As hydrometeors fall within or from a cloud, they reach a terminal velocity because of friction with the air through which they settle. This friction has previously been shown to result in significant vertically integrated dissipation of energy, but the nature and vertical profile of this dissipation warrant further investigation. Here, its energetic origin is discussed. It is confirmed explicitly that the dissipated energy originates from the conversion of hydrometeor potential energy during settling as suggested in an earlier study by Pauluis and Held. The magnitude of this heating is then analyzed in a cloud-resolving model simulation of tropical, aggregated convection. Maximum heating from hydrometeor friction reaches $\sim 10 \text{ K h}^{-1}$. The simulation is compared to one without hydrometeor frictional heating. For the case simulated, hydrometeor frictional heating results in a drier mean state, greater cloud cover, lessened convective mass flux, and a warmer atmosphere throughout much of the troposphere. It is suggested that the heating imparted to the atmosphere by dissipation allows the air to recover most of the energy previously expended in lofting hydrometeors.

1. Introduction

As air rises in a moist atmosphere, liquid or frozen hydrometeors often form. They are gravitationally accelerated and begin to settle toward Earth’s surface. Because these hydrometeors fall through a column of air, they are affected by friction, and their translation is slower than it would be in a vacuum. Friction is so high that many hydrometeors experience an almost total cessation of their sedimentation and form long-lived clouds while some manage to fall at a substantial speed and may eventually reach the ground.

Recently, there has been some interest in quantifying how much frictional heating precipitating hydrometeors impose on the column of air through which they fall (Pauluis et al. 2000; Pauluis and Held 2002a; Bannon 2012; Makarieva et al. 2013; Pauluis and Dias 2012). Previous work broadly suggests the mean transfer of power (from frictional heating) between falling hydrometeors to dry air is on the order of a few watts per square meter in the tropics. While this value pales in comparison to the heating provided to the dry atmosphere from the condensation and freezing of hydrometeors, which is $\sim 70 \text{ W m}^{-2}$ (L’Ecuyer et al. 2015), it nevertheless represents a conceptually significant source of mostly overlooked atmospheric heating from clouds. One of the few physical predictions made to date that considered hydrometeor frictional heating on cloudy systems was that of Sabuwala et al. (2015), who used a simple heat engine analogy to estimate that hydrometeor frictional heating could weaken tropical storm strength by as much as 20% because of increased entropy production within the storm. Such a large estimated impact suggests that more research regarding the impacts of frictional heating in cloudy systems is needed.

We will note that “frictional dissipation,” the transfer of useable energy to frictional heating (Thomson 1852), should not be confused with the commonly discussed effects of “hydrometeor loading,” the downward acceleration of air caused by hydrometeors dragging air as they fall or confused with the subsidence of air cooled by intense evaporation of hydrometeors. Dissipation, as we will discuss it here, is a frictional heating whereas loading is an acceleration.

Herein, we derive a simple energetic formulation of the frictional dissipation of energy within clouds. Our goal contrasts with previous efforts in two ways. Generally, frictional dissipation has been derived from dynamic equations in the past; and previous discussion (e.g., Pauluis and Held 2002a; Romps 2008) has mostly

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Corresponding author: Matthew R. Igel, migel@ucdavis.edu

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focused on the role of hydrometeor dissipation on larger-scale atmospheric energy and entropy budgets. Once the energetic nature is explored, we move on to discuss the impacts of this irreversible work on convective clouds by examining simulations of deep convection in a cloud-resolving model that includes hydrometeor energy dissipation.

2. Nature of dissipated energy

In principle, the dissipation of energy from friction between a rigid body and a fluid through which it moves is already well understood. It results from the frictional heating between fluid molecules and a moving object. The basic equations of motion of a single, spherical particle falling through a steady-state fluid were derived a century ago (Basset 1888; Boussinesq 1903; Oseen 1927). More recently, Kiger and Lasher (1997) and Maxey and Riley (1983) have extended these basic solutions to more complex flow fields. Yet, most atmospheric models entirely neglect the heating effects of this friction, although some exceptions exist (Pauluis et al. 2000; Pauluis and Held 2002b; Romps 2008; Bannon 2012; Makarieva et al. 2013; Pauluis and Dias 2012; Sabuwala et al. 2015). Therefore, we reintroduce it.

