Condensational power of air circulation in the presence of a horizontal temperature gradient

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Physics Letters A 378 (2014) 294-298.
doi: http://dx.doi.org/10.1016/j.physleta.2013.11.019

Abstract

From the condition of hydrostatic equilibrium and energy conservation a general expression is derived for the power of air circulation induced by water vapor condensation in the presence of a horizontal gradient of temperature. It is shown that the obtained expression for circulation power agrees with the continuity equation. The impact of droplets that form upon condensation on the circulation power is evaluated. Theoretical estimates are compared with observational evidence.

1 Introduction

Phase transitions of water vapor in the air bring about pressure changes and lead to the appearance of pressure gradients that drive air circulation. Condensation of saturated water vapor occurs when the moist air ascends and, consequently, cools, as well as when moist air flows horizontally towards areas of lower temperature. The condition of hydrostatic equilibrium does not allow large vertical velocities to develop, such that all the condensational power released in the ascending air is translated to the power of horizontal winds [1–3].

Dissipation of the horizontal air flow due to friction on the Earth's surface leads to the formation of vortices. Turbulent diffusion enhances evaporation and mixing of water vapor in the air. Therefore, if the air moves towards an area where temperature and humidity are high enough to ensure sufficient evaporation, an increase in the partial pressure of water vapor by evaporation can significantly impede or even fully arrest the condensational air circulation.

In this work, we formulate a general equation for the power of condensational circulation of air in the case of arbitrary temperature gradients. We show that this equation is consistent with the continuity equation for air in the presence of the phase transitions of water vapor. It is also shown that, on average, formation of condensate particles reduces the power of condensational circulation by a small relative magnitude. A qualitative explanation of the observed peculiarities for global circulation of the atmosphere of Earth is given.

2 Dynamic equation of condensational air circulation

In the ascending air all its gaseous components, including water vapor, share the same vertical velocity $\mathbf{w}$. In hydrostatic equilibrium, the increase of potential energy of the

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ascending air in unit volume is equal to the decrease of air pressure p, $(-\partial p/\partial z = \rho g)$, where ρ is air density, g is the acceleration of gravity, $\mathbf{w}$ and z-axis are directed opposite to $\mathbf{g}$. This means that work is not produced on the ascending air, and the kinetic energy of air does not change. In the absence of condensation of water vapor, the relative partial pressures p_i/p of all air gases remain constant and independent of z, because the ascending gases share the same vertical velocity. It follows that partial pressures of the air gases, which all have different molar masses M_i, are equally distributed over height. Their distribution coincides with that of the air as a whole:

$$-\frac{1}{p_i}\frac{\partial p_i}{\partial z} = -\frac{1}{p}\frac{\partial p}{\partial z} = \frac{1}{h}, \qquad h \equiv \frac{RT}{Mg}, \qquad p = \rho g h, \tag{1}$$

where $M = \sum_i M_i p_i/p$ is molar mass of air, R is the universal gas constant, h is the scale height of the vertical distribution of all air gases (height of a uniformly dense air column). The last equality in (1) is equation of state for the ideal gas. The fulfillment of Eq. (1) for the ascending moist air with pressure p is referred to below as the condition of hydrostatic equilibrium.

In a stationary flow, when moist air ascends and cools, the water vapor condenses and leaves the gaseous phase above the horizontal surface (cloud base) where the water vapor is saturated. The vertical distribution of water vapor shrinks towards the Earth's surface and does not conform to Eq. (1):

$$-\frac{\partial p_v}{\partial z} = \frac{p_v}{h_c} \gg \frac{p_v}{h}, \quad h^{-1} \ll h_c^{-1} = h_T^{-1}\xi, \quad h_T^{-1} \equiv -\frac{1}{T}\frac{\partial T}{\partial z}, \quad \xi \equiv \frac{L_v}{RT}, \tag{2}$$

where h_c is determined by the Clausius-Clapeyron equation, L_v is the energy of vaporization (latent heat released during condensation) [1,3], $\xi \gg 1$ is dimensionless.

