Atmos. Chem. Phys. Discuss.,
doi:10.5194/acp-2016-203-AC4, 2016

Atmospheric Chemistry and Physics Discussions
Open Access
EGU

Interactive comment on "Quantifying the global atmospheric power budget" *by* A. M. Makarieva et al.

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Received and published: 11 July 2016

Here we reply to Comment 1 of Referee 3 who suggests that, contrary to our claims, our main results and specifically Eqs. 20-22 are not original. Presumably there is some misunderstanding involved so we have revised the text clarifying how our results relate to previous work. In particular, we now show that Eqs. 20-22 could not *in principle* be formulated by Pauluis et al. 2000, because their basic assumptions are not consistent with either Eqs. 20-22 or with Eq. 4 of Pauluis and Held 2002. We acknowledge the value in making this claim clear and explicit as it is precisely because Eqs. 20-22 were not published previously that the global gravitational power of precipitation W_P could also not be estimated from re-analyses until now. Eqs. 20-22 are distinct in showing that W_P can be estimated directly from air velocity and pressure gradient without any knowledge of the atmospheric moisture content or precipitation rates.

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Discussion paper

The revised Section 4 (now Section 3) of our manuscript can be found below[1], where subsection 3.3 is devoted to comparing our results with those of Pauluis et al. This is followed by the referee's comment and our reply to it.

[1] In the revised manuscript this section follows the revised Section 2 published in our previous comment [doi:10.5194/acp-2016-203-AC3]. The relevant equations from the previous sections are as follows:

$$W_I \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{M}} \mathbf{v}\cdot\nabla p \alpha d\mathcal{M} = -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} \mathbf{v}\cdot\nabla p d\mathcal{V}, \tag{1}$$

$$W_{II} \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} \mathbf{u}\cdot\nabla p d\mathcal{V}, \tag{2}$$

$$W_{III} \equiv \frac{1}{\mathcal{S}}\int_{\mathcal{V}} p\nabla\cdot\mathbf{v} d\mathcal{V}, \tag{3}$$

$$W_{IV} \equiv \frac{1}{\mathcal{S}}\int_{\mathcal{M}} p\frac{d\alpha}{dt} d\mathcal{M}, \tag{4}$$

$$\frac{dX}{dt} \equiv \frac{\partial X}{\partial t} + \mathbf{v}\cdot\nabla X, \tag{5}$$

$$\frac{\partial \rho}{\partial t} + \nabla\cdot(\rho\mathbf{v}) = \dot{\rho}, \tag{6}$$

$$W_p \equiv \frac{p}{\tilde{V}}\frac{d\tilde{V}}{dt}, \tag{7}$$

$$W \equiv \frac{1}{\mathcal{S}}\sum_{i=1}^{n} W_{pi}\tilde{V}_i = \frac{1}{\mathcal{S}}\int_{\mathcal{V}} W_p d\mathcal{V} = \frac{1}{\mathcal{S}}\int_{\mathcal{V}} p(\nabla\cdot\mathbf{v}) d\mathcal{V}. \tag{9}$$

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Discussion paper

3 Revisiting the current understanding of the atmospheric power budget

3.1 The boundary condition for vertical velocity at the surface

Noting that $p(\nabla \cdot \mathbf{v}) = \nabla \cdot (p\mathbf{v}) - \mathbf{v} \cdot \nabla p$ and using the divergence theorem (Gauss-Ostrogradsky theorem) we can see that $W = W_{III}$ (9) coincides with W_I (1),

$$W = W_{III} = \frac{1}{\mathcal{S}} \int_{\mathcal{V}} p(\nabla \cdot \mathbf{v}) d\mathcal{V} = -\frac{1}{\mathcal{S}} \int_{\mathcal{V}} (\mathbf{v} \cdot \nabla p) d\mathcal{V} + I_T + I_s \equiv W_I + I_T + I_s, \tag{12}$$

if the following integrals are zero:

$$I_T \equiv \frac{p_T}{\mathcal{S}} \int_{z=z(p_T)} (\mathbf{v} \cdot \mathbf{n}) d\mathcal{S} = 0, \tag{13}$$

$$I_s \equiv \frac{1}{\mathcal{S}} \int_{\mathcal{S}} p_s (\mathbf{v} \cdot \mathbf{n}) d\mathcal{S} = 0. \tag{14}$$

Integral (13) is taken over the upper boundary $z = z(p_T)$, where $z(p_T)$ is the altitude of the pressure level $p = p_T$ defining the top of the atmosphere. Since the distribution of pressure versus altitude is exponential and I_T is proportional to p_T, by choosing a sufficiently small p_T it is possible to ensure that I_T (13) is arbitrarily small compared to W. For $p_T = 0.1$ hPa we estimate $I_T \sim 10^{-4} W$ (see Fig. 6d in Appendix C). So it is safe to assume that $I_T = 0$.