As the basis of a thought experiment, consider a newly formed hydrometeor at some height above the ground that is initially at rest relative to the ground (as a note, the reference frame we will use throughout this discussion is one that is stationary relative to Earth’s surface where the height is 0 m). This hydrometeor would accelerate under the force of gravity toward the surface until it reaches its size-and-shape-determined terminal velocity. At its terminal fall speed, acceleration ceases, and the hydrometeor falls at a constant velocity. Since the gravitational force is proportional to the mass of the hydrometeor, the frictional force exerted in this balanced state is exactly equal and opposite to the weight. That is,

\[ F_f = g \rho_H, \]  

where \( F_f \) is the frictional force (of hydrometeors on air), \( g \) is acceleration due to gravity, and \( \rho_H \) is the mass density of one or many hydrometeors in a unit volume of air. This equality is convenient since calculating the frictional force directly usually involves a complex computation that depends on the size and shape of the hydrometeor and poorly constrained empirical constants. Because power (or dissipation, the time rate of change of energy) is force times velocity, the vertically integrated (from the surface to height \( Z \) at the top of the atmosphere) frictional heating by the hydrometeor is

\[
D = \int_0^Z v_T F_f \, dz = \int_0^Z v_T g \rho_H \, dz, \tag{2}
\]

where \( D \) is the frictional dissipation and \( v_T \) is the terminal velocity of the hydrometeor. Equation (2) follows the convention of Pauluis et al. (2000) and Makarieva et al. (2013). Equation (2) is formulated from the point of view of the air and is exact if no other forces are acting. It should also be noted for completeness that (1) very slightly overestimates \( F_f \) because the displacement of air by the hydrometeor has been neglected. This displacement yields a very small upward buoyant force on the falling hydrometeor that is several orders of magnitude smaller (i.e., one over the density of water or ice) than \( F_f \) in (1). Also, real-world applications of (2) will consider a spectrum of drops rather than the single drop we used for narration.

Since it is formulated dynamically, (2) fails to provide much insight into the origin of the energy being dissipated. Equation (2) implies that the energy comes from friction or perhaps gravity, but this is a somewhat unsatisfying conclusion. It is not clear whether the energy going into dissipation originates from intrinsic or extrinsic energy of the hydrometeors themselves or from the air since \( D \) would be zero without either. Thus, we seek to calculate \( D \) in terms of energy (instead of through force by way of power) for the totally general case of moving and accelerating hydrometeors and air. Performing this calculation from an energy perspective is a novel technique in its own right and is worth pursuing simply on mathematical grounds. Energy solutions are of greater generality in principle. And, as we will show, using energy considerations will allow us to most precisely discuss \( D \).

a. Energy formulation

To calculate the dissipation \( D \) from energy principles, we need to consider all the components of energy that could change as a result of relative or absolute motion of air and hydrometeors. There are three such components to consider. Air and hydrometeor potential energies \( (P_A \) and \( P_H, \) respectively) and kinetic energy of the air \( K_A \) could change because of falling hydrometeors. Note that possible changes in \( K \) from hydrometeors are being ignored. [This is the usual assumption in the formulation of both model equations and the analytic momentum equations (Romps 2008). The assumption is not correct strictly, but it is required in order to assume additionally that hydrometeors are always falling at their terminal velocities.] The total internal energy \( I \) of the air–hydrometeor system could also change. We will assume that the mass of air and hydrometeors remains constant. We are concerned with the time rate of change of each of
these quantities, which is physically equivalent to the power discussed in Pauluis et al. (2000). Dissipation is a vertical integral of heating over a layer occurring at a constant rate. Therefore, its vertical derivative is the rate of internal heating. Upon assuming no sources of \( I \) other than \( P_A, P_H, \) and \( K_A \), the balance becomes

\[
\frac{d}{dz} D = \frac{d}{dt} I = -\frac{d}{dt}(P_A + P_H + K_A). \tag{3}
\]

Equation (3) states that \( D \) is due to the imbalance of the time rate of change of the sum of the relevant energies. Assuming conservation of mass, the basic form of the three terms on the rhs of (3), where \( v_A \) is the vertical velocity of air, are

\[
\frac{d}{dt} P_A = \frac{d}{dt} \rho_A g z_A = \rho_A g v_A, \tag{4}
\]

\[
\frac{d}{dt} P_H = \frac{d}{dt} \rho_H g z_H = \rho_H g (v_A - v_T), \tag{5}
\]

\[
\frac{d}{dt} K_A = \frac{d}{dt} \frac{1}{2} \rho_A v_A^2 = \rho_A v_A \frac{d}{dt} v_A. \tag{6}
\]