The fact that the vertical distributions (2) and (1) for the water vapor are different implies that when the ascent is accompanied by condensation, work is performed on the air that should generate kinetic energy with the following power [1–3]:

$$s = w p_v \left(\frac{1}{h_c} - \frac{1}{h}\right) = -w\left(\frac{\partial p_v}{\partial z} - \gamma\frac{\partial p}{\partial z}\right), \quad \gamma \equiv \frac{p_v}{p}. \tag{3}$$

Equation (3) can be re-written in another convenient form:

$$s = w p_v \frac{1}{h_\gamma} = -wp\frac{\partial \gamma}{\partial z}, \quad -\frac{\partial \gamma}{\partial z} = \frac{\gamma}{h_\gamma}, \quad h_\gamma^{-1} \equiv h_c^{-1} - h^{-1}. \tag{4}$$

The power generated by water vapor with partial pressure p_v is of the order of $\gamma \ll 1$ as compared to the power generated by air as a whole with the total pressure p, for example, when the air is filling vacuum. By integrating power s (3) over time ($wdt = dz$), we find that total work corresponding to complete condensation of water vapor in the ascending air is of the order of $p_v = \gamma p$ where p_v and p are values at the Earth's surface ($z = 0$). This work is equal to the kinetic energy $\rho v^2/2 \sim \gamma p$ that would be acquired in the absence of friction by the air volume upon complete condensation of water vapor it contains ($\mathbf{v} = \mathbf{u} + \mathbf{w}$, where $\mathbf{u}$ is the horizontal velocity component). For $\gamma = 3 \times 10^{-2}$, $p = 10^5$ Pa and $\rho = 1.3$ kg m^{-3} this corresponds to velocity $v = (2\gamma p/\rho)^{1/2} \approx 70$ m s^{-1}. Depending on the geometry of the circulation this kinetic energy produced by condensation corresponds to either vertical or horizontal motion or both.

In large condensation areas with horizontal size L significantly exceeding condensation height, the latter being of the order of the atmospheric scale height h, the observed annual mean vertical velocities of the ascending air are of the order of 1 mm s^{-1} [4]. This indicates that any significant vertical acceleration of the air is absent and that condition (1) of the hydrostatic equilibrium is preserved for the air as a whole with total pressure p. In this case, energy conservation and the condition of hydrostatic equilibrium dictate that power s (3) corresponds to the power of the horizontal air flow $-\mathbf{u}\nabla p$, where $\mathbf{u}$ is the vector of horizontal velocity [2]. Flow stationarity for circulating air corresponds to the equality between the characteristic times of the horizontal and vertical air motions (see Sections 3 and 4 below). Thus, the equality between the power released during the vertical motion of the ascending air (3) and the power of the horizontal air motion is equivalent to the equality of the corresponding amounts of work (energy conservation) and takes the form of the following dynamic relationship:

$$s = -\mathbf{w}(\nabla p_v - \gamma \nabla p) \equiv -p\mathbf{w}\nabla\gamma = -\mathbf{u}\nabla p. \tag{5}$$

Equality (5) holds true, if the horizontal air motion is not accompanied by phase transitions of water vapor for constant z, i.e. when $\mathbf{u}\nabla p_v = 0$ and $\mathbf{u}\nabla T = 0$ and the condition of a horizontal isothermality is fulfilled [2, 3]. In this case, the surfaces of constant relative humidity (in particular, the cloud base surface where relative humidity is unity) remain horizontal in the entire circulation area.

It is however not difficult to generalize (5) to the case of a non-zero horizontal temperature gradient. In this case, to the left-hand side of the equality (5) one should add the power related to the change of water vapor partial pressure caused by evaporation or condensation in the horizontal air flow. This power is equal to $-\mathbf{u}\nabla p_v$. The general form of the dynamic equation for the power of condensational circulation (it was obtained from different physical considerations in [6]) is then given by the following relationship:

$$s = -p\mathbf{w}\nabla\gamma - \mathbf{u}\nabla p_v = -\mathbf{u}\nabla p. \tag{6}$$

A horizontal gradient of temperature can enhance condensation, if $-\mathbf{u}\nabla p_v > 0$, or weaken it by evaporation, if $-\mathbf{u}\nabla p_v < 0$ [6].

