Organization of Tropical Convection in Low Vertical Wind Shears: The Role of Water Vapor

ADRIAN M. TOMPKINS*
Max-Planck-Institut für Meteorologie, Hamburg, Germany

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ABSTRACT

A modeling study is conducted to gain insight into the factors that control the intensity and organization of tropical convection, and in particular to examine if organization occurs in the absence of factors such as vertical wind shear or underlying sea surface temperature (SST) gradient. The control experiment integrates a cloud-resolving model for 15 days using a 3D domain exceeding 1000 km in length, with no imposed winds, and horizontally uniform SST and forcing for convection. After 2 days of random activity, the convection organizes into clusters with dimensions of approximately 200 km. Convective systems propagate through the clusters at speeds of 2–3 m s$^{-1}$, while the clusters themselves propagate at minimal speeds of around 0.5 m s$^{-1}$.

Examining the thermodynamic structure of the model domain, it is found that the convective free bands separating the clusters are very dry throughout the troposphere, and due to virtual temperature effects, are correspondingly warmer in the lower troposphere and boundary layer. This suggests a positive feedback between convection and water vapor, where convective moistening of the local atmosphere renders it more favorable to future convection. The existence of this feedback is demonstrated by experiments in which the free-tropospheric water vapor is perturbed in convective regions, and it is found that the lower-atmospheric water vapor is most critical in controlling convection, most likely through the role of downdrafts. Examination of the boundary layer in the control experiment also indicated that convectively generated cold pools also play a key role in the organization of convection, possibly by their influence on the boundary layer water vapor field.

In order to see how the water vapor feedback modifies established convective organization, a further experiment was conducted with an SST gradient imposed, which established a mock Walker cell type circulation, with ascending motion over the warmest SSTs. After 5 days, the SST gradient is reversed to see how the convection would establish itself over the new SST maximum. This highly idealized experiment therefore represents a surrogate for the atmospheric response to SST "hotspots," that observations have shown to form under the descending branch of large-scale tropical circulations such as the Madden–Julian oscillation, due to increased incident solar radiation and decreased latent heat fluxes at the surface. It is found that the convection does not spontaneously initiate over the new SST maximum, but instead must propagate toward it. After a further 5 days, much longer than the boundary layer adjustment timescale, the warmest SSTs are still completely free from convection. This is directly due to the dryness of the atmosphere caused by the initial period of subsidence.

A further set of experiments examines the robustness of the feedback in cases of imposed vertical wind shear. It is found that strong wind shears prevent the feedback by effectively mixing water vapor. However, the feedback is still very important in cases of weak wind shears.

1. Introduction

In order to understand the tropical climate it is essential to gain an insight into the factors that control the intensity and organization of convection. The organization of convection has a large effect on the vertical transport of momentum, heat, and water vapor (Moncrieff and Klinker 1997), and coupling via surface wind stress can also influence tropical ocean dynamics (Webster 1994). For example, clouds in moist atmospheres have an increased precipitation efficiency (Anthes 1977; Raymond 1995). When convection is clustered, clouds must occur in a moister atmosphere compared to randomly distributed convection, and therefore the clustering must also reduce mean atmospheric moisture content [observed in simulations by Tompkins (2000)].

One simple measure often used as a predictor for the location and intensity of convective activity is Convective Available Potential Energy (CAPE) (e.g., Emanuel 1994), calculated by integrating the buoyancy of a hypothetical parcel up to its level of neutral buoyancy, and which can be related to the parcel’s theoretical maxi-
mum velocity. Maximum CAPE values are usually associated with parcels originating in the boundary layer, and thus CAPE depends only on the boundary layer equivalent potential temperature ($\theta_e$), and the atmospheric temperature profile (ignoring the virtual temperature and water loading effects of the environmental air). Thus one can expect the sea surface temperature (SST) to play an important role in the location and strength of convection since it directly affects the boundary layer temperature and moisture properties through the action of surface fluxes. Observations show that convection is strongly correlated with SST (e.g., Graham and Barnett 1987; Zhang 1993; Arking and Ziskin 1994). Similarly, large-scale atmospheric waves have also been suggested as having a role in suppressing or enhancing convection by altering the upper-tropospheric temperature profile (Emanuel et al. 1994). However, other factors that are not accounted for in the standard CAPE calculation can also alter the distribution of convection, either spatially or temporally. For example, vertical wind shear can organize convection into squall line systems and convective clusters (e.g., Rotunno et al. 1988; LeMone et al. 1998). Large-scale SST gradients or wind shears are frequently regarded as “external” organizing influences for atmospheric convection. This stems from the fact that our efforts to understand convection have frequently been motivated by the requirement to parameterize its effects in general circulation models (GCMs), which regard the large-scale flow and SST gradients as being well resolved by the GCM and more slowly evolving than convection.

