The Incredible Lightness of Water Vapor

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ABSTRACT

The molar mass of water vapor is much less than that of dry air. This makes a moist parcel lighter than a dry parcel of the same temperature and pressure. This effect is known as the vapor buoyancy effect and has often been overlooked in climate studies. We propose that the vapor buoyancy effect increases Earth's outgoing longwave radiation (OLR) and that this negative radiative effect increases with warming, stabilizing Earth's climate. We illustrate this mechanism in an idealized tropical atmosphere, where there is no horizontal buoyancy gradient in the free troposphere. Temperature increases toward dry atmosphere columns to compensate reduction of vapor buoyancy, increasing OLR by $O(1\ \mathrm{W/m^2})$ at the reference climate. In warmer climates, the temperature difference between moist and dry columns would increase due to increasing atmospheric water vapor, leading to enhanced radiative effect and thereby stabilizing Earth's climate. We estimate that this feedback strength is about $O(0.2\ \mathrm{W/m^2/K})$ in the idealized atmosphere, which compares with cloud feedback and surface albedo feedback in the current climate. We further show evidence from observations and real-gas radiative transfer calculations for a significant radiative effect of vapor buoyancy in the tropical atmosphere.

1. Introduction

How fast would Earth's climate respond to increasing CO_2 (Manabe and Wetherald 1975; Flato et al. 2013; Collins et al. 2013)? Why is tropical climate more stable than extratropical climate (Holland and Bitz 2003; Polyakov et al. 2002; Pierrehumbert 1995)? What sets the inner edge of the habitable zone of Earth-like planets (Yang and Abbot 2014; Pierrehumbert 2010)? Understanding and accurately estimating climate feedbacks are key to address these pressing questions.

The importance of water vapor seems to be widely recognized in the literature of climate feedbacks (Manabe and Wetherald 1967; Ingersoll 1969; Held and Soden 2000; Flato et al. 2013). Previous studies have focused on three basic effects of water vapor: E1) water vapor is a greenhouse gas; E2) water vapor can condense to liquid water and release latent heat; E3) saturation vapor pressure increases with temperature exponentially. The combination of E1 and E3 gives rise to the water vapor feedback, the dominant positive climate feedback (Manabe and Wetherald 1967; Held and Soden 2000; Flato et al. 2013). Increasing temperature leads to more water vapor, which leads to an enhanced greenhouse effect, warming the planet further. The water vapor feedback could even lead to a runaway greenhouse state when the atmosphere is sufficiently opaque to longwave radiation that the outgoing longwave radiation (OLR) is insensitive to surface temperature (Ingersoll 1969). The combination of E2 and E3 gives rise to the (tropical) lapse rate feedback, a negative climate feedback in the tropical atmosphere (Flato et al. 2013). Increasing temperature leads to more water vapor, which leads to less steep lapse rate in the tropical atmosphere. This effect increases upper troposphere temperature more than the lower troposphere, leading to higher emission of outgoing longwave radiation (OLR), which cools the planet. At higher latitudes, temperature lapse rate is no longer controlled by moist convection, so the lapse rate feedback is less constrained. Both feedbacks are among the five most important climate feedbacks in the Intergovernmental Panel on Climate Change (IPCC) reports and have been extensively evaluated in general circulation models (GCMs) (Flato et al. 2013).

However, the lightness of water vapor has been completely overlooked in the context of climate feedbacks. In Earth's atmosphere, buoyancy is represented by the virtual temperature [*e.g.*, see (5) in Yang (2018a)]. We define the virtual temperature

$$T_v = T\left(\frac{1+r/\varepsilon}{1+r}\right), \tag{1}$$

where T is temperature, r is water vapor mixing ratio, and $\varepsilon = M_v/M_d$. The molar mass of water vapor M_v is 18 g/mol, significantly lighter than that of dry air M_d, which is 29 g/mol. This makes a moist parcel lighter than a dry parcel of the same temperature and pressure (Emanuel 1994). Here we refer to this as the vapor buoyancy effect, though it is also referred to as the virtual effect (Yang 2018a,b). The vapor buoyancy effect makes T_v slightly higher than T by O(1 K) in the tropical atmosphere, which is of the same order of magnitude as typical temperature fluctuations there.

We propose that the vapor buoyancy effect can increase Earth's OLR and helps stabilize Earth's climate by regulating the atmosphere's thermal structure. Figure 1 shows temperature and virtual temperature (buoyancy) fields in the moisture space from 2^oS to 2^oN using NASA AIRS data. In the free troposphere ($p < 850$ hPa), buoyancy is horizontally uniform because of the small Coriolis parameter and efficient gravity waves (Charney 1963; Bretherton and Smolarkiewicz 1989; Sobel et al. 2001; Yang 2018a). However, temperature increases toward dry columns due to the vapor buoyancy effect. Moving toward the dry columns, moisture and its associated vapor buoyancy are reduced. To maintain uniform buoyancy, temperature has to increase. We propose that the temperature tilt would increase with climate warming due to increasing atmospheric moisture, leading to enhanced OLR over the dry area. This is a negative feedback and can stabilize Earth's climate.