Integral (14) is taken over the Earth's surface (p_s is surface pressure). In a dry atmosphere we have

$$\mathbf{v} \cdot \mathbf{n}|_{z=0} = w_s = 0. \tag{15}$$

Here w_s is the surface value of the vertical velocity of air $\mathbf{w}$. As we discuss below, for a moist atmosphere with surface evaporation Eq. (15) also holds, such that $W = W_{III} = W_I$, see Eqs. (9), (3), (1), (12).

Evaporation from the Earth's surface represents a flux of water vapor molecules. As far as the relative humidity at the surface is always less than or equal to unity, the local value of this flux at $z = 0$ is never negative. The evaporating water molecules may have a mean vertical velocity $w_E > 0$ of the order of sound velocity only until they collide with the other air molecules at a distance of about one free path length $l_f \sim 10^{-7}$ m from the surface. Molecular collisions ensure that the mean velocity of evaporated molecules rapidly approaches the parcel's mean due to molecular collisions. Molar density N_E of evaporating molecules is obtained from evaporation rate E (kg m^{-2} s^{-1}) as $N_E w_E = E/M_v$, where M_v is molar mass of water. If l_p is the linear size of this parcel, we have for the mean vertical velocity of all molecules in the parcel $w_s = w_E N_E l_f/(N_s l_p)$, where $N_s = p_s/(RT_s)$ is molar density of air at the surface, T_s is surface temperature. Then in Eq. (14) we have $p_s w_s = (l_f/l_p)ERT_s/M_v$.

Since, as we will see in Sections 5 and 6, global atmospheric power is of the order of PRT_s/M_v, where $P = E$ is global mean precipitation and evaporation, the surface term $p_s w_s = (l_f/l_p)ERT_s/M_v$ can be neglected in Eq. (12) if $l_f/l_p \ll 1$, i.e. on any macroscopic scale. This reflects the fact that the atmosphere *does not* circulate because of being "pushed upwards" by surface evaporation. Notably, Eq. (15) is not in contradiction with the existence of an inflow of water vapor into the atmosphere at $z = 0$. It just means that water vapor should be considered as arising by evaporation within the surface air parcels, the latter having zero vertical velocity at their lower boundary.

We emphasize that the boundary condition (15) is vital for the equality between $W = W_{III}$ (9), (3) derived from the thermodynamic definition of work and W_I (1) of Lorenz (1967). In the last decades W_I was used by various researchers to evaluate the atmospheric energy cycle. For example, Laliberté et al. (2015) used W_I as their thermodynamic definition of atmospheric work output thus assuming that Eq. (15) holds. If one ignored Eq. (15) and defined $w_s > 0$ for $z = 0$ from the upward flux of water vapor as $\rho_{vs} w_s = E$, where ρ_{vs} is water vapor density at the surface (see, e.g., Eq. 3 of Pauluis et al., 2000, to be discussed below), one would have obtained unphysical results. With $E \sim 10^3$ kg m^{-2} year^{-1} and $\rho_{vs} \sim 10^{-2}$ kg m^{-3} we would have $w_s \sim 3 \times 10^{-3}$ m s^{-1} and $I_s = p_s w_s = 3 \times 10^2$ W m^{-2}. Then from Eq. (12) we would conclude that total power of atmospheric circulation $W = W_I + I_s > I_s$ exceeds the incoming flux of solar radiation.

