Reassessing thermodynamic and dynamic constraints on global wind power

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Abstract

Starting from basic physical principles, we present a novel derivation linking the global wind power to measurable atmospheric parameters. The resulting expression distinguishes three components of the atmospheric power (the kinetic power associated with horizontal and vertical motion and the gravitational power of precipitation) and highlights problems with previous approaches. Focusing on Laliberté \textit{et al.} (2015), we show how inappropriate treatment of material derivatives in the presence of phase transitions leads to significant errors in wind power analyses. We discuss the physical constraints on global wind power and the opportunities provided by considering the dynamic effects of water vapor condensation.

1 Introduction

Atmospheric scientists have long sought fundamental principles which can explain global circulation power. This search has gained renewed significance in view of the need to reconcile an apparent mismatch between model predictions and observed trends (e.g., Kociuba and Power, 2015). While models suggest circulation should slow as global temperatures increase, independent observations indicate that in fact global wind power is increasing (e.g., de Boisséson \textit{et al.}, 2014). Without a coherent theoretical framework such results cannot be readily reconciled. Meanwhile no expression for global atmospheric power in a moist

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atmosphere has been derived from fundamental physical principles. Here we
derive such an expression. This permits us to examine the determinants of the
atmospheric power budget and to identify and address some widespread misun-
derstandings. In particular we argue that equilibrium thermodynamics offer little
insight into the constraints on global circulation power.

2 What is atmospheric power?

2.1 Total power

Work in the atmosphere is done by expanding air. Atmospheric water in its solid
and liquid form is incompressible and neither performs work nor occupies any
appreciable volume (Pelkowski and Frisius, 2011). Work per unit time (power) of
an air parcel containing \( \tilde{N} \) moles and occupying volume \( \tilde{V} \) (m\(^3\)) is

\[
p_{\frac{d\tilde{V}}{dt}} = p\tilde{N} \frac{dV}{dt} + pV \frac{d\tilde{N}}{dt} = \tilde{N} \left(-V \frac{dp}{dt} + R \frac{dT}{dt}\right) + RT \frac{d\tilde{N}}{dt}.
\]

Here we have used the ideal gas law, \( pV = RT \) and \( pdV = RdT - Vdp \), where
\( T \) is temperature, \( V \equiv N^{-1} \) is the atmospheric volume occupied by one mole of
air, \( N \) is air molar density (mol m\(^{-3}\)), \( \tilde{V} \equiv \tilde{N}V \), \( p \) is air pressure and \( R = 8.3 \)
J mol\(^{-1}\) K\(^{-1}\) is the universal gas constant.

In Eq. (1) the expression in braces represents part of the parcel’s power per
mole that is independent of the rate of phase transitions \( d\tilde{N}/dt \). To express it per
unit volume we use the relationship \( \tilde{N} \equiv V/N \). Then total power \( W_{tot} \) (W) in
an atmosphere containing \( n \) air parcels with total fixed volume \( V = \sum_{i=1}^{n} \tilde{V}_i = \int_V d\mathcal{V} \) (subscript \( i \) refers to the \( i \)-th air parcel) is

\[
W_{tot} \equiv \sum_{i=1}^{n} p_i \frac{d\tilde{V}_i}{dt} = \int_V \left(-\frac{dp}{dt} + NR \frac{dT}{dt} + RT \frac{d\tilde{N}}{dt}\right) d\mathcal{V}.
\]

Here \( \tilde{N} \) is the molar rate of phase transitions per unit volume (mol m\(^{-3}\) s\(^{-1}\)). Its
integral over volume \( \mathcal{V} \) is equal to the total rate of phase transitions in all the \( n \) air
parcels: \( \int_V \tilde{N}d\mathcal{V} = \sum_{i=1}^{n} d\tilde{N}_i/dt \).

