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Interactive Comment

Interactive comment on "Biotic pump of atmospheric moisture as driver of the hydrological cycle on land" *by* A. M. Makarieva and V. G. Gorshkov

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In response to the comments of Y. Dovgaluk:

Dr. Dovgaluk raises the question of how the notion of biotic pump and the underlying physics (evaporative force) relate to the widely accepted paradigms of the modern meteorological thought, namely (1) convective instability for the vertical air movements and (2) temperature-related barometric gradient for the horizontal air movements. In our response we first show that the proposed approach quantitatively accounts for the major observed parameters of atmospheric circulation and then discuss the advantages of this approach as well as the critical points in the traditional meteorological

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paradigm.

1. Vertical air movements

1.1. The evaporative force and the dynamic and diffusional fluxes of sensible and latent heat

The evaporative force rises air masses at velocity w to the characteristic height $h_{\mathrm{H_2O}}$ (p. 2634, Eq. 10), which describes the vertical distribution of atmospheric water vapor. The dynamic flux of sensible heat from the Earth's surface to the atmosphere, which is associated with this upward transport of the warm surface air to the upper cooler atmosphere, is equal to $A_c = wC_pN_s\Gamma_{ob}h_{\mathrm{H_2O}}$. (In this expression we have taken into account that that the scale height of atmospheric air $h >> h_{\mathrm{H_2O}}$ (p. 2634, lines 7, 10), so the integral of N from the surface to $h_{\mathrm{H_2O}}$ is approximately equal to $N_sh_{\mathrm{H_2O}}$). Here $C_p = 29$ J mol^{-1} K^{-1} is the molar heat capacity at constant pressure, $N_s = 45$ mol m^{-3} is air molar density at the surface ($N_s = \rho_s/M$, ρ_s is air mass density at the surface, $M = 29$ g mol^{-1} is air molar mass), $\Gamma_{ob} = 6.5$ K km^{-1}, $h_{\mathrm{H_2O}} = 2.4$ km (see p. 2634, line 1; p. 2637, line 7; p. 2638, line 20). The dimension of A_c is W m^{-2}.

The dynamic flux of latent heat from the surface to the vapor-saturated atmosphere, driven by the evaporative force, is equal to $A_w = wN_{\mathrm{H_2O}s}Q_{\mathrm{H_2O}} = w\gamma_sN_sQ_{\mathrm{H_2O}}$, where $\gamma_s \equiv N_{\mathrm{H_2O}s}/N_s \sim 2\times 10^{-2}$ at the global mean surface temperature of 15 $^{\mathrm{o}}$C (p. 2641, line 11), $Q_{\mathrm{H_2O}} = 44$ kJ mol^{-1} (p. 2634, line 20). For the ratio between the fluxes of sensible and latent heat (Bowen ratio) we thus obtain:

$$B = \frac{A_c}{A_w} = \frac{h_{\mathrm{H_2O}}C_p\Gamma_{ob}}{\gamma_sQ_{\mathrm{H_2O}}} = 0.01\gamma_s^{-1} \sim 0.5. \tag{1}$$

This theoretical estimate obtained from the consideration of the evaporative force agrees satisfactorily with observations (e.g., Palmen and Newton, 1969). Moreover, since γ_s grows with surface temperature, formula (1) provides theoretical grounds for the observed pattern that mean Bowen ratio is lowest at the equator and grows towards

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the poles (Palmen and Newton, 1969).