Previous studies implied that vapor buoyancy could make temperature increase toward dry columns in the tropical atmosphere (Tompkins 2001; Bretherton and Smolarkiewicz 1989; Bretherton et al. 2005; Yang 2018b,a). However, they have often considered this effect to be small and negligible, simplifying the dynamics according to a weak temperature gradient approximation (Sobel et al. 2001). These studies, therefore, did not consider that its radiative effect is significant, which is the novelty of this study.

In Section 2, we explain our hypothesis in detail. We first illustrate how the vapor buoyancy effect increases Earth's OLR (a negative radiative effect) and then explain why this effect strengthens with climate warming. In Section 3, we derive a simple model for the radiative effect and feedback strength of the vapor buoyancy effect. We then use the simple model to make order-of-magnitude estimates for the radiative effect and feedback strength. In Section 4, we estimate the radiative effect by using tropical sounding profiles with a real-gas radiative transfer model. In Section 5, we conclude and discuss implications on the climate stability of Earth and other planets.

2. Hypothesis

We propose that the vapor buoyancy can increase OLR (a negative radiative effect) due to a clear-sky effect, and that the radiative effect increases with climate warming. Figure 2 illustrates our hypothesis by comparing OLR from two stand-alone atmospheres with overturning circulations: one considers the vapor buoyancy effect (control), the other does not consider this effect. The overturning circulation is analogous to the Walker Circulation or convective self-aggregation in the tropics (Bretherton et al. 2005; Pritchard and Yang 2016; Yang and Ingersoll 2013, 2014). The upwelling branch of the circulation is associated with deep convection and moist air, and the downwelling branch is associated with clear sky and dry air. For illustrative purposes, we make a few simplifications: S1) the two atmospheres are non-rotating; S2) the two atmospheres sit above ocean surface with the same, uniform surface temperature; S3) the two atmospheres have the same water vapor distribution. The first two simplifications are relevant to the tropical atmosphere as the rotation effect and surface temperature gradient are both weak in the tropics. The third simplification is often required when calculating the radiative effect.

Figure 2 shows that the control atmosphere emits more OLR than the no-vapor-buoyancy atmosphere due to higher temperature in the dry area. OLR is primarily a function of temperature and water vapor mixing ratio r. When r remains the same in the two atmospheres (S3), the OLR difference would come from temperature differences between the two atmospheres. Here we provide physical intuition on why there should be temperature differences, leaving detailed derivation in Section 3e. The temperature profiles of moist areas in the two atmospheres are set by convective plumes. Because these convective plumes rise from the same surface temperature, the temperature profiles should be almost identical in the two moist areas. Temperature profiles in the dry areas, however, differ significantly, leading to differences in OLR. According to long-accepted results in geophysical fluid dynamics, the horizontal buoyancy gradient is negligible in the free troposphere without rotation because gravity waves can effectively smooth out buoyancy anomalies (Charney 1963; Sobel et al. 2001). We refer to this effect as the weak buoyancy gradient (WBG) approximation (Yang 2018a). In the control atmosphere, buoyancy is a function of both temperature and and water vapor mixing ratio r due to the vapor buoyancy effect. The horizontal moisture gradient then leads to horizontal temperature gradient: dry air is warmer than moist air. In the no-vapor-buoyancy atmosphere, temperature is uniformly distributed in the free troposphere, as buoyancy is a function of temperature only. The dry column of the control atmosphere, therefore, is warmer than that of the no-vapor-buoyancy atmosphere by $O(1\text{ K})$, leading to enhanced OLR. The spectra of H_2O in the longwave is also sensitive to temperature. However, this impact is likely small.

In warmer climates, the vapor buoyancy effect would become more significant due to increasing water vapor. Therefore, we expect that the radiative effect due to the vapor buoyancy also increases with climate warming. This is a negative climate feedback (Fig. 2b). The proposed mechanism relies on ample atmospheric water vapor, so it would be most effective in stabilizing the tropical climate. In principal, this feedback should have been represented by climate models. However, it has not been evaluated nor even discussed.

We will construct a simple model of the proposed feedback mechanism. This will give an order-of-magnitude estimate of the associated radiative effect and the rate at which it increases with climate warming.

3. A simple model

We construct a simple model based on the schematic diagram (Fig. 2). Each atmosphere with overturning circulations is represented by a dry column and a moist column (Pierrehumbert 1995). Because the moist columns would have the same temperature profiles, the OLR difference primarily comes from the dry columns, which we will focus on. Again, we aim to estimate the "radiative effect" due to the vapor buoyancy effect. Therefore, we assume that all basic dynamic (*e.g.*, circulation and pressure) and thermodynamic features (*e.g.*, moisture) are the same in the two atmospheres–one with the vapor buoyancy effect, and the other without it.

