The influence of wind speed on shallow marine cumulus convection

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ABSTRACT
The role of wind speed in regulating shallow marine cumulus convection is explored using Large-Eddy Simulations and concepts from bulk theory. Focusing on cases that are characteristic of regions influenced by the trade-winds it is found that the equilibrium trade-wind layer is deeper at stronger winds, with larger surface moisture fluxes and smaller surface heat fluxes. The opposing behavior of the surface fluxes is caused by more warm and dry air being mixed to the surface as a result of the deepening of the cloud layer. This leads to little difference in equilibrium surface buoyancy fluxes and mass fluxes at cloud base. Shallow cumuli are deeper at stronger winds, but not more numerous or more energetic. The deepening response is necessary to resolve a contradiction that arises in the sub-cloud layer when wind speed changes. With an increase in winds, a moistening of the sub-cloud layer tends to lower cloud base height, if taken as the level at which parcels condensate. The increase in surface moisture and buoyancy fluxes however tend to raise cloud base height, if taken as the height where the buoyancy flux reaches a fixed fraction of its surface value. Because this is demonstrated using simple bulk concepts, it is suggested that the internal dynamics of clouds and how they mix with the environment are only of secondary importance to the deepening response.

1. Introduction
The ‘trades’ owe their name to the easterly surface winds that prevail over subtropical oceans and that once made foreign commerce flourish. Within the trades, widespread fields of shallow cumulus dominate the various forms of moist convection. These ‘trade-wind’ cumuli remain shallow as their growth is halted by the trade inversion, limiting their depth to typically 1 or 2 km (Malkus 1956; Stevens 2005). It is not uncommon however to find cumuli that reach up to 3 or 4 km (Medeiros et al. 2010) and that produce considerable rain showers (Nuijens et al. 2009).

In studying what controls the behavior of these clouds, the role of the trade-winds, from which the region derives its original meaning, has received far less attention compared to other features, such as large-scale divergence, sea surface temperature and the thermodynamic structure of the lower troposphere. Although it is understood that the winds are crucial, by help inducing evaporation from the ocean’s surface, which drives turbulence and moist convection, their persistent nature may have encouraged us to neglect their effect. Exactly how and how much the weakening and strengthening of the trades, as measured by wind speed, affects the clouds that they embody has not been extensively studied. In fine-scale modeling studies of shallow cumulus convection for instance, wind speed is commonly prescribed as a constant forcing.

The objective of our work is to gain more insight into the influence of wind speed on cloud properties, turbulent fluxes and the structure of the trade-wind layer. For cases characteristic of the trades, Large-Eddy Simulation (LES) is used to study the response of convection to perturbations in wind speed. In addition, we use bulk theoretical concepts to help us understand the behavior observed in the simulations.

The motivation for having a closer look at wind speed comes from a previous study that used observations collected during the Rain in Cumulus over the Ocean (RICO) field campaign to study relationships between clouds, precipitation and the large-scale meteorological environment (Nuijens et al. 2009). Results from that study suggest that wind speed, in addition to subsidence, plays a major role in regulating variability in boundary layer humidity, hence in cloudiness and rainfall. More specifically, stronger winds are observed to correspond to higher humidities throughout the entire boundary layer and an increase in area rainfall. The causality of this relationship, which we like to test here, is that stronger winds lead to enhanced evaporation, which increases humidity in the lower sub-cloud layer, leads to more upward mixing of moisture by clouds, hence higher humidities in the cloud layer. This favors the development of deeper clouds with more liquid water, that may rain more.

Relationships between wind speed, humidity and rainfall are not new in studies of deep convection and the ideas
explaining such relationships are similar to what is proposed for shallow convection. By compositing sounding profiles over the island of Nauru in the western tropical Pacific, Holloway and Neelin (2009) show that variability in deep convective rainfall is mainly linked to variability in free tropospheric humidity. They attribute the increase in rainfall with humidity to an increase in the buoyancy of cloudy updrafts, via entrainment. Similar findings are discussed in Bretherton et al. (2004) and Back and Bretherton (2005), who also show that wind speed explains a significant part of the variability in daily rainfall in the Pacific ITCZ from four years of satellite retrieved data.

One idea proposed in these studies is that enhanced evaporation under stronger winds lead to a greater number of shallow cumuli, where each cumulus contributes to a moistening and deepening of the boundary layer (here including both the sub-cloud and cloud layer), which increases the chance that a deeper precipitating system develops. Viewing the RICO results in perspective of these ideas, a few interesting questions arise, for instance: does an increase in wind speed just lead to a greater number of shallow cumuli or to a deepening of cumuli in general? How does that impact the moistening versus the deepening of the boundary layer? Given that the layer close to the ocean surface moistens, can the enhanced evaporation be sustained over a longer period of time?

The approach we use in addressing these questions is inspired by the observations that motivated our study. The vertical structure observed during RICO at a single point reflects the history of air masses that have travelled for a few days through the trades at a certain wind speed. During that time an equilibrium may have been established between the different processes that act on the layer, the surface fluxes of heat and moisture, subsidence and radiation, among which clouds are an important link. The increase in boundary layer humidity with wind speed may thus be considered as an equilibrium response to stronger winds. These ideas are further explained in section 2 and precede our simulations that we use to describe not only the transient response to wind speed perturbations, but also the equilibrium response (section 3). We focus on shallow cumulus cases that have an idealized thermodynamic structure, not far from what is typically observed within the trades, for which wind shear and rain microphysics are neglected. Excluding rainfall may seem at odds with our motivation, but our goal here is to understand the response of clouds to wind speed first, and rain may blur part of that response. Following the simulations, we then apply concepts from bulk (equilibrium) theory, such as those described in Betts and Ridgway (1989) (BR89), to help understand how the underlying conservation equations constrain the equilibrium response of the trade-wind layer (section 4). The results are discussed in section 5 and conclusions are summarized in section 6.

2. The idea

Imagine that the vertical profiles in the top panel of Figure 1 represent the vertical structure of air masses that have travelled a few days through the trades. If the wind shear that is present in the profile of $u$ is ignored, one assumes that the air masses have been advected in their entirety by a mean (surface) wind speed $U$. An idealized picture of the vertical structure at low $U$ (dashed lines) is shown in the bottom panel. It exhibits the structure that is typical for the trades, with a well-mixed layer of depth $\eta$ above the ocean’s surface, a less rapidly mixed layer in which clouds are embedded, topped by an inversion layer at $h$, which separates the turbulent boundary layer from the free tropospheric air above it.