At the Earth's surface the dynamic fluxes of sensible and latent heat are supplemented by the diffusional flux a_c of sensible heat, that is proportional to the vertical lapse rate of air temperature, and by the diffusional flux a_w of latent heat dictated by the non-equilibrium state of atmospheric water vapor. Both fluxes arise due to atmospheric turbulence. Eddy flux of sensible heat a_c is expressed via the coefficient of eddy diffusivity (kinematic viscosity) ν as $a_c = \nu C_p N_s \Gamma_{ob}$. Putting $\nu = w h_{\mathrm{H_2O}}$ (p. 2642, line 28) we obtain $a_c = A_c$. That is, total flux of sensible heat from the surface to the atmosphere, which is equal to the sum of the dynamic and diffusional fluxes, is equal to two times the dynamic flux. The eddy flux of latent heat is $a_w = -\nu \left[\frac{\partial N_{\mathrm{H_2O}}}{\partial z} - \left(\frac{\partial N_w}{\partial z}\right)_0\right] Q_{\mathrm{H_2O}}$, where low index "0" refers to the concentration gradient of water vapor in hydrostatic equilibrium. Using the equation of state for water vapor, $N_{\mathrm{H_2O}} = p_{\mathrm{H_2O}}/(RT)$, we obtain $a_w = \dfrac{\nu}{h_{\mathrm{H_2O}}} \gamma_s N_s Q_{\mathrm{H_2O}} \left(1 - \dfrac{h_{\mathrm{H_2O}}}{h_w}\right) = \dfrac{\nu}{h_{\mathrm{H_2O}}} \gamma_s N_s Q_{\mathrm{H_2O}} \times 0.82$. Putting $\nu = w h_{\mathrm{H_2O}}$ we have $a_w \approx A_w$, so the total flux of latent heat is also approximately equal to twice the dynamic flux of latent heat A_w; thus, the estimated Bowen ratio (1) for total fluxes remains unchanged.

Expression $\nu = w h_{\mathrm{H_2O}}$ that we suggested for the eddy diffusion coefficient ν, is based on dimensional considerations: we propose that atmospheric turbulence is caused by the evaporative force, which determines two dimensional parameters, w (mean velocity of the vertical movement of air masses) and $h_{\mathrm{H_2O}}$ (scale height of the non-equilibrium distribution of atmospheric water vapor). The relationship $\nu = w h_{\mathrm{H_2O}}$ should be true to the accuracy of a dimensionless multiplier of the order of unity. We have put this multiplier equal to unity using the global mean velocity $\bar{w}$ estimated from the global mean value of the latent heat flux (mean global flux of evaporation $\bar{E}$). When estimating $\bar{w}$ (p. 2641, Eq. (15)) we equated the dynamic flux of latent heat A_w to the available estimate of the mean global latent flux, $A_w = Q_{\mathrm{H_2O}} \bar{E}$. Our present account of the eddy flux $a_w \approx A_w$ of latent heat, $a_w + A_w \approx 2A_w = Q_{\mathrm{H_2O}} \bar{E}$, leads to a two-fold reduction of

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the resulting estimate for $\bar{w}$, from 2.5 mm s^{-1} (p. 2641, Eq. (15)) to 1.3 mm s^{-1}. In the result, the estimate for $\bar{\nu}$ decreases from 5 m^2 s^{-1} at $h_{\mathrm{H_2O}} \sim$ 2 km to $\bar{\nu}$= 3.1 m^2 s^{-1} at $h_{\mathrm{H_2O}} \sim$ 2.4 km (p. 2637, line 7). This theoretically obtained eddy diffusion coefficient coincides to the accuracy of a dozen per cent with the phenomenological values used in general circulation modelling (e.g., 3.5 m^2 s^{-1} in the modelling of Hadley circulation (Fang and Tung, 1999)).

In accordance with the second law of thermodynamics both diffusional and dynamic fluxes of sensible heat are directed upward, from the warmer to the colder atmospheric layers. Eddy flux of atmospheric water vapor is also always directed upward as dictated by the non-equilibrium vertical distribution of water vapor. In the meantime, the dynamic (mass) fluxes of air and atmospheric water vapor can, generally, be directed either up or down (upwelling and downwelling regions of the atmosphere). Thus, in the regions of downwelling the upward diffusional and the downward dynamic fluxes of water vapor approximately coincide in magnitude at the surface and are of opposite sign. This means that the total flux of water vapor in the regions of downwelling, which is equal to the sum of the dynamic and diffusional fluxes, is close to zero. The order of magnitude of this flux cannot be estimated within the accuracy of the above considerations, we can only predict that it is much less than the absolute magnitude of the dynamic flux of water vapor.