The goal of this simple model is to provide an order-of-magnitude understanding of our hypothesis. Therefore, we employ a two-band radiative transfer model. The two-band model is more realistic than a gray atmosphere model by allowing two absorption bands with distinct absorption coefficients, leading to different emission levels. The two band model is, on the other hand, much simpler than a real-gas radiative transfer model, so the results are easier to interpret.

a. The two-band model

We consider a plane-parallel atmosphere. Only the clear-sky longwave (IR) radiation is considered, and the IR opacity is mainly due to water vapor. Here we parameterize the water vapor absorption spectrum by two broad bands that occupy roughly equal fractions of blackbody emission at Earth-like temperatures (Beucler and Cronin 2016): one with a strong absorption coefficient (κ_S) and the other with a weak absorption coefficient (κ_W).

We first consider one absorption band with any given κ. OLR is defined as

$$OLR^{\kappa} \equiv F^{\uparrow}(p=0) - F^{\downarrow}(p=0), \tag{2}$$

where $F^{\uparrow}$ and $F^{\downarrow}$ are upward and downward longwave radiative fluxes. We know that $F^{\downarrow}(0) \approx 0$, so a primary focus is to solve for $F^{\uparrow}(0)$ in the gray atmosphere, which is given by

$$\frac{dF^{\uparrow}}{d\tau} = F^{\uparrow} - \sigma T^4, \tag{3}$$

where T is temperature, τ is optical depth, and σ is the Stefan-Boltzmann constant. We integrate (3) and get

$$F^{\uparrow}(0) = e^{-\tau_s} F^{\uparrow}(\tau_s) + \int_0^{\tau_s} \sigma T^4 \times e^{-\tau'} d\tau'. \tag{4}$$

The OLR is then given by

$$OLR^{\kappa} = e^{-\tau_s} \sigma T_s^4 + \int_0^{\tau_s} \sigma T^4 \times e^{-\tau'} d\tau', \tag{5}$$

where A_s represents the surface value of A, and we have used $F^{\uparrow}(\tau_s) = \sigma T_s^4$. This equation shows that *OLR* has two components: one is the surface contribution, and the other is the atmosphere contribution.

We now use (5) to calculte the OLR difference between the two atmospheres, each containing one moist and one dry columns. We remind the readers that T_s, r and thereby τ of the two atmospheres are identical, so the OLR difference primarily comes from dry columns, in which there is significant air temperature difference. The OLR difference of the dry column is given by

$$\Delta OLR^{\kappa} \equiv OLR_{v}^{\kappa} - OLR_{nv}^{\kappa} \approx \int_0^{\tau_s} 4\sigma T_d^3 \Delta T \times e^{-\tau'} d\tau' \approx \int_0^{\tau_s} 4\sigma T_m^3 \Delta T \times e^{-\tau'} d\tau', \tag{6}$$

where OLR_v and OLR_{nv} represent OLR in the atmosphere with and without the vapor buoyancy effect. In the last equal sign, we assumed that $(T_d - T_m)/T_m \ll 1$. Because the strong and weak absorption bands occupy equal portions of the spectrum, the total OLR difference is given by

$$\Delta OLR = 0.5 \times (\Delta OLR^{\kappa_S} + \Delta OLR^{\kappa_W}). \tag{7}$$

To compute ΔOLR, we need information of T_m, ΔT, τ, and thereby r, which is the mixing ratio of water vapor.

b. Temperature

In the Earth's tropical atmosphere, temperature profiles can be approximated by power-law relations of pressure:

$$T = T_s \left(\frac{p}{p_s}\right)^{R_d \Gamma_M / g}, \tag{8}$$

where T_s is the surface temperature, p_s is surface pressure, R_d is the gas constant for dry air, Γ_M is the moist adiabatic lapse rate, and g is gravity acceleration. This has been referred to as the "all-troposphere model" by Pierrehumbert (2010), as the lapse rate is entirely determined by moist convection. Equation (8) fits the observed temperature profiles in the tropical troposphere, but introduces significant biases in the stratosphere (Beucler and Cronin 2016). Earth's OLR is dominated by tropospheric contributions, which justifies the use of (8).

c. Moisture

The water vapor mixing ratio r is the ratio of the mass of water vapor to the mass of dry air and is given by

$$r = RH \times r^*(T,p), \tag{9}$$

where RH is the relative humidity, and r^* is the saturation mixing ratio. For the moist column, we assume that $r_m = r_m^*$ ($RH = 1$) at all vertical levels; For the dry column, we have $r = \beta \cdot r_m^*$, where $0 < \beta < 1$. Here β is a more convenient parameter than the relative humidity of dry columns (RH_d). This is because, at given T_s, the two dry columns are of different temperatures, so they would have different RH values corresponding to same mixing ratio. In reality, β could have complicated vertical structures, which requires multiple parameters to describe. For the purpose of illustrating the proposed mechanism with minimal parameters, we take β as a constant at all vertical levels. This simplification may affect the results quantitatively but will not affect the results qualitatively.