In coordinate axes x and z, where x is directed along $\mathbf{u}$, the horizontal pressure gradient in (6) has the form (see (1), (2) and (4) in [6]):

$$-\frac{\partial p}{\partial x} = \frac{p_v}{h_\gamma}\frac{w}{u}(1-A), \quad A \equiv -\frac{h_\gamma}{h_c}\frac{u}{w}\frac{\partial T/\partial x}{\partial T/\partial z} = \xi\frac{u}{w}\frac{h_\gamma}{T}\frac{\partial T}{\partial x}. \tag{7}$$

All variables in (7) can be retrieved from observations. As shown in [5] with the empirically measured values entered into the right-hand side of (7), this equation yields an estimate of the mean horizontal pressure gradient in the Amazon river basin that agrees well with observations.

3 Continuity equation in the presence of phase transitions of water vapor

In the stationary case, the continuity equations for the water vapor and the dry air constituents have the form

$$\nabla\mathbf{v}N_v \equiv N_v\nabla\mathbf{v} + \mathbf{v}\nabla N_v = -S, \quad \mathbf{v} = \mathbf{u} + \mathbf{w}, \tag{8}$$

$$\nabla\mathbf{v}N_d \equiv N_d\nabla\mathbf{v} + \mathbf{v}\nabla N_d = 0, \quad N = N_d + N_v, \tag{9}$$

where N_v, N_d and N are the molar densities of water vapor, dry air constituents and moist air as a whole, respectively. Air velocity $\mathbf{v}$ is equal to the sum of the horizontal $\mathbf{u}$ and vertical $\mathbf{w}$ velocity components. The quantity S (mol m^{-3} s^{-1}) represents volume density of the rate at which molar density N_v of water vapor is changed by phase transitions. By multiplying (9) by $\gamma_d \equiv N_v/N_d$ and excluding $N_v\nabla\mathbf{v}$ from (8), we obtain

$$\mathbf{v}(\nabla N_v - \gamma_d\nabla N_d) = -S. \tag{10}$$

Using the ideal gas equation of state (see Eq. (1)):

$$p = NRT, \quad p_v = N_v RT, \quad p_d = N_d RT, \quad \gamma_d = \frac{p_v}{p_d}, \tag{11}$$

it is possible to replace molar densities N_i in (10) with partial pressures p_i $(i = v, d)$ and the rate of phase transitions S with the power of phase transitions s:

$$\mathbf{v}(\nabla p_v - \gamma_d \nabla p_d) = -s, \quad s = SRT. \tag{12}$$

Note that owing to the universality of the gas constant R the temperature gradient ∇T cancels during this procedure and does not appear in the final equation (12). The kinematic continuity relationship (12) should be fulfilled for any s. The magnitude of power s of water vapor phase transitions is not determined by the continuity equation (12). It must be specified independently by using dynamic physical principles [7].

Such a dynamic physical principle is the equality between the power of phase transition s and the power of horizontal circulation:

$$s = -\mathbf{u}\nabla p. \tag{13}$$

Substituting (13) into (12) and taking into account the easily verifiable identity

$$\nabla p_v - \gamma_d \nabla p_d \equiv (1 + \gamma_d)(\nabla p_v - \gamma \nabla p), \quad \gamma \equiv \frac{\gamma_d}{1 + \gamma_d}, \tag{14}$$

we obtain the following relationship for (12):

$$(\mathbf{w} + \mathbf{u})(\nabla p_v - \gamma \nabla p) = \frac{1}{1 + \gamma_d} \mathbf{u}\nabla p.$$

By transferring $\gamma \mathbf{u}\nabla p$ to the right-hand side of the last relationship and taking into account relationship (14) between γ and γ_d, we obtain:

$$p\mathbf{w}\nabla\gamma + \mathbf{u}\nabla p_v = \mathbf{u}\nabla p, \quad p\nabla\gamma \equiv \nabla p_v - \gamma \nabla p, \tag{15}$$

which coincides with Eq. (6).