3.2 Kinetic power and the gravitational power of precipitation

We now show how total atmospheric power (12) is comprised of three distinct terms. In hydrostatic equilibrium we have

$$\nabla_z p = \rho \mathbf{g}. \tag{16}$$

In the real atmosphere due to the presence of non-gaseous water the distribution of air deviates from Eq. (16) such that we instead have

$$\nabla_z p = (\rho + \rho_c)\mathbf{g}. \tag{17}$$

Using (17) we have in (12) $-\mathbf{v} \cdot \nabla p = -\mathbf{u} \cdot \nabla p - \mathbf{w} \cdot \nabla p = -\mathbf{u} \cdot \nabla p - (\rho + \rho_c)\mathbf{w} \cdot \mathbf{g}$. Here $-\rho \mathbf{w} \cdot \mathbf{g}$ represents the vertical flux of air: it is positive (negative) for the ascending (descending) air. Recalling that

$$\mathbf{g} = -g\nabla z \tag{18}$$

and using the divergence theorem and the stationary continuity equation (6) we have

$$W_P \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} \rho(\mathbf{w} \cdot \mathbf{g}) d\mathcal{V} = \frac{g}{\mathcal{S}}\int_{\mathcal{V}} \rho(\mathbf{v} \cdot \nabla z) d\mathcal{V} = \frac{g}{\mathcal{S}}\int_{\mathcal{S}} \mathbf{n} \cdot (\mathbf{v}\rho z) d\mathcal{S} - \frac{1}{\mathcal{S}}\int_{\mathcal{V}} gz\dot{\rho} d\mathcal{V}. \tag{19}$$

The surface integral in (19) is taken at the Earth's surface (here it is zero because $z = 0$) and $z = z(p_T)$ (here it is also zero, because $\rho \mathbf{n} \cdot \mathbf{v} = 0$).

It is natural to call W_P the "gravitational power of precipitation". Indeed, in the last integral in (19) gz represents potential energy of a unit mass in the Earth's gravitational field. Therefore, when evaporation occurs at the surface $z = 0$ and the condensate falls from height z where it originated, it is clear from Eq. (19) that W_P (19) is equal to $Pg\mathcal{H}_P$, where $\mathcal{H}_P$ is the global mean height of condensation.

The stationary power budget for a hydrostatic atmosphere can be written as

$$W = -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} \mathbf{v}\cdot\nabla p d\mathcal{V} \equiv W_K + W_P, \tag{20}$$

$$W_K \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} (\mathbf{u}\cdot\nabla p) d\mathcal{V} + W_c \approx -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} \mathbf{u}\cdot\nabla p d\mathcal{V}, \quad W_c \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} \rho_c(\mathbf{w}\cdot\mathbf{g}) d\mathcal{V}, \tag{21}$$

$$W_P \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} \rho\mathbf{w}\cdot\mathbf{g} d\mathcal{V} = -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} gz\dot{\rho} d\mathcal{V} = Pg\mathcal{H}_P, \quad P \equiv -\frac{1}{\mathcal{S}}\int_{z>0} \dot{\rho} d\mathcal{V}. \tag{22}$$

Equations (20)-(22) and their derivation have not been previously published (see the next section). These equations clarify the physical meaning of the atmospheric power budget.

Term W_c in Eq. (21) describes the impact of condensate loading. It represents kinetic energy generation on the vertical scale of the order of the atmospheric scale height $\mathcal{H} \equiv -p/(\partial p/\partial z) = RT/(Mg) \sim 10$ km. This energy is generated because the vertical air distribution deviates from the hydrostatic equilibrium (16). Hydrometeors falling at terminal velocity exert a force on the air equal to their weight. The condensate thus acts as resistance preventing the pressure difference $\Delta p \sim \rho_c g\mathcal{H}$ from converting to the kinetic energy of a vertical wind. In the atmosphere on average $\rho_c/\rho \sim 10^{-5}$ (Makarieva et al., 2013). Without hydrometeors, the non-equilibrium pressure difference $\Delta p \sim 10^{-5}\rho g\mathcal{H} \sim 1$ Pa would produce maximum vertical velocity of about $w_m \sim 1$ m s^{-1} ($\rho w_m^2/2 = \Delta p$). This is two orders of magnitude larger than the characteristic vertical velocities $w \sim 10^{-2}$ m s^{-1} of large-scale air motions. (Hydrometeors thus inhibit vertical motion in a similar way to how turbulent friction at the surface inhibits horizontal air motion. For example, the observed meridional surface pressure differences of the order of $\Delta p_h \sim 10$ hPa in the tropics, if friction were absent, could have produced maximum horizontal air velocities of about $u_m \sim 40$ m s^{-1} ($\rho u_m^2/2 = \Delta p_h$).) Quantitatively, $-\rho_c \mathbf{w} \cdot \mathbf{g}$ is less than 1% of W and can be neglected: its volume integral taken per unit surface area is less than $\rho_c g \mathcal{H} w \sim 10^{-5} p w \sim 10^{-2}$ W m^{-2}, where $p = \rho g \mathcal{H} = 10^5$ Pa is air pressure at the surface.

