Response of Atmospheric Convection to Vertical Wind Shear: Cloud-System-Resolving Simulations with Parameterized Large-Scale Circulation. Part I: Specified Radiative Cooling

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ABSTRACT

It is well known that vertical wind shear can organize deep convective systems and greatly extend their lifetimes. Much less is known about the influence of shear on the bulk properties of tropical convection in statistical equilibrium. To address the latter question, the authors present a series of cloud-resolving simulations on a doubly periodic domain with parameterized large-scale dynamics based on the weak temperature gradient (WTG) approximation. The horizontal-mean horizontal wind is relaxed strongly in these simulations toward a simple unidirectional linear vertical shear profile in the troposphere. The strength and depth of the shear layer are varied as control parameters. Surface enthalpy fluxes are prescribed.

The results fall in two distinct regimes. For weak wind shear, time-averaged rainfall decreases with shear and convection remains disorganized. For larger wind shear, rainfall increases with shear, as convection becomes organized into linear mesoscale systems. This nonmonotonic dependence of rainfall on shear is observed when the imposed surface fluxes are moderate. For larger surface fluxes, convection in the unsheared basic state is already strongly organized, but increasing wind shear still leads to increasing rainfall. In addition to surface rainfall, the impacts of shear on the parameterized large-scale vertical velocity, convective mass fluxes, cloud fraction, and momentum transport are also discussed.

1. Introduction

Environmental wind shear can organize atmospheric moist convection into structures that contain mesoscale motions with a range of horizontal scales, from a few kilometers to thousands of kilometers. Severe convective storms, supercells, squall lines, mesoscale convective systems (MCS), tropical cloud clusters, trade cumulus, tropical cyclones, and the Madden–Julian oscillation are all influenced by wind shear, in various ways (e.g., Cotton and Anthes 1989; Houze 1993). Understanding the role of the vertical wind shear in the formation of thunderstorms, squall lines, and other mesoscale features has been a longstanding problem, studied since the availability of upper-air soundings (e.g., Newton 1950). Simulations of squall lines with numerical models (e.g., Hane 1973; Moncrieff 1981; Thorpe et al. 1982; Rotunno et al. 1988; Weisman et al. 1988; Liu and Moncrieff 2001; Robe and Emanuel 2001; Weisman and Rotunno 2004) show that environmental wind shear is crucial to their organization. Rotunno et al. (1988) and Weisman et al. (1988), widely known as RKW and WKR, respectively, emphasized that the cold pool–shear interaction may greatly prolong the lifetimes of squall lines and enhance their intensities.

In an environment without shear, convective down-drafts bring low-entropy air from the midtroposphere...
down to near the surface, which then forms a surface-based cold pool, where it is detrimental to further convection. On the other hand, the cold pool can propagate horizontally as a density current, generating circulations that lift environmental boundary layer air to its level of free convection and triggering new convective cells. When the environment is sheared, the circulation associated with the shear may balance the circulation associated with the cold pool on the downshear side, promoting deeper lifting. The case where the cold pool is roughly in balance with the shear has been called the optimal state, meaning a state in which the system maintains an upright updraft and repeatedly generates new cells on its leading edge (Xu et al. 1996, 1997; Xue 2000; Robe and Emanuel 2001).

Despite its importance, the effect of vertical wind shear has not been included explicitly in parameterization schemes for large-scale models. This may cause biases in quantities that directly regulate energy balance, such as cloud fraction, optical depth, and radiative fluxes. These quantities are related to convective organization, which is modulated by shear (e.g., Liu and Moncrieff 2001). The aim of this study is not to simulate squall lines or other specific mesoscale storm types per se, but rather to investigate the influence of vertical shear on mean precipitation and convective organization in the tropics in statistical equilibrium. We investigate this in cloud-resolving simulations of a small tropical region in which the interaction of that region with the surrounding environment is represented through a simple parameterization of the large-scale circulation.