Under no external forces other than gravity, the acceleration of air laden with hydrometeors is

\[
\frac{d}{dt} v_A = -g \left( \frac{\rho_H}{\rho_A} + 1 \right). \tag{7}
\]

Note that \( \rho_H/\rho_A \) is the hydrometeor mixing ratio, which is the usual quantity used to express hydrometeor loading. From the perspective of the air, (7) is composed of the viscous drag between gravitationally settling hydrometeors and air ("loading") and gravitational attraction between the air and Earth. From an external perspective, the mixed volume of air and hydrometeors have a total acceleration dependent on the total mass contained in the volume. With (7), (6) becomes

\[
\frac{d}{dt} K_A = -\rho_A v_A g \left( \frac{\rho_H}{\rho_A} + 1 \right) = -v_A g (\rho_H + \rho_A). \tag{8}
\]

Upon collecting terms, (3) reduces to

\[
\frac{d}{dz} D = \frac{d}{dt} I = -\left[ \frac{A}{\rho_A v_A} + \frac{A^N}{\rho_H} + \frac{C}{A g v_T (\rho_H + \rho_A)} \right] \tag{9}
\]

This derivation of (9) recovers the relevant physical quantities of the integrand of (2), but this formulation suggests where the energy for the frictional dissipation originates. The change in \( K_A \) [see (8)] is composed of changes caused by gravitational attraction of the air and suspended hydrometeors with Earth. This term in (8) perfectly (and reversibly) offsets the changes in \( P \) caused by air and hydrometeor motion, term \( B \), within the gravitational field. That is, term \( A + A' = 0 \) and term \( B + B' = 0 \) in (9). The change in \( P_H \) caused by hydrometeors falling at their \( v_T \) [term \( C \) in (9)] is not offset by any change in \( K_A \). This result has been suggested and alluded to using the simpler static case in previous work (e.g., Pauluis et al. 2000), but it is shown here explicitly for the dynamic case of moving and accelerating air and hydrometeors.

So, \( D \) arises from a specific portion of hydrometeor potential energy (and possibly just a small fraction of the total change in \( P_H \) per time) being converted indirectly to heat. There are two primary sources of \( P_H \) in a convective column. 1) In the case of a newly formed hydrometeor, \( P_H \) is extracted directly from \( P_A \). Mass is transferred from moist air to the hydrometeor (i.e., condensation occurs) at a particular height such that total \( P, P_A + P_H \), is conserved during the transfer. 2) In the case of \( v_A > 0 \) with a hydrometeor formed at some lower level and lofted upward, \( P_H \) is imparted from \( K_A \). The value of \( K_A \) is lessened by hydrometeors whose \( P_H \) is increasing [see (8)]—this is the familiar hydrometeor loading effect. In either case, energy is extracted from the air and given to hydrometeors; it is then dissipated back to the air in the form of heat as hydrometeors fall (see also Pauluis et al. 2000).

But what portion of the total energy given to hydrometeors is subsequently given back to the air in a column? A net loss of energy from the hydrometeor–air mixture occurs when energy is laterally fluxed out of the volume of this mixture. The only possible loss of energy occurs when energy is fluxed between the hydrometeor–air mixture and the surface of Earth. In the case at hand, this only occurs when hydrometeors impact the surface. The net flux of energy is equal to the total \( K_P \) (hydrometeor kinetic energy) at the surface (impact). Because the maximum \( v_T \) is approximately 10 m s\(^{-1}\), total \( K_P \) at the surface is approximately equivalent to \( P_H \) of a hydrometeor with a height of only 5 m. Most hydrometeors originate orders of magnitude higher in the atmosphere than this so surface \( K_P \) and equivalent net atmospheric loss are insignificant compared to the total \( P_H \) dissipated to the air (i.e., surface \( K_H \) divided by initial \( P_H \) is very small). Thus, (conversion of \( P_H \) to \( D \)) represents a mechanism by which storms recapture/convert energy expended in lofting moisture or hydrometeors. Precipitating hydrometeors have effectively no net energetic effect on the column through which they rise and fall since they lose only a tiny fraction of their energy to the surface and return the rest to the air, in the form of \( I \). Hydrometeors merely vertically redistribute and...
convert energy from one type to another. For conceptual simplicity, our arguments exclude the work done by water vapor expansion, which is discussed elsewhere (Pauluis and Held 2002a; Romps 2008) as being a related process that transports energy downward but is dislocated in space and time from hydrometeor frictional dissipation at the scales of concern here.