Thus, the main dynamic equation (6) is in agreement with the kinematic stipulation stemming from the continuity equations (8) and (9) without assuming the existence of a horizontal isothermality. The main dynamic relationship for atmospheric circulation (6) (and (15)) is derived from two physical principles: preservation of hydrostatic equilibrium in the presence of condensation and the law of energy conservation. The latter principle, due to the equality between the characteristic times of the vertical and horizontal motion for the circulation air, corresponds to the equality of powers. These principles are consistent with the continuity equation.

In analysis of air circulation, it is sometimes erroneously assumed (see, e.g., works [8,9]) that condensation rate S in (8) and (10) (and, consequently, condensation power s in (12)) is fully determined by the continuity equation (8) such that its specification does not require any additional physical principles (like energy conservation and the condition of hydrostatic equilibrium). As a justification for such an approach it is considered that S differs from zero only for saturated water vapor, for which the values of N_v and p_v depend exclusively on temperature T as dictated by the Clausius-Clapeyron equation. The temperature, in its turn, is determined from the first law of thermodynamics, which in the numerical models of atmospheric circulation plays the role of the energy-conservation law [10]. However, the first law of equilibrium thermodynamics does not take into account the kinetic energy of the air motion. In the troposphere, the kinetic energy of wind per unit volume does not exceed the value of water vapor partial pressure p_v (J m^{-3}). The ratio of p_v to the total potential energy, which is contained in air pressure p, is a small value $\gamma = p_v/p \sim 10^{-2}$. Temperature T in the entire atmospheric column, where air pressure p changes by a hundred per cent, can be obtained from the first law of equilibrium thermodynamics with the necessary accuracy of the order of γ only after the magnitudes of S and s are specified from independent dynamic physical principles [2,3,6].

4 Impact of condensate particles on the power of circulation

In most cases, the horizontal size L of the condensation area significantly exceeds the mean height from which condensate particles fall down to the surface, which is of the order of h [11]. In such cases, condensate particles (raindrops, snowflakes, hailstones) that are falling in the rest reference frame of air with a constant terminal velocity $\mathbf{W}$, $\nabla\mathbf{W} = 0$, determined by the equality between the Stokes friction force and particle weight [12], increase air pressure in the condensation/precipitation area. The increment of pressure near the Earth's surface is equal to the cumulative weight of all droplets in the atmosphere in the condensation area. Therefore, for $L/h \gg 1$, local increase of pressure gradient owing to the presence of falling droplets is on average close to the weight of droplets $\rho_l g$ in a unit volume. Mass density of droplets $\rho_l = N_l M_v$, where N_l (mol m^{-3}) is the molar density of droplets and $M_v = 18$ g mol^{-1} is the molar mass of water, can be estimated using the continuity equation for droplets:

$$\nabla(\mathbf{v} + \mathbf{W})N_l = S, \quad \mathbf{v} = \mathbf{u} + \mathbf{w}. \tag{16}$$

For the mean absolute velocity of the ascending air, we have $w \sim 1$ mm s^{-1}, while the mean absolute velocity of falling droplets is $W \sim 1$ m s^{-1}, i.e. we have $W/w \sim 10^3 \gg 1$. As far as $\gamma_d = N_v/N_d \ll 1$, the magnitude of N_d is influenced by condensation insignificantly – by a small relative amount of the order of γ_d. Therefore, N_d can be factored out from the sign of differentiation in the left-hand part of (9). Then Eq. (9) becomes $\nabla\mathbf{v} = 0$ and can be re-written as $\partial x/\partial u = -\partial z/\partial w$ or $L/u \sim h/w$. This reflects the equality between the characteristic times of the horizontal and vertical air motions, which was used when the powers of these motions were equated in Eqs. (5) and (6). Taking into account that $\nabla\mathbf{v} = 0$, we have $N_l\nabla\mathbf{u} \sim N_l\nabla\mathbf{w} \ll \mathbf{W}\nabla N_l$ ($\nabla\mathbf{w} \sim w/h$, $|\nabla N_l| \sim N_l/h$). To the relative accuracy of the order of w/W, we can re-write relationship (16) using (5) and (11) as follows