We will now use the definition of material derivative for \( X = \{p, T, N\} \):

\[
\frac{dX}{dt} \equiv \frac{\partial X}{\partial t} + \mathbf{v} \cdot \nabla X,
\]

the continuity equation:

\[
\nabla \cdot (N\mathbf{v}) = \dot{\tilde{N}} - \frac{\partial N}{\partial t},
\]

2
and the divergence theorem:

\[
\int_V N \mathbf{v} \cdot \nabla T dV = \int_S NT (\mathbf{v} \cdot \mathbf{n}) dS - \int_V T \nabla \cdot (N \mathbf{v}) dV = - \int_V T \nabla \cdot (N \mathbf{v}) dV. \tag{5}
\]

Here \( v \) is air velocity, \( n \) is unit vector perpendicular to the unit surface of area \( dS \); the integral is taken over the surface enclosing the atmospheric volume. Since the air does not leave the atmosphere, we have \( \mathbf{v} \cdot \mathbf{n} = 0 \), and the integral over \( S \) in (5) is zero. With help of Eqs. (3)-(5) and the ideal gas law \( \partial p/\partial t = R \partial (NT)/\partial t \) we obtain from Eq. (2)

\[
W_{tot} = - \int_V (\mathbf{v} \cdot \nabla p) dV = - \int_V (\mathbf{u} \cdot \nabla p + \mathbf{w} \cdot \nabla p) dV, \tag{6}
\]

where \( \mathbf{u} \) and \( \mathbf{w} \) are the horizontal and vertical air velocities, \( \mathbf{v} = \mathbf{u} + \mathbf{w} \).

### 2.2 Kinetic power

In Eq. (6) the expression \(- (\mathbf{v} \cdot \nabla p)\) represents work performed per unit time per unit air volume by the pressure gradient. The horizontal pressure gradient generates the kinetic energy of the horizontal wind. The vertical pressure gradient generates the kinetic energy of the vertical wind plus it changes the potential energy of air in the gravitational field.

In hydrostatic equilibrium of air we have

\[
\nabla_z p = \rho g, \tag{7}
\]

where \( \rho = NM \) is air density (kg m\(^{-3}\)), \( M \) is air molar mass (kg mol\(^{-1}\)). In the real atmosphere due to the presence of non-gaseous water the air distribution deviates from Eq. (7) such that we have \( \nabla_z p = (\rho + \rho_l) g \) and in Eq. (6) \(- \mathbf{w} \cdot \nabla p = -\rho \mathbf{w} \cdot g - \rho_l \mathbf{w} \cdot g\), where \( \rho_l \) is mass density of the non-gaseous water in the air.

Term \(- \rho \mathbf{w} \cdot g\) represents the vertical flux of air: it is positive (negative) for the ascending (descending) air. Recalling that \( g = -g \nabla z \) and using the divergence theorem and the stationary continuity equation, see (4), we can write

\[
W_P \equiv - \int_V \rho \mathbf{w} \cdot g dV = \int_S \mathbf{n} \cdot (\rho \mathbf{w} g z) dS - \int_V g z \nabla (\rho \mathbf{w}) dV = - \int_V g z \dot{\rho} dV. \tag{8}
\]

For a dry atmosphere where \( \dot{\rho} = 0 \), this last integral in Eq. (8) 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. In a moist atmosphere, evaporation \( \dot{\rho} > 0 \) makes a negative contribution to \( W_P \), while condensation \( \dot{\rho} < 0 \) makes a positive contribution. This is because \( W_P \) reflects the 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. If condensation occurs below where evaporation occurs, the water vapor must compress to reach the evaporation site; thus, the work is negative. When evaporation occurs at the Earth’s surface \( z = 0 \), \( W_P \) is equal to \( P gH_P S_E \), where \( H_P \) is the mean height of condensation, \( P \) is global mean precipitation at the surface (kg m\(^{-2}\) s\(^{-1}\)) and \( S_E \) is Earth’s surface area (Makarieva et al., 2013a). It is natural to call \( W_P \) the ”gravitational power of precipitation”.

