Surface Moisture Exchange Under Vanishing Wind in Simulations of Idealized Tropical Convection

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Abstract Under radiative-convective equilibrium (RCE), surface moisture fluxes drive convection, while convection-driven winds regulate surface fluxes. Most simulations of RCE do not resolve the boundary-layer turbulence that drives near-surface winds due to too coarse grid spacing and instead parameterize its effects by enforcing a minimum wind speed in the computation of the ocean-atmosphere exchange. We show from RCE simulations with fully resolved boundary-layer turbulence that capturing wind dynamics at low speeds impacts the spatially averaged surface moisture flux, as well as its spatial distribution. A minimum wind speed constraint of only 1 m s⁻¹ leads to ~10% increase in spatially averaged surface flux in the evolution towards RCE and reduces the surface flux differences between windy and calm regions with more than a factor of two. Hence, the ability of simulations to let wind vanish is key in representing the wind-induced surface heat exchange feedback and is potentially important in convective self-aggregation.

1. Introduction

Radiative-convective equilibrium (RCE) is an idealized representation of the tropical atmosphere in which cooling due to radiation is balanced by heating through convection. Simulation of RCE is often used to study the tropical climate and to assess the impact of climate change therein (e.g., Bretherton et al., 2005; Held et al., 1993; Manabe & Strickler, 1964; Popke et al., 2013; Tompkins & Craig, 1998a). Spontaneous self-aggregation of convection can occur in such simulations (Held et al., 1993) leading to a drastic decrease in the radiative forcing due to clouds (Tobin et al., 2013; Wing & Cronin, 2015) as large cloud-free regions effectively emit radiation back to space. While self-aggregation is a potentially important negative feedback mechanism in global warming, understanding the underlying physics of its onset (Coppin & Bony, 2015; Muller & Held, 2012) and development (Holloway & Woolnough, 2016; Muller & Bony, 2015; Wing & Emanuel, 2014) remains challenging.

The surface-flux feedback has been introduced as a potential enhancer of self-aggregation (Wing & Emanuel, 2014). In RCE, surface moisture fluxes are the source of water and energy for deep convection, and in turn, the convection-driven winds, either as part of overturning cells (Coppin & Bony, 2015) or cold pools, (Schlemmer & Hohenegger, 2015; Zuidema et al., 2017) control the surface fluxes. In regions of strong convective activity, the enhancement of surface moisture fluxes due to increased wind speed outweighs a possible reduction in fluxes as an effect of the decrease in the atmosphere-ocean specific humidity gradient due to additional moistening (Wing et al., 2017). Hence, the surface-flux feedback acts as a positive feedback in the development of self-aggregation.

The sea-surface temperature plays a major role in the presence, speed, and intensity of self-aggregation in simulations (Bretherton et al., 2005; Hohenegger & Stevens, 2016; Reed et al., 2015), but it remains unclear whether the simulations capture ocean-atmosphere exchange with sufficient detail. This is partially the effect of the horizontal resolution of typical RCE simulations (Δx ~ 1 km) being sufficient to capture the circulation due to the largest clouds but insufficient to resolve boundary-layer turbulence and shallow circulation (Naumann et al., 2017). In the absence of (fully) resolved boundary-layer turbulence, cloud-resolving and large-scale models parameterize its effects by introducing a minimum wind speed in the surface-exchange formulation. This added wind speed permits the widely used Monin-Obukhov similarity theory-based surface exchange formulations to be applied under conditions of free convection (Beljaars, 1995), such as RCE (section 2).
The aim of this research is to study the influence of ocean-atmosphere exchange at very low wind speeds on the magnitude and spatial structure of the surface moisture fluxes in RCE simulations. We do this through large-eddy simulations following the Radiative-convective equilibrium model intercomparison project (RCEMIP) description (Wing et al., 2018) but with an increased spatial resolution ($\Delta x = 100m$) in order to fully resolve the boundary-layer turbulence at the expense of the domain size. Although the chosen domain size ($\sim 150$ km) prevents self-aggregation from occurring, it permits us to study the subtle details of ocean-atmosphere exchange at wind speeds close to vanishing. We study to what extent RCE simulations are sensitive to imposing a minimum wind speed in the surface-flux formulation. Our main hypothesis is that not allowing wind to vanish could result in an underestimation of the surface-flux feedback strength, as latent heat fluxes are increased in calm regions, thus reducing this flux contrast between calm and convective regions. It is likely that this effect is stronger in idealized setups such as RCEMIP than over the actual ocean where large-scale disturbances in the ocean and atmosphere are additional drivers of horizontal wind.

