2O}$, water vapor is in aerostatic equilibrium and is saturated only near the ground.

At $G > G_{H_2O}$, water vapor is not in aerostatic equilibrium. A negative gradient of the partial pressure of water vapor minus the weight of a unit volume of water vapor results in the appearance of the evaporation force $f = -(dp_{H_2O}/dz + p_{H_2O}/h_w) = (\beta - \beta_0)\rho_a\gamma g$, where $\beta \equiv h_a/h_w = 0.62$, $\gamma \equiv p_{H_2O}/p_a$, and $\rho_a = p_a/gh_a$ is the air density. At the observed $G_{ob} = 6.5$ K/km, $h_{H_2O} \approx 2$ km, $\beta \approx 3.5$. The force f is directed upwards, causes upstream air flows, and maintains the cloud cover at a fixed altitude. The work of this force over the distance dz is equal to the change in the kinetic energy of a unit volume of air: $\frac{1}{2}\rho_a dw^2 = fdz$ (Euler's equation), where w is the vertical velocity of humid air. The steady-state velocity of the air ascent is $w = E/\gamma_s\rho_{as}$, where E is the evaporation. The global mean $\overline{w} = 2.5$ mm/s at $\overline{E} = 1$m/year, $\gamma_s = 2 \times 10^{-2}$, $\rho_{as} \approx 1$ kg/m^3 (L'vovich, 1970). The evaporation force disappears as evaporation ceases. Then the velocity w becomes zero as the energy of air flows dissipates.

The difference in the evaporation rate between two adjacent areas of Earth's surface leads to a difference in the evaporation force (f) and the velocity of upstream air flows. This results in horizontal air flows in the lower atmospheric layers from the area with a smaller evaporation to the area with a greater one. The velocity of this horizontal air flow (u) is related to the difference between the vertical velocities (w) in these areas according to the law of conservation of matter (the continuity equation): $u = wL/h_{H_2O}$, where L is the linear size of the area with a greater evaporation. The velocity

u is determined by the condition of coincidence of the power of air mass ascent caused by the evaporation force and the power of the dissipation of the horizontal air mass flow caused by the force of friction against the ground, which may be expressed as $(\beta - \beta_0)\gamma g \sim u^2/l$, where l is the vertical size of the area where the friction force acts. This size corresponds to the fastest change in the velocity u with altitude z and is of the same order of magnitude as the height of the forest cover (Gorshkov and Makar'eva, 2006). At $\gamma_s \sim 10^{-2}$, $l \sim 50$ m and $u \sim \sqrt{\gamma_s g l} \sim 6$ m/s. At $w > \bar{w}$, water the forest biotic pump can bring water from the ocean over a distance of $L = h_{H_2O} u/w > 4000$ km.

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