However, it is possible that other process interactions exist that also cause convective organization. For example, Emanuel (1987) suggests a mechanism where convectively enhanced winds could increase local surface fluxes of heat and moisture and act to destabilize the large-scale circulation. On a smaller spatial scale, Tompkins and Craig (1998a) found in their cloud-resolving modeling study that wind sensitive surface fluxes were part of the cause for the localization of convection observed in their investigation of convecting quasi-equilibrium states. They also demonstrated that the horizontal variation in atmospheric radiative heating rates between the convective regions and their surrounding clear-air environment can act to strongly organize convection. The influence of cloud–radiation interactions on the local tropospheric dynamics was shown by Sherwood (1999) to increase convergence into cloudy areas in a simple model, which also appears to be observed over the Pacific (Bergman and Hendon 2000), and Raymond (2000a) has suggested that radiative–convective interactions could have a large influence on the Hadley and Walker circulations. Cold pools generated from convective downdrafts can also act to localize convection by lifting air through the negative inhibition layer and triggering new daughter cells (e.g., Simpson 1980; Thorpe et al. 1982; Fovell and Tan 1998). This issue of the importance of triggering energy has recently been focused upon by Mapes (1997), and Mapes (2000) has attempted to include this effect into a simple model of the Tropics.

There are therefore a number of thermodynamical and dynamical mechanisms that are candidates for convective organization, and which are difficult to isolate from the influence of SST gradients and wind shear in observations. To conduct an investigation into these effects it is therefore useful to use models so that controlled experiments can be conducted, removing the interaction with SST, for example, by imposing a horizontally homogeneous lower boundary condition. Su et al. (2000) have previously investigated the organization of convection using a mesoscale model operated with a three-dimensional domain exceeding 1000 km on each side, using horizontally homogeneous forcing and boundary conditions. They refer to the “spontaneous” manifestation of organization as “self-aggregation,” and they found that organization did indeed occur depending on the forcing profile used. However, the grid size of the mesoscale model (15 km) made the use of a convective parameterization scheme necessary.

It is not clear if mesoscale or global-scale models can resolve all of the process interactions that could cause organization or include representations of the processes in their parameterization schemes. Therefore, we use a cloud-resolving model (CRM), which resolves cloud-scale dynamics, to investigate this issue of convective organization further. The CRM is operated on a sufficiently large domain to reveal the development of mesoscale organization, and for a period of time sufficiently long enough to allow structure in the water vapor field to establish itself. The cloud model and setup used are described in the following section, and section 3 describes the organization that occurs in the main control experiment. On the basis of this experiment it is suggested that the organization that appears is the result of a positive feedback between water vapor and convection, where convective moistening of the atmosphere makes it more favorable to future convection. The evidence for this feedback from sensitivity tests is presented in section 4. Further experiments are then conducted in sections 5 and 6 to see how this feedback influences organization by SST gradients, and if it can operate in strong and weak wind shear conditions.

## 2. Model and setup

A detailed description of the CRM is contained in Tompkins and Craig (1998a) and references therein. The CRM includes a microphysical scheme that integrates prognostic equations for rain, snow, cloud water, cloud ice, and graupel amounts, and also the ice crystal concentration number (Brown and Swann 1997). The surface fluxes are provided using similarity theory that gives realistic fluxes in low wind conditions without the requirement for an artificial minimum surface wind speed.
Regarding the setup of the model domain, Tompkins (2000) demonstrated that two-dimensional simulations artificially impose a large-scale organization on convection. On the other hand, limited 3D domains of $O(100^2 \text{ km}^2)$ are not able to reveal organization of convection that may occur on the mesoscale or larger. Current computing resources are unfortunately not sufficient to allow simulations on domains of $O(1000^2 \text{ km}^2)$ with explicitly resolved convection, and therefore a compromise is necessary. The intermediate solution is to select a rectangular domain that is 1024 km long to allow organization to occur, but with a limited third dimension of 64 km. The horizontal resolution is 2 km in each direction, the horizontal boundary conditions are periodic, and no Coriolis force is applied. For future ease of notation, the 1024 km axis will sometimes be referred to as the $x$ axis.

It has recently been shown that a tropospheric resolution of at least 25 hPa is necessary to resolve the processes that determine the water vapor profile (Tompkins and Emanuel 2000), and therefore the model uses 50 vertical layers, with the resolution stretching from 100 m in the boundary layer to 500 m at a height of 6 km. A sponge layer is applied from a height of 17 km to the domain lid at 21 km, which damps all prognostic variables to their horizontal mean values.

Unless otherwise stated, the surface is assumed to be an ocean of fixed horizontally homogeneous temperature in all experiments. Observations and CRM studies have shown that the relationship between SST and convection is in fact due to the SST-gradient-induced circulation (Lau et al. 1994; Bony et al. 1997; Lau et al. 1997; Chaboureau et al. 1998; Tompkins and Craig 1999). Over a horizontally homogeneous surface, convection and cloud properties are relatively insensitive to the absolute value of SST (Tompkins and Craig 1999), which allows the arbitrary value of 300 K to be chosen without raising concerns that the model results will be dependent on the lower boundary condition imposed.