We have set up a simulation experiment in order to test this hypothesis (section 3). Based on the results of the experiment (section 4), we discuss the implications for RCE simulations and studies to convective self-aggregation (section 5).

2. Computation of Surface Fluxes Under Vanishing Wind

The exchange of heat, water, and momentum between surface and atmosphere is commonly parameterized using Monin-Obukhov similarity theory (MOST) in atmospheric models (chapter 10, Wyngaard, 2010). MOST predicts surface fluxes of moisture, temperature, and momentum from their respective near-surface gradients using empirical functions fitted to observational data. The surface latent heat flux is formulated as

$$\text{LHF} = -\rho_0 L \beta q_*,$$

where $\rho_0$ is the reference density, $L$ the latent heat of vaporization, and $u_*$ and $q_*$ are the friction velocity and the moisture scale, respectively. MOST provides the framework to relate $u_*$ and $q_*$ to the values of wind speed and specific humidity at the first model level. We show $u_*$ as an example, $q_*$ is defined analogously, with its own roughness length and MOST functions.

$$u_* = \kappa \left(U_1 - U_{SS}\right) \left[\log \left(\frac{z_1}{\zeta_{m0}}\right) - \Psi_m \left(\frac{z_1}{L}\right) + \Psi_m \left(\frac{\zeta_{m0}}{L}\right)\right]^{-1},$$

where $\kappa$ is the Von Kármán constant, $U_1 - U_{SS}$ is the wind speed difference between atmosphere and ocean, $z_1$ is the height of the first model level, $\zeta_{m0}$ is the roughness length for momentum, and $L$ the Obukhov length, which is a measure of the stability of the near-surface atmosphere. The function $\Psi_m$ is the integrated form of the empirical MOST relationship that relates the near-surface dimensionless vertical wind gradient to the surface momentum flux. Throughout the years, many studies have presented best fits of the MOST relationships to observational data (e.g., Högström, 1988) but most have the wrong asymptotic limit under free convective conditions with vanishing winds. More recently, Wilson (2001) provided alternative fits that have an overall good match with observational data and, in addition, have the correct asymptotic behavior, which is a key quality for RCE simulations.

The common approach to deal with free convection is to prevent $U_1$ in equation (2) from approaching 0 by introducing convection-driven winds into its definition. The most basic method is application of a minimum wind speed $U_1$ at height $z_1$.

$$U_1 = \max(U_1, U_{\text{min}}).$$

which is the proposed method in the RCEMIP specification with $U_{\text{min}} = 1 \text{ m s}^{-1}$ (Wing et al., 2018).

Large-scale models often use the method proposed by Beljaars (1995), in which the Deardorff free convective velocity scale $\omega_c$ is added to the calculation of $U_1$.

$$U_1^2 = u_*^2 + \nu_*^2 + (\beta w_*)^2.$$

This extra term represents wind induced by convective eddies, where constant $\beta$ is fitted on large-eddy simulation data. This method is, however, questionable for large-eddy simulation as the turbulence is resolved explicitly, and adding a convective velocity to the total wind speed leads to (partial) double counting of convection-driven winds.
### 3. Methods