Thus for the stationary total power we have

\[
W_{tot} = W_K + W_P, \tag{9}
\]

\[
W_K = - \int_V (\mathbf{u} \cdot \nabla p + \rho_l \mathbf{w} \cdot \mathbf{g}) dV \approx - \int_V \mathbf{u} \cdot \nabla p dV, \tag{10}
\]

\[
W_P = - \int_V gz \dot{\rho} dV \approx PgH_P S_E, \quad P \equiv - \int_{z>0} \dot{\rho} dV / S_E. \tag{11}
\]

Per unit area, global \( W_K \) and \( W_P \) were estimated at 2.5 and 0.8 W m\(^{-2}\), respectively (Huang and McElroy, 2015; Makarieva et al., 2013a).

The term \(-\rho_l \mathbf{w} \cdot \mathbf{g}\) in Eq. (10) is not related to the gravitational power of precipitation. It describes kinetic energy generation on the vertical scale of the order of the atmospheric scale height \( H = -p/(\partial p/\partial z) = RT/(Mg) \approx 10 \text{ km} \). This energy is generated because the vertical air distribution deviates from the hydrostatic equilibrium (7). Hydrometeors act as resistance not allowing the non-equilibrium pressure difference \( \Delta p \sim \rho_l g H \) to be converted to the kinetic energy of the vertical wind. In the atmosphere on average \( \rho_l / \rho \sim 10^{-5} \) (Makarieva et al., 2013a). Without hydrometeors, a pressure difference \( \Delta p \sim 10^{-5} \rho g H \sim 1 \text{ hPa} \) would produce a vertical velocity of about \( w \sim 1 \text{ m s}^{-1} \). This is two orders of magnitude larger than the characteristic vertical velocities \( w \sim 10^{-2} \text{ m s}^{-1} \) of large-scale air motions. Hydrometeors thus have an effect similar to turbulent friction at the surface which does not allow horizontal velocities to develop. For example, the observed meridional surface pressure differences of the order of \( \Delta p \sim 10 \text{ hPa} \) in the tropics, if friction were absent, would have produced horizontal air velocities of about 40 m s\(^{-1}\) instead of the observed 7 m s\(^{-1}\). The term \(-\rho_l \mathbf{w} \cdot \mathbf{g}\) is less than 1% of \( W_{tot} \) and can be neglected: its volume integral taken per unit surface area is less than \( \rho_l g H w \sim 10^{-5} \rho w \sim 10^{-2} \text{ W m}^{-2} \), where \( p = \rho g H = 10^5 \text{ Pa} \) is air pressure at the surface.
3 Revisiting the current understanding of the atmospheric power budget

3.1 The physical meaning of $W_{\text{tot}}$, $W_K$ and $W_P$

To our knowledge, Gorshkov (1982) was the first to estimate $W_P$ for land assuming $H_P = 2$ km. In the meteorological literature, Pauluis et al. (2000) defined precipitation-related frictional dissipation as $W_P + \int_V \rho_l \mathbf{w} \cdot \mathbf{g} dV$ and estimated its value for the tropics. This estimate was later revised by Makarieva et al. (2013a). Pauluis et al. (2000) stated that in the stationary case total power of atmospheric motions must be equal to the sum of the dissipation rate at the microscale around the hydrometeors and the rate of dissipation of motions on convective scale. The latter was approximated it as the power of the buoyancy force that is proportional to vertical velocity (Pauluis et al., 2000, their Eq. 8). Derivation of Eqs. (9)-(11) have not been previously published.

These equations provide insight into the controls on circulation power. First, we can see that $W_P$ does not depend on the interaction between the air and the falling drops. This term would be present in the atmospheric power budget even if drops were experiencing free fall and did not interact with the air at all (such that no frictional dissipation on drops occurred in the atmosphere). Second, the small term $-\rho_l \mathbf{w} \cdot \mathbf{g}$ describes generation of kinetic energy at the convective scale $H$ (while this energy dissipates on hydrometeors) and thus should be included into the total rate of kinetic energy generation $W_K$. Finally, in the absence of a derivation relating Eq. (11) and Eq. (9), it is not clear a priori whether precipitation makes an independent contribution to the total power budget. Indeed it could appear as a form of turbulent dissipation of the kinetic energy generated at larger scales (i.e. water is lifted by turbulent diffusion), and thus be comprised in the expression for the total kinetic power.