When the large-scale circulation is parameterized, the model itself can determine the occurrence and intensity of deep convection. This is as opposed to more traditional methods, in which otherwise similar numerical experiments are performed with specified large-scale vertical motion, strongly constraining the bulk properties of convection a priori (e.g., Mapes 2004). The large-scale parameterization here is the weak temperature gradient (WTG) method (e.g., Sobel and Bretherton 2000; Raymond and Zeng 2005; Wang and Sobel 2011, 2012; Wang et al. 2013). In this method, horizontal-mean temperature anomalies in the free tropical troposphere, relative to a prescribed target profile, are strongly relaxed toward zero. This represents the effects of compensating vertical motion, communicated to large distances by gravity waves (e.g., Bretherton and Smolarkiewicz 1989) so as to maintain weak horizontal pressure and temperature gradients. The parameterized adiabatic cooling by the large-scale vertical velocity approximately balances the diabatic heating explicitly simulated by the cloud-system-resolving model (CRM). Parameterizations of large-scale dynamics have been used for a range of idealized calculations. Among these, Sessions et al. (2010) studied the response of precipitation to surface horizontal wind speed with a fixed sea surface temperature (SST) using a 2D CRM under WTG. They demonstrated the existence of multiple equilibria corresponding to precipitating and non-precipitating states for the same boundary conditions but different initial conditions, corroborating the findings of Sobel (2007) in a single-column model with parameterized convection. Wang and Sobel (2011) showed the equilibrated precipitation as a function of SST using both a 2D and a 3D CRM. Both of these studies show monotonic increases in surface precipitation rate (in the precipitating state, where it exists) with either increasing surface wind speed at fixed SST or vice versa. Here we extend these studies to examine the response of convection to vertical shear of the horizontal wind.

This paper is organized as follows. We describe the model setup and experiment in section 2. In section 3 we show how convective organization, mean precipitation, and other thermodynamic and dynamic quantities vary with the magnitude of shear when the depth of the shear layer is fixed and approximately equal to the depth of the troposphere. In section 4 we show the effect of variations in the depth of the shear layer. We conclude in section 5.

2. Experiment design and model setup

a. Experiment design

The model used here is the Weather Research and Forecast Model (WRF), version 3.3, in three spatial dimensions, with doubly periodic lateral boundary conditions. The experiments are conducted with Coriolis parameter $f = 0$. The domain size is $196 \times 196 \times 22$ km$^3$ and the horizontal resolution is 2 km. We use 50 vertical levels, with 10 levels in the lowest 1 km. Vertically propagating gravity waves are absorbed in the top 5 km to prevent unphysical wave reflection off the top boundary by using the implicit damping vertical velocity scheme (Klemp et al. 2008). Microphysics scheme is the Purdue–Lin bulk scheme (Lin et al. 1983; Rutledge and Hobbs 1984; Chen and Sun 2002). This scheme has six species: water vapor, cloud water, cloud ice, rain, snow, and graupel. The 2D Smagorinsky first-order closure scheme is used to parameterize the horizontal transports by subgrid eddies. The surface fluxes of moisture and heat are parameterized following Monin–Obukhov similarity theory. The Yonsei University (YSU) first-order closure scheme is used to parameterize boundary layer turbulence and vertical subgrid-scale eddy diffusion.
In this scheme nonlocal counter gradient transport (Troen and Mahrt 1986) is represented, and the local Richardson number, temperature, and wind speed determine the depth of the boundary layer.

Radiative cooling is set to a constant rate of $1.5 \text{ K day}^{-1}$ in the troposphere. The stratospheric temperature is relaxed toward 200 K over 5 days, and the radiative cooling in between is matched smoothly depending on temperature as in Pauluis and Garner (2006):

$$Q_{\text{rad,cooling}} = \begin{cases} -1.5 \text{ K day}^{-1} & \text{for } T > 207.5 \text{ K} \\ 200 - T & \text{elsewhere} \end{cases} \frac{5 \text{ K day}^{-1}}{5 \text{ days}}$$

The vertical wind shear is maintained by a term in the horizontal momentum equation that relaxes the horizontal-mean zonal wind to a prescribed profile $U(z)$ with a relaxation time scale of 1 h. Figure 1a shows $U(z)$ for a constant shear-layer depth of 12 km and varying shear magnitude, specified by varying the wind speed at the top of the shear layer from 10 to 40 m s$^{-1}$ in increments of 10 m s$^{-1}$ while maintaining $U = 0$ m s$^{-1}$ at the surface. We refer to these experiments as U0, U10, U20, U30, and U40. Figure 1b shows profiles in which we vary the depth of the shear layer while fixing the wind speed at the top of the layer at 20 m s$^{-1}$. In all cases, we relax the horizontal-mean meridional wind to zero.