In contrast, the gravitational power of precipitation W_P does not depend on air-condensate interactions. (For example, this term would be present in the atmospheric power budget even if the condensate disappeared immediately upon condensation or experienced free fall not interacting with the air at all.) This is because W_P (19) reflects the net work of water vapor as it travels from the level where evaporation occurs (where water vapor arises) to the level where condensation occurs (where water vapor disappears). When condensation occurs above where evaporation occurs, the water vapor expands as it moves upwards towards condensation, and the work is positive irrespective of what happens to the condensate. For a dry atmosphere where $\dot{\rho} = 0$, the last volume integral in Eq. (19) is zero and $W_P = 0$: indeed, in this case at any height z there is as much air going upwards as there is going downwards.

Next, Eqs. (20)-(22) clarify the relationship between the two formulations of atmospheric power W_I (1) and W_{II} (2): $W_{II} \approx W_K$ coincides with $W = W_I$ in the absence of phase transitions only, i.e. when $W_P = 0$. This resolves some confusion in the literature, whereby in some publications it is total atmospheric power $W = W_I$ that is referred to as *generation of kinetic energy* (e.g., Robertson et al., 2011, their Eq. 1), while in others the same term is applied to W_K, which is estimated from horizontal velocities (see, e.g., Boville and Bretherton, 2003; Huang and McElroy, 2015). At the same time, in such studies W_K is confused with the total atmospheric power W: i.e. in the total power budget the gravitational power of precipitation, W_P, is overlooked (e.g., Huang and McElroy, 2015, their Fig. 10). We also note that the gravitational power of precipitation W_P has not been explicitly identified in past studies assessing the Lorenz energy cycle (see, e.g., Kim and Kim, 2013, and references therein).

Finally, Eqs. (20)-(22) show that the gravitational power of precipitation can be estimated from air velocity and pressure gradient alone as $W_P = W - W_K$ without any knowledge of the atmospheric moisture content or precipitation rates. This allows global W_P to be estimated from re-analyses data, see Section 6. So far, the only global estimate of W_P was that of Makarieva et al. (2013) based on $W_P = Pg\mathcal{H}_P$. Pauluis et al. (2000) used $Pg\mathcal{H}_P$ to estimate precipitation-related dissipation in the tropics.

3.3 Our results compared to Pauluis et al. 2000

Our assessment of the atmospheric power budget started from the thermodynamic definition of work (7). Integrated over atmospheric volume Eq. (7) yielded total atmospheric power $W = W_{III}$ (9), (3). The boundary condition (15) turned W_{III} into the commonly used W_I (12). Then we used the continuity equation (6) and hydrostatic equilibrium (17) to separate the kinetic energy generation W_K from the gravitational power of precipitation W_P, $W = W_K + W_P$, in Eqs. (20)-(22).

Here we compare our results with those of Pauluis et al. (2000) who likewise identified two distinct terms in the atmospheric power budget. Pauluis et al. (2000) were presumably aware of the fact that total atmospheric power is equal to W_{III}, since Eq. (3) was listed by Pauluis and Held (2002). However, as we show below, several inconsistencies in their basic assumptions did not permit obtaining results equivalent to Eqs. (20)-(22).

Noting that condensate is falling at terminal velocity v_T experiencing resistance force $\rho_c g$, Pauluis et al. (2000) defined the precipitation-related frictional dissipation as follows (we added factor $1/\mathcal{S}$ to enable comparison with our results):

$$W_P^* \equiv \frac{1}{\mathcal{S}} \int_{\mathcal{V}} \rho_c g v_T d\mathcal{V}. \tag{23}$$

Here $v_T \equiv w - w_c$ is the difference between the vertical velocities air and condensate. Assuming that *at any level* $z = z_0$ in the atmosphere the upward flux of water wapor is balanced by the downward flux of condensate,

$$\int_{z=z_0} \rho_c w_c d\mathcal{S} + \int_{z=z_0} \rho_v w d\mathcal{S} = 0, \tag{24}$$

Pauluis et al. (2000, see their Eq. 3) obtained[2]

$$W_P^* = \frac{1}{\mathcal{S}} \int_{\mathcal{V}} (\rho_c + \rho_v) w g d\mathcal{V} \equiv W_P + W_c, \tag{25}$$

where W_P and W_c are defined in Eqs. (22) and (21). Thus, W_P^* lumps together two terms with distinct meaning, with W_c depending on the interaction between the condensate and air and W_P independent of it.

