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Hydrology and Earth System Sciences Discussions

Interactive comment on "Comment on "Biotic pump of atmospheric moisture as driver of the hydrological cycle on land" by A. M. Makarieva and V. G. Gorshkov, Hydrol. Earth Syst. Sci., 11, 1013–1033, 2007" *by* A. G. C. A. Meesters et al.

A. Makarieva

elba@peterlink.ru

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The biotic pump theory raises interest and questions quite widely, which is not surprising given the importance of its potential implications. One stimulating question that we recently received from a research group bears directly to the present DP (p. 412, lines 13-29), so we believe it would be relevant to provide a response to it within the present discussion. This response might also help the DP authors understand why condensation is not a "gradual process" (S174), which rate will "remain limited" because "the vapor pressure equals its saturation value, which is a function of temperature alone."

(S173) The latter statements of the DP authors logically complement what is already there in the DP (S1).

The question was: "We are particularly interested in modelling how the strength of the biotic pump varies with distance inland and forest transpiration rates. But, although you discuss the importance of forest canopy evaporative fluxes (p. 1025) in your paper, we cannot see which equations this number feeds into to influence the vertical and horizontal velocity of the air movements. Does it ultimately influence A_E or A_f? Our concern is that none of the terms in these equations (p. 1026) appear directly linked to forest evaporation."

Let us consider two regions, the donor region (ocean) and the acceptor region (forest) with lengths L_d and L_a, respectively. (A clarifying figure is provided at www.bioticregulation.ru/pump/pump3-3.php#fig.) Let us for simplicity imagine the atmosphere above the two regions as two boxes of one and the same height h and width D and lengths L_d and L_a, respectively. Mean water vapor concentration within these boxes is N_d and N_a, respectively (dimension mol H_2O/m^3).

Let E_a and E_d (dimension mol H_2O m^{-2} s^{-1}) be the regional mean evaporation from the forest cover and the ocean, respectively. Our task is to find the link of E_a to the horizontal air velocity u delivering atmospheric moisture from ocean to the forest.

To do so, we have to consider several mass balance equations.

The first one is for atmospheric air that crosses the ocean-forest interface with horizontal velocity u. All air that comes into the forest box via vertical cross-section of area hD ascends within this box at vertical velocity w_a over the area L_aD, $uhD = w_aL_aD$, so we have:

$$uh = w_aL_a \tag{1}$$

It might be helpful to keep characteristic figures in mind: with a river basin length L_a = 2000 km, $h \sim 8$ km and horizontal velocity u = 5 m/s, the vertical velocity w_a of

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large-scale air ascent is 2 mm/s.

The second mass balance equation is for water vapor that leaves the donor region (the ocean box). In the stationary case (assuming that there is no precipitation in the donor region) all water that evaporates above the ocean per unit time (which is $E_d L_d D$) is taken away with the air flux to the acceptor region at a rate $uN_d hD$, so we have:

$$E_d L_d D = uN_d hD = \text{Runoff} \quad (2)$$

Term $uN_d hD$ taken per unit surface of the forest river basin is conventionally termed as moisture convergence (e.g., Marengo 2005). In the stationary case all this amount of water returns to the ocean as runoff. So, the stationary runoff is directly proportional to the stationary horizontal velocity u.

Now we will see what happens with water vapor in the forest region. Obviously, there is an ascending flux F_a of water vapor (dimension mol H_2O/s) that consists of water vapor evaporated from the forest AND water vapor brought from the donor region:

$$F_a = E_a L_a D + uN_d hD \quad (3)$$

Note that F_a is equal to total regional precipitation. On the other hand, we know that in the forest region atmospheric air ascends at velocity w_a and water vapor ascends together with it. Additionally, water vapor that has a highly non-equilibrium distribution is, unlike the other gases, intensely transported by eddy diffusion (see discussion on p. 1023 of Makarieva and Gorshkov 2007), so we have for F_a:

$$F_a = (N_a w_a + \nu dN_a/dz) L_a D \quad (4)$$

We note that $dN_a/dz = N_a/h_{\mathrm{H_2O}}$, where $h_{\mathrm{H_2O}}$ is the exponential scale height of water vapor. Additionally, according to the biotic pump physics, eddy diffusion coefficient ν is of the order of $\nu \sim h_{\mathrm{H_2O}} w_a = c h_{\mathrm{H_2O}} w_a$ (p. 1023), where c is a parameter of the order of