We note explicitly that the sign of \( (d/dt)I \) is unambiguously positive in (9). Each quantity in term C is always greater than zero. So regardless of the dynamics or state of the atmosphere, frictional heating is always positive. Even when hydrometeors are being lifted (i.e., \( v_A > v_T \)) and their potential energy is increasing, hydrometeors continue to heat the atmosphere at the same rate as they would in quiescent air.

b. Frictional heating rate

Equation (9) can be used to calculate a heating rate by recognizing that \( dI = \rho_A c_p dT \)

\[
\frac{d}{dt} T = v_T \frac{g}{c_p} \rho_H
\]  

Equation (10) depends only on the terminal velocity of hydrometeors, the mean dry stratification \( g/c_p \), and the hydrometeor mixing ratio.

While the local frictional heating rate depends crucially on \( v_T \), the total amount of heating in a layer (after a hydrometeor has fallen all the way through the layer) does not. The total layer frictional heating only depends on the hydrometeor mixing ratio. Although this is obvious from our discussion leading to (9), we show this result explicitly at a macroscopic level by integrating (10) through the time \( \Delta A \) it takes for a hydrometeor to fall through a shallow (such that \( \rho_A \) is nearly constant) layer \( \Delta Z \):

\[
\Delta T = \int_0^{\Delta A} \frac{\Delta Z}{\Delta A} \frac{g}{c_p} \frac{\rho_H}{\rho_A} dt = \Delta Z \frac{g}{c_p} \frac{\rho_H}{\rho_A}, \tag{11}
\]

where \( v_T \) has been replaced by \( \Delta Z/\Delta A \). Equation (11) shows that the time-integrated change in \( T \) of a shallow layer of constant depth \( \Delta Z \) only depends on the depth of the layer and \( \rho_H \), not \( v_T \) of hydrometeors in that layer. That is, the change in temperature is determined by the change in \( P_H \) across a layer (i.e., its depth) (which is formulaic now that we showed explicitly that \( D \) originates from \( P_H \)).

3. Frictional heating magnitude

Next, we seek to understand the maximum magnitude of (10) to determine what local impacts the irreversible effects of falling hydrometeors might have on clouds and cloudy atmospheres. Previous studies have suggested that (2) should have a mean magnitude of a few watts per square meter over the tropics (Pauluis et al. 2000; Pauluis and Held 2002a; Pauluis and Dias 2012; Makarieva et al. 2013). While a few watts per square meter is larger than dry eddy diffusion by deep convection (Pauluis et al. 2000), it is only 1%-10% of other diabatic processes such as condensational or radiative heating (L’Ecuyer et al. 2015). But, all previous analyses of dissipation arising purely from hydrometeor frictional heating did not provide an indication of the vertical distribution of \( (d/dt)I \), which might be important to convective storms.

a. Simple scaling

First, we examine the maximum possible instantaneous, localized magnitude of (10). If we assume that, to leading order, \( \rho_H \approx 10^{-3} \text{kg m}^{-3}, 10^{-1} < \rho_A < 10^4 \text{kg m}^{-3}, v_T \approx 10 \text{m s}^{-1}, g \approx 10 \text{m s}^{-2}, \) and \( c_p \approx 10^3 \text{J kg}^{-1} \text{K}^{-1}, \) then the frictional heating rate could vary between \( \sim 10^{-3} \) and \( \sim 10^{-4} \text{K s}^{-1} \) (or \( \sim 4 \) and \( 0.4 \text{K h}^{-1} \)). A few kelvins per hour is smaller than the possible maximum heating caused by, again, latent, radiative heating, or even advective heating by hydrometeors, but it is not so small that it can be assumed to be inconsequential. Also, recall that frictional dissipation is the mechanism by which hydrometeors return energy to the atmosphere that was gained from hydrometeor loading and lofting. These latter processes are included in most models, so for the sake of model energy conservation, the frictional dissipation should also be included (as should all potential, kinetic, and internal energy terms).

b. Simulated frictional heating

To thoroughly examine the potential magnitude of (10), we added it to a cloud-resolving model. There are many possible impacts of hydrometeor frictional heating on different types of clouds. The impacts depend on the vertical distribution of precipitating hydrometeors and the evolution of that distribution throughout the lifetime of the cloud. We cannot, of course, provide an estimate of the importance of hydrometeor frictional heating in all possible meteorological environments. Instead, we conducted simulations that allow us to suggest that hydrometeor frictional heating can be consequential to the development and evolution of some cloudy systems.