Rather than using Eq. (18) and the continuity equation for water vapor similar to what was done in Eq. (19), Pauluis et al. (2000) further *assumed* that W_P^* is *proportional to the precipitation rate P at the surface, which is given by the surface integral $(1/\mathcal{S}) \int_{z=0} \rho_c v_T d\mathcal{S}$.* This formulation of surface precipitation via v_T yielded

$$W_P^* = P g \mathcal{H}_P \tag{26}$$

for the case when no re-evaporation occurs in the atmosphere. In reality, however, $P = -(1/\mathcal{S}) \int_{z=0} \rho_c w_c d\mathcal{S}$, so Eq. (26) is not consistent with Eq. (25).

The two integrals coincide, $-\int_{z=0} \rho_c w_c d\mathcal{S} = \int_{z=0} \rho_c v_T d\mathcal{S}$, only if $w|_{z=0} \equiv w_s = 0$. But this is inconsistent with the key assumption (24), since for $w_s = 0$ and $w_{cs} \neq 0$ Eq. (24) does not hold for $z = 0$. Indeed, for $z = 0$ Eq. (24) contradicts the boundary condition (15) $w_s = 0$. In particular, for the case when local evaporation equals local precipitation, Eq. (24) gives $w_s = -\rho_{cs} w_{cs} / \rho_{vs} > 0$.

[2]Using the continuity equations for dry air $\nabla \cdot (\rho_d \mathbf{v}) = 0$ and water vapor $\nabla \cdot (\rho_v \mathbf{v}) = \dot{\rho}$, where ρ_d and ρ_v are densities of dry air and water vapor, we find from Eq. (19) that $W_P = -(1/\mathcal{S}) \int_{\mathcal{V}} \rho_v \mathbf{w} \cdot \mathbf{g} d\mathcal{V}$.

This inconsistency between Eq. (24) for $z = 0$ and the equality $W_I = W_{III}$, see Eq. (12), precludes a straightforward derivation of W (9) from W_{III} (3) of Pauluis and Held (2002). Instead, Pauluis et al. (2000) assumed that *total mechanical work by resolved eddies* W_{tot} is equal to the sum of the *frictional dissipation associated with convective and boundary-layer turbulence* W_D and the *total dissipation rate due to precipitation* W_P^*:

$$W_{tot} = W_D + W_P^*. \tag{27}$$

Since no general specification for turbulent processes exists, this formulation *per se*, unlike Eqs. (20)-(22), cannot guide an assessment of W_{tot} from observations. However, W_D can be retrieved from the equation of motion as the volume integral of $-\mathbf{F}\cdot\mathbf{v}$, where $\mathbf{F}$ is turbulent friction force (cf. Lorenz, 1967, Eq. 101).

For the moist air (gas) moving under the action of a pressure gradient force, gravity, condensate loading and turbulent friction force, the scalar product of the equation of motion with velocity $\mathbf{v}$ reads (see, e.g., Huang and McElroy, 2015, for more details):

$$\rho\frac{dK}{dt} = -\mathbf{v}\cdot\nabla p + \rho\mathbf{w}\cdot\mathbf{g} + \rho\mathbf{w}\cdot\mathbf{g} + \mathbf{F}\cdot\mathbf{v}. \tag{28}$$

Here $K \equiv v^2/2$ is the kinetic energy of air per unit air mass. By virtue of relationship (17) the sum of the first three terms in the right-hand side of (28) is equal to $-\mathbf{u}\cdot\nabla p$. So, integrating (28) over volume, using the divergence theorem, continuity equation (6), boundary condition (19) and Eq. (21), we obtain

$$W_D \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}}(\mathbf{F}\cdot\mathbf{v})d\mathcal{V} = W_K - W_c - \overline{\dot{K}}, \quad \overline{\dot{K}} \equiv \frac{1}{\mathcal{S}}\int_{\mathcal{V}} K\dot{\rho}d\mathcal{V}. \tag{29}$$

Now, if we assume that $W_{tot} = W_I$ as in Eq. (20) and use W_D (29), the correct formulation for the atmospheric power budget in terms of Pauluis et al. (2000) would be

$$W_{tot} = W_D + W_P^* + \underline{\overline{\dot{K}}}, \quad W_P^* = Pg\mathcal{H}_P + \underline{W_c}. \tag{30}$$

The underlined terms differentiate the correct equations (30) from the formulations (26) and (27) of Pauluis et al. (2000). Term $\overline{\dot{K}}$ describes the sink of the kinetic energy of air caused by condensation. For example, if condensation occurs in the middle troposphere where air velocity is of the order of 25 m s^{-1}, the magnitude of this term estimates as $Pv^2/2 \sim 10^{-2}$ W m^{-2}. This is of the same order as the condensate loading term W_c, which was retained by Pauluis et al. (2000) in the definition of W_P^* (25).