d. Optical depth

Outgoing longwave radiation is observed from space, so it would be convenient to define the optical depth τ as an increasing function as pressure. We thus require that $\tau(p=0) = 0$. We write the optical depth as

$$d\tau = \kappa \cdot r \cdot \frac{dp}{g}. \tag{10}$$

The optical depth would have different values for the two absorption bands: $d\tau = \kappa^s \cdot r \cdot dp/g$ for the strong band, and $d\tau = \kappa^w \cdot r \cdot dp/g$ for the weak band. In this model, we ignore the pressure-broadening effect and treat the absorption coefficients as constant: $\kappa^s = 1.66 \times 0.1$ (m^2/kg) and $\kappa^w = 1.66 \times 0.02$ (m^2/kg), where the factor of 1.66 is referred to as the diffusivity factor (Pierrehumbert 2010). Here the absorption coefficients are consistent with previous modeling studies of similiar complexity (Ingersoll 1969; Pierrehumbert 2010; Beucler and Cronin 2016). We can integrate (10) to obtain the optical depth at an arbitrary pressure level

$$\tau(p) = \int_0^p \kappa \cdot r \cdot \frac{dp'}{g}. \tag{11}$$

At surface $p = p_s$, we then have $\tau_s \equiv \tau(p_s) = \int_0^{p_s} \kappa \cdot r \cdot dp'/g$. Here τ_s measures the optical depth at the surface, which is total optical depth of the atmospheric column.

e. The WBG approximation and ΔT

Buoyancy is horizontally homogenized in the tropical free troposphere (Fig. 1). We refer to this constraint as the weak buoyancy gradient (WBG) approximation (Yang 2018a). This is an improvement of the weak temperature gradient (WTG) approximation, which neglects the vapor buoyancy effect (Charney 1963; Sobel et al. 2001). In a moist atmosphere, buoyancy is related to the virtual temperature, which is given by

$$T_v = T\left(\frac{1+r/\varepsilon}{1+r}\right), \tag{12}$$

where $\varepsilon = M_v/M_d$, where M_v and M_d represent the molar mass of water vapor and dry air, respectively. In the free troposphere, uniform buoyancy requires the virtual temperature to be uniform across the moist and dry area:

$$T_m\left(\frac{1+r_m/\varepsilon}{1+r_m}\right) = T_d\left(\frac{1+r_d/\varepsilon}{1+r_d}\right). \tag{13}$$

We substitute $T_d = T_m + \Delta T_{WBG}$ into (13) and get

$$\Delta T_{WBG} = T_m\left(\frac{1+r_m/\varepsilon}{1+r_m} - \frac{1+r_d/\varepsilon}{1+r_d}\right)\left(\frac{1+r_d}{1+r_d/\varepsilon}\right). \tag{14}$$

Equation (14) is derived without approximations about the amount of water vapor and the amplitude of ΔT. Although this form is quite accurate, we would like to simplify it by assuming water vapor is a trace gas: $r \ll 1$. This is a good assumption for the current climate and may still be good till surface temperature reaches 320 K, at which temperature $r^*(p_s) = 73$ g/kg. With this approximation, we get

$$\Delta T_{WBG} = T_m \left(\frac{1}{\varepsilon} - 1 \right) \left(r_m - r_d \right). \tag{15}$$

This simplified equation clearly tells that ΔT depends on the contrast, not just absolute values, of mixing ratio and molar mass.

The above calculation is more accurate in the free troposphere, where gravity waves efficiently smooth out buoyancy anomalies. Although there is no such constraints in the boundary layer, we can assume that $\Delta T = 0$ at the surface temperature because of the uniform sea surface temperature (SST). We, therefore, require ΔT equals ΔT_{WBG} in the free troposphere but smoothly decays to 0 at surface:

$$\Delta T = \Delta T_{WBG} \times \left[1 - \left(\frac{p}{p_s} \right)^n \right], \tag{16}$$

where n controls the decay rate with pressure. The p/p_s term would decay faster (slower) with large (small) n, so different n could potentially result in different amplitudes and altitudes of the maximum temperature difference. We, however, find that the values of ΔOLR and its sensitivity to surface temperature only change by 50% while we vary n over an order of magnitude, from 5 to 50. Therefore, we conclude the results are robust to the choice of n, and we take $n = 30$ in the following calculation. Figure 3a shows ΔT profiles with different β values at different surface temperatures: ΔT decreases with β and increases with T_s. At 300 K surface temperature, the maximum of ΔT is about 0.45 K when $\beta = 0.8$; the maximum of ΔT is about 1.1 K when $\beta = 0.5$; the maximum of ΔT is about 1.7 K when $\beta = 0.2$. We find that ΔT increases faster with T_s in drier columns, which would be used to explain the sensitivity of ΔOLR to T_s.