It has been demonstrated by Tompkins and Craig (1998a) that convection is strongly organized by radiative–convective feedbacks. Thus, to prevent this possibly swamping or obscuring other organizational processes, we apply a fixed horizontally homogeneous radiative forcing, consisting of a constant 2 K day$^{-1}$ cooling from the surface to 400 hPa, which is then linearly decreased to zero at 200 hPa. The simulation lasted for 15 days, enough to achieve a quasi-equilibrium state with the forcing imposed (Tompkins and Craig 1998b), and sufficient to allow any relevant organization to become apparent.

3. Control run

a. General organization

In order to examine any organization that may develop in the convection in the three-dimensional domain, Fig. 1 shows a Hovmöller diagram of surface rainfall during the entire experiment. The surface rainfall is summed across the short 64-km axis. After convection is initiated, it is seen to be randomly distributed for the first 2 days. After 2 days of simulation, the first signs of organization appear, with small areas free of convection. These grow in size until, by day 6, the organization becomes clearly visible. The domain is divided into regions of convective activity, that propagate in both easterly or westerly directions at speeds of approximately 0.5 m s$^{-1}$. In advance of the convection, there are bands that are completely free of convection, with widths varying between 50 and 120 km.

Within the convecting areas, lines of convection can be seen that propagate in the opposite direction at speeds between approximately 1.6 and 2.2 m s$^{-1}$. Since the graph has collapsed the 3D domain onto two dimensions, the true propagation speeds are likely to be higher since the convection will be propagating in directions nonparallel to the $x$ axis. The speed of propagation approximately corresponds to the average downdraft outflow velocity in 3D (Tompkins 2000).

b. Thermodynamic structure

In order to examine the thermodynamic structure more clearly, Figs. 2 and 3 show the water vapor and potential temperature perturbation, respectively, at var-
Fig. 2. Horizontal slices showing potential temperature perturbation in K about the mean value at each level. Slices are taken at heights of 50, 610, 3.5, 8.25, and 13.2 km, on day 15 of the control simulation.

Fig. 3. As for Fig. 2 but showing total water vapor mass mixing ratio, in kg kg$^{-1}$.

ious heights throughout the troposphere. The graphs are “snapshots” taken on day 15 of the simulation.

Examining the temperature graph, no distinct variations can be seen at 13.2 km. Excluding points found within convective updraft and downdraft cores, the temperature variation is everywhere less than $\pm 0.1$ K. Temperature variations are greater at lower levels in the atmosphere, with much structure apparent below 2 km. In particular, examining the panel for the height of 610 m, one can see that the convection free areas in Fig. 1 centered at 170, 450, and 730 km, respectively, correspond to areas where the air is significantly warmer. This feature is also visible at the bottom model level at 50 m, which also reveals the cold patches due to cold pool outflow from previous convective events.

In contrast, the plot of moisture reveals structure at all levels in the atmosphere. The upper panel clearly shows the local moistening effect of convection, with
large moist areas indicating the spreading anvil clouds where convection is detraining. At these levels, the differences between the dry and moist regions are huge; over one order of magnitude.

The large horizontal variation is the direct result of the very different timescales associated with the moistening and drying of the atmosphere by convective activity. The tendency of the clear air water vapor budget due to cumulus mass flux is given by

$\frac{\partial q}{\partial t} = (q_e - q) \frac{\partial M}{\partial z} - M \frac{\partial q}{\partial z}$, \hspace{1cm} (1)

where $M$, $q_e$, and $q$ represent the net cumulus mass flux, the in-cloud total water mixing ratio, and the environmental water vapor mixing ratio, respectively (e.g., Emanuel 1991). The first term on the right represents the moistening of the environment due to convective detrainment and the second term represents the drying due to the compensating subsidence in the environment. Since these two effects of convection are associated with different physical processes, each has its own horizontal propagation timescale. The drying is associated with the environmental subsidence, and spreads out at the propagation speed of gravity waves (Bretherton and Smolarkiewicz 1989). On the other hand, the atmospheric moistening by convective detrainment of water vapor and cloud condensate progresses at much slower advection speeds. Thus, even if an ensemble of cumulus clouds had no net effect on the water vapor at a particular height, they must moisten their local environment while drying more distance regions. This has been seen directly in tropical satellites observations of Liao and Rind (1997), who showed that in the first few hours after the onset of convection the local atmosphere dries, which they associated with the subsidence induced by convection, which is then replaced by a moistening trend, associated with the detrainment. Udellhofen and Hartmann (1995) also noted atmospheric moisture content dropping off with distance from convective clouds.