Surface fluxes of sensible heat (SH) and latent heat (LH) are prescribed and held constant at each grid point and each time step. Three sets of values are used for the total heat flux and its individual components: 160 W m$^{-2}$ (low), partitioned as $SH = 15$ W m$^{-2}$ and $LH = 145$ W m$^{-2}$; 206 W m$^{-2}$ (moderate), partitioned as $SH = 22$ W m$^{-2}$ and $LH = 184$ W m$^{-2}$; and 280 W m$^{-2}$ (high), partitioned as $SH = 30$ W m$^{-2}$ and $LH = 250$ W m$^{-2}$. These represent a range of deviations from the vertically integrated radiative cooling, which is 145 W m$^{-2}$. We consider only positive deviations from the radiative cooling, as we are most interested in equilibrium cases featuring significant convection. The parameterized circulations in our model tend to result in positive gross moist stability, so that the mean precipitation rate increases with the excess of surface fluxes over radiative cooling (e.g., Wang and Sobel 2011, 2012).

We prescribe surface fluxes, rather than just SST (as in previous work) because for different shear profiles, convective momentum transport induces significant changes in surface wind speeds, even with a strong relaxation imposed on the surface winds. This causes surface fluxes to vary significantly with shear if the fluxes are calculated interactively. Varying fluxes by themselves will change the occurrence and properties of deep convection, apart from any effects of shear on convective organization. We wish to completely isolate the direct effect of shear, without considering this modulation of surface fluxes. Also, prescribing surface fluxes will eliminate any feedback from cold pools.

The model is run for 40 days for each simulation, and the output data are sampled every 12 h. After the first 5 days, the simulations reach approximate statistical equilibrium (see Fig. 3 below). The analysis is performed over this equilibrium period only, neglecting the first 5 days.

### b. WTG

We use the WTG method to parameterize the large-scale circulation, as in Wang and Sobel (2011), whose methods in turn are closely related to those of Sobel and Bretherton (2000) and Raymond and Zeng (2005). Specifically, we add a term representing large-scale vertical advection of potential temperature to the thermodynamic equation. This term is taken to relax the horizontal-mean potential temperature in the troposphere to a prescribed profile:

$$\frac{\partial \theta}{\partial t} + \cdots = -\frac{\bar{\theta} - \theta_{\text{RCE}}}{\tau},$$

where $\theta$ is potential temperature, $\bar{\theta}$ is the mean potential temperature of the CRM domain (the overbar indicates the CRM horizontal domain average), and $\theta_{\text{RCE}}$ is the target potential temperature taken from a 6-month-long radiative–convective equilibrium (RCE) run. Based on previous work in this configuration with this model, we
are confident that the initial conditions do not influence the statistical properties of the results, except that we avoid any possible dry equilibrium state (Sobel 2007; Sessions et al. 2010) by starting with a humid, raining state. Also $\tau$ is the Newtonian relaxation time scale, taken to be 3 h in our simulations. As $\tau$ approaches zero (a limit one may not be able to reach owing to numerical issues), this becomes a strict implementation of WTG and the horizontal-mean free-tropospheric temperature must equal $\theta_{\text{RCE}}$. In general, $\tau$ is interpreted as the time scale over which gravity waves propagate out of the domain, thus reducing the horizontal pressure and temperature gradients (Bretherton and Smolarkiewicz 1989). Finite $\tau$ allows the temperature to vary in response to convective and radiative heating.

The large-scale vertical circulation implied by this relaxation constraint is the WTG vertical velocity:

$$W_{\text{WTG}} \frac{\partial \bar{\theta}}{\partial \eta} (-\rho g/\mu) = \frac{\bar{\theta} - \theta_{\text{RCE}}}{\tau}, \quad (3)$$

where $\rho$ is the density, $\eta$ is the mass-based vertical coordinate of WRF, and $-\rho g/\mu$ is part of the coordinate transformation from $z$ to $\eta$ [$\eta = (p_d - p_T^d)/\mu$, where $p_d$ is the dry pressure, $p_T^d$ is a constant dry pressure at the model top, and $\mu$ is the dry column mass]. Within the boundary layer, following Sobel and Bretherton (2000) we do not apply (3) but instead obtain $W_{\text{WTG}}$ by linear interpolation from the surface to the PBL top. Unlike in previous studies with our implementation of WTG, the PBL top is not fixed but is diagnosed in the boundary layer parameterization scheme (discussed below). The scheme determines a PBL top at each grid point, and we use the maximum value in the computational domain at each time step as the PBL top for the computation of $W_{\text{WTG}}$.

Transport of moisture by the large-scale vertical motion introduces an effective source or sink of moisture to the column. The moisture equation is updated at each time step by adding the following terms associated with the WTG vertical velocity:

$$\frac{\partial q}{\partial t} + \cdots = -W_{\text{WTG}} \frac{\partial \eta}{\partial \eta} (-\rho g/\mu), \quad (4)$$

where $q$ is the moisture mixing ratio. The right-hand side of (4) is the advection by the large-scale vertical velocity. We assume that the moisture field is horizontally uniform on large scales, thus neglecting horizontal advection.