If the condensate does not interact with the air but experiences free fall, then, despite $\rho_c \neq 0$, the condensate loading (i.e. terms proportional to ρ_c) is absent from the equation of motion as well as from Eqs. (17), (21) and (28). However, since W_c is present in the definition of W_P^* (23), in this case the correct equation for W_{tot} becomes $W_{tot} = W_D + W_P^* - \underline{W_c} + \underline{\overline{\dot{K}}}$, i.e. Eqs. (23) and (27) additionally overestimate the actual W_{tot} by the term equal to condensate loading. Thus, the formulations of Pauluis et al. (2000) are generally consistent with Eqs. (20)-(22) if only the condensate loading term and $\overline{\dot{K}}$ are neglected. Nevertheless, even in this case there remains a discrepancy between $W_{tot} = W_I$ of Pauluis et al. (2000) and $W_{tot} = W_{III}$ (3) of Pauluis and Held (2002), both representing the total atmospheric power, see Eq. (12). Since Eq. (24) of Pauluis et al. (2000) assumes $w_s > 0$ and $I_s > 0$, it follows from the divergence theorem that $W_{tot} = W_I + I_s \neq W_{III}$. In other words, Pauluis et al. (2000) could not demonstrate that their W_{tot} *defined* by Eq. (27) is equivalent to the thermodynamic definition of work W_{III} (3) (see discussion in the end of Section 3.1). The derivation of Eqs. (20)-(22) is free from this contradiction.

Reply to the Referee

Comment 1 of Referee 3 [doi:10.5194/acp-2016-203-RC3]:

1. Appropriation in the main result:

The manuscript states pretty explicitly that the main contribution here is

"Starting from the definition of mechanical work for an ideal gas, we present a novel derivation linking global wind power to measurable atmospheric parameters. The resulting expression distinguishes three components: the kinetic power associated with horizontal motion, the kinetic power associated with vertical motion and the gravitational power of precipitation."

as it is stated in the abstract. This claim is repeated on multiple occasions. I assume that this specifically refer to the equation (20-22), which the authors claim that "Equations Eqs. (20)-(22) and their derivation have not been previously published."

These equations are presented in Pauluis etal. (2000) (See equations (2), (4), (8) and (10). See also equations (4) and equation (6) of Pauluis and Held (JAS, 2002)). It is very troublesome that the authors fail to mention that equations (20-22) are presented in Pauluis etal. (2000) despite the fact that this pa

[the referee's comment continues below after our reply]

We revised the text having added a separate "Section 3.3 Our results compared to Pauluis et al. 2000". Right below Eqs. (20)-(22) we explain why in our view these equations are original. Furthermore, we also explicitly refer the readers to Section 3.3 where these results are compared with Pauluis et al. 2000 by noting: "Equations (20)-(22) and their derivation have not been previously published (see the next section)." Readers can judge our claims for themselves.

As a separate point, we note that Eqs. (20)-(22) make it clear that W_P can be estimated from the data on air velocity and pressure gradient with no information required about moist processes. As can easily be verified by examining the texts in question, this message is absent from the works cited by the referee (or indeed in any previous publications of which we are aware). To facilitate this comparison we list the equations mentioned by the referee below together with our Eqs. 20-22 from the submitted manuscript.

Pauluis et al (2000), Eqs. (2), (4), (8) and (10), respectively:

$$W_p = \int_\Omega g\rho_c v_\mathsf{T}, \tag{c1}$$

$$W_p = \int_\Omega g\rho_t w, \tag{c2}$$

$$W_D = \int \bar{\rho} g w \left[\frac{\Theta'}{\overline{\Theta}} + \left(\frac{R_v}{R_d} - 1 \right) \frac{\rho_v}{\bar{\rho}} - \frac{\rho_c}{\bar{\rho}} \right], \tag{c3}$$

$$W_\mathsf{tot} = \int w g \left[\bar{\rho} \frac{\Theta'}{\overline{\Theta}} + \rho_v \frac{R_v}{R_d} \right], \tag{c4}$$

where $\rho_t = \rho_c + \rho_v$.