We perform large-eddy simulations using MicroHH (van Heerwaarden et al., 2017). Time integration is done with a fourth-order Runge-Kutta scheme, and scalars are advected with a third-order near-monotonic scheme, referred to as $2i3$. A fourth-order variant, $2i4$, is used in a sensitivity experiment. Subgrid-scale turbulent transport is handled by a Smagorinsky scheme. We use a two-moment microphysics scheme (Seifert & Beheng, 2005). As discussed in section 2, surface fluxes are calculated using a MOST-based parameterization with a correct free convection limit (Wilson, 2001). The ocean has a constant sea-surface temperature with specific humidity at saturation at the sea surface. A sponge layer is applied above the troposphere. The simulation setup follows the case description given by RCEMIP for an sea-surface temperature of 300 K. Rotation and large-scale forcings are absent. Initial random noise is applied to the $\theta$ field in the lowest model levels with a maximum amplitude of 0.1 K. As the cost of radiation computations would render this study impossible, we use a prescribed radiative cooling profile. This profile is based on the mean net radiative cooling rate over Days 70–100 in a preliminary System for Atmospheric Modeling (Khairoutdinov & Randall, 2003) simulation of the 300 K RCEMIP case (courtesy of Allison Wing). Our results therefore describe the convective adjustment towards radiative cooling.

A series of simulations (Table 1) is set up that captures the initial stage of convection development under radiative cooling with total durations varying between 20 and 30 days. A reference simulation ($ldr\_reference$) is performed on a domain of size $153.6^2 \text{km}^2$ by 19.1 km, which is sufficiently large to capture the convective cells that are typical for the early RCE development. The minimum wind speed input for the surface layer formulation, $U_{\text{min}}$, is set to 0.1 m s$^{-1}$ in this reference run. A horizontal resolution of 100 m is combined with a vertical grid that stretches from $\Delta z = 30$ m at the lowest level to $\Delta z \approx 350$ m at 19 km. In our setup, the lowest level $z_1$ of the $u$ and $v$ horizontal wind components is at a height of 15 m. The high spatial resolution results in a large increase of the computational costs in comparison to the RCEMIP specification. Furthermore, a set of runs is performed on a smaller domain of $38.4^2 \text{km}^2$ by 18.7 km in order to study the effect of $U_{\text{min}}$ and test the importance of domain size for our results. This set features a reference ($sdr\_reference$) and forced ($sdr\_forced$) simulation, which are run with $U_{\text{min}} = 0.1\text{ m s}^{-1}$ and 1 m s$^{-1}$, respectively, for 30 days. Additional runs are performed on a small domain at $\Delta x = 200$ m (Table 1) to test the sensitivity to horizontal resolution and advection scheme.

### 4. Results

#### 4.1. The Spatial Characteristics of Surface Fluxes and Near-Surface Wind

The cross-section of the equivalent potential temperature ($\theta_e$) combined with the near-surface wind speed mask (Figure 1) provides insight in the structure and magnitude of the convection-driven winds. The main structure is determined by cold pools, recognizable by low values of $\theta_e$ that mark the midtropospheric air that...
Figure 1. Example horizontal cross sections of $\theta_e$ at $z_1$ (15 m) for a 30-km by 30-km subsection of ldr_{reference}. $\theta_e$ is colored in gray scale linearly from white (339 K) to black (346 K). This cross-section is taken at $t = 8.8$ days. The hatched, transparent gold overlay marks the grid points where $U_1$ is below either 0.1 or 1 m s$^{-1}$ for (a) and (b), respectively. The fraction of the domain covered by these masks is 2.1% for (a) and 77% for (b).

has been transported towards the surface by the downdrafts. The presence of the cold pools is in line with observations suggesting that cold pools form when rain rates exceed approximately 2 mm hr$^{-1}$ (Zuidema et al., 2017). The high spatial resolution of our simulations also reveals the structure of the shallow circulation that is the precursor of new rain cells to be formed in the near future. For example in the region surrounding $(x, y) = (25, 20)$, thin warm and moist bands of high $\theta_e$ mark the new convection.