This thermodynamic structure results from a subtle balance between the processes that act on the layer and as such it may be useful to consider the equilibrium state of the layer. This eliminates the time derivative from the underlying conservation laws and allows one to directly link the gradients of temperature and humidity to the processes that play a role. The equilibrium state follows from the tendency equations of the conserved variables $\theta_l$ and $q_l$ (the liquid water potential temperature and total specific humidity):

\begin{equation}
0 = -\overline{\mathbf{w}} \frac{\partial \theta}{\partial z} - \frac{1}{\rho c_p} \left( \frac{\partial F_r}{\partial z} - \frac{\partial F_q}{\partial z} \right)
\end{equation}

\begin{equation}
0 = -\overline{\mathbf{w}} \frac{\partial q}{\partial z} - \frac{\partial F_q}{\partial z}
\end{equation}

where for convenience $q_t$ is written as $q$, $\theta_l$ is written as $\theta$ and the time derivatives $\partial \theta / \partial t$ and $\partial q / \partial t$ are put to zero. It is assumed that the air mass is horizontally homogeneous (which eliminates the horizontal advection terms) and that there is no significant lateral advection into the air mass. $F_r$ denotes the net radiative flux at any height, $\overline{\mathbf{w}}$ is the subsidence velocity ($\overline{\mathbf{w}} < 0$) and $F_{\theta_r} = \overline{w' \theta'}$ and $F_{q_r} = \overline{w' q'}$ denote the turbulent fluxes of $\theta$ and $q$ respectively. In equilibrium, conservation of heat (or in this case conservation of $\theta_l$) requires that at any height between the surface $s$ and the inversion $h$, warming due to subsidence (acting on the temperature gradient) and due to the turbulent heat flux divergence is balanced by radiative cooling (with the radiative cooling rate $Q_r = -\frac{1}{\rho c_p} \frac{\partial F_r}{\partial z}$). Moisture conservation requires that drying due to subsidence is balanced by moistening due to the turbulent moisture flux divergence.

Wind speed enters the above equations in the surface fluxes. These are typically modeled with bulk formulae (Fairall et al. 2003), in which the flux of any quantity is the product of wind speed and the difference of that quantity between the surface and the sub-cloud layer (for $\theta$ and $q$):

\begin{equation}
F_{\theta,s} = C_D U (\theta_s - \theta_m)
\end{equation}

\begin{equation}
F_{q,s} = C_D U (q_s - q_m)
\end{equation}
Fig. 1. Top panel shows composite profiles of $\theta$, $q$, $u$ and $\theta_e$ conditioned on periods with little rainfall (dashed lines) and periods with moderate rainfall (solid lines) as observed during RICO (adapted from Figure 7 in Nuijens et al. (2009)). Bottom panel shows an idealization of the $\theta$ and $q$ profiles of the first composite (with little rainfall and lower zonal wind speeds), along with subsidence $w$, radiative cooling $Q_r$ and the surface fluxes of heat and moisture $F_{\theta,s}$ and $F_{q,s}$.

Here, $s$ denotes the surface, $m$ denotes the well-mixed subcloud layer, $C_D$ is a surface transfer coefficient determined from similarity theory and $U = \sqrt{u_m^2 + v_m^2}$ is the wind speed above the surface. $\theta_s$ is the sea surface temperature (SST) and $q_s$ is the saturation specific humidity at SST. The surface fluxes combine into the surface buoyancy flux $F_{b,s}$:

$$F_{b,s} \simeq \frac{q}{\theta_{v,0}} (F_{\theta,s} + \epsilon \theta F_{q,s})$$

where $g = 9.81 \text{ m s}^{-2}$ is the gravitational acceleration, $\theta_{v,0}$ is the basic state virtual potential temperature and $\epsilon = \frac{R_v}{R_d} - 1 = 0.608$ where $R_v = 461.5 \text{ J kg}^{-1}$ and $R_d = 287 \text{ J kg}^{-1}$ are the gas constants for water vapor and dry air.

Using Equation 3 and 4 one can see that if $U$ is perturbed by an amount of $\delta U$, the surface fluxes change directly. This affects the warming and moistening rate of the layer (Equation 2 and 2), which changes $\theta_m$ and $q_m$ and asks for a further adjustment of the surface fluxes. If it warms and moistens enough close to the surface (decreasing $\theta_s - \theta_m$ and $q_s - q_m$) the effect of a stronger wind speed on the fluxes can be offset. This leads to little difference in equilibrium surface fluxes: $\delta F_q, \delta F_{\theta} \approx 0$. If moistening and warming is not sufficient, the surface fluxes in a new equilibrium will differ from their original values: $\delta F_q, \delta F_{\theta} \neq 0$.

In the next section, LES provides a detailed picture of these changes. One of our main findings is that changes in wind speed are accompanied by changes in the depth of the layer $h$, which majorly affect the response of the surface fluxes. The response of the layer to stronger winds is different than hypothesized in our previous study (Nuijens et al. 2009), in which we assumed that $h$ does not change, and as such the layer would warm and moisten enough for $\delta F_q, \delta F_{\theta} = 0$. In section 5 we return to some of the above bulk theoretical concepts to explore why the equilibrium solution whereby the depth of the layer $h$ remains unchanged is not seen in the simulations. It will be demonstrated that such a solution is inconsistent and that changes in the surface winds are necessarily accompanied by changes in $h$. 


3. Large-Eddy Simulation

a. Simulation set-up

All simulations are performed with the UCLA-LES code (Savic-Jovcic and Stevens 2008) for cases that develop typical shallow cumulus convection after a few hours of spin-up. The first case explored is the RICO LES inter-comparison case, that derives its initial profiles (Figure 2) from measurements during an undisturbed three week period during RICO, and that is described in Van Zanten et al. (2010). By varying the initial wind profiles of RICO by 50%, the following results emerge after a simulation time of 60 hours (Figure 3). At stronger winds, clouds are deeper (panel c), the boundary layer has deepened, but the thermodynamic gradients have changed little (panel a and b). The subcloud layer has warmed and moistened (panel a and b), but the surface fluxes from case to case are different (panel e and f). Interestingly, little difference in the surface buoyancy flux and cloud-base mass flux is observed (panel g and h).

To get at the heart of the issue and to evaluate the extent to which the response observed during RICO does not depend on specific features of the case, it makes sense to use a simpler framework. Some of the features of the RICO case we prefer to exclude, at least for the moment, are its non-stationarity (the case continues to deepen during the simulation because subsidence is relatively weak), the presence of wind shear (the initial profile of $u$ is sheared by 8 m s$^{-1}$ over 4 km), and the development of rain. The non-stationarity prevents us from evaluating possible new equilibria, and both wind shear and rainfall provide additional mechanisms that can affect cloud development, turbulence and surface fluxes. A final point for not focusing on the RICO case are the prescribed horizontal advection of heat and moisture, derived from regional downscaling of meteorological analyses. These principally support an Eulerian reference frame, whereas the background of our ideas lends itself better for a Lagrangian reference frame.