Pauluis and Held (2002), Eqs. (4) and (6), respectively:

$$W = \int_\Omega p\partial_i V_i, \tag{c5}$$

$$D_p = \int_\Omega g\rho_c V_T = \int_\Omega \rho q_t g w, \tag{c6}$$

where V_i is the ith component of the velocity, $\partial_i = \partial/\partial x_i$ is the partial derivative in the i direction, ρ_c is the mass of falling hydrometeors per unit volume, q_t is mass of total water per unit mass of moist air, V_T is the terminal velocity of the falling hydrometeors, and w is the vertical velocity of the air.

Equations (20)-(22):

$$W = -\frac{1}{\mathcal{S}} \int_\mathcal{V} \mathbf{v} \cdot \nabla p d\mathcal{V} \equiv W_K + W_P, \tag{c7}$$

$$W_K \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}}(\mathbf{u}\cdot\nabla p)d\mathcal{V} + W_c \approx -\frac{1}{\mathcal{S}}\int_{\mathcal{V}}\mathbf{u}\cdot\nabla p d\mathcal{V}, \quad W_c \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}}\rho_c(\mathbf{w}\cdot\mathbf{g})d\mathcal{V}, \quad \text{(c8)}$$

$$W_P \equiv -\frac{1}{\mathcal{S}}\int_{\mathcal{V}}\rho\mathbf{w}\cdot\mathbf{g}d\mathcal{V} = -\frac{1}{\mathcal{S}}\int_{\mathcal{V}} gz\dot{\rho}d\mathcal{V} = Pg\mathcal{H}_P, \quad P \equiv -\frac{1}{\mathcal{S}}\int_{z>0}\dot{\rho}d\mathcal{V}. \quad \text{(c9)}$$

Note that $\rho = \rho_d + \rho_v \neq \rho q_t$; $\mathbf{v} = \mathbf{u} + \mathbf{w}$ is air velocity (horizontal and vertical).

The referee continues:

The appropriation is not limited to the equations, but extends to some of the arguments presented. For instance, the authors relate the claim

"The meaning is that hydrometeors perform work at the expense of their potential energy. To acquire this energy, a corresponding amount of water vapor must be raised by air parcels. We can also see that WP does not depend on the interaction between the air and the falling hydrometeors. This term would be present in the atmospheric power budget even if hydrometeors were experiencing free fall and did not interact with the air at all (such that no frictional dissipation on hydrometeors occurred)."

This points is made previously (and more clearly) in Pauluis etal. JAS (2000, p. 991):

"The dissipation by precipitation can be thought as proceeding in two steps. First, water is lifted by the atmospheric circulation, increasing its potential energy. Then, during precipitation, the potential energy of condensed water is transferred to the ambient air where it is dissipated by molecular viscosity in the microscopic shear zone around the hydrometeors."

To put it bluntly, the authors are presenting as their own an analysis that was done by others, and in doing so, are misleading their reader.

We show in the revised text (see the last paragraph in Section 3.3) that the above statement of Pauluis et al. (2000, p. 991) is not consistent with their own analysis, because their definition of precipitation-related dissipation hinges on the interaction between the condensate and the air. This definition, besides the gravitational power of precipitation to which the above comment correctly refers, also includes the condensate loading term W_c which has a different meaning. As demonstrated by our Eqs. 20-22 and the text below them, this term (but not W_P) does depend on the interaction between the condensate and the air[3].

References

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[3]For the record, the first estimate of the gravitational power of precipitation of which we are aware is Gorshkov (1982, p. 6). Considering possible sources of renewable energy to be harnessed on land, Gorshkov (1982) estimated the gravitational power of terrestrial precipitation by analogy with hydropower, for which the formulation $Pg\mathcal{H}_P$ is commonly used. He used global mean precipitation on land of 0.5 m year^{-1} as given by L'vovitch (1979) and assumed that it rains on average from somewhere in the middle of the troposphere to obtain 10^{14} W (for an English citation, see Gorshkov (1995, p. 30)).

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Interactive comment on Atmos. Chem. Phys. Discuss., doi:10.5194/acp-2016-203, 2016.