3. Response to a shear layer of fixed depth

a. Convective organization and precipitation

Figure 2 shows randomly chosen snapshots of hourly surface rain for different values of shear strength and surface fluxes. In the case of low surface fluxes (Fig. 2a), convection in the unsheared flow is random in appearance and has a “popcorn” structure. For small shear (cases U10 and U20), convection starts to organize but
does not maintain a particular pattern. In some snapshots it has lines normal to the shear direction, while in others it loses this structure and appears to form small clusters. For strong shear, lines of intense precipitation form parallel to the shear direction.

In the case of moderate surface fluxes (Fig. 2b), the unsheared flow’s convection is random and distributed uniformly across the domain but less so than in the case of low surface fluxes, as there are arcs and semicircular patterns. Under weak shear (case U10), there are organized convective clusters in part of the domain. As the shear increases further, lines of intense precipitation (brown shading) are trailed by lighter rain (blue shading), which propagate downshear (eastward). The organization in all cases is three dimensional.

For high surface fluxes (Fig. 2c), convection is organized in linear forms in the absence of vertical wind shear, loses linear structure with moderate shear (U10), and transforms into “aggregated” states in which intense precipitation only occurs in a small part of the domain. Moncrieff and Liu (2006) simulated three-dimensional propagating shear-perpendicular MCSs in unidirectional quasi-constant vertical shear. Mesoscale downdrafts/density currents were important, and synoptic waves played a part in the intermittent occurrence of the MCS episodes. The small domain setup under WTG approximation in this study and lack of synoptic-scale variability likely suppress intermittency and may explain the lack of shear-perpendicular systems as prevalent convective regime in our simulations.

Shear-parallel bands were observed in the eastern tropical Atlantic during Global Atmospheric Research Program (GARP) Tropical Atlantic Experiment (GATE) and were reproduced in simulations of that experiment by Dudhia and Moncrieff (1987). Although there are a number of differences between those simulations and ours, there are a number of similarities between their results and those found here, including downgradient convective momentum transport discussed later in section 3c(3).

To further quantify the impact of the shear on precipitation, Fig. 3 shows time series of the domain-mean daily rain rate for the different shear strengths. Statistical equilibrium is reached in the first few days. The magnitude of the temporal variability is minimized in the unsheared case and increases with increasing surface fluxes, while it is maximized in the cases U30 and U40 for low surface fluxes (Fig. 3a) and in U20 for higher surface fluxes (Figs. 3b,c). In many of the simulations the variability is quasi periodic, with periods on the order of
days. We are interested here primarily in the time-averaged statistics and have not analyzed these oscillations in any detail.

To define a quantitative measure of convective organization, we first define a “blob” as a contiguous region of reflectivity greater than 15 dBZ in the vertical layer from 0 to 2 km (Holder et al. 2008). Figure 4a shows a normalized probability density function (PDF) of the total (as the sum of all times for which we have data) number density of blobs, and Fig. 4b shows the normalized PDF of the total area of blobs in the domain for the moderate surface fluxes case. The number of blobs decreases as convection clusters into aggregated structures with stronger shear. The areal coverage of convection increases with stronger shear as the tail of the PDF spreads toward larger areas. The other cases of surface fluxes are qualitatively similar and not shown.

b. Mean precipitation, thermodynamic budget, and large-scale circulation

1) MEAN PRECIPITATION

A key question that we wish to address is the dependence of the mean precipitation in statistical equilibrium as a function of the shear. Figure 5 shows the domain- and time-mean precipitation as a function of
shear for three cases of surface fluxes as a direct model output (red) and as diagnosed from the energy budget (blue) as we will discuss in section 3b(4). The most striking feature is that for low and moderate surface fluxes, the mean precipitation is a nonmonotonic function of the shear (Figs. 5a,b). For low surface fluxes, small shear brings the precipitation below that in the unsheared case, achieving its minimum at U20. For strong shear (U30 and U40), however, precipitation increases not only relative to the unsheared case but also above the surface fluxes, which indicates moisture import by the large-scale circulation. For moderate surface fluxes, the structure of the nonmonotonicity differs from that in the low-surface-fluxes case. The minimum precipitation now shifts to U10, above which the behavior is monotonic. Again, strong shear is required to bring precipitation above that in the unsheared case.