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unity. This allows us to re-write Eqs. (3) and (4) together (use also Eq. (1) to express u in terms of w_a) as

$$(N_a w_a + c w_a N_a) L_a D = E_a L_a D + (w_a L_a / h) N_d h D \tag{5}$$

From Eq. (5) and Eq. (1) we obtain the following relationships between forest evaporation E_a and vertical and horizontal air velocities w_a and u:

$$w_a (N_a - N_d + c N_a) = (uh/L_a)(N_a - N_d + c N_a) = E_a \tag{6}$$

Assuming that concentration of water vapor is approximately equal in the forest and oceanic atmosphere, $N_a \approx N_d$, and recalling that $c \sim 1$, we obtain a simple approximate relation $E_a \sim w_a N_a$.

Now using Eqs. (1) and (2) we can see that the stationary surface-specific runoff R (i.e. the amount of moisture brought to the forest from the ocean per unit area of the forest river basin, dimension mol H_2O m^{-2} s^{-1}) is directly proportional and close in the order of magnitude to forest evaporation:

$$R = u N_d h D / (L_a D) = w_a N_d = E_a N_d / (N_a - N_d + c N_a) \sim E_a \tag{7}$$

From Eq. (7) and the well-known mass balance equation $P = E_a + R$, where P is precipitation in the river basin, it follows that runoff should constitute approximately one half of precipitation. This is consistent, for example, with the data for the Amazon river basin (Marengo J.A. Climate Dynamics (2005) 24: 11-22), for which observations give $P \approx 6$ mm/day and $R = P/2 = 3$ mm/day. Note the conversion factor: 1 mm/day = 0.64×10^{-3}mol H_2O m^{-2} s^{-1}.

At this point we have resorted to the biotic pump physics once – when we used the expression for eddy diffusivity to go to Eq. (5) from Eq. (4). Now we turn to the biotic pump physics to know about the important constraints on the sizes of L_a and L_d of the stationary donor and acceptor regions. **Not all regions but only those of a particular size can enjoy a stationary circulation, high evaporation and runoff.**

We now write the power of the evaporative force f_E as $A_E = \Delta p w_a D L_a$ (in the notations of Makarieva and Gorshkov (2007) $L_a = L$, $w_a = w_f$, $\Delta p = f_E h_{\mathrm{H_2O}}$). Another physically transparent interpretation of this magnitude is the rate at which potential energy is released during condensation of water vapor ascending with the air at velocity w_a. It is more correct to write this power as $\Delta p(1+c) w_a D L_a$ to take into account the above discussed two terms of the water vapor transport, Eq. (4). The condition of stationarity consists in the statement that the power released during water vapor condensation is spent to counteract the power of surface friction, which we represent as $A_f = \rho g z_T u D L_a$, where z_T is surface roughness scale and $\rho g z_T$ has the meaning of turbulent friction of rest per unit surface area (stress), see (Makarieva et al. (2009) ACPD 8: S11275, http://www.cosis.net/copernicus/EGU/acpd/8/S11275/acpd-8-S11275.pdf and Makarieva and Gorshkov 2009 HESSD 6: S59, http://www.cosis.net/copernicus/EGU/hessd/6/S59/hessd-6-S59.pdf) for discussion of these notions. (Note that in the work of Makarieva and Gorshkov (2007) A_f was represented in a different way. A more accurate treatment was given later in ACPD and HESSD.) The value of z_T is dictated by the properties of forest cover. We thus have in the stationary case dropping DL_a in both sides, $A_E = A_f$:

$$\Delta p(1+c) w_a = g z_T u \tag{8}$$

Now combining Eqs. (1) and (8) we obtain the horizontal linear scale where the stationary circulation can take place:

$$L_a = h \Delta p(1+c)/(\rho g z_T) \tag{9}$$

This result means that if natural forest cover is reduced so that the length of the forest region is smaller than L_a, no stationary circulation will be possible. What consequences

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will it have for the forest basin?