As stated earlier, the use of an interactive radiation scheme, that responded to water vapor and cloud condensate, would probably amplify the water vapor differences between the dry and moist regions, by causing additional convergence into convecting regions (Tompkins and Craig 1998a).

Examining the other height levels in Fig. 3, the dry bands are seen to extend throughout the troposphere into the boundary layer. Here, a circular pattern is visible, with cold pool dry patches surrounded by rings that appear to be moister than the boundary layer mean. Apart from new convective events, the majority of cold pools appear to reach a radius of about 10–20 km.

At lower levels there is a direct correspondence between the temperature and moisture perturbations, with the dry bands corresponding exactly with warm temperatures. The temperature at low levels simply reflects the virtual temperature effect of water vapor. If the buoyancy and virtual potential temperature fields are examined (not shown), they are found to be very uniform (with the obvious exception of convective cores themselves). This is expected, since gravity waves remove buoyancy perturbations very efficiently in the Tropics. The lack of temperature variation in the upper troposphere is simply a reflection of the small water vapor amounts there, which do not have a significant virtual temperature effect. Such virtual temperature perturbations associated with tropospheric dryness were observed by Mapes and Zuidema (1996).

The vertical structure of the bands in the water vapor is seen more clearly in Fig. 4, which displays the normalized water vapor perturbation, averaged across the short axis, which is defined as $q' / \sigma(q)$, where $\sigma(q)$ is the horizontal standard deviation of $q$ and $q'$ represents the perturbation about the horizontal mean. Snapshots are shown once a day for the last 5 days of the simulation. The overlaid contours represent the total cloud amount (liquid + ice), also averaged over the short axis, hence the use of a low contour value of $10^{-7}$ kg kg$^{-1}$. The high values of moisture associated with the cirrus detrainment zones are clearly visible at the 12–14 km height. The dark dry bands and adjacent convective zones are approximately uniform in width throughout the free troposphere, indicating very limited lateral mixing between the two zones. Not surprisingly, the regions that were very moist in the upper troposphere, for example, between 800 and 900 km on day 15, correspond to regions of very active deep convection. However, the dry bands are not totally free from convection. For example, the majority of the dry band centered at 730 km on day 15 is actually subject to shallow convection, with only a small region on its far left being totally convection free. The figure also shows how the two bands centered at approximately 200 and 760 km on day 11 on the simulation, propagate east and west, respectively, while the convective cluster that originated between them is suppressed sharply as the other clusters approach. There is evidence of a reverse circulation in the lower stratosphere, with a moist lower stratosphere apparent above the dry subsidence regions.

Various thermodynamic soundings are shown in Fig. 5 to highlight the differences that can occur between regions. The bottom panel shows the water vapor for the bottom model layer for a 256-km subsection of the domain, with four letters marking the positions at which the soundings are taken. The first sounding, A, represents an arbitrary sounding. The boundary layer is seen to be dry adiabatic and well mixed in moisture, and the temperature profile in the free troposphere above is seen to be roughly moist adiabatic. The water vapor profile exhibits large variation in the vertical, with two “bulges” in the profiles centered at 250 and 600 hPa, respectively, most likely the result of previous nearby deep convective events. Profile B is taken in the middle of a cold pool area, which is reflected in the very dry boundary layer values of moisture. The convective event...
that was the source of this cold pool is in the dissipating stage, and the atmosphere above 550 hPa is seen to be very humid, and is saturated virtually throughout the layer between 400 and 120 hPa. Although the temperature profile is virtually identical throughout most of the troposphere to the profile examined in A, the boundary layer temperatures are lower.

The third profile at C is taken at a position close to B, at the very edge of a cold pool originating from a much earlier convective event situated approximately 20 km “north” of B. Even though the separation of B and C is only 14 km, the water vapor in the free troposphere shows no signs of advective moistening by the convective cloud at B yet. Close examination of the boundary layer reveals that the air on the edge of this cold pool appears to be much moister than the profile taken at A, as seen in the boundary layer moisture slice. Thus, if these parcels were to be lifted to their level of free convection one would expect them to have higher values of CAPE than an average background parcel. Calculations of CAPE do indeed reveal a reversible CAPE value of 2750 J kg⁻¹ at D, compared to a domain mean of 1440 J kg⁻¹. For comparison point A has a CAPE value of 1420 J kg⁻¹, whereas the downdraft sounding at B has zero CAPE. The last sounding D is taken in the middle of the first dry band centered at around 170 km east. What is remarkable with this location is the dryness of the boundary layer, comparable to that in the middle of a convective cold pool, and this point has a reversible CAPE value very close to zero. As expected over the oceans (e.g., Emanuel et al. 1994), the convective inhibition energy (CIN) is generally limited, with a value of only 11.6 J kg⁻¹ on average, but with a value exceeding 100 J kg⁻¹ at D, for example, and having a domain maximum value of 300 J kg⁻¹ at this point in time. Comparing the tephigram soundings, the horizontal variation that occurs in the water vapor field is again emphasized. Between these four soundings, the relative humidity at 13 km ranges from complete saturation to a little over 10% (with respect to ice).