We disagree with the recent commentary of Pauluis (2015) on the work of Laliberté et al. (2015). To provide some context, an ideal atmospheric Carnot cycle consuming heat flux $F = 100$ W m$^{-2}$ ($F$ is limited from above by solar power flux reaching the planetary surface) at surface temperature $T_{\text{in}} = 300$ K and releasing heat at $T_{\text{out}} = T_{\text{in}} - \Delta T$ with $\Delta T = 30$ K being roughly the magnitude of the Earth’s greenhouse effect, would generate kinetic energy at a rate of $W_C = F(\Delta T/T_{\text{in}}) = 10$ W m$^{-2}$. Laliberté et al. (2015) estimated total atmospheric power $W_{\text{tot}}$ per unit area to be around 4 W m$^{-2}$. Comparing this result with $W_C$, Pauluis (2015) notes that ”estimates for the rate of kinetic energy production by atmospheric motions are about half this figure”. Confusion has apparently arisen between total atmospheric power $W_{\text{tot}}$ and kinetic power $W_K$. Indeed, Pauluis (2015) continues that ”the difference is very likely due to Earth’s
hydrological cycle, which reduces the production of kinetic energy in two ways”, one of which is the gravitational power of precipitation \( W_P \). However, as is clear from Eq. (9), \( W_P \) is equal to the difference between \( W_{tot} \) and \( W_K \) and is unrelated to the difference \( W_C - W_{tot} \).

Since the relationship between \( W_C, W_P \) and \( W_K \) was discussed by Pauluis et al. (2000) (see their Eq. (7)), the confusion may stem from a misinterpretation of the magnitude estimated by Laliberté et al. (2015) as atmospheric power,

\[
\int_M (1/\rho)(dp/dt)dM = \int_V (dp/dt)dV,
\]

for \( W_K \) – while it was in fact \( W_{tot} \). Notably, neither Pauluis et al. (2000) nor Laliberté et al. (2015) identified the fact that \( W_K \) depends on horizontal and not vertical velocities. This is essential for comparing theory and observations. Horizontal pressure gradients and wind velocities are observed, while the vertical velocities are inferred with significant uncertainty.

There are other studies where kinetic energy generation is estimated from horizontal velocities as \( W_K \) (see, e.g., Boville and Bretherton, 2003; Huang and McElroy, 2015). At the same time, \( W_K \) is sometimes confused for the total atmospheric power: 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).

### 3.2 The analysis of Laliberté et al. (2015)

Equations (11)-(11) also allow us to identify errors in analyses of \( W_{tot} \). To estimate the atmospheric power budget, Laliberté et al. (2015) adopted the thermodynamic identity

\[
\frac{T}{\rho} \frac{ds}{dt} = \frac{dh}{dt} - \frac{1}{\rho} \frac{dp}{dt} \mu \frac{dq_T}{dt},
\]

where \( s \) is entropy, \( h \) is enthalpy, \( \mu \) is chemical potential (all per unit mass of wet air), \( 1/\rho \) is specific air volume and \( q_T \) is water mixing ratio. When integrating Eq. (12) over atmospheric mass, Laliberté et al. (2015) noted that the enthalpy term vanishes,

\[
\int_M (dh/dt)dM = 0,
\]

because the atmosphere is in a steady state. However, Eqs. (3) and (4) indicate that for any scalar quantity, in our case enthalpy \( h \), in the stationary case, \( \partial h/\partial t = 0 \), we have

\[
\int_M \frac{dh}{dt}dM = \int_M h \frac{d\rho}{dt} \rho dV = - \int_M h \frac{d\rho}{dt} dV.
\]

Here \( \dot{\rho} \) is the mass rate of phase transitions per unit volume (kg m\(^{-3}\) s\(^{-1}\)), \( \nabla \cdot (\rho \nu) = \dot{\rho} \). This integral is not zero in an atmosphere where phase transitions take place.