For high surface fluxes, the behavior is monotonic, and small shear suffices to bring precipitation above that in the unsheared case. The difference between the minimum and maximum precipitation in the high-surface-flux case exceeds that in the lower-surface-flux cases, which is also (as one would expect) manifested in parameterized $W_{WTG}$ (as we will show in Figs. 8c,f). There is no obvious relationship, however, between mean precipitation and organization. While small shear can organize convection from a completely random state, the mean of that more organized convection need not be larger than that in the unsheared random state (Figs. 2a,b).

To examine the role of the cold pool on the relationship between mean precipitation and shear, we have also performed a set of similar simulations but with cold pools suppressed by setting evaporation of precipitation to zero at levels below 1000 m. Although, as expected, mean precipitation increases in these simulations relative to those in which precipitation can evaporate at all altitudes, the dependence of mean precipitation on the shear remains the same (not shown). This suggests that the cold pools are unlikely to be a key factor controlling the mean precipitation–shear relationship.

2) MOIST STATIC ENERGY BUDGETS

The moist static energy (MSE) budget can be a useful diagnostic for precipitating convection as MSE is approximately conserved in adiabatic processes. It is the sum of thermal, potential, and latent heat terms:

$$h = c_p T + gz + L_q q,$$

where $c_p$, $L_q$, and $q$ are the heat capacity of dry air at constant pressure, latent heat of condensation, and water vapor mixing ratio, respectively.

Fig. 5. Time- and domain-mean model output precipitation (red), derived precipitation (blue), and surface fluxes (black) as a function of the shear: (a) low, (b) moderate, and (c) high surface fluxes.
To define the basic state from which perturbations will be computed, the equilibrium vertical profiles of temperature, water vapor mixing ratio, and MSE for the unsheared case of moderate surface fluxes of 206 W m$^{-2}$ are shown in Fig. 6 (other unsheared cases for low and high surface fluxes are very similar). Because of the fixed radiative cooling, the tropopause is near 14 km, and the melting level is near 4 km.

Figure 7 shows the time- and domain-mean profiles of temperature, water vapor mixing ratio, and moist static energy, all expressed as perturbations from the same quantities in the unsheared experiment. There is a weak sensitivity to the shear in the temperature profile for low surface fluxes, becoming larger as surface fluxes increase. Recall that the temperature is continually being strongly relaxed back toward the target profile. The temperature perturbation is positive in the middle to upper troposphere and negative in the boundary layer in the strong-shear case.

The moisture profile is more sensitive to the shear than temperature is, but varies little with increasing surface fluxes. Above the boundary layer the strong shear case is moister than the unsheared case by about 0.2 g kg$^{-1}$ and drier in the boundary layer by about 0.5 g kg$^{-1}$. The weak-shear case is drier than the unsheared case at all levels for all three cases of surface fluxes. This increasing dryness with shear is also apparent in the snapshots of surface rainfall (Fig. 2). The moist static energy variations with shear are dominated by the moisture term, are very similar for all three cases of surface fluxes, and depend nonmonotonically on the shear. The nonmonotonicity remains the same for all cases of surface fluxes, unlike the large-scale vertical velocity and mean precipitation.

3) LARGE-SCALE CIRCULATION

Figure 8 shows the vertical profiles of large-scale vertical velocity (Figs. 8a–c) and its maximum value with respect to the vertical (Figs. 8d–f) for the three cases of surface fluxes. For low surface flux, $W_{\text{WVG}}$ is very weak (less than 1 cm s$^{-1}$), and its absolute value varies little with shear. As the surface fluxes increase, $W_{\text{WVG}}$ increases, but the effect of the shear is much more significant in the highest surface fluxes case, where $W_{\text{WVG}}$ increases by 4 cm s$^{-1}$ from the unsheared case to the case with the strongest shear.

4) NORMALIZED GROSS MOIST STABILITY

To diagnose the mean precipitation from the point of view of the moist static energy budget, we use the diagnostic equation for precipitation as in, for example, Sobel (2007) or Wang and Sobel (2011); see also Raymond et al. (2009):

![Fig. 6. Vertical profiles of (top) temperature, (middle) water vapor mixing ratio, and (bottom) MSE for the unsheared case of moderate surface fluxes.](image-url)
where \( \langle \cdot \rangle = \int_{Z_0}^{Z_H} \rho dZ \) is the mass-weighted vertical integral from the bottom to the top of the domain and \( P, E, H, \) and \( Q_R \) are precipitation, latent heat flux, sensible heat flux, and radiative cooling rate, respectively. We have also utilized the large-scale vertical velocity and the moist static energy introduced in the above two subsections to define \( M = \langle W_{W_aTG}(\partial h/\partial z) \rangle / \langle W_{W_aTG}(\partial s/\partial z) \rangle \) as the normalized gross moist stability, which represents the export of moist static energy by the large-scale circulation per unit of dry static energy export. Here, \( s \) is the dry static energy (sum of the thermal and potential energy only, without the latent heat term), and the overbar is the domain mean.
The above equation is not predictive because we use the model output to compute the terms; in particular, we do not have a theory for $M$. Nonetheless, it gives a quantitative, if diagnostic, understanding of the role played by large-scale dynamics in controlling precipitation. We must first verify that our moist static energy budget closes well enough that the precipitation calculated from (6) agrees well with the model output precipitation. This is the case, as shown in blue in Fig. 5.