1) MODEL AND SETUP

We use the Regional Atmospheric Modeling System (RAMS) (Cotton et al. 2003). The RAMS double-moment bulk microphysics scheme (Saleeby and van den Heever 2013) includes eight liquid or frozen hydrometeor species.
cloud, drizzle, rain, hail, graupel, pristine ice, snow, and aggregates. The size distribution of each species is assumed to conform to a three-parameter gamma probability distribution function (PDF). The mass-weighted terminal velocity (upon which the frictional heating rate depends) for each species is calculated using the Mitchell (1996) power laws. The mass-weighted terminal velocity will depend on the parameters of the gamma PDF. In a two-moment scheme, there is one free parameter, the shape parameter (Igel et al. 2015). The shape parameter for all precipitating hydrometeors was set to 2. Advection by hydrometeors of sensible heat is handled explicitly in RAMS for rain, graupel, and hail by including the internal energies of these hydrometeors as prognostic variables (Walko et al. 2000). The simulations discussed below were conducted on the National Center for Atmospheric Research’s “Yellowstone” supercomputer (CISL 2012).

We seek to diagnose the impacts of hydrometeor frictional heating in a tropical environment [in a state of radiative–convective equilibrium (RCE)]. For this purpose, RAMS is set up with a fixed ocean surface temperature of 301 K and diurnally varying incoming solar radiation but is otherwise unforced. For more information on the basic setup and other model schemes, see Igel et al. (2017). A grid size of 1 km is used over a domain of 200 km by 200 km. RAMS is run for 60 days with and without hydrometeor frictional heating included. Both simulations take about 30 days to reach an approximate equilibrium (e.g., Fig. 2) and then maintain a quasi–steady state thereafter. Except where obvious, results from the simulations reflect their final 20 days (480 h). Convection aggregates in both simulations. The two simulations are referred to as “HEATING” and “CONTROL.” Our discussion of model results progresses from the magnitude of frictional heating to its various impacts.

As a note, convection in RCE simulations conducted at high surface temperature and high resolution tends to aggregate. This will likely serve to locally concentrate hydrometeor frictional heating and decrease its total occurrence.

2) SIMULATION RESULTS

First, it is of interest to compare the simulation results with previous estimates of the dissipation rates from frictional heating. The maximum magnitude of hydrometeor frictional heating over the final 480 h of HEATING is 9.8 K h$^{-1}$. This is twice the estimated maximum frictional heating in section 3a because both the mixing ratio of hydrometeors and their fall velocity were slightly underestimated above. Mean heating rates are much lower and tend to be higher in strong convective updrafts (discussed below) where large mixing ratios of precipitating hydrometeors occur are more common. Finally, the mean $D$ is 2.7 W m$^{-2}$, which falls within the range of previous estimates of $\sim 2$–5 W m$^{-2}$ (Pauluis et al. 2000; Pauluis and Held 2002b; Romps 2008; Pauluis and Dias 2012).

Figure 1 shows a cross section of hydrometeor frictional heating rate (K h$^{-1}$) from an example deep convective storm in HEATING. Colored contours extend to the cloud boundary.
5.02 mm day\(^{-1}\) in CONTROL. One may estimate the water vapor residence time by dividing the mean precipitable water by the mean rain rate. In HEATING, the water vapor residence time is 8.83 days; in CONTROL, it is 9.28 days. This difference would be approximately equivalent to the change in residence time caused by a surface warming of 0.3 K (Igel et al. 2014). Together these results indicate a modestly, but consequentially, different hydrological cycle between the simulations. The average increase in atmospheric heating from hydrometeor friction (\(13\) W m\(^{-2}\)) and the increase in surface turbulent fluxes (\(13\) W m\(^{-2}\)) result primarily in an increase in outgoing longwave radiation (\(+5\) W m\(^{-2}\)) and a small increase in downward longwave radiation (+1 W m\(^{-2}\)).