c. Boundary layer

The boundary layer water vapor (Fig. 5) shows the clustering of cold pools, and that more recent convective events appear to be situated on the borders of previously existing cold pools. For example, the convection centered at B is found on the boundary of the cold pool from which the sounding C was taken, and two smaller unmarked cold pools to the south. This is not surprising, since cold pools have long been associated with the triggering of new convection (e.g., Simpson 1980; Thorpe et al. 1982; Rotunno et al. 1988), especially in relation to systems occurring in sheared wind environments such as squall lines. The convective triggering is usually regarded as being the result of the work done by cold pools in dynamically lifting environmental air through the negative inhibition layer to its level of free
Fig. 5. The lower panel shows a subsection of the 50-m water vapor from Fig. 3, on which four positions are marked A, B, C, and D. The vertical thermodynamic soundings at each of these positions are plotted on tephigrams in the upper panels.
convection. However, the cold pool centered at X has taken a period of more than 2 h to reach its present radius of 15 km, and by this point, the vertical velocity for this region of the domain is found to be much more uniform than the water vapor field (Fig. 6). The event marked as B in Fig. 5 is clearly seen as a ring of downward velocity, peaking at \(-0.5 \text{ m s}^{-1}\). Around the periphery, a light colored ring is visible, indicating uplifted air. The upward velocities do not exceed \(0.2 \text{ m s}^{-1}\) anywhere and are more typically on the order of \(0.1 \text{ m s}^{-1}\).

The larger cold pool situated at “X,” which was clearly visible in the water field, is in fact barely discernible in the velocity field. On the western and southern front, however, there are signs of weakly uplifted air, but the velocities are limited. On the east side of the front there is no sign at all of velocities perturbations, and indeed the two smaller cold pools to the south of B are also virtually absent in the velocity field.

Figure 5 highlights another aspect of cold pool evolution that has been relatively ignored; the role of moisture. CAPE is calculated for each gridpoint in the domain, assuming a reversible parcel ascent. Since the CAPE calculation does not take into account any lateral mixing of environmental air into the parcel during the ascent, nearly all of the horizontal variation in CAPE is due to the changing boundary layer \(\theta_e\), and CAPE can essentially be considered as a boundary layer property. The distribution of CAPE (Fig. 7a) produces a replica of the boundary layer water vapor pattern. The moist air found at the boundaries of the cold pool is associated with the greatest CAPE values found anywhere within the domain. This indicates that the air on the outer boundaries of downdraft cold pools is perhaps the most likely to sustain deep convection. Figure 7b shows the corresponding values of CIN for the bottom model level and reveals a strong negative correlation with CAPE as expected. The dry bands are regions of large CIN, as are the centers of convective cold pools. More remarkable are the rings of almost zero CIN air on the borders of the cold pools, exactly where the highest CAPE values are observed.

The parcel sounding used to calculate CAPE and CIN was calculated using the column values of temperature and moisture to calculate parcel buoyancy, rather than horizontally averaging over a wide area to give the mean “environmental” sounding. It is interesting to note the reason for this. Using a mean background sounding had very little effect on the value calculated for CAPE, but made a crucial difference to the distribution of CIN, which, due to its limited magnitude over the oceans, is very sensitive to small virtual temperature changes. It was found that using an average environmental sounding, removing the significant horizontal variations in water vapor amounts, resulted in unrealistic zero CIN values for a not insignificant proportion of the model domain. This was true for both reversible and pseudoadiabatic parcel ascent assumptions.

In the dry, convection free bands, CAPE is almost uniformly zero, whereas in the convective areas, the mean CAPE is around 1500 J kg\(^{-1}\). This appears to indicate that CAPE is positively correlated with convective activity, and additionally that horizontal variations in convective activity can be explained purely in terms of variations in boundary layer thermodynamic properties. This positive correlation between CAPE and convective activity seems to conflict with the findings of McBride and Frank (1999), who found an inverse relationship, which they attributed to the direct cooling and drying effect that convective downdrafts have on the boundary layer. But in fact the findings are consis-
tent, since within the convecting moist bands, CAPE
does indeed decrease in response to convective wakes;
it is simply that the very dry bands far away from con-
vection are dryer still, and therefore have almost zero
values of CAPE. Thus it is noted that CAPE–convective
correlations ultimately depend on the temporal and spa-
tial scale, that one observes. In this experiment, CAPE
is correlated with convection on small spatial and tem-
poral scales since convection seems to erup from the
moist rings of high CAPE values. However, examined
on a larger spatial scale (or over a longer timescale),
convection is negatively correlated with CAPE, since
over the lifetime of a complete convective system,
the injection of midtropospheric dry and cold air into the
boundary layer reduces CAPE, as noted by Mapes and
Houze (1992) and McBride and Frank (1999). Viewed
over still larger spatial scales CAPE is once again cor-
related with convective activity in these experiments,
due to the boundary layer dryness in the convective free
bands.