$$(\mathbf{v} + \mathbf{W})\nabla N_l = S = -N\mathbf{w}\nabla\gamma. \tag{17}$$

For $w \ll u \leqslant W$ (i.e., when the angle of the direction of motion of falling droplets to the vertical is small), the first term in the left-hand part of (17) does not exceed the second one. Accounting that $\mathbf{W}$ and $\mathbf{w}$ are of opposite directions in the area where the air ascends and condensation occurs and using the absolute magnitudes of these velocities w and W, we have from (17):

$$W\frac{\partial N_l}{\partial z} \leqslant Nw\frac{\partial\gamma}{\partial z}. \tag{18}$$

Owing to the smallness of $\gamma \sim \gamma_d \ll 1$ the vertical air flow Nw is changed by a relative magnitude of the order of γ, i.e. it remains practically independent of height, in the condensation area $z_1 \leqslant z \leqslant z_2$, where $\partial\gamma/\partial z < 0$, and γ is changed by a relative magnitude of the order of hundred per cent [11]. Therefore, after integrating (18) over z, we obtain for ρ_l:

$$\rho_l(z) = \rho_v(z)\frac{w}{W}\left(1 - \frac{\gamma(z_2)}{\gamma(z)}\right), \tag{19}$$
$$\rho_l = N_l M_v, \quad N_l(z_2) = 0, \quad \rho_v = \gamma N M_v, \quad \gamma(z_2) \leqslant \gamma(z) \leqslant \gamma(z_1).$$

Taking into consideration the power of the droplets' impact on the ascending air, $w\rho_l g$, we obtain from (19) the following expression for total condensation power (6) (see

Eq. (4)):

$$s = -(1-\alpha)p\mathbf{w}\nabla\gamma - \mathbf{u}\nabla p_v, \quad \alpha \equiv \frac{w}{W}\frac{h_\gamma}{h_v}\left(1 - \frac{\gamma(z_2)}{\gamma(z)}\right), \tag{20}$$

$$h_\gamma^{-1} \equiv h_c^{-1} - h^{-1}, \quad h_v \equiv \frac{RT}{M_v g} = \frac{p_v}{\rho_v g}, \tag{21}$$

where $h_v = 13.5$ km is the exponential scale height of the unsaturated water vapor in motionless air. Height h_c describes the vertical distribution of the saturated water vapor and, depending on surface temperature, varies between 2.4 km and 5 km [2, 6]. Height $h = 8.4$ km at 300 K determines the vertical distribution of air as a whole and the unsaturated water vapor in the ascending air. For any temperature $h_\gamma/h_v < 1$. The magnitude enclosed in round brackets in (20) is also less than unity. The smallness of α is mainly determined by the ratio $w/W \sim 10^{-3}$. Thus, for $w/W \ll 1$, the reduction of the condensational power (6) due to precipitating condensate particles is small and can be neglected.

Velocity of falling droplets W becomes smaller than air velocities u and w only when condensation is spatially concentrated like in the windwall of hurricanes ($w \ll W < u$) and tornadoes ($W < u$, $W \ll w$). In hurricanes, the horizontal size of the windwall still significantly exceeds the condensation height [13], such that the vertical air velocity remains small: we have $w \lesssim 10$ cm s^{-1} and $w/W \lesssim 10^{-1}$. In tornadoes, the horizontal size of the windwall is of the same order of magnitude as the height of condensation [14], such that in the windwall the air velocities u and w both exceed the velocity of falling droplets W. In this case, however, the impact of droplets on the air cannot be described as $\rho_l g$, i.e. as the local impact of all droplets in a unit air volume on the same unit air volume. The force imposed by droplets on the air, which corresponds to a pressure gradient, decreases quadratically with increasing distance r from the droplet. The potential of this force, which corresponds to pressure, decreases linearly with growing r. For a stationary jet of droplets, this potential (pressure) decreases logarithmically with increasing distance r from the jet (similarly to the potential of a charged wire), if $r \leqslant h$, where h is the height of the jet.