1.2. The evaporative force and constant mixing ratio of dry air

As already noted, the evaporative force rises air masses at velocity w; this leads to formation of the dynamic air flux $F_w = wN$. Dynamic flux of each i-th gas in the air mixture is $F_{wi} = wN_i$, where N_i is molar concentration of the i-th gas. Within this dynamic flow common for all gases there appear additional eddy diffusional fluxes of each gas. These fluxes, as we mentioned in our response to Dr. Sherman, are proportional to the deviation of the concentration gradients of air gases from their hydrostatic

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equilibrium values:

$$F_{ei} = \nu \left[\frac{\partial N_i}{\partial z} - \left(\frac{\partial N_i}{\partial z} \right)_0 \right] = \nu N_i \left(\frac{1}{h} - \frac{1}{h_i} \right) = \nu \frac{N_i}{h} (1 - \beta_i), \tag{2}$$

$$h = \frac{RT}{Mg}, \quad h_i = \frac{RT}{M_i g}, \quad \beta_i = \frac{h}{h_i} = \frac{M_i}{M}, \quad N_i = \frac{p_i}{RT},$$

$$-\frac{\partial p_i}{\partial z} = \frac{p_i}{h}, \quad \left(-\frac{\partial p_i}{\partial z} \right)_0 = \frac{p_i}{h_i}.$$

The first term $\partial N_i / \partial z$ in (2) is calculated taking into account the observed constant mixing ratio of the major dry air constituents and the constant molar mass M of atmospheric air, which correspond to a single scale height h for all air gases except water vapor. The second (equilibrium) term $(\partial N_i / \partial z)_0$ corresponds to Boltzmann's distributions for different molar masses M_i of each gas. Using relationship $\nu = w h_{\mathrm{H_2O}}$ it is possible to quantify the ratio between the magnitudes of eddy and dynamic fluxes of each gas:

$$\varepsilon_i \equiv \frac{F_{ei}}{F_{wi}} = \frac{1 - \beta_i}{\beta}, \quad \beta \equiv \frac{h}{h_{\mathrm{H_2O}}} \quad (\text{see p. } 2636, \text{ line } 26 \text{ Eq. } 12).$$

Using the known β_i for air gases we obtain:

$$\varepsilon_{\mathrm{N_2}} = 0.01; \quad \varepsilon_{\mathrm{O_2}} = -0.03; \quad \varepsilon_{\mathrm{CO_2}} = -0.15; \quad \varepsilon_{\mathrm{H_2}} = 0.19. \tag{3}$$

Eddy fluxes (turbulent mixing) of atmospheric gases work to restore the state of hydrostatic equilibrium for each gas, when different gases would have different scale heights and the mixing ratio of dry air and its molecular mass would be changing with height.

However, as can be seen from (3), for all gases their dynamic fluxes are much greater than the eddy diffusional fluxes. Hence, it can be concluded that the constant mixing ratio of dry air can be explained by the small relative value of eddy fluxes as compared to the dynamic fluxes of each gas maintained by the evaporative force.

It should be noted that for CO_2 at $\nu = wh_{H_2O}$ the ratio of eddy to dynamic fluxes (they are opposite in direction) is equal to 15%. This means that if atmospheric turbulence were the same at all heights in the troposphere, then the deviation of atmospheric CO_2 from the observed constant mixing ratio would have been around 15%, which does not agree with observations (370 ppm for CO_2 at any height in the troposphere). This unambiguously suggests that the eddy diffusion coefficient ν (kinematic viscosity) drops rapidly with height. That this is indeed so is supported by the existence of geostrophic winds in the upper atmosphere that blow along isobars at practically zero viscosity.

1.3. Conclusions on vertical air movements

In the traditional theoretical consideration of convective instability outlined by Dr. Dovgaluk the velocity w of the vertical air movements and the eddy diffusion coefficient for atmospheric mixing remain undetermined; for modelling purposes they have therefore to be postulated phenomenologically. In contrast, the developed physical approach based on the evaporative force and the non-equilibrium distribution of atmospheric water vapor allows one to quantify the major observed parameters of the atmospheric transport of sensible and latent heat (e.g., Bowen ratio), as well as of the vertical distribution and mass fluxes of air gases. In particular, it numerically explains the observed constancy of the mixing ratio of dry air and estimates the global mean vertical velocity of air movements and the eddy diffusion coefficient (kinematic viscosity) of the atmosphere.

In our paper we demonstrate that the evaporative force exists in the presence of a liquid hydrosphere and moist soil surface (the inherent property of the planetary surface of Earth). The evaporative force (and the associated air circulation) arise when the vertical lapse rate of air temperature exceeds 1.2 K km^{-1} (p. 2635, lines 13, 19). This value is much less than the moist adiabatic lapse rate of 6 K km^{-1}, which is a critical parameter in the conventional consideration of convection. Hence, at the observed vertical lapse rate of air temperature of 6.5 K km^{-1} a consistent theoretical description of meteorological phenomena should necessarily include the evaporative force.