In the stationary case an intensely evaporating forest must possess a high store of soil moisture. This persistent store of moisture is the major prerequisite for a high evaporation. But moist soil loses water due to gravity. These losses must be compensated and soil moisture recharged by precipitation. However, the rate of soil moisture recharge is limited and not correlated with the precipitation rate. If precipitation is too intense, most part of it will be lost to the ocean producing floods rather than stored in the soil. That is, very well moistened soil can only be recharged by an amount of water it has lost.

If the original forest cover is reduced below L_a, no balance between the evaporative force and surface friction will be observed any longer. Air will accelerate up to much higher velocities than those corresponding to the stationary evaporation rates E_a and E_d. In the result, water vapor accumulated in the atmosphere will be spent nearly instantaneously producing intense storms and floods. Most part of precipitated water will be lost to the ocean. Then, prescribed by the energy conservation law, a long period of drought will follow during which the depleted store of atmospheric moisture will be replenished via the slow process of evaporation. However, during this dry time the runoff will continue! Soil will continue to lose water. This loss will remain uncompensated because the water intended to balance it came with too intense a thunderstorm and was wasted to the ocean.

(On the dependence of the rate of condensation on vertical velocity: due to the presence of a sufficiently large vertical lapse rate of air temperature, an air parcel lifted to about 2 km above the ground surface will have most of its water condensed. Therefore, the rate of condensation is directly proportional to vertical air velocity - the more rapidly air parcels ascend, the higher the rate of condensation and, consequently, precipitation. In hurricanes, for example, condensation rate can reach 15 mm/hour (Miller B.I. 1964 Mon Wea Rev 92: 389-406), which is over a hundred of times a higher rate than the global mean evaporation rate of 1 m/year. Unlike the evaporation rate that is limited by solar energy, condensation can occur at an arbitrarily high rate depending on vertical air velocity.)

The uncompensated gravitational loss of water to the ocean will be making the soil even drier. The soil will be unable to sustain the original high flux of evaporation. Due to this positive feedback, this will lead to the diminishment of the incoming moisture store from the ocean, further depletion of moisture store and so on. The region becomes arid. Ultimately the circulation reverses its direction. The particular time scale of these events will depend on the local properties of soil moisture, character of deforestation, geophysical parameters of the region.

Another theoretical problem to be quantitatively solved within the biotic pump theory is to illustrate the stability of circulation directed towards the region with higher evaporation E_a from the region with lower evaporation E_d. Since the dynamic power for air circulation comes from condensation, it is in principle possible that if an intense condensation event incidentally occurs in the region with lower evaporation, the air will start moving there. However, such circulation will be unstable. That this is so is most evident in the case of desert and ocean. If we artificially organize an intense condensation event above the desert surface, moist air will initially move to the desert from the ocean sustaining condensation and, hence, the ocean-to-desert circulation itself. However, as far as the only source of moisture is the one brought horizontally (local evaporation is zero), any occasional decrease of the horizontal velocity will diminish the incoming water vapor flux, diminish the condensation power and, consequently, further diminish the power of circulation and horizontal velocity. These effects will cascade until the circulation ceases. If, in contrast, there is a significant flux of local evaporation (like in the forest), this flux serves as a guarantee that the condensational power that generates the circulation never falls below a certain non-zero value. An exact theoretical treatment of this problem is to be presented elsewhere.

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