Instead we therefore use the framework introduced by Bellon and Stevens (2010), where large-scale forcings and initial profiles are taken as typical for the subtropics, but somewhat idealized (Table 1 and Figure 2). Warming due to large-scale subsidence is assumed to balance a fixed radiative cooling rate $Q_r$ of 2.5 K d$^{-1}$ in the free troposphere. This, combined with the profile of the subsidence rate, provides the free tropospheric temperature profile. The subsidence rate $\overline{w}$ is given an exponential form:

$$\overline{w}(z) = \overline{w}_0[1 - e^{-z/H}]$$

where $\overline{w}_0$ is equal to 7.5 mm s$^{-1}$ (case S7.5) or 8.5 mm s$^{-1}$ (case S8.5) and $H$ represents a scale height, taken to be 1 km (Figure 2 and Table 1).

Initial temperature and humidity profiles are well-mixed and constant with height up to 1 km, with $\theta = 298 \text{ K}$ and $q = 13 \text{ g kg}^{-1}$, topped by an inversion layer that extends up to 1.6 km. In the free troposphere, the profile of $\theta$ follows from the assumed balance between subsidence warmth-
Fig. 3. The vertical profile of liquid water potential temperature $\theta$, specific humidity $q$, liquid water $q_l$, zonal wind speed $u$, turbulent heat flux $F_\theta$, turbulent moisture flux $F_q$, buoyancy flux $F_b$ and mass flux $M$ for the RICO case, averaged from hour 52 to hour 60 of simulations that start with the control wind speed profile (green), \( u + 5 \) m s\(^{-1}\) (dark green) and \( u - 5 \) m s\(^{-1}\) (yellow-green).

\[
\frac{d\theta}{dz}(z) = \frac{Q_r}{\overline{w}}(z)
\]

whereas humidity in the free troposphere is simply constant at \( 4 \) g kg\(^{-1}\). The initial wind profile equals the geostrophic wind that is constant with height, with a zonal component of \( 10 \) m s\(^{-1}\) and a meridional component that is \( 0 \) m s\(^{-1}\) (Figure 2). The simulated airmass is assumed to be horizontally homogeneous in its mean properties and horizontal advection is hence excluded.

Surface turbulent fluxes are parametrized using bulk aerodynamic formulae (Equation 3,4) with a prescribed sea surface temperature (SST) of 300 K, and a slip/no-penetration condition at the surface. The domain is 12.8 x 12.8 x 5 km with a grid-spacing of 50 m in the horizontal and 25 m in the vertical, stretching by a factor 1.02 in the region where \( z > 4 \) km. Time stepping is performed with a Runge-Kutta-3 scheme with an adaptive time-step limited to be no larger than 1 s.

As motivated earlier, the simulations are performed without rain microphysics. Since clouds and especially precipitation exhibit a significant sensitivity to the numerical representation of the flow (Matheou et al. 2010), performing simulations without rain removes at least one uncertainty. To reduce numerical dissipation and to lower the CFL number, a Galilean transform is applied with the Galilean velocity equal to the geostrophic wind velocity. In order to not develop systematic numerical differences between the simulations, the Galilean velocity is always changed along with a wind speed perturbation. The only differences in numerical advection speed are then located in the sub-cloud layer where shear develops (with more shear present in simulations with stronger winds). Test simulations with different CFL numbers show that such differ-
ences in structure between S\textsubscript{5} and S\textsubscript{7} due to the surface buoyancy flux \( F_{b,s} \), is weak and strongly influenced by the moisture flux \( F_{q,s} \). Because the fluxes do not change noticeably, we assume the sub-cloud layer is in equilibrium at hour 60 and perturb the zonal wind speed by \( \delta u = +5 \) (U\textsubscript{15}) and \( \delta u = -5 \) (U\textsubscript{5}). All simulations, including U\textsubscript{10}, are continued for two more days after the perturbation.

The most apparent changes occur in the six hours after the perturbation, then followed by a more gradual relaxation to equilibrium (Figure 5). Similar to the RICO simulations, the layer deepens at stronger winds, with a sudden increase in surface moisture flux and an increasingly negative surface heat flux. Cloud fraction and the mass flux are initially perturbed, but quickly relax backward to their original values. That the surface fluxes evolve more gradually throughout the simulations, suggests that their evolution reflects changes in the layer structure (\( \theta_m \), \( q_m \)), and not just the direct effect of a perturbation in wind speed.

The overall response is not completely linear i.e., subtracting wind does not give the exact opposite response to adding wind. Some secondary features develop in U\textsubscript{15}, such as more shear in the sub-cloud layer and increased cloudiness at upper levels (Figures 6 h and c), which changes its character compared to U\textsubscript{5} and U\textsubscript{10}. The main response is however unaffected by these differences. One can note that the behavior of the surface heat, moisture and buoyancy flux, as well as the mass flux, is similar to the RICO case (Figures 3 e-h), which supports the idea that the response we are seeing is general, and motivates the use of the idealized case for developing our ideas. In the following analyses we focus mainly on S\textsubscript{7.5}, even though it does not reach true equilibrium at higher wind speeds. We do so simply because clouds in that case get deeper and are better resolved, and in terms of the general response to a change in wind speed, S\textsubscript{7.5} and S\textsubscript{8.5} are very similar.

**Surface flux and deepening response**

Whereas \( F_{q,s} \) responds immediately to the perturbation, a direct response of \( F_{\theta,s} \) is less apparent, simply because it is approximately zero to begin with (Figure 5). Within the next hour however, \( F_{\theta,s} \) starts to decrease in U\textsubscript{15} (and increase in U\textsubscript{5}), a response that was somewhat unexpected. The reason for \( F_{\theta,s} \) turning negative in U\textsubscript{15} is because the layer is warming from enhanced entrainment at the top of the sub-cloud layer, mixing potentially warmer cloud layer air into the sub-cloud layer (note that \( F_{\theta,s} \) is zero initially, hence the warming cannot be caused by additional surface input). The enhancement of turbulent entrainment right after the perturbation is caused by more vigorous turbulence in the sub-cloud layer, a result of a

<table>
<thead>
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<th>Case</th>
<th>S\textsubscript{7.5} / S\textsubscript{8.5}</th>
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<tbody>
<tr>
<td>Forcings</td>
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<tr>
<td>( \overline{w}_0 ) (m s\textsuperscript{-1})</td>
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<tr>
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<td>( Q_r ) (K d\textsuperscript{-1})</td>
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<tr>
<td>( u_g ) (m s\textsuperscript{-1})</td>
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<tr>
<td>( v_g ) (m s\textsuperscript{-1})</td>
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**Initial and boundary conditions**