2. Horizontal air movements

When water vapor is out of hydrostatic equilibrium and undergoes condensation in the atmospheric column, hydrostatic equilibrium of moist air as a whole is impossible as contradicting Dalton's law and the kinetic theory of gases, the latter supported by all existing empirical data. In the presence of the evaporative force it is only dry air as a whole that can be in hydrostatic equilibrium; we discussed the physical meaning of this equilibrium in our response to the comments of Dr. Sherman (pp. S1130-S1132). So below under hydrostatic equilibrium we understand hydrostatic equilibrium of dry air.

Indeed, as is well-known and also pointed out by Dr. Dovgaluk, the decrease of surface temperature from the equator to the poles changes the hydrostatic equilibrium of atmospheric air. Scale height h (p. 2634, Eq. (8)), which describes the equilibrium distribution of air, changes proportionally to the surface temperature (p. 2634, lines 10-11). This creates horizontal gradient of air pressure (barometric gradient). From the equation of state $p = NRT$ we have:

$$\frac{1}{p}\frac{dp}{dx} = \frac{1}{T}\frac{dT}{dx} + \frac{1}{N}\frac{dN}{dx},$$

where x is distance counted along the meridian, $x = R\varphi$, φ is longitude and R is Earth's radius. Taking that mean surface temperature $T \sim$ 300 K changes by $\sim$30 K over the distance $\sim$10^4 km from equator to the poles and neglecting the relative change of air molar density, one can conclude that the relative pressure gradient, equal to the relative temperature gradient, is around 10^{-5} km^{-1}. Thus, at mean air pressure of 10^3 bar, mean horizontal gradient of air pressure would be indeed 1 bar (100 km)$^{-1}$, which

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agrees with observations. However, such an estimate presumes that atmospheric pressure at the poles should be about 10% less than at the equator. This is not supported by observations – atmospheric pressure at the poles is approximately the same or even slightly higher than at the equator. This means that it is inappropriate to neglect the relative gradient $dN/(Ndx)$ of air molar density when estimating the horizontal gradient of air pressure.

On the other hand, assuming equal atmospheric pressure at the equator and the poles it is easy to show that the horizontal barometric gradient related to the horizontal gradient of surface temperature is maximized at a height close to height h (indeed, the barometric gradient is put zero at the surface; air pressure exponentially declines with height; and there is no other height scale except h). At such heights the kinematic viscosity is low; consequently, air mixing working to equate air temperature could have led to atmospheric circulation in the upper atmospheric layers only, with little impact on the lower troposphere. Such pattern apparently contradicts the observations. Generally, problems with the conventional account for general circulation based on the equations of hydrodynamics where the kinematic viscosity is borrowed from observations, are not uncommon in the literature. For example, one of such problems is the problem of the unsatisfactory theoretical representation of Hadley circulation (e.g., Fang and Tung, 1999).

As shown in our paper, the conventional meteorological approach cannot explain the existence of the biotic pump of atmospheric moisture that was demonstrated on the basis of precipitation data (see Section 2, pp. 2625-2633 in the paper). Neither can it account for the absence of a monsoon-like climate in deserts (p. 2671, Fig. 2) or for the exponential decline of precipitation with distance from the ocean in the non-forested areas (pp. 2625-2628; 2671; Fig. 2). Consideration of the evaporative force predicts that Hadley circulation (trade winds) can exist even at zero gradient of sea surface temperatures in the tropical zone. Our approach also yields theoretical estimates of wind speeds in such atmospheric structures as hurricanes and tornadoes (see p. 2641

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and our response to Dr. Sherman).

In the last several decades years many attempts have been made to describe the observed atmospheric circulation with use of various numerical general circulation models. These models incorporate the equations of hydrodynamics with all the known forces (Coriolis force, friction force, centripetal force) (McGuffie and Henderson-Sellers, 2001), but the evaporative force has been ignored.

References

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

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

Palmen, E. and Newton, C. W.: Atmospheric circulation systems, their structure and physical interpretation, Academic Press, New York and London, 1969.

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