Thus, the dynamic impact of droplets that are concentrated in the windwall of radius r_0 and thickness $\triangle r \sim r_0$ spreads over the entire condensation area. In hurricanes, condensation height is $h_\gamma \sim 4$ km, windwall radius is $r_0 \sim 40$ km, radius of condensation area is $R_0 \sim 400$ km [13]. Pressure of droplets that are precipitating in the windwall declines linearly with growing distance from the windwall and spreads over the entire condensation area. In the result, the pressure of droplets in the windwall decreases and becomes of the order of $r_0/R_0 \sim 10^{-1}$ compared to the weight of droplets in the windwall. Therefore, the magnitude of α (20) in hurricanes is of the order of $\alpha \sim wr_0/(WR_0) \sim 10^{-2}$. Owing to the high vertical velocity of the ascending air in the windwall of tornadoes, where $h \sim r_0$, $r_0/R_0 \sim 10^{-1}$ [14], the condensate particles are distributed practically uniformly over the entire condensation area, such that there is no precipitation in the windwall. Therefore, in tornadoes, $\alpha \sim (r_0/R_0)^2 \sim 10^{-2}$ as well as in hurricanes. To summarize, in all types of intense circulation the impact of condensate particles on the condensational power (6) is practically negligible.

In some papers (see, e.g., [15, 16]), on the basis of an incorrect interpretation of Newton's third law about the equality of the action and reaction between two interacting bodies it is stated that the force with which droplets act on a unit air volume is always equal to the force by which this air volume acts on the droplets enclosed within this volume, i.e. $\rho_l g$. In contrast to a set of droplets non-interacting with each other, the unit air volume, where the droplets are located, is not a body with which these droplets are interacting. Such a body is represented instead by the atmospheric air as a whole, with the atmospheric air pressure determined by the gravitational interaction of the air with the Earth. The impact of droplets on the air is determined by the elastic properties of the air. It declines with increasing distance from the considered volume of droplets according to the physical relationships described above.

5 Discussion

The condensational power of wind is determined by Eq. (6). Fog and dew, i.e. condensation phenomena not related to vertical air motions, contribute a very small proportion to the total precipitation, which is largely determined by condensation in the ascending air. Consequently, wind power should be proportional to precipitation. This relationship can be obtained from formula (5) for the case of a horizontal isothermality. By integrating (5) over height z and using the equation of state (11), we obtain the power of wind per unit area of the Earth's surface Π (W m^{-2}):

$$\Pi = \int_0^\infty s dz = -\int_0^\infty pw\frac{\partial\gamma}{\partial z}dz \approx PRT, \quad P \equiv N_v w|_{z=z_1}, \tag{22}$$