\(^1\)The unconventional sign at the chemical potential term follows from \( \mu \) being defined in Eq. (12) relative to dry air: hence, when the dry air content diminishes (while \( dq_T/dt > 0 \)) this term is negative. For details see p. 8 in the Supplementary Material of Laliberté et al. (2015).
This omission creates a significant error. To obtain an approximate estimate we can assume that all evaporation occurs at the surface, while all condensation occurs at a global mean height of $H_P = 2.5$ km, where about half of surface water vapor has condensed in the ascending air (see Makarieva et al., 2013a). Then we have $\int_{M}(dh/dt)dM/S_E \approx -P[h(0) - h(H_P)]$, see Eq. (11), where $h(0)$ and $h(H_P)$ are mean enthalpies at the surface ($z = 0$) and at mean condensation height ($z = H_P$). With $h \approx c_p T + L q$, where $q$ is water vapor mass mixing ratio, $c_p = 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$ is heat capacity of air at constant pressure, $L = 2.5 \times 10^6 \text{ J kg}^{-1}$ is latent heat of vaporization, $T(0) - T(H_P) \approx 15 \text{ K}$, $q(0) - q(H_P) = 0.5q(0) \approx 0.5 \times 10^{-2}$, $P = 10^3 \text{ kg m}^{-2} \text{ yr}^{-1}$, we find that $\int_{M}(dh/dt)dM/S_E \sim -1 \text{ W m}^{-2}$. This figure is about one quarter of the total atmospheric power $W_{tot}/S_E \approx 4 \text{ W m}^{-2}$ estimated by Laliberté et al. (2015) for the MERRA re-analysis and the CESM model.

As Eq. (12) is an identity, whether the incorrect assumption has increased or decreased the value of atmospheric power depends on the particular method of calculating it. Laliberté et al. (2015) first calculated the mass integral of $Tds/dt$ from Eq. (12), then calculated $\mu dq_T/dt$ from atmospheric parameters and then used the obtained values to estimate the total power $-\int_{M}^{\rho}(dp/dt)dM$. In such a procedure, putting $\int_{M}(dh/dt)dM = 0$ resulted in an overestimate of the atmospheric power by about 1 W m$^{-2}$. Since the omitted term depends on precipitation rate, its omission is crucial not only for a correct estimate of the mean value of $W_{tot}$, but also for the determination of any trends related to precipitation, which is a focus of Laliberté et al. (2015). Thus the quantitative conclusions of Laliberté et al. (2015) appear invalid.

4 Dynamic versus thermodynamic approach

Under an assumption of local thermodynamic equilibrium Eq. (12) defines $ds/dt$ via measurable atmospheric variables. As such, it does not carry any additional information about the atmosphere besides that contained in the set of local values of pressure, temperature, relative humidity and air velocity. Total atmospheric power can be estimated from Eqs. (9)-(11) without involving entropy. This fact illustrates a limitation of the thermodynamic approach in that it cannot explain why a particular thermodynamic cycle exists: i.e. why does the atmosphere generate power.

Indeed, the kinetic power $W_K$ can be viewed as a measure of the dynamic disequilibrium of the Earth’s atmosphere. In equilibrium, for example, under conditions of hydrostatic and geostrophic or cyclostrophic balance, no power is generated and $W_K = 0$. Considerations of entropy allow one to constrain the maximum work that can be extracted from a given thermodynamic cycle. How-
ever, equilibrium thermodynamics cannot predict whether the considered system will be in equilibrium. It thus provides no insight as to if or why \( W_K \) on Earth differs from zero. The obvious upper limit – the global efficiency of solar energy conversion into useful work, which amounts to about 90% (Wu and Liu, 2010; see Pelkowski, 2012 for a rigorous theoretical discussion) – appears irrelevant for constraining \( W_K \): \( W_K \) is far below this thermodynamic limit, being about 1% of incoming solar radiation.