There is therefore an open question of how exactly
cold pools are triggering new events in this simulation,
since it seems that the high CAPE values found at the
periphery of cold pools combined with the low vertical
velocities could indicate that cold pool triggering is
more thermodynamical than dynamical in nature. The
origin of the high water vapor band is not clear; although
it is obvious that it is not the result of dynamical lifting
air, since the high moisture values are present through-
out the well-mixed boundary layer. Therefore the only
possible sources of moisture are horizontal advection,
surface fluxes, or evaporation of precipitation falling
through the parcel. This role of cold pools is important
to understand and is the subject of a companion paper
to this one (Tompkins 2001b).

4. The role of water vapor

The control experiment of the previous section
showed the spontaneous formation of convective me-
oscale clusters separated by clear sky bands that were
drier than average throughout the troposphere into the
boundary layer. The question arises as to what extent
the water vapor actively plays a role in organizing the
convection. There has been recent observational evi-
dence of dry air episodes greatly inhibiting convection
(Numaguti et al. 1995; Yoneyama and Fujitani 1995;
Mapes and Zuidema 1996; Brown and Zhang 1997; Par-
sons et al. 2000), which were usually found to be
associated with the advection of air from the extratropics,
which regularly penetrates as far as the equator (Yoney-
amaya and Parsons 1999). Convection is inhibited by dry
troposphere since the entrainment of dry air reduces in-
cloud buoyancy, and also because convective down-
drafts will inject dry air directly into the boundary layer
e.g., Raymond 1995; Michaud 1998; Raymond 2000b).

A positive feedback of this kind involving shallow con-
vection preconditioning the lower troposphere for future
deep convection via its moistening effect was previously
suggested by Esbensen (1978), Johnson (1978), and
Nicholls and Lemone (1980), for example.

On the other hand, even if water vapor did not affect
convective intensity directly, the observed water vapor
distribution would still result, since convective activity
is the only source for the free troposphere. Thus it is
not clear if the water vapor distribution is merely the
product of convection organized by other means, or
whether the water vapor actively plays a role in the
organization. In the case where a feedback does exist
between the convection and water vapor, the relative
role of the dryness in the boundary layer and free tro-
posphere should be quantified (i.e., low CAPE pre-
venting convective initiation or entrainment of dry air
reducing updraft buoyancy). Therefore, further experi-
ments are conducted, in which a water vapor pertur-
bation is introduced at day 15 of the control experiment
to observe the effect on the convective activity and or-
ganization. The perturbation applied is a reduction
of 60% in the water vapor amount everywhere between
800 and 1024 km on the x axis. This region is chosen
to coincide with one of the most convectively active
parts of the model domain on day 15. In the first sen-
sitivity test the perturbation is applied throughout the
free troposphere between the 900 and 300 hPa levels.
An additional experiment then applies the perturbation
only between 700 and 300 hPa. Since a compensating
temperature adjustment is not applied, the vapor per-
turbation also represents a virtual temperature pertur-
bation, but this will be removed quickly on a gravity
wave timescale, and therefore it is the water vapor
changes that must provide any long-term influence for
convection.

The rainfall pattern is shown in Fig. 8 for the 5 days
after the perturbation is applied. In the case where the
perturbation is applied above 900 hPa, it almost im-
mediately cuts off all convection in the perturbation
region, which remains the case for the following 5 days.
However, the region free from convection becomes nar-
rower throughout this period, as convection encroaches
from the edges, highlighting once again the importance
of downdrafts, and the localized moistening of convec-
tive outflow air.

At the standard subsidence rate associated with a ra-
diative cooling rate of 2 K day$^{-1}$, the dry air at 900 hPa
would take 15 h to subside to the top of the boundary
layer, but the convection is extinguished almost im-
mediately. The total absence of even shallow convection
and the speed at which it is extinguished indicates that
the dry perturbation affects the boundary layer almost
immediately, highlighting the central role of convective
scale downdrafts, as previously emphasized by Emanuel
(1995) and Raymond (1995). One can see this by ex-
amining the relative humidity at the bottom model layer
in various segments of the domain (Fig. 9), which show
a large decrease in relative humidity to less than 70%-
within the first 6 h. After this time, different points
Fig. 8. As for Fig. 1 for (a) an extension of the control run, and the two sensitivity tests with a water vapor perturbation applied between (b) 700 and 300 hPa, and (c) 900 and 300 hPa. The perturbation is applied on day 15 between \( x = 800 \) and \( 1024 \) km. See text for details.

within the dry perturbation have contrasting behavior. At the edges of the dry perturbation, relative humidity starts to recover after a day, due to horizontal advection and increased surface fluxes due to nearby convection. This contrasts to the behavior at the center of the perturbation, which starts to decline further after about 15 h, coinciding exactly with the time required for the dry air to subside to the top of the boundary layer.