The mean structure of clouds is also different (Fig. 3). The profile of mean cloud fraction, defined as the number of grid cells at any height with a total hydrometeor mixing ratio greater than 10\(^{-5}\) kg kg\(^{-1}\) normalized by the total number of grid cells at that height, shows relatively large cloud fractions at the top of the boundary layer and in the upper troposphere. Cloud fraction is 8% higher (38% higher when using a percentage difference formula) in HEATING than CONTROL. This difference is comparable to the change in cloud fraction because of a doubling of aerosol (Kaufman et al. 2005; van den Heever et al. 2011). The mean boundary layer cloud fraction in HEATING is more than one standard deviation higher than the mean in CONTROL. Boundary layer clouds also appear to be shifted toward lower heights in HEATING although they exhibit approximately the same distribution of thicknesses (not shown) and the same height in peak cloudiness (Fig. 3).

The upper-tropospheric cloud fraction peak is 1% higher (4% higher when using a percentage difference formula) in HEATING and occurs one model level higher than in CONTROL. Upper-level clouds also appear slightly thinner in HEATING.

In Fig. 4a, we show the average convective mass flux profiles for both CONTROL and HEATING using three different thresholds on the minimum vertical velocity: >1, >5, and >10 m s\(^{-1}\). For each threshold and for every height below 16 km, the mean convective mass flux is less in HEATING. In an average sense, this is the result of a more stable troposphere overall (see next paragraph), but in a local sense, the decrease in mean convective mass flux is likely the result of local stability generation below 6 km (positive vertical gradient in heating; Fig. 4b) by hydrometeor frictional heating. Despite instability generation above 10 km at all thresholds, the initially weaker convecting parcels in HEATING never attain the magnitude of mean mass flux exhibited in CONTROL.

Finally, we show the difference in the temperature profiles between HEATING and CONTROL (Fig. 5). The atmospheric temperature is warmer everywhere between the mean height of boundary layer cloud base and 16 km. The maximum difference reaches 1.3 K near 10 km. Interestingly, the subcloud layer is cooler in HEATING despite hydrometeor frictional heating being always positive (or zero). This appears to be associated with the fact that boundary layer clouds tend to occur at lower heights and be more plentiful. The causal relationship between a relative increase in inversion strength and a depression of boundary layer cloud tops is impossible to determine, but these properties are clearly related.
4. Summary and conclusions

The primary goal of this paper was to revisit the functional form of a missing energy source in many models—that from falling hydrometeors frictionally heating the air through which they fall. This new derivation was used to confirm that frictional heating from hydrometeor fallout is due to imperfect conversion of hydrometeor potential energy to kinetic energy.

The second goal was to determine if hydrometeor dissipation might be a significant contributor to atmospheric heating rates in some circumstances and whether it might affect the properties and behavior of clouds and precipitation. Through both simple scaling and model simulations of a state of RCE, the maximum hydrometeor frictional heating was shown to be on the order of a few kelvins per hour in tropical deep convective clouds. We showed mathematically that the time-integrated heating of a shallow layer from hydrometeor friction depends only on the hydrometeor mixing ratio, not on their terminal velocity. In a tropical RCE simulation, hydrometeor frictional heating resulted in a drier RCE state and slightly higher precipitation rates, which resulted in a shorter water vapor residence time. Shallow boundary layer clouds became more plentiful and shifted toward lower heights. Upper-tropospheric clouds thinned slightly and exhibited marginally higher coverage. Convective mass flux was lessened. The atmosphere was warmed by hydrometeor frictional heating at all levels in the troposphere (by up to 1.3 K) except in a thin layer near the surface.

We interpret these results as suggesting that heating caused by hydrometeor frictional heating can be consequential to the dynamics of clouds. While the frictional heating rates shown above are relatively weak compared to other diabatic contributions, they are high enough to suggest that this term may have significant and systematic impacts in some instances. We suggest that...
modelers should consider including it in their simulations. It is included by default [e.g., Das Atmosphärische Modell (DAM) in Romps (2008)] or by choice [e.g., Cloud Model version 1 (CM1) in Bryan and Fritsch (2002)] in very few models of which the authors are aware. Equation (10) could be easily added to most (if not all) cloud-resolving models that do not already include it with little additional computational cost for completeness, if for no other reason. Admittedly though, for the inclusion of (10) to be an asset, the parameterizations resulting in precipitation generation and fallout must be of sufficient quality. Finally, our derivation above suggested that hydrometeor frictional heating represents a recovery mechanism of the energy used to loft hydrometeors within convective clouds. Models commonly include the hydrometeor loading term, and therefore for consistency, we suggest that models should also include hydrometeor frictional heating, the mechanism by which this energy is recovered.

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