The superimposed masks of regions with wind below 0.1 m s$^{-1}$ (2.1% of domain in Figure 1a) and 1 m s$^{-1}$ (77% of domain in Figure 1b) show that three quarters of the domain has near-surface wind speeds below 1 m s$^{-1}$. Only in the vicinity of cold pools centers (light colors in Figure 1) the 1-m s$^{-1}$ threshold is exceeded. These regions correspond to the leading edges, or gust fronts, of actively spreading cold pools. Increased latent heat fluxes and moisture convergence make these gust front regions both favorable for new convection to form and are also the main driver of the horizontal redistribution of moisture (Schlemmer & Hohenegger, 2014, 2015). The threshold of 0.1 m s$^{-1}$ is exceeded in nearly the entire domain due to the wind speeds generated by boundary-layer turbulence and shallow convection in our simulations. Setting a minimum wind speed value of 1 m s$^{-1}$ in the computation of the surface fluxes thus constrains the interaction between boundary-layer turbulence and surface fluxes and consequently narrows the range of wind speeds over which the surface-flux feedback can be active.

While suppressing surface-flux feedback dynamics in such a large part of the domain is cause for concern, its impact can only be assessed through a quantitative analysis. MOST dictates that the surface flux is proportional to the wind speed and the moisture difference between atmosphere and ocean. Therefore, we have constructed the joint probability density function (PDF) of these two quantities based on data from Days 12 to 20 of our reference simulation ldr_{reference} and combined it with cumulative density functions of the wind speed in the ldr_{reference} and sdr_{forced} simulations (Figure 2). In order to relate the occurrence of combinations of wind speed and moisture differences to the flux they produce together, a weighting of the probability by the surface moisture flux is applied. The cumulative density functions show that approximately two thirds of the surface moisture flux is generated in regions where surface wind speeds are below 1 m s$^{-1}$, whereas the unweighted line shows that 76% of the total surface area has a wind speed below 1 m s$^{-1}$. Furthermore, the peak in the PDF is well defined and located at a wind speed of 0.5 m s$^{-1}$, indicating that the variability in this range needs to be well resolved in order to capture the surface-flux feedback in all its detail. The structure of the PDF is indicative of a positive surface-flux feedback because the loss of moisture difference between ocean and atmosphere is small for increasing wind speed. For instance, a fivefold increase in wind speed from the center of the PDF (0.5 m s$^{-1}$) to its tail (2.5 m s$^{-1}$) leads to a less than 10% (0.5 g kg$^{-1}$) moisture difference reduction. Therefore, new convection is likely to form nearby an already active region.
To elaborate the impact of boundary-layer turbulence on the spatial structure of the surface moisture flux, we have quantified its partitioning between calm ($U_1 < 1$ m s$^{-1}$) and convective ($U_1 \geq 1$ m s$^{-1}$) regions based on 4 days of data for all experiments in Table 1 (Figure 3). Each experiment contains a reference and a forced simulation. The forced simulations display significant differences in surface moisture flux between calm and convective regions as well as in the domain mean compared to the reference simulations. In forced simulations, the mean flux in calm regions is approximately 10 W m$^{-2}$ (25%) higher than in reference simulations and covers a slightly smaller area (2 to 4 percentage points). This relationship is reversed in the convective region of the domain due to the decreased near-surface moisture gradient. Here, the mean flux is approximately 10 W m$^{-2}$ (15%) lower in forced simulations. Since this area is also three times smaller than the calm region, the total effect is a significant net increase in total surface latent heat flux of ~10% and a shift of flux towards the calm region throughout the transition towards RCE. The results are robust under a reduction of the horizontal grid spacing to 200 m ($\text{s}_r\_200$) as well as under a change in advection scheme ($\text{s}_r\_214$).