Equations (6-10, 16) form the complete set of this model. With proper parameter values, we can estimate the magnitude of ΔOLR and its change with surface temperature.

f. Results

Our calculation shows that the vapor buoyancy effect can significantly impact Earth's energy balance and future climate changes. Figure 4a shows that ΔOLR is of O(4 W/m^2) for a wide range of parameter values. In the reference climate ($T_s = 300$ K), ΔOLR is about 2.5 W/m^2 with $\beta = 0.5$, a similar magnitude to the radiative effect due to doubling CO_2. We then understand the sensitivity of ΔOLR to T_s and β according to (6).

- ΔOLR increases with T_s at given β. This is mainly because ΔT increase with warming, as will be quantified in Figs. 4b & 4c.
- ΔOLR is small at both moist and dry limits. In the moist limit ($\beta \to 1$), ΔT is small according to (15). In the dry limit ($\beta \to 0$), although ΔT maximizes, ΔOLR is dominated by surface emission, insensitive to ΔT. The OLR difference, therefore, peaks at intermediate β values.
- The ΔOLR peak shifts toward smaller β in warmer climates. This is because, at high temperatures, ΔT increases faster with warming in the small-β columns (Fig. 3a) and also because the large-β columns become increasingly opaque to IR emission (Fig. 3b-c).

Consistent with our hypothesis, ΔOLR increases with T_s, showing a negative climate feedback. To quantify the feedback strength, we define feedback parameters

$$\lambda_t = \frac{d\Delta OLR}{dT_s}, \tag{17}$$

$$\lambda_{vb} = \frac{d\Delta OLR}{dT_s}\Big|_{T,\tau}, \tag{18}$$

where λ_t is the total sensitivity of ΔOLR to T_s, and λ_{vb} is the vapor-buoyancy feedback parameter, which only concerns $d\Delta T/dT_s$. Figure 4b shows that λ_t is of O(0.2 W/m^2/K) in the reference climate, which compares with the feedback strength due to clouds and surface albedo. The feedback parameter keeps increasing with surface temperature and reaches about 1.4 W/m^2/K at 320 K, suggesting that the vapor buoyancy effect becomes increasingly important in future climates.

Figure 4c shows that λ_{vb} is of similar magnitude to λ_t, suggesting the vapor-buoyancy feedback dominates the entire ΔOLR sensitivity to T_s. We find that λ_{vb} is small at the moist and dry limits. This is because $\Delta T \to 0$ when $\beta \to 1$ at all surface temperatures, and ΔOLR is dominated by surface emission when $\beta \to 0$ at all surface temperatures, not feeling ΔT and its changes. In addition, we find that the peak of λ_{vb} moves towards small-β columns with warming because ΔT increases faster with warming at small-β columns (Fig. 3a), and also because large-β columns become increasingly opaque at high temperature (Fig. 3b-c), insensitive to changes of ΔT that peaks in the lower troposphere.

The overall results do not depend on the assumed ΔT profiles in the boundary layer. Figure 4d-f shows ΔOLR, λ_t, and λ_{vb} for the free troposphere ($p < 900$ hPa). The free-troposphere results almost reproduce the full-column results, with amplitudes of 10 - 15% weaker than the full-column calculation. This suggests that the vapor-buoyancy radiative effect and feedbacks occur primarily in the free troposphere.

4. Observation: A case study

We estimate the radiative effect due to the vapor buoyancy using *in-situ* observed temperature and moisture profiles from the Nauru Atmospheric Radiation Measurement (ARM) site during the period from April 2001 to August 2013 (Fig. 5). The method of this calculation is based on the two column model developed in Fig. 2 and in Section 3. We assume that, in the free troposphere, buoyancy of this atmospheric column is the same as that of a saturated, convecting atmospheric column, and then we solve for the temperature profile of the convecting column using (13). This derived temperature profile corresponds to that of an atmosphere without the vapor buoyancy effect (Fig. 2a) and is about 0.8 K colder than the observed temperature profile in the lower troposphere (Fig. 5d).

We then compute the clear-sky OLR of the two columns by using the Rapid Radiative Transfer Model [RRTM v3.3, Mlawer et al. (1997)]. The model was run using 28 vertical levels between 1013 hPa to 55 hPa. The CO_2 mixing ratio is set at 400ppm, and all other trace gases, including ozone, were set to zero. The clear-sky OLR of the observed atmospheric column is 300.4 (W/m^2); the clear-sky OLR of the no-vapor-buoyancy column is 299.3 (W/m^2). Therefore, the vapor buoyancy effect is responsible for about 1.1 W/m^2 increase in clear-sky OLR for a typical atmospheric column in the deep tropics. This result is encouraging as it agrees well with our simple model calculations at 300 K surface temperature. If we shift the boundary layer top from 900 hPa to 800 hPa, this OLR difference would be 0.9 W/m^2, which remains significant. This sensitivity test suggests that the radiative effect of vapor buoyancy primarily originates from the free troposphere.