where z_1 is the height where relative humidity reaches unity (the cloud base height), P (mol H_2O m^{-2} s^{-1}) is precipitation flux at the Earth's surface, T is the mean temperature in the condensation layer $z_1 \leqslant z \leqslant z_2$, where $\partial\gamma/\partial z < 0$. The integral in Eq. (22) is calculated, taking into account that in the condensation layer $z_1 \leqslant z \leqslant z_2$ the vertical air flow Nw changes little, by no more than a small relative magnitude $\gamma \ll 1$, while the relative change of $\partial\gamma/\partial z$ is of the order of unity. The smallness of changes in Nw follows from the continuity equation (sum of (8) and (9)) with an account of the fact that S in (8) and (10) is a small magnitude of the order of γ. Putting the observed value for global mean annual precipitation into (22) $(\overline{P}M_v)/\rho_{\rm w} = 1$ m y^{-1} $\approx 10^{-1}$ mm hr^{-1} [17], $\rho_{\rm w} = 10^3$ kg m^{-3} is liquid water density, $R = 8.3$ J $mol^{-1}K^{-1}$, $T \simeq 300$ K, we obtain $\Pi = 4$ W m^{-2} in good agreement with empirical estimates [3, 18–20]. In the hurricane windwall, we have $(\overline{P}M_v)/\rho_{\rm w} \approx 20$ mm hr^{-1}, such that Eq. (22) yields $\Pi \sim 1$ kW m^{-2}. Thus the wind power is generated in the regions of condensation. Relationship (5) means that in those areas, where condensation is absent, we have $\mathbf{u}\nabla p = 0$.

Release of latent heat during condensation diminishes the cooling rate of the ascending air by a magnitude equal to the difference between the dry (9.8 K km^{-1}) and moist ($\sim$ 4 K km^{-1}) adiabatic lapse rates of air temperature [2]. The decrease of cooling rate does not change the amount of condensed water vapor and the associated decline of mean pressure in the atmospheric column and at the Earth's surface. At heights $z > z_2$, where condensation ceases, the air flow changes its direction from ascending to horizontal motion towards the area of descent. This transition from vertical motion to horizontal motion towards the area of descent occurs at the expense of forces of air elasticity in the gravitational field, which form the pressure gradient that is perpendicular to the direction of air motion (similar to the forces that keep satellites on their circular orbits). No additional power is generated during such transitions. While some part of the latent heat released in the condensation area is radiated into space and some propagates via turbulent heat flows, the remainder is transferred towards the area of descending air motion, where condensation is absent. The descending air warms at a dry adiabatic lapse rate (9.8 K km^{-1}). If this area of descending air motion is located over the hydrosphere (ocean, sea), the warming causes enhanced evaporation and increases the amount of atmospheric water vapor, which then travels in the lower atmosphere with the air flow towards the condensation area thus intensifying the circulation.

Accounting for the horizontal temperature gradients becomes essential at the ocean-land interface. A forest-covered river basin during the season of photosynthesis is characterized by lower pressure than the adjacent ocean or unforested territories (the latter having a higher surface temperature than the forest) [5,21]. In this case, the flow of moisture evaporated from the ocean towards the interior of the forest-covered river basin is spatially and temporarily homogeneous such that either floods or droughts are absent [1,6]. When the oceanic air arrives to the shore of an unforested continent, which has a higher temperature than the ocean, the value of A in (7) can approach unity. This fully suppresses condensation-induced dynamics as the corresponding pressure gradient turns to zero. The moist air flow towards the continental hinterland is stopped at the ocean-land border. This leads to an increase in the concentration of the saturated water vapor, intensification of precipitation in the coastal zone and possible floods [22].

The cause-effect relationships between the vertical motion of moist air (deep convection) and air convergence in the lower atmosphere are actively investigated (see, e.g., work [23] for a discussion). Here we have shown that condensational circulation power (5) and the horizontal air convergence can exist in the absence of a horizontal temperature gradient. Friction forces transform this power into the power of turbulent eddies. The resulting intense turbulent diffusion and turbulent heat conductivity work to erase the horizontal temperature inhomogeneities that are not related to condensation. The remaining temperature gradients associated with latent heat release are the consequence of condensation rather than the cause of the wind power generation.

The impact of condensate particles on wind power generation becomes significant only in the case of formation of the smallest droplets falling with their terminal velocity W, which are smaller than the vertical velocity w of the ascending air, $W < w$. If such smallest droplets form over a large horizontal area that exceeds in its linear size the height of condensation, the value of α (20) may approach unity. Under such conditions, condensation power comes close to zero, the intensity of the ascending air motion and condensation greatly diminishes, and a thin layer of cloudiness composed of the smallest droplets can be maintained in the atmosphere for a long time with the horizontal wind practically absent.

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