\( SST \) (K) | 300 |
| \( q \) (g kg\textsuperscript{-1}) | \begin{align*} z < 1 \text{ km} &: 13 \\ z \geq 1.6 \text{ km} &: 4 \end{align*} |
| \( \theta \) (K) | \begin{align*} z < 1 \text{ km} &: 298 \\ z \geq 1.6 \text{ km} &: \frac{\partial \theta}{\partial z} = Q_r/\overline{w} \end{align*} |

**Domain and resolution**

\( \Delta t \) (s) | 1 |
| \( \Delta x, \Delta y, \Delta z \) (m) | 50 \times 50 \times 25^\circ |
| \( n_x, n_y, n_z \) (-) | 256 \times 256 \times 190 |
| domain size (km) | 12.8 \times 12.8 \times 5 |

Table 1. Initial and boundary conditions, forcings and other specifications. Details are given in the text.* At \( z > 4 \) km the vertical grid is stretched uniformly by a factor of 1.02.
larger $F_{q,s}$ and $F_{b,s}$. In $U_5$ the opposite takes place: entrainment (warming) decreases and is less able to compensate the unchanged radiative cooling. This leads to a gradual cooling of the sub-cloud layer accompanied by a gradual increase in $F_{\theta,s}$.

At stronger winds, more shear develops in the sub-cloud layer. This occurs in a few hours after the perturbation (on a time scale much shorter than the adjustment time of the surface fluxes), and differences in near-surface wind speed between $U_5$, $U_{10}$ and $U_{15}$ are therefore less than 5 m s$^{-1}$ (see also the vertical profiles of $u$ and $v$ and of $S = (u'u''\omega_x^2) + (v'v''\omega_x^2)$ taken at the end of the simulation (Figures 6 d and h)). Because the convective forcing in the subcloud layer is not strong to begin with, shear-generation of turbulence may begin to play a more important role in $U_{15}$. Shear near cloud base has been shown to positively influence the rate of entrainment (Moneng and Sullivan 1994; Pino et al. 2003; Conzemius and Federovich 2006), which would imply more drying, and require a larger surface moisture flux in equilibrium. Because $F_{q,s}$ in $U_{15}$ shows more evidence of a relaxation back to its value of $U_{10}$, a point to which we return in the discussion (section 4), we believe such a shear effect is small and does not majorly affect the behavior of the sub-cloud layer.

Evidently, the wind speed perturbation, followed by a change in the surface buoyancy flux, changes the rate of mass exchange between the cloud layer and the free troposphere, as measured by $dh/dt - \overline{w_h}$. The sudden increase in $F_{q,s}$ and $F_{b,s}$ allows deeper clouds to develop, that are associated with a greater flux of liquid water into the inversion. Upon mixing with the overlying dry air, liquid water evaporates, cools and moistens the inversion, thereby slowly deepening the layer as a whole (Betts 1973; Stevens 2007). As clouds penetrate further into the free troposphere, the net amount of dry and warm free tropospheric air that is mixed into the cloud layer increases. This additional warming and drying is felt closer to the surface, through entrainment across the sub-cloud and cloud layer interface. In equilibrium, this results in more input of moisture at the surface and less input of heat. The latter explains why the surface fluxes act in opposite ways and force the surface buoyancy flux to relax backward to its original value (most evident for $U_5$ and $U_{10}$).

Into the second day after the perturbation, the cases therefore have a similar surface energy input. Cloudy updrafts are moister, but also warmer, so that the liquid water flux at each level is very similar in each case. Because the layers have different depths, and the liquid water flux carried into the inversion scales with the depth of the layer, their deepening tendencies remain different (Stevens 2007).

**Cloud and mass flux response**

Along with the sudden change in moisture and buoyancy flux right after the perturbation, the fraction of buoyant updrafts that reach their saturation level changes. This is evident from the sudden increase in (total) cloud cover, $cc$, for $U_{15}$, and the decrease in $cc$ for $U_5$. The cloud cover, defined as the number of cloudy columns in the domain, remains altered during the simulation, but cloud fraction at

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**Fig. 4.** The vertical profile of liquid water potential temperature $\theta$, specific humidity $q$, liquid water $q_l$ and zonal and meridional wind speed $u$ and $v$ for case $S_{7.5}$ and $S_{8.5}$, averaged from hour 52 to hour 60 of the simulations.
cloud base (not shown), cloud-core fraction $a_{co}$ and the mass flux at cloud base $M$ quickly approach values similar to that of $U_{10}$. The change in total cloud cover thus reflects mainly clouds getting deeper, and because they have irregular shapes and do not necessarily fill the same column at any given height (i.e., they do not obey the maximum overlap rule) $cc$ can change even if the vertical profile of cloud fraction does not change.

Why cloud (core) fraction changes little can be understood from the structure and dynamics of the sub-cloud layer. The saturation level of updrafts is typically somewhat higher than the depth of the well-mixed layer ($\eta$) and the two are separated by a thin stable layer often called the transition layer. The transition layer is marked by a small negative jump in humidity and positive jump in temperature and reflects the penetrative nature of dry convection in the sub-cloud layer. When a sudden perturbation, such as an increase in wind speed, lowers the saturation level because the well-mixed layer moistens, the number of cloudy updrafts (as measured by cloud core fraction $a_{co}$) increases and along with it the mass flux:

$$M = w_{co} a_{co} \tag{8}$$

where $w_{co}$ is the cloud-core vertical velocity. More mass removed from the layer in turn lowers $\eta$:

$$\frac{d\eta}{dt} = E - M + \overline{w}_\eta \tag{9}$$

where $\overline{w}_\eta < 0$ and $E$ is the rate at which cloud layer air is entrained into the sub-cloud layer ($E > 0$). As $\eta$ is lowered, it is soon brought close to the saturation level again, which re-establishes the thin transition layer and ensures that $a_{co}$ and $M$ quickly relax to their original values. Note that also $w_{co}$ is similar for each case, though most evidently for $U_5$ and $U_{10}$ (Figure 6 j), which can be understood if assumed that it scales with the convective velocity scale $w^*$ (Stevens 2006; Neggers et al. 2006) so that:

$$w_{co} \propto w^* = (\eta F_{b,s})^{1/3} \tag{10}$$

In our results, the saturation level and $\eta$ do not change much with wind speed overall (note that the layer also warms), and neither does $F_{b,s}$, because $F_{q,s}$ and $F_{\theta,s}$ respond oppositely to wind speed changes.

c. Towards a new equilibrium

The surface fluxes are roughly constant throughout the second day after the perturbation (Figure 5) and the sub-cloud layers in $U_5$ and $U_{10}$ of $S_{7.5}$, as well as in $U_5$, $U_{10}$ and $U_{15}$ of $S_{8.5}$, are in equilibrium. Except for the two deepest cases, where boundary layer growth is not zero, the cloud and inversion layer are also in equilibrium. Substantial differences in boundary layer depth have developed at that point, and both the sub-cloud and cloud layer are warmer and moister at stronger winds (Figure 6). Because the surface fluxes remain different from case to case, the sub-cloud layer $\theta$ and $q$ evidently have not increased enough to offset the wind speed change.

The larger moisture fluxes and larger (negative) heat fluxes throughout the cloud layer (Figure 6 e and f) reflect more moist and (potentially) cold air being mixed upwards, as well as more dry and warm air being mixed downwards, at stronger winds. The profiles of $F_b$, $w_{co}$ and buoyancy excess in cloud cores (Figure 6 g, j and l) show that updrafts are not more energetic at stronger winds, nor are clouds more numerous or mass fluxes larger (Figure 6 i and k). Because the mass fluxes are so similar, there is no evidence that the rate at which buoyant thermals overshoot and entrain cloud layer air, $E$, is larger at stronger winds. This can be seen from Equation 9 where in equilibrium the left-hand side vanishes and $E$ must balance $M - \overline{w}_\eta$. Because $\eta$ (and therefore $\overline{w}_\eta$) and $M$ do not vary much, in particularly not between $U_5$ and $U_{10}$, we can assume $E$ is also not that different. This implies that the larger turbulent fluxes across the transition layer (the entrainment fluxes) are just the result of larger differences in $\theta$ and $q$ across the transition layer ($\Delta \theta, \Delta q$). In other words, the cloud layer has warmed and moistened more relative to the sub-cloud layer.

It is important to note that the cloud layer has been mixed homogeneously i.e., the gradients of $\theta$ and $q$ in the cloud layer are approximately unchanged. The amount of drying and warming that the cloud layer experiences due to large-scale subsidence acting on the gradients is therefore the same. In equilibrium this constrains the divergence of the turbulent fluxes throughout the cloud layer (Equation 2 and 2), so that the difference in surface fluxes from low to high wind speed reflect the difference in fluxes at the inversion that develop.

4. An explanation for the deepening response from bulk theory

To what extent does the response to wind speed rely on how clouds mix with their environment and on the internal structure of the layer? If there would not have been an increase of mass flux and liquid water into the inversion right after the perturbation, would the layer still have deepened? Using the concepts from bulk (equilibrium) theory introduced in section 2, we here demonstrate that from an energetic point of view, the boundary layer $h$ has to be deeper at stronger winds, regardless of how clouds are mixing. Our strategy is as follows: we explore whether a new equilibrium is possible that is consistent with $h$ not changing, which is done for two hypothetical situations, one whereby equilibrium surface fluxes remain unchanged, and one whereby they do not.
In developing our arguments we make use of Figure 7 that displays a simplified vertical structure of the trade-wind layer, similar to Figure 1. The top panel (black dashed lines) shows the profiles of temperature and humidity and their corresponding fluxes before wind speed is perturbed. The sub-cloud layer is well-mixed in temperature $\theta_m$ and humidity $q_m$. The sub-cloud layer depth $\eta$ is assumed to equal the saturation level. Jumps in temperature and humidity at the top of the sub-cloud layer (the transition layer) are ignored. The profiles are linear in the cloud layer with a discontinuity at the inversion $h$ represented by $\Delta h \theta$ and $\Delta h q$. The surface fluxes $F_{\theta,s}$ and $F_{q,s}$ are modeled using bulk aerodynamic formulae (Equation 3,4).

The fluxes at $h$ can be linearly related to the surface fluxes using the flux difference across the layer $\Delta F = F_s - F_h$:

$$F_{q,h} = F_{q,s} - \Delta F_q = C_D U (q_s - q_m) - \Delta F_q \quad (11)$$

$$F_{h,h} = F_{\theta,s} - \Delta F_\theta = C_D U (\theta_s - \theta_m) - \Delta F_\theta \quad (12)$$

where $\Delta F_\theta, \Delta F_q > 0$. In equilibrium, $F_{q,h}$ and $F_{\theta,h}$ must equal the large-scale drying flux, respectively warming flux, at $h$. These can be represented by the product of the subsidence velocity at $h$ and the jump of temperature and humidity across $h$:

$$\overline{w}_h \Delta_h q = C_D U (q_s - q_m) - \Delta F_q \quad (13)$$

$$\overline{w}_h \Delta_h \theta = C_D U (\theta_s - \theta_m) - \Delta F_\theta \quad (14)$$

with $\overline{w}_h < 0$, $\Delta_h q < 0$, $\Delta_h \theta > 0$. Note that we have anticipated that the radiative cooling $Q_r$ in the expression for the heat balance cancels out when comparing low and high wind speeds, because it is assumed constant.

Now let us assume a wind speed perturbation $\delta U > 0$ and a new equilibrium whereby $h$ has not changed. The profiles in red (middle panel) and green (bottom panel) in Figure 7 resemble possible new equilibria. In both situations the layer has moistened homogeneously by an amount of $\delta q$. The situation in red implies that the layer moistens enough by $\delta q$ to offset the increased contribution of wind speed to the surface flux, hence $\delta F_q = 0$. In green we have sketched a situation where the moistening is not enough so that $\delta F_q \neq 0$. Note that for both we mimic the situation in LES whereby $F_{\theta,s}$ is zero. This, as follows below, implies that $\delta \theta = 0$ (based on the assumption that $h$ does not change, which is not true for the simulations and explains why the heat flux response we outline here is different from the LES).

The change in surface moisture flux $\delta F_q$ is defined as:

$$\delta F_q = C_D \delta U (q_s - q_m) - C_D U \delta q \quad (15)$$

If $F_{q,s}$ does not change, and neither does $\Delta F_q$ (because the gradients do not change), then $F_{q,h}$ does not change either (Equation 11). Such a situation however is inconsistent with the large-scale drying flux at $h$ that has increased by an amount of $\overline{w}_h \delta q$:

$$\overline{w}_h (\Delta h q - \delta q) \neq F_{q,h} \quad (16)$$

A situation with unchanging $h$ as well as unchanging surface fluxes is thus not valid.

For the other possibility, in green, we expect $\delta F_q > 0$ if $\delta U > 0$. Rewriting Equation 14 gives the new flux balance at $h$:

$$\overline{w}_h (\Delta h q - \delta q) = F_{q,h} + \delta F_q$$

$$= F_{q,h} + C_D \delta U (q_s - q_m) - C_D U \delta q$$

$$= F_{q,h} + (\delta U/U) F_{q,s} - C_D U \delta q \quad (17)$$

which appears a consistent and plausible equilibrium. The expression can be re-arranged to solve for $\delta q$:

$$\delta q = \frac{F_{q,s} (\delta U/U)}{C_D U - \overline{w}_h} \quad (18)$$

and a similar derivation can be done for temperature, for the more general case where $F_{q,s} \neq 0$:

$$\delta \theta = \frac{F_{\theta,s} (\delta U/U)}{C_D U - \overline{w}_h} \quad (19)$$

In both expressions the denominator is $> 0$, so that a situation with positive surface fluxes initially will lead to a net moistening and warming when wind speed increases. The expressions are valid even if moistening and warming within the cloud layer is not homogeneous i.e., the gradients $dq/dz$ and $d\theta/dz$ can differ locally, as long as $\delta q$ and $\delta \theta$ at the surface equal $\delta q$ and $\delta \theta$ at $h$. They also do not depend on $\eta$, but they do constrain it.

Thermodynamically $\eta$ is controlled by maintaining a height close to the saturation level (Equation 9 in section 3b), which in turn is controlled by sub-cloud layer humidity and temperature, so that moistening and cooling lead to a decrease in $\eta$. In the case considered here with $\delta \theta = 0$, then $(\delta \eta)_{T} < 0$, where we use $T$ to denote the thermodynamic control on $\eta$.

This effect on $\eta$ however is inconsistent with the changes that take place in the buoyancy budget of the sub-cloud layer. In equilibrium, the divergence of the buoyancy flux $F_b$ within the sub-cloud layer must equal the radiative cooling rate $Q_r$:

$$\frac{dF_b}{dz} = \frac{F_{b,n} - F_{b,s}}{\eta} = Q_r \quad (20)$$

where $Q_r < 0$. Using the common closure whereby the flux at the top of the sub-cloud layer is a fixed fraction $\kappa$ of the surface flux, the expression becomes:

$$\kappa F_{b,s} = F_{b,s} + Q_r \eta \quad (21)$$
with $\kappa = -0.2$. In response to a wind speed perturbation of $\delta U$, the buoyancy flux changes by $\delta F_b$ and the new equilibrium is:

$$\kappa [F_{b,s} + \delta F_b] = [F_{b,s} + \delta F_b] + Q_r (\eta + \delta \eta)$$ (22)

where we assume $\delta Q_r = 0$. This may not be a bad assumption given that cloud fraction does not change much with wind speed in the simulations. The equation can then be rewritten to:

$$\kappa \delta F_b = \delta F_b + Q_r \delta \eta$$ (23)

This relationship demonstrates that a situation in which the sub-cloud layer depth does not change with wind speed ($\delta \eta = 0$) can only be true if either $\kappa = 1$ or if $\delta F_b = 0$. In other words: if the sub-cloud layer depth does not change, the surface buoyancy flux cannot change and vice versa.

If $\delta \eta \neq 0$ however, we can write:

$$(\delta \eta)_D = -\delta F_b \frac{1 - \kappa}{Q_r}$$ (24)

where $(\delta \eta)_D$ refers to changes in sub-cloud layer depth that are constrained dynamically by variations in the surface buoyancy flux $F_{b,s}$. Replacing $F_b$ with $F_b + \epsilon \theta \delta F_q$ (Equation 5) gives:

$$(\delta \eta)_E = -[\delta F_b + \epsilon \theta \delta F_q] \frac{1 - \kappa}{Q_r}$$ (25)

For the situation present in the simulations, $F_{\theta,s} = 0$, Equation 19 predicts $\delta \theta = 0$ and as such $\delta F_b = 0$. Hence, only the surface moisture flux changes by an amount of $\delta F_q > 0$. Equation 25 then predicts that the sub-cloud layer depth must increase: $(\delta \eta)_D > 0$ (recall that both $\kappa$ and $Q_r < 0$).

The contradiction between $(\delta \eta)_D$ and $(\delta \eta)_T$ can be resolved by deepening the layer as a whole i.e., increasing $h$ as we observed in the simulations. This increases the amount of warm and dry air from the overlying free troposphere that is mixed into the layer and allows for $\delta \eta \approx 0$. This extra warming and drying causes the surface heat and moisture flux (in the simulations) to respond oppositely, leading to $\delta F_{b,s} \approx 0$.

How dependent are our findings on the assumption that the surface heat flux is zero when the perturbation takes place? If the analysis is performed for cases where $F_{\theta,s} > 0$ or $< 0$, similar contradictions arise. For instance, if $F_{\theta,s} < 0$, Equation 19 predicts a cooling of the layer which tends to lower $\eta$ just as moistening does, so that $(\delta \eta)_T < 0$. Equation 25 however requires $(\delta \eta)_D = 0$ because of the opposite response of $\delta F_b$ and $\delta F_q$. In that situation an increase in $h$ can resolve the contradiction as well. Only when $F_{\theta,s} < 0$ as well as $F_{b,s} < 0$ initially, Equation 18, 19 and 24 all predict $\delta \eta < 0$. However, an initial buoyancy flux that is negative is rather unrealistic.

The concepts we used above may be considered a simplified approach of the one-layer bulk model developed by Betts and Ridgway (1989) (BR89), in which an explicit mixing line representation is used to model the thermodynamic properties of the cumulus layer (Betts 1985). The main difference is that their closure at cloud base height is based on temperature, and not buoyancy: $F_{\theta,b} = \kappa F_{\theta,b}$. Because in the situations we consider here, a substantial fraction of the buoyancy flux comes from the moisture flux, differences in the closure used will matter in terms of how $\eta$ behaves with wind speed. If we had used the closure of BR89, the term $\epsilon \theta \delta F_q$ would disappear from Equation 25. This would imply that in equilibrium the heat flux determines $\eta$, and only the additional warming (not the drying) that results from deepening $h$ can resolve the inconsistency in $\delta \eta$ that is present if $F_{\theta,b} = 0$ or $> 0$ initially. Solutions are then more likely to predict a decrease in $\eta$ when $\delta U > 0$, which is indeed true for the experiments described in BR89.

5. Discussion

Bulk concepts have provided an explanation for the deepening response to an increase in wind speed. It appears that inconsistencies in the energy budget of the layer develop, which can be resolved by deepening the boundary layer, and that those inconsistencies do not strongly depend on the detailed internal structure of the layer. This suggests that the way clouds are mixing with their environment is only of secondary importance to how the layer as a whole responds towards a new equilibrium. In doing our analysis, we have assumed that moistening and warming occurs homogeneously throughout the layer i.e., the gradients of $\theta$ and $q$ do not change. This of course constraints how the turbulent fluxes adjust, because the constant gradients imply that the divergences of the large-scale drying and warming flux do not change. We make this assumption based on the idea that turbulence and clouds will mix the layer in similar ways independent of whether the winds are weak or strong. To what extent is that a valid idea and if not, would it change our results?

First, we should point out that the bulk analysis principally does not exclude the possibility that gradients in $\theta$ and $q$ vary locally within the layer even though we have drawn the illustration as such. As long as $\delta \theta$ and $\delta q$ at the surface and at $h$ are equal, the equations remain valid. Second, even if we allow the gradients to change, for instance, we assume that humidity and temperature at the top of the cloud layer change much less than in the sub-cloud layer, contradicting responses in the sub-cloud layer depth are present.

The assumption that the way clouds are mixing with the environment does not change with wind speed is also used by BR89. In their model it is assumed that bulk mix-
ing line parameters are constant with a change in large-scale conditions. This is done by fixing the partitioning between clear and cloudy air *a priori*, a parameter to which the surface fluxes and sub-cloud layer depth are sensitive (Stevens 2006). This may not be a bad assumption because cloud fraction does not vary much with wind speed in the simulations. Bellon and Stevens (2010) also show that bulk mixing line parameters when estimated from a set of Large-Eddy Simulations that are run to equilibrium, are relatively insensitive to a range in sea surface and free tropospheric temperature values. This suggests that simple models, such as the mixing line model, can successfully capture the behavior of the cumulus-topped mixed layer under varying conditions. The extent to which this is also true for wind speed and other large-scale forcings, and how important clouds and their mixing and entrainment processes are, is an area of current research.

We have used the bulk concepts to address the deepening response, but not explored what a change in *h* implies for the equilibrium surface fluxes. The simulations show that under stronger winds, stronger surface moisture fluxes are maintained also towards a new equilibrium. The U_{15} simulation of case S7.5 however shows a somewhat deviating behavior (Figure 5) as the flux appears to relax back to its value at a 10 m s^{-1} wind speed. We may explain this behavior using the same bulk concepts as before. Equation 2 shows that the turbulent moisture flux divergence must balance the drying due to large-scale subsidence. The large-scale drying term \( \pi d\bar{q}/dz \) can be written as a flux divergence: \( dF_{q,LS}/dz \). Simplified profiles of both fluxes are shown in the bottom panels of Figure 8, where \( F_q \) (in black) can be considered a generalization of the moisture flux profile for U_{5}, U_{10} and U_{15} (Figure 6 f). Because turbulent fluxes vanish above \( h \), the moisture flux divergence at the inversion \( F_{q,h} \) must equal the divergence of \( F_{q,LS} \) at \( h \) in equilibrium: \( \pi \bar{q} \Delta_q \) (note this is equivalent to Equation 11).

Because the boundary layer from U_{5} to U_{10} and U_{15} gets deeper, but the humidity gradient in the free troposphere is zero, \( \Delta q_{h} \) decreases with wind speed (top panels of Figure 8). The subsidence rate at the inversion \( \bar{\omega}_h \) however also changes with height (see the prescribed subsidence profile in Figure 2 a). A simplified version of this profile is sketched in red on the bottom left in Figure 8, with an increase in subsidence rate up to a certain height \( h_{w,max} \), and a constant rate above \( h_{w,max} \). The amount of drying experienced by clouds near \( h \), as measured by \( \pi \bar{q} \Delta_q \), can thus either increase with wind speed, stay constant or even decrease, depending on how deep the layer is.

Clouds in U_{10} for instance experience more large-scale drying at \( h \) than those in U_{5}. This implies that in U_{10} a larger drying flux is present within the cloud layer and the top of the sub-cloud layer, which in equilibrium implies larger turbulent moisture fluxes everywhere, including at the surface, so that \( F_{q,s}(U_{5}) < F_{q,s}(U_{10}) \). Once \( h \) exceeds \( h_{w,max} \), as is true for U_{15} in which the boundary layer continues to deepen throughout the simulation, large-scale drying starts to decrease. This is because of the combined effect of a constant subsidence rate, but a decreasing \( \Delta q_{h} \). As such, less (turbulent) moisture flux is needed to have a zero net moistening at \( h \). In other words, in simulation U_{15}, \( F_{q,h} \) starts to decrease as the layer gets deeper, which may explain why \( F_{q,s} \) decreases throughout the second day after the perturbation. This effect is absent in the heat flux response, because \( d\theta/dz \) in the free troposphere increases with height (maintaining a constant \( \Delta \theta_h \) and a constant large-scale warming beyond \( h_{w,max} \)). The decrease in \( F_{q,s} \) is also absent in the strongest wind speed case of the RICO simulations (not shown), likely because \( dq/dz \) in the free troposphere is negative there.

These ideas suggest that if a wind speed (or any) perturbation leads to a deepening of the boundary layer into a region where large-scale drying increases with height, the surface moisture fluxes in the new equilibrium may be larger. If large-scale drying decreases or remains constant with height, surface moisture fluxes may relax back to their original values. The profile of subsidence rate and the free tropospheric gradients of \( \theta \) and \( q \) will therefore play an important role in regulating surface fluxes on longer time scales.

Here we have taken Large-Eddy Simulation as a reference of what response to expect when wind speed changes. Because the simulations rely on a few simplifications, it is useful to consider what aspects of the simulated response are in fact realistic and can be expected in observations downstream of trade-wind trajectories. Important to consider first is that from an Eulerian point of view, the deepening that takes place when traveling a distance \( X \) is very similar for weak \( U \) and strong \( U \) + \( \delta U \) winds. This is because the amount of deepening depends on the amount of moisture transported into the inversion, which is a function of the surface moisture flux, as well as the time period travelled, \( T = X/U \). From the simulations, the deepening can be estimated as the tendency to deepen the layer \( (dh/dt - \bar{\omega}_h) \) times \( T \) and the ratio of this deepening for \( U \) + \( \delta U \) versus \( U \) is in fact very similar:

\[
\frac{(dh/dt - \bar{\omega}_h)_{U+\delta U}}{(dh/dt - \bar{\omega}_h)_{U}} = \frac{U}{U+\delta U} = \frac{10}{10+4} = \frac{5}{6} \tag{26}
\]

Differences \( h \) observed at a single point downstream are therefore likely small, unless of course they are accompanied by changes in other parameters such as subsidence.

Indeed, the simulations do not take into account the SST gradients and changes in subsidence present along trade-wind trajectories. Although these are important for the absolute response, the question of relevance here is mainly how these co-vary with wind speed. For instance, we may hypothesize that a stronger large-scale circulation...
implies stronger trade winds as well as stronger subsidence, which would significantly reduce the deepening of the layer at strong winds compared to weak winds. We may also expect that stronger north-south SST gradients, present at times of stronger (zonal) surface wind speeds, relate to more wind shear (from the thermal wind equation) and therefore little difference in the depth of the layer over which easterly winds prevail.

Rainfall is known to reduce the deepening rate of the layer (Stevens 2007; Stevens and Seifert 2008; Van Zanten et al. 2010) and is neglected in the simulations performed here. Including rain microphysics likely reduces the differences in boundary layer depth between the different wind speed cases. Evaporation of rainfall within the sub-cloud layer in turn can lower the surface moisture flux and increase the surface heat flux. This effect on the surface fluxes may be further amplified by less deepening of the boundary layer when it rains more, which implies less warming and drying due to downward mixing of overlying free tropospheric air. Rainfall during RICO was on average 30 W m$^{-2}$, roughly half of the change in $F_{\text{q,s}}$ for 6U = 3 to 5 m s$^{-1}$, so that rain effects may not be insignificant. Note that rainfall would bring the simulations closer to the observations, which show little change in the surface fluxes, and less evident differences in boundary layer depth (Nuijens et al. 2009).

6. Conclusions

The influence of wind speed on shallow cumulus convection and the trade-wind layer is explored using Large Eddy Simulations and bulk equilibrium theory. Beginning with a basic state that is characteristic of the trades, we find that:

i. The deepening of the cloud layer with a strengthening of the winds is a necessary part of the adjustment of the trade-wind layer to a new equilibrium.

ii. This deepening is required because the sub-cloud layer depth experiences contradicting tendencies when wind speed changes. Thermodynamically, an increase in wind speed tends to lower cloud base height, when associated with the saturation level. Dynamically, an increase in wind speed tends to increase cloud base height, when associated with the height where the buoyancy flux falls to a fixed fraction of its surface value.

iii. At stronger winds clouds are deeper, but not more numerous or more energetic. This is because the equilibrium surface buoyancy flux varies little with wind speed, as the surface heat and moisture flux respond in opposite ways to increasing winds.

iv. The adjustment towards a deeper cloud layer is accomplished by the transient response of the mass flux, which initially increases with wind speed as the surface moisture flux increases. The deepening brings drier and warmer air to the surface, which maintains a larger input of moisture at the surface, but reduces the surface sensible heat fluxes. The surface buoyancy fluxes are hence reduced and the initial increase in cloud base mass fluxes is damped.

v. The equilibrium response of the surface fluxes weakly depends on the thermodynamic structure of the free troposphere.

vi. Equilibrium solutions in which boundary layer growth is zero appear present for only a narrow range of wind speeds.

vii. Our analysis follows from (simple) bulk concepts, which suggests that the internal dynamics of the layer i.e., how clouds are mixing with the environment, are only secondary to the deepening response to a change in wind speed (or a change in surface forcing in general).

A key finding of our work is that the equilibrium trade-wind layer is deeper at stronger winds. Differences in the depth of the boundary layer however may be less apparent when observed at a single point downstream, that is, assuming the layers reach equilibrium. Larger wind speeds imply less travel time, hence less time to develop a deeper layer. Deeper clouds may also rain more along their way, limiting boundary layer growth. Moreover, as wind speed increases, shear becomes increasingly difficult to neglect and it may mediate the further response of the layer. Another point worth investigating is whether wind speed generally co-varies with subsidence or perhaps the thermodynamic structure of the free troposphere i.e., other forces part of the large-scale circulation may either amplify or reduce the effect of wind speed. Such interactions will determine whether the wind speed effect is felt just locally, on cloud and boundary layer dynamics, but also deeper into the tropics, feeding back to the large-scale circulation.

When moving on to more realistic (and more complex) studies, that focus on interactions between clouds, meteorology and precipitation, the influence of wind speed on the aerosol and hence the microphysical development of clouds need to be considered. Of relevance is the positive correlation between strong winds and concentrations of sea-salt aerosol (Clarke et al. 2003; Woodcock 1953), that is attributed to sea spray from breaking waves. In addition the apparent result that stronger winds correspond to stronger cloud base updrafts during most of the RICO flights (Colón-Robles et al. 2006), a result for which we do not find evidence in our simulations, requires attention. Such an effect would influence peak supersaturations in moist updrafts and hence cloud droplet concentrations and rain formation.
The trades are a sensitive regime, in which small variations in the structure of the boundary layer as well as in the forces that act upon it may have significant effects. This may come as no surprise to the casual observer within the tropics, where small changes in the mean state of the atmosphere can be accompanied by substantial changes in the structure and organization of clouds and precipitation.

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Fig. 5. Time series before and after the wind speed perturbation for case $S_{7.5}$. Panels show (from top to bottom): the surface sensible heat flux $F_{\theta,s}$, the surface latent heat flux $F_{q,s}$, the surface buoyancy flux $F_{b,s}$, the boundary layer height $h$, the deepening tendency $dh/dt - \bar{w}_h$, the fraction of cloudy columns $cc$ and the mass flux at cloud base $M$ for simulations $U_{10}$ (blue), $U_5$ (grey) and $U_{15}$ (dark blue).
Fig. 6. Profiles of (a) liquid water potential temperature, (b) specific humidity, (c) liquid water $q_l$, (d) zonal and meridional wind speed $u, v$, (e) the turbulent heat flux $F_{\theta,s}$, (f) the turbulent moisture flux $F_q$, (g) the buoyancy flux $F_b$, (h) shear production of TKE $= \overline{u'w'}, \overline{v'w'}$, (i) cloud core fraction $a_{co}$, (j) cloud core vertical velocity $w_{co}$, (sk) mass flux $M$, (l) cloud core excess of virtual potential temperature $(\theta_{v,co} - \overline{\theta_v})$. Profiles are averaged over hour 100-108 for $U_5$ (grey), $U_{10}$ (blue) and $U_{15}$ (dark blue) for $S_{7.5}$, as well as for $S_{8.5}$ (orange, red, dark red).
Fig. 7. Cartoon illustrating the possible response of the $\theta$ and $q$ profiles and the heat $F_\theta$, moisture $F_q$ and buoyancy flux $F_b$ profile to a wind speed perturbation of $\delta U$, given the assumptions that the boundary layer does not deepen, that $\theta$ and $q$ gradients do not change, and that radiative cooling $Q_r$ does not change. The initial equilibrium state (here with $F_{b,s} = 0$) is indicated with dashed lines and the possible new equilibria with solid red and green lines. The meaning of the different situations and parameters is described in section 4.
Fig. 8. Idealized sketch of the moisture balance at the inversion $h$ for different wind speeds, denoted by $U - \delta U$, $U$, and $U + \delta U$. Top panels show simplified humidity profiles from Figure 6. Bottom panels show profiles of the turbulent moisture flux $F_{q,h}$ (black) and the large-scale drying flux $F_{q,LS}$ (red). The bottom left profile corresponds to an idealized subsidence profile.
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