This case study is a first step toward quantifying the radiative effect of vapor buoyancy in Earth's atmosphere. Future analyses should expand this study by using global-scale data sets, providing a more accurate estimate of the radiative effect of vapor buoyancy over the entire tropical atmosphere.

5. Conclusion and discussion

The conventional wisdom is that the vapor buoyancy effect is small, so its impact on temperature is negligible in the free troposphere. However, using NASA AIRS observations, we have demonstrated that the vapor buoyancy effect could lead to about 1.5 K horizontal temperature difference in the lower troposphere from the driest column to the moistest column (Fig. 1), which has a significant impact on Earth's radiative balance.

Based on the novel observation, this paper proposes that the vapor buoyancy effect can increase Earth's OLR by increasing the air temperature in the dry columns. We have developed a simple model that computes the OLR difference between two atmospheres: one with the vapor buoyancy effect, and the other without this effect. We show that the magnitude of this effect is of O(1 W/m^2) at T_s = 300 K, which is then confirmed by observations, and that it increases rapidly with climate warming due to an exponential increase of atmospheric water vapor, leading to a negative climate feedback (Fig. 2b). We further show that the feedback strength λ is of O(0.2 W/m^2/K), the amplitude of which compares with major climate feedback, including cloud feedback and surface albedo feedback. Therefore, faithful representation of the vapor buoyancy effect in climate models is necessary for accurate estimates of climate sensitivity and reliable predictions for future climate changes.

The vapor buoyancy effect may help explain why tropical climate has been more stable than extratropical climate (Holland and Bitz 2003; Polyakov et al. 2002; Pierrehumbert 1995). The strength of the vapor buoyancy feedback depends on water vapor contrast between moist and dry columns, which in turn depends on water vapor abundance and thereby temperature of the atmosphere. This effect, therefore, operates more efficiently in the tropics and less efficiently at higher latitudes. This spatial pattern may explain why fluctuations of sea surface temperature in the tropics are much smaller than that of higher latitudes in the past 100 million years (Pierrehumbert 1995).

The vapor buoyancy effect helps extend the inner edge of the habitable zone, in particular, for tidally locked exoplanets. Tidally locked planets are often slowly rotating, so their free troposphere could be in the WBG regime globally (Koll and Abbot 2016; Mills and Abbot 2013). These planets have one fixed diurnal hemisphere and one nocturnal hemisphere, corresponding to the moist and dry columns of our model, respectively. When the tidally locked planets are approaching the inner edge of the habitable zone, their surface temperature could be significantly higher than Earth's tropical SST, providing an ideal environment for the vapor buoyancy feedback to work efficiently. However, previous studies have neglected the vapor buoyancy effect and assumed WTG (Yang et al. 2013; Yang and Abbot 2014; Pierrehumbert 2010), which could lead to considerably narrower habitable zones. Therefore, we suggest that the vapor buoyancy effect should be accurately represented not only in GCMs but also in low-order models that are used to study climate habitability.

To focus on order-of-magnitude understanding, we have inevitably introduced simplifications to our model that only considers the clear-sky longwave radiation. An important one is that we use the two-band radiative transfer model, lacking detailed representation of water vapor's absorption spectrum. We have also assumed that β is uniform in altitude, whereas β often has complicated vertical structures in the real atmosphere. However, a suite of cloud-resolving model (CRM) simulations has shown similar estimates of ΔOLR and λ. The CRM uses a comprehensive radiation scheme and explicitly simulates atmospheric circulation and water vapor dynamics. The CRM results have also shown that the vapor buoyancy effect does not affect the short-wave radiation budget and that the clear-sky effect dominates the OLR response. The CRM results, therefore, justify our simplifications and will be presented in a companion paper (Seidel and Yang 2018).

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LIST OF FIGURES

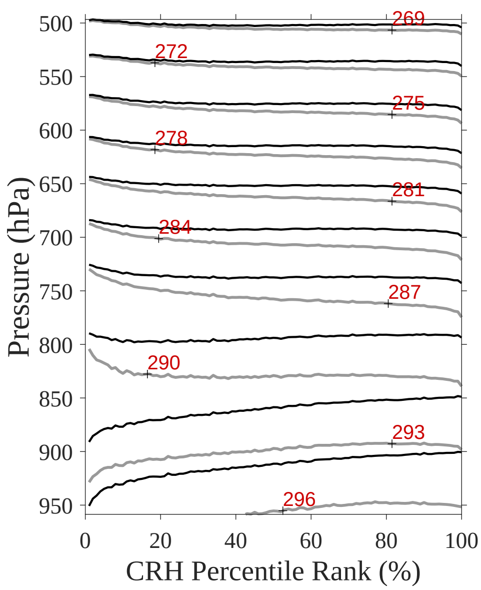


FIG. 1. Temperature and virtual temperature fields in the moisture space using NASA AIRS data from 2^oS to 2^oN for the calendar year 2017. Black contours correspond to virtual temperature, and gray contours correspond to temperature. The *x*-axis is column relative humidity (CRH) rank, where CRH was calculated as precipitable water divided by saturation precipitable water above the altitude of 850 hPa. The driest columns are to the left, and the moistest columns are to the right. The temperature contours deviate from virtual temperature contours due to the vapor buoyancy. In the free troposphere, the horizontal buoyancy gradient is weak. Therefore, virtual temperature contours are approximately flat and temperature increases toward dry columns. In the planetary boundary layer, surface drag can sustain significant pressure gradients, and thereby buoyancy gradients. There, both virtual temperature and temperature increase toward the moist column, forming a low pressure center over the up-welling branch of the atmospheric circulation.

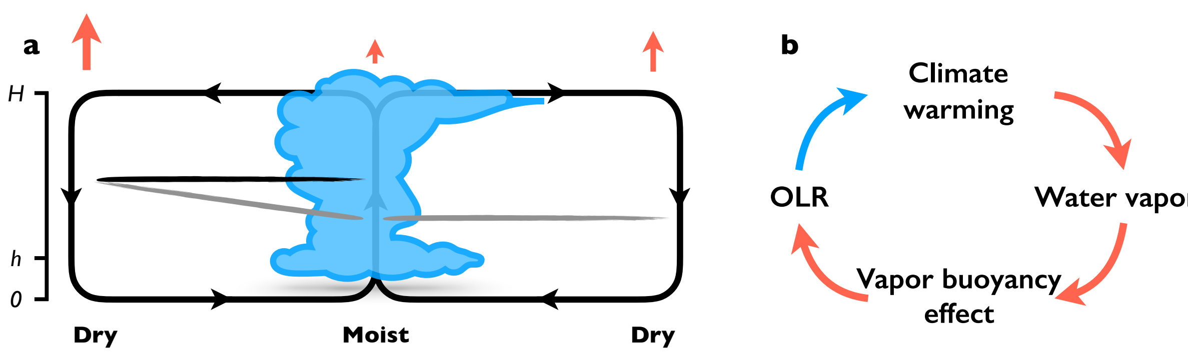


FIG. 2. Schematic diagrams. **a)** The vapor buoyancy effect increases OLR in the tropical atmosphere. This figure depicts two stand-alone atmospheres: the control atmosphere (left); no-vapor-buoyancy atmosphere (right). The horizontal axis is x or CRH; the vertical axis is height (h = boundary layer height, H = tropopause height). The gray lines represent temperature contours, and the black line represent buoyancy or virtual temperature contour. The orange arrows represent OLR emission: large (small) arrow corresponds to more (less) OLR. **b)** The negative climate feedback. Orange arrows represent an increase effect; the blue arrow represents a decrease effect.

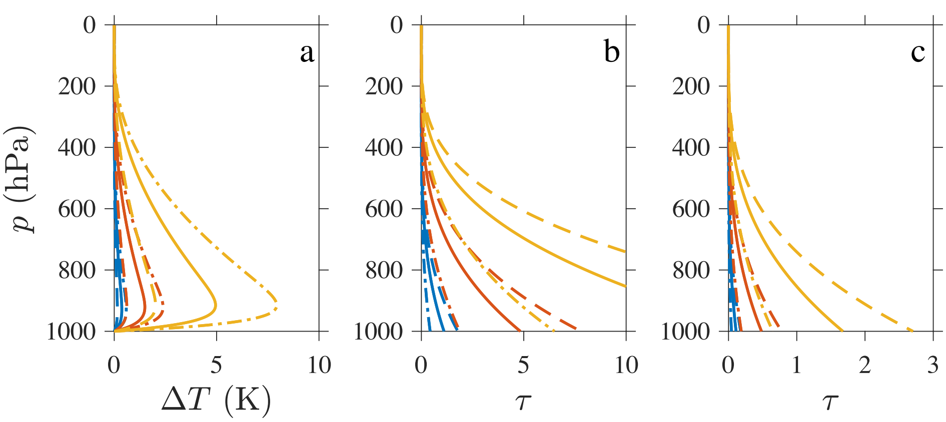


FIG. 3. (a) ΔT profiles (K). (b) τ profiles for the strong absorption band. (c) τ profiles for the weak absorption band. Blue: 280 K, red: 300 K, yellow: 320 K. Dot-dashed: $\beta = 0.2$; solid: $\beta = 0.5$; dashed: $\beta = 0.8$.