When the perturbation is only applied above 700 hPa, convection is still affected, but to a much lesser extent. The effect of the perturbation appears to initiate on a much slower timescale, with convection almost extinguished by day 20, by which time the subsidence has brought the upper-level perturbation into the lower troposphere. This again highlights the importance of lower troposphere water vapor. It should be pointed out that since entrainment into convective clouds is generally higher in the lower troposphere, whereas greater detrainment is expected at upper levels, the higher sensitivity to lower-tropospheric perturbations cannot be necessarily attributed entirely to the action of downdrafts. It is possible that the entrainment of dry air into updraft cores also has a much more significant effect at these heights. That said, Gilmore and Wicker (1998) show that downdrafts are more sensitive than updrafts to environmental dry air.

The conclusion that the influence of dry air on entrainment is less important than the action of downdrafts should be made with caution, however, due to the course 2-km resolution of these experiments. At this resolution, nearly all of the turbulence mixing between core and

Fig. 9. Time series of relative humidity for the bottom model level at 50 m, calculated for the three regions within in the area subjected to a moisture perturbation, and the whole of the region unaffected by the perturbation.
environment is handled by the subgridscale mixing parameterization, since the cores are at best represented by a few grid points. Therefore it is highly possible that the effect of moisture on core buoyancy upon entrainment is severely misrepresented and that upper-tropospheric vapor is possibly much more (or less) important in controlling convection than indicated here.

5. Convective organization over SST gradients

As stated in the introduction, convection is usually affected by a number of organizing processes, such as underlying SST gradients. The experiments in this paper so far have demonstrated that the feedback between convection and water vapor can organize convection into clusters. With this knowledge it is now possible to investigate if this feedback can alter the organization due to large-scale SST gradients. An experiment is conducted in which an idealized SST gradient is applied, taking the form of a sine-wave of amplitude 0.5 K, equivalent to a fairly typical Pacific SST gradient of approximately $2 \times 10^{-3}$ K km$^{-1}$. It is applied along the long $x$ axis of the 1024-km domain, similar in fashion to the 2D experiments of Grabowski et al. (2000). The highest SST of 300.5 K is situated at $x = 256$ km. This gradient was held constant for a period of 5 days to allow a Walker cell type of circulation to establish, with mean ascent over the warmest SSTs, and subsidence balancing radiation over the coolest SSTs.

Within the warm pool, the increased surface incident shortwave radiation and suppressed surface latent heat fluxes in the clear sky, subsidence region are likely to cause a positive SST tendency (e.g., Lau and Sui 1997), possibly resulting in a SST “hot spot” (Waliser 1996). To see how the thermally direct circulation adjusts to evolving SSTs, we introduce a surrogate hot spot by reversing the SST gradient on day 5, so that the hottest SSTs are now at $x = 768$ km.

The rainfall pattern for the entire 10-day experiment is shown in Fig. 10, which reveals surprising behavior. During the first 5 days of the experiment the convection establishes itself as one would expect, with net upward motion over the warm SSTs and convection suppressed over the cooler SSTs.

After the SST reversal causes convection to die out over the lowest SSTs within a day or so. But instead of spontaneously flaring up over the new warm SSTs, convection instead slowly propagates toward the new SST hot spot. After five further days of simulation, the hottest part of the domain is still completely free of convection. The experiment was limited to 10 days by available computing resources but the propagation rates suggest that the timescale for convection to propagate to the new SST hot spot is of the order of 2 weeks.

Thus the central role of the water vapor feedback with convection is highlighted by this experiment. Figure 11 shows the normalized variance of water vapor for the last 4 days of the experiment. The striking feature is the dryness of the atmosphere over the region centered at $x = 768$ km, the result of the lack of convection during the first 5 days of the experiment. The feedback between water vapor and convection means that after the SST reversal, convection cannot flare up over the warm SSTs, since it is suppressed by the atmospheric dryness. Convection remains over medium SSTs, which are more favorable to convection due to the preexisting moist atmosphere. Thus a quasi stable circulation exists, with convection propagating toward the new SST maximum at the rate at which it can moisten the atmosphere. The moistening occurs in two ways. First, the upper-level cirrus outflow is advected into the dry region, especially apparent on days 9 and 10 between 832 and 896 km. In addition the surface easterly flow between $x = 768$ and 896 km is moistened by surface fluxes, resulting in a low-level moist band, within which shallow convection is visible. As the air moves eastward, surface fluxes increase the moisture content of this air and the shallow convective layer deepens. However, deep convection is prevented by the dry layer above. This effective capping of deep convection allows high boundary layer moisture values to build up, and provides the reason for which the clusters propagate in the control run with homogeneous SSTs, instead of a stationary pattern being established. As the capped, very moist boundary layer air reaches the convective cluster, a region locally moistened by convection, the lid is removed. Thus the clusters will preferentially propagate toward the dry regions. The boundary layer dry band...
between 800 and 900 km on day 11 in Fig. 3 shows a clear example of this.