As shown in Fig. 5, the variations of $M$ with respect to shear (themselves shown in Fig. 9) explain those in the mean precipitation. Since both surface fluxes and radiative cooling are constant in each plot in Fig. 5, variations in $M$ must correspond to variations in dry static energy export. Since dry static energy export changes can only be balanced by changes in the rate of condensation (radiative cooling, again, being fixed) and, hence, in the precipitation rate. For the case of low surface fluxes, $M$ exceeds 1 for the cases U10 and U20; this is because of the low-level subsidence and upper-level ascent in the $W_{WTG}$ profile, with midlevel inflow bringing in relatively low-moist static energy air.

c. Convective cloud properties

1) Cloud Fraction

The vertical profile of cloud fraction for the three values of surface fluxes is shown in Fig. 10. The cloud is
defined as the set of grid points where the mixing ratio of cloud hydrometeors (water and ice) is greater than 0.005 g kg$^{-1}$. Generally, there are three peaks: a shallow peak at 1 km, a midlevel peak related to detrainment at the melting level at about 4 km, and a peak associated with deep convection at 9–12 km. When the cloud fraction in the deep convective peak becomes large, we interpret it as stratiform cloud.

Fig. 9. Normalized gross moist stability as a function of the shear: (a) low, (b) moderate, and (c) high surface fluxes.

Fig. 10. Cloud fraction: (a) low, (b) moderate, and (c) high surface fluxes.
In all three cases, the shear has almost no effect on the shallow convection. For low surface fluxes (Fig. 10a), the effect of strong shear is seen mainly in the midlevels, where cloud fraction increases from 30% in the unsheared flow to 60%. A modest stratiform layer forms as cloudiness increases from 10% to 30%.

For moderate surface fluxes (Fig. 10b), the strong shear, while causing an increase of more than 40% in the midlevel peak over the unsheared case, also leads to 100% cloud cover in the layer from 9 to 12 km. It is also interesting to notice a transition from a single peak in the high clouds at 9 km in the unsheared and weak shear cases to a cloudy layer from 5 to 12 km deep in the strong shear. In Fig. 10c near-total deep cloudiness is determined by the strong, persistent convection associated with the high surface fluxes. The shear has no effect on the cloud fraction at upper levels, since it is 100% even with no shear. The effect of shear on midlevel cloudiness is clear, however, as cloud fraction increases over 35% from the unsheared to the strongest shear flow. The cloud cover in the midlevel increases monotonically with shear, but the deep cloudiness does not (Figs. 10a,b), behaving instead more like mean precipitation.

2) CONVECTIVE MASS FLUX

Figure 11 shows the convective mass fluxes in both updrafts and downdrafts. Updrafts are defined here as including all grid points with liquid water and ice content greater than 0.005 g kg\(^{-1}\) and vertical velocity greater than 1 m s\(^{-1}\), normalized by the total number of grid points. The downdrafts are defined including all grid points with vertical velocity less than −1 m s\(^{-1}\). The updraft mass fluxes are about twice as large as the downdraft mass fluxes. Both have two peaks, in the lower and upper troposphere, both of which strengthen by about a factor of 3 from lowest to highest surface fluxes.

Despite different convective organization for different shear profiles, the vertical profiles of the updrafts and downdrafts have similar shapes. For the downdrafts, the lower-tropospheric peak exceeds the upper-tropospheric one for all three cases of surface fluxes and is similar for different shear strengths. The largest difference occurs in the case of highest surface fluxes, in which the downdraft in the lower peak for the smallest shear is stronger than that for the strongest shear by about 20%. The effect of the shear on the upper-tropospheric peak, however, is more prominent. The strongest shear causes the downdraft mass flux to strengthen by 50%.

The updraft mass fluxes in the low-surface-fluxes case are greater in the lower troposphere than in the upper troposphere, because of the dominant role of the shallow...
convection, while in the moderate-surface-fluxes case both peaks are almost comparable, and in the high-surface-fluxes case the upper peak is larger owing to deep clouds. Although with high surface fluxes there is an increase of 25% in the upper-tropospheric updraft from the weakest to strongest shear, the effect of the shear on the updraft is smaller than in the downdraft and has negligible impact on the lower-tropospheric peak, and, unlike the downdraft, is nonmonotonic, similarly to cloud fraction.

3) MOMENTUM FLUXES

Another collective effect of the cloud population and its organization on the large-scale flow results from vertical fluxes of horizontal momentum, including both convective momentum transport and gravity wave momentum transport (Lane and Moncrieff 2010; Shaw and Lane 2013). We evaluate these momentum fluxes as a function of the shear strength. Figure 12 shows the vertical profiles of the time- and domain-mean zonal momentum fluxes $\rho u'w'$ where $u'$ and $w'$ are the zonal and vertical velocity perturbations from the horizontal average, respectively, and the overbar is the horizontal mean. The magnitude of the net momentum flux is a monotonic function of the shear. It is negligible for all unsheared cases, as expected, and negative (directed down the gradient of the mean zonal wind) and peaks close to the top of the shear layer for the sheared cases. This negative momentum flux would act (if it were not strongly opposed by the imposed relaxation) to accelerate the mean horizontal wind in the lower troposphere and decelerate it in the upper troposphere and the stratosphere. The amplitude of the momentum transport also increases with surface fluxes. Interestingly, for the highest surface fluxes, the momentum flux in the U10 case (Fig. 12c) is upgradient in the lower troposphere and peaks at about 4 km, while it is downgradient in the upper troposphere.

In the 2D analytic steering-level model of Moncrieff (1981), the updrafts tilt downshear and precipitation falls into the updrafts, weakening the cold pool and causing upgradient momentum transport. With the exception of the lower troposphere in the U10 case for the largest surface fluxes, our momentum fluxes are primarily downgradient throughout the troposphere. This may be due to the fact that none of our simulations show truly long-lived, quasi-2D shear-perpendicular squall lines as in Moncrieff’s model, as discussed further below.

4. Response to different shear depths

In this section we investigate the effect of varying the depth of the shear layer. We repeat the above
calculations but with shear depths of 1500, 3000, 4500, 6000, and 9000 m. In each case, the wind speed is held fixed at 20 m s$^{-1}$ at the top of the layer (Fig. 1b). Surface latent and sensible heat fluxes are held fixed at the intermediate values from the previous section, totaling 206 W m$^{-2}$.

As of convective organization, Fig. 13 shows snapshots of hourly surface precipitation for a period of seven consecutive hours, chosen from the last 7 h of each simulation. Each row corresponds to a different shear case: 1500, 3000, 4500, 6000, and 9000 m (from top to bottom). Most of these snapshots are a mixture of linear squall lines and supercell-like 3D entities, which have strong rotational components and tend to split into left and right movers (e.g., the 4500-m case in the third row of Fig. 13). For shallow shear of depth 1500 m, convection is to some extent organized in lines normal to the shear (as seen in some of the snapshots). Nonetheless, the shear-perpendicular squall lines prevalent in analytical models (e.g., Moncrieff 1992) and 2D numerical simulations (e.g., Liu and Moncrieff 2001) do not become fully established even in the shallow-shear case. Midlevel shear (3000–4500 m) is associated with quasi-linear structure parallel to the shear with intense precipitating cores trailed by lighter precipitation. As the shear depth increases above that—approaching the deep-shear cases analyzed in the preceding sections—the quasi-linear regime is still less prominent. For deeper shear of 9000 m, convection aggregates in clusters.

Figure 14 shows time- and domain-mean precipitation as a function of the shear depth. The shallowest shear layer, with depth 1500 m, produces the least precipitation; in this sense, these intermediate shear-layer depths are “optimal.” Although the convection is in statistical equilibrium, this is in agreement with studies of transient storms of Weisman and Rotunno (2004). While it was not the case that the optimal state produces the greatest surface precipitation in their original papers and later work, Bryan et al. (2006) demonstrated that optimal state is indeed associated with the greatest precipitation in several meso-scale models. Bryan et al. (2006) further attributed this discrepancy to numerical artifacts in the model used by Weisman and Rotunno (2004). Increasing the shear depth above 4500 m reduces the precipitation again, with the values for 6000- and 9000-m depths being smaller than those for the intermediate depths.

We have performed simulations analogous to those in Fig. 14 but with low and high surface fluxes (not shown). In these simulations, we find a similar dependence of the mean precipitation on the shear depths. The relationship shown in Fig. 14 between mean precipitation and shear depths appears to be qualitatively independent of the magnitude of the mean precipitation, at least in the regime studied here.

The vertical profiles of the large-scale vertical velocity as a function of the shear-layer depth are shown in Fig. 15. As we expect, the amplitude differences are largely consistent with those in the mean precipitation. The peak value, for instance, is smallest for the shallow-shear case and largest for the midlevel-shear case. The cases of 3000 and 4500 m are almost identical, while there is a small increase in precipitation for 4500 m over that at
3000 m. This is due to the more abundant moisture in the lower to midtroposphere for the case of 4500 m, as shown in Fig. 16. Dominated by moisture, moist static energy has a maximum for the 4500-m case.

Figure 17 shows the mass fluxes in updrafts and downdrafts for the different shear depths. Although the maximum mean precipitation occurs for the shear depth of 4500 m, the maximum updraft mass flux occurs at 3000-m shear depth. This demonstrates that convective mass flux need not vary in the same way as the precipitation rate. The response of the downdrafts is much weaker than that of the updrafts. In fact, the downdraft mass flux for the shallow shear is almost identical to that for the deeper shear layers with depths 6000 and 9000 m.

Finally, Fig. 18 shows the domain-averaged time-mean zonal momentum fluxes for different shear depths. Similar to the analytic model of Moncrieff (1992), momentum flux here is down the mean gradient for all shear depths, as in Fig. 12, and achieves the minimum at the top of the shear layer in each profile. This is likely due to momentum mixing, consistent with shear-parallel lines (Dudhia and Moncrieff 1987). The downward momentum transport is greatest when the shear is shallow: 1500 and 3000 m deep, and becomes smaller for deep shear (9000 m), similar to the cases with uniform shear depth (Fig. 12). This momentum flux dependence on the shear depth differs from the variations in the mean precipitation. As with other measures of organization, momentum flux apparently does not have a direct relationship to precipitation in our simulations.
5. Summary and discussion

We have quantified the effect of vertical wind shear on atmospheric convection in a series of 3D CRM simulations with large-scale dynamics parameterized according to the weak temperature gradient approximation. We varied the shear while holding surface heat fluxes fixed. We repeated the calculations for three different values of surface fluxes. As surface fluxes are increased, the strength of the simulated convection increases, as measured by domain-averaged precipitation. Precipitation increases more rapidly than the surface latent heat flux because of parameterized large-scale moisture convergence. The three cases thus correspond to weak, moderate, and strong convection.

The response of mean precipitation to shear in statistical equilibrium for the lower two surface flux cases is nonmonotonic in the magnitude of the shear, with minimum precipitation at an intermediate shear value. The shear value at which the minimum precipitation occurs is somewhat different for low and moderate surface fluxes. Only for strong shear does the precipitation exceed that in the unsheared flow. In the case of the highest surface fluxes used, the response to shear is monotonic and a small amount of shear is enough to cause the precipitation to exceed that of the unsheared flow.

The dependences of large-scale vertical velocity and the moist static energy on shear are both similar to the dependence of precipitation on shear. The dependence on shear of the normalized gross moist stability, which combines these two quantities in a way relevant to the column-integrated moist static energy budget, is found to explain the variations in mean precipitation with shear well.

Despite the relative smallness of the changes in the mean precipitation with shear, shear has a strong effect on convective organization. Weak shear is sufficient to organize convection even under weak surface fluxes when the large-scale vertical motion is negligible. Stronger shear can organize convection into convective clusters and squall-line-like structures with lines of intense precipitation trailed by lighter rain. As the surface fluxes and domain-averaged precipitation increase, however, convection is found to be organized even in the absence of vertical wind shear.

The wind shear also has a significant effect on cloud cover, which increases with the shear (especially high clouds), convective mass flux (in particular the downdraft mass flux), and momentum fluxes, as more momentum is transported downward when the shear increases.

When the total change in velocity over the shear layer is fixed but the layer depth is varied, midlevel-shear depths are found to be optimal for producing maximum mean precipitation, maximum large-scale vertical velocity, and maximum column moist static energy. Convective organization for this shear depth is very similar to that for deeper shear layers, but with more intense lines of precipitation. Confining the shear in a shallow
layer dries out the atmospheric column, leaving pre-
ceptualization at a minimum.

We have neglected the complications of cloud-radiation feedback by fixing the radiative cooling rate in a simple relaxation scheme. In reality, longwave radiative cooling can be suppressed by the abundant upper-tropospheric ice cloud in the deep tropics, which by itself is controlled by the shear. This effect will be investigated in a separate, future study.

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