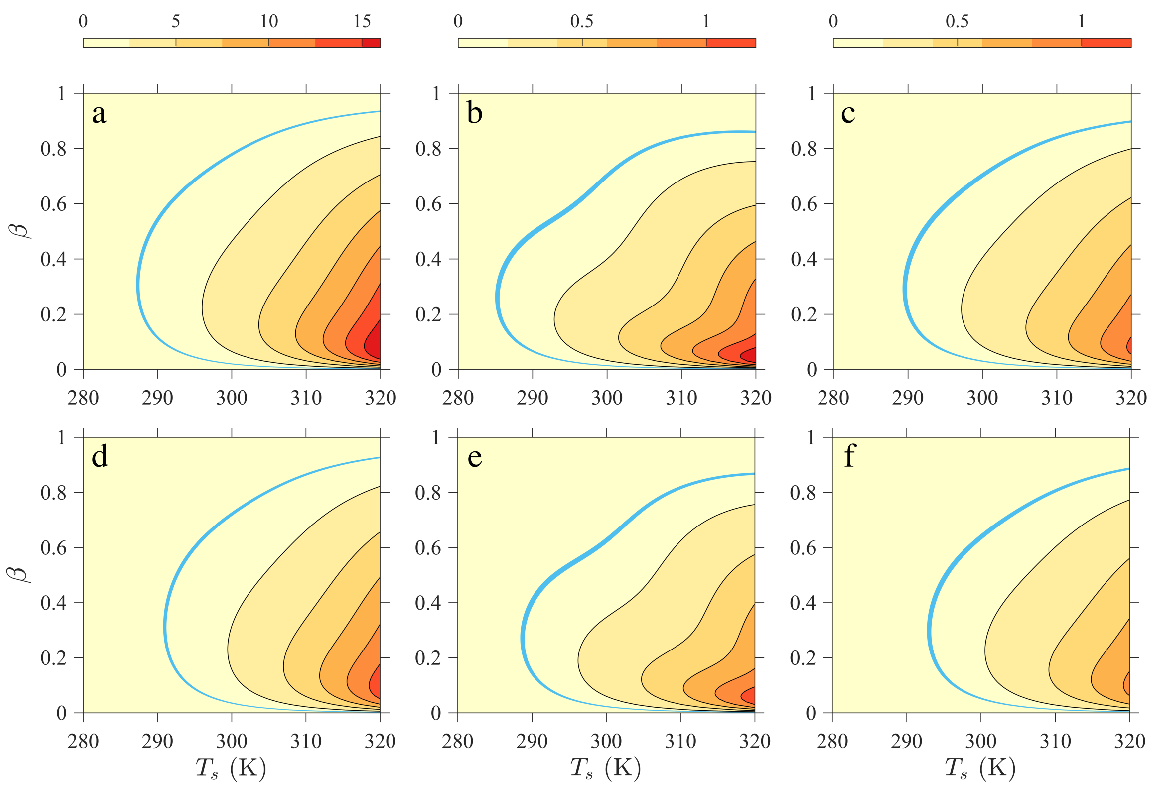
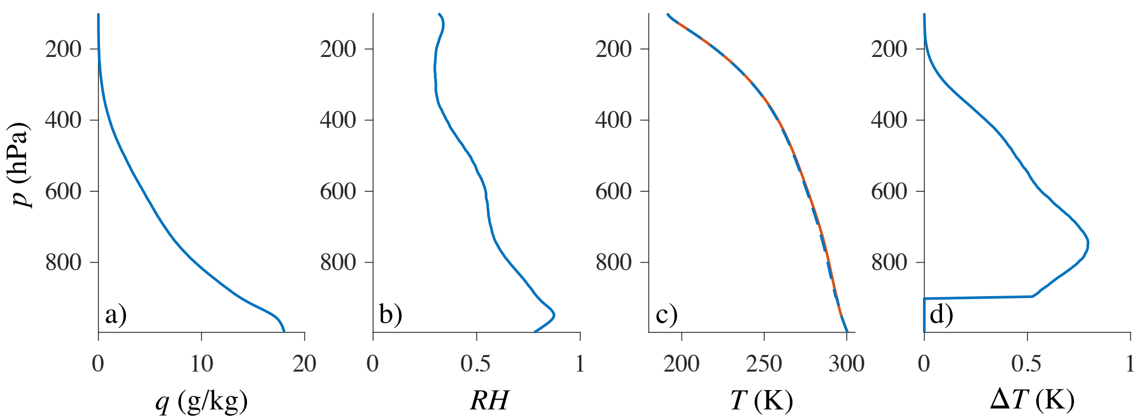


FIG. 4. (a) The radiative effect ΔOLR (W/m^2) due to the vapor buoyancy effect. The blue curve corresponds to $\Delta OLR = 1$ W/m^2. (b) The total feedback parameter λ_t (W/m^2/K). (c) The feedback parameter λ_{vb} (W/m^2/K) of the vapor buoyancy feedback. The blue curve corresponds to $\lambda = 0.1$ W/m^2/K in (b-c). (d-f) ΔOLR, λ_t, and λ_{vb} for the free troposphere ($p < 900$ hPa), respectively.



503 FIG. 5. Time-averaged temperature and moisture profiles at the Nauru ARM site during the period from April 2001 to August 2013. (a) Specific humidity. (b) Relative humidity. (c) Temperature profiles. The red curve corresponds to the observed temperature profile; the blue curve corresponds to the derived temperature profile of a convecting column. (d) Temperature difference, which is defined as red minus blue. Here we define the boundary layer top as 900 hPa, consistent with the definition in the simple model.