Examining Fig. 11 further, it is also apparent how the region centered at 256 km steadily dries out from the upper atmosphere downward after the convection extinguishes around day 6. This developing dryness, especially in the mid- to upper troposphere, in convectively suppressed phases has already been highlighted in observations by Brown and Zhang (1997).

Thus the significance of the water vapor feedback with convection is made very apparent by these investigations. It is suggested that on a different planet with a dry atmosphere, surface temperature perturbations would result in the rapid occurrence of convection directly over the temperature maximum and Mapes (1997) demonstrated clearly that the large-scale atmospheric circulation would evolve quickly on the gravity wave propagation speed timescale. With a moist atmosphere the situation is entirely different. The water vapor field, unlike buoyancy, evolves on a slower advective timescale. Thus water vapor stamps a memory on established large-scale circulations, making them persistent and slow to evolve. Convection cannot spontaneously occur over the highest SSTs, but instead must propagate toward them at the rate at which it can moisten the atmosphere.

Based on this experiment, Tompkins (2001a) suggests that this positive feedback between water vapor and convection has significant implications for the interaction between tropical atmospheric dynamics and the ocean temperature distribution.

6. Effect of vertical wind shears

Since water vapor appears to be central in the organization of convection, it is possible that the application of a vertical wind shear, while organizing convection into long-lived systems, will at the same destroy the localization due to water vapor via its lateral mixing effect. In their 2D studies of radiative convective equilibrium, Held et al. (1993) found that applying a vertical wind shear did indeed destroy the localization of convection, and Tompkins (2000) showed that introducing a mean shear made the humidity field much more horizontally uniform. Here, the wind shear is applied by relaxing the mean winds in the x direction to a target profile, using a relaxation timescale of 1 h. Two wind profiles are used, representing both strong and weak vertical wind shears. Both simulations started from the last day of the control run, with the convection already organized.

a. Strong shears

In the first experiment, the shear applied is almost identical to that of Tompkins (2000), with a surface velocity of 8 m s$^{-1}$ increasing to 12 m s$^{-1}$ at 1 km, then changing linearly to $-10$ m s$^{-1}$ at 12 km, thereafter reducing to zero at 14.5 km. Examining the surface rainfall (not shown) it is clear that the application of the wind shear drastically changes the nature of the convection as one would expect. The convection does indeed organize into long-lived systems almost imme-
diately, associated with much larger rainfall rates, which propagate through the domain at the steering velocity of about 11 m s\(^{-1}\). The systems are seen to be evenly spaced, with no obvious signs of other scales of organization in existence, other than the squall line type organization imposed by the wind shear itself. This is unsurprising, since the water vapor field (Fig. 12) does not reveal any banded structure seen in the earlier zero wind runs. In the free troposphere above 3 km, the water vapor is very uniform in the horizontal away from the moist spots of the convective updrafts themselves. This is particularly the case at the convective detrainment level at 13.2 km.

\textit{b. Weak shears}

In the second case an analogous, but weakly sheared, wind profile is applied. The surface velocity is zero, increasing to 1 m s\(^{-1}\) at 1 km, then changing linearly to −1 m s\(^{-1}\) at 11 km, thereafter reducing to zero at 15 km. In this case, the plot of rainfall reveals that the organization persists with weak wind shears (Fig. 13). Several properties of the convective organization can be observed. First, after the application of the wind shear, the propagation of convection within the clusters is easterly only, in the direction of the low-level wind. The clusters themselves remain fairly stationary for 4–5 days, keeping the same scale of organization. On about day 20, several clusters join, the scale of the organization widens, and the clusters themselves propagate and become more robust. The interaction of the wind shear with the organization is more clearly seen in the plot of normalized water vapor and cloud contours, shown in Fig. 14. The figure reveals that the coherence of the water vapor structure is not destroyed by the weak wind shear. The interaction with the wind shear and the occurrence of convection at any particular location is also apparent. When the wind shear advects dry air at low levels over a very moist boundary layer it effectively caps the shallow convective activity and prevents
the occurrence of deep convection. A clear example is seen on day 23.5 at $x = 256$ km. The modulation of deep convective activity by upper-level moisture variations is also visible.

Thus, in contrast to the case where convection was organized by SST, it seems that the organization of convection via the water vapor field is not significant when strong wind shears are imposed, since these efficiently
mix water vapor. However, the feedback between convection and water vapor remains robust when smaller, more typical, vertical wind shears are