In contrast, the dynamic approach identifies situations when production of kinetic energy cannot be zero in principle. Saturated water vapor in the gravitational field above the liquid ocean is a unique dynamic system, because its pressure is controlled by temperature alone rather than by temperature and molar density as for the non-condensable ideal gases. Specifically, while dry air can rise in hydrostatic equilibrium, the saturated water vapor cannot. In an atmosphere composed of pure saturated water vapor the water vapor rises (returning to the Earth as precipitation) and kinetic energy is produced at a local rate 

\[
-w \left( \frac{\partial p_v}{\partial z} - \rho_v g \right) > 0,
\]

where \( \rho_v \) is mass density of water vapor (Makarieva et al., 2015b). If non-condensable gases are added to such an atmosphere, it becomes possible to arrange a hydrostatic equilibrium in the vertical plane, such that all the kinetic power that derives from condensation is now generated in the horizontal plane:

\[
-u \cdot \nabla p = -w \left( \frac{\partial p_v}{\partial z} - \rho_v/H \right) = -wRTN_{\partial \gamma/\partial z}.
\]

Since 

\[
-\int_{z>0} wN_{\partial \gamma/\partial z} dV \approx P/M_v,
\]

where \( M_v \) is molar mass of water vapor, the global kinetic power that can be derived from condensation is about 4 W m\(^{-2}\) (assuming mean temperature of condensation \( T \sim 270 \) K and \( P = 10^3 \) kg m\(^{-2}\) yr\(^{-1}\)) (Makarieva et al., 2013a). This theoretical estimate is about 40% higher than the most recent estimate of \( W_K/S_E = 2.5 \) W m\(^{-2}\) obtained using the MERRA re-analysis (Huang and McElroy, 2015).

The difference is likely related to an insufficient spatial and temporal resolution. Even when in real eddies there is no evaporation in the descending air (i.e. for \( w < 0 \) we have \( \partial \gamma/\partial z = 0 \) and \( N = 0 \)), since in the rising saturated air we always have \( \partial \gamma/\partial z < 0 \), the spatial and temporal averaging produces a non-zero vertical gradient \( \partial \gamma/\partial z < 0 \) everywhere. This results in "spurious" evaporation \( \hat{N} \approx w\partial \gamma/\partial z > 0 \) for \( z > 0 \). It diminishes the integral of 

\[
-\int_{z>0} \hat{N} dV
\]

and all the related quantities, including the kinetic power derived from condensation\(^2\). Our theoretical estimate \( W_K/S_E = 4 \) W m\(^{-2}\) suggests that with increasing spatial and temporal resolution the atmospheric power estimated from observations should grow. In agreement with these ideas, Kim and Kim (2013) using the daily averaged MERRA data obtained \( W_K/S_E = 2 \) W m\(^{-2}\), which is 20% less than the value obtained by Huang and McElroy (2015) using 3-hour observations (Huang

\(^2\)This spurious evaporation also diminishes the integral of \( dh/dt \) \((\ref{eq:dhdt})\), a problem not discussed by Laliberté et al. (2015).
J, personal communication).

In global circulation models a non-zero rate of kinetic energy generation is achieved by introducing an ad hoc intensity of turbulent diffusion which determines the rate at which kinetic energy is dissipated (and, in the steady state, generated) (see Makarieva et al., 2015a). It is this parameterization that postulates a certain value of $W_K$ and controls its behavior. Since this parameterization is unrelated to the hydrological cycle (i.e. one and the same turbulent diffusion coefficient can be used in both dry and moist models), comparing $W_K$ across models with varying intensity of the hydrological cycle does not shed light on the actual role of water vapor. Meanwhile the condensation-induced atmospheric dynamics outlined above (for further details see Makarieva et al., 2013b; 2015b and references therein) suggests that in the absence of the hydrological cycle the atmospheric power on Earth would have been negligible compared to what it is today. We thus urge attention to the dynamic effects of condensation.

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