### 4.2. The Temporal Evolution of Convection

We now study the time evolution of convection because idealized studies of convection over a constant temperature surface have shown a strong coupling between the rate of change of most of the bulk boundary-layer properties to the magnitude of the surface fluxes (van Heerwaarden & Mellado, 2016). Figure 4 shows the temporal evolution towards RCE, the Bowen ratio, $u_*$, and the specific humidity at the lowest model level for reference and forced simulations on a small and large domain (Table 1). The first strong cold pools form after ~48 hr, resulting in cascading release of convective available potential energy, which is particularly evident in $\text{ldr}_\text{reference}$. The surface enthalpy flux is consistently higher in the forced simulation for at least the first 25 days, after which, it starts to converge with the reference simulation. Most of the enthalpy flux comes...
Figure 4. Time series of six domain-mean variables: (a) surface enthalpy flux (SEF), (b) Bowen ratio, (c) precipitation expressed in W m$^{-2}$, (d) rate of change of the vertically integrated moist static energy $\dot{\hat{h}}$, (e) surface friction velocity, and (f) specific humidity at the lowest model level. A 24-hr centered moving average is applied to all time series, except for rain rate that is averaged with an 8-day window. This is done to more clearly see differences between simulations.

from the latent heat flux, given the low Bowen ratio throughout the simulations (Figure 4b), resulting in a marginal increase in precipitation in the forced simulation (Figure 4c). The convergence of the surface fluxes is a direct consequence of our experimental setup as we are simulating adjustment to a prescribed radiative cooling profile in order to keep the simulation computationally affordable. This means that the sensible heat flux is decreased in the forced simulation compared to the reference (Figure 4b). Nonetheless, our simulations demonstrate that forced simulations consistently have a higher $u^*$ for the same flux and consequently a lower moisture difference between ocean and atmosphere. Therefore, there is a moister atmosphere, with the strongest moistening in the boundary layer (Figure 4f), caused by the enhanced surface moisture fluxes over the calm regions (section 4.1). The potential for radiative cooling over dry areas is thus suppressed in the forced simulation. This most likely lowers the potential for self-aggregation because the radiative cooling differences between dry and wet areas drive up-gradient moist static energy transport by shallow circulation (Naumann et al., 2019). In addition, the forced simulation adjusts faster to the radiative forcing as is indicated by its higher surface enthalpy flux (Figure 4a). This is in line with earlier findings that the time scale of convective adjustment is inversely proportional to the magnitude of the surface fluxes (Tompkins & Craig, 1998b, their Equation 12; Cronin & Emanuel, 2013, their Equation 7; and van Heerwaarden & Mellado, 2016, their Equation 6). The faster adjustment makes radiative cooling a relatively less powerful mechanism to enhance contrasts between dry and moist areas.

5. Summary and Perspective

In this study, we demonstrate the importance of ocean-atmosphere moisture exchange at vanishing wind in simulations of idealized tropical convection. We have performed a series of high-resolution large-eddy simulations based on the RCEMIP (Wing et al., 2018) but with an increased spatial resolution in order to resolve boundary-layer turbulence. Our results show that enforcing a minimum wind speed to account for boundary-layer turbulence and shallow circulation, which is the common practice in large-scale and cloud-resolving models, has a profound influence on the ocean-atmosphere moisture exchange. Imposing a minimum wind speed of 1 m s$^{-1}$ enhances the mean surface latent heat flux with more than 10% compared to a simulation in which the wind can vanish in the surface-flux computation. Furthermore, as the flux...
enhancement mostly happens in calm regions at the expense of the surface flux in the convective regions, the contrast in latent heat flux magnitude between calm regions and convective regions loses more than half of its magnitude if a 1 m s\(^{-1}\) minimum wind speed is imposed.

Our simulations give therefore a new perspective on the wind-induced surface heat exchange feedback. In our simulation setup, the ability of the model to reduce surface fluxes under vanishing wind is more important than the enhancement of fluxes in regions of high wind speeds. Furthermore, the simulations in which the wind can vanish have lower surface moisture fluxes; hence, the ability of radiation to enhance contrasts between moist convective regions and dry calm regions is expected to be larger. Consequently, the surface-flux feedback in self-aggregation is potentially stronger than assumed until now. This conjecture can only be confirmed or rejected after performing RCE simulations on large domains with fully resolved boundary-layer turbulence and interactive radiation, but this will remain a computational challenge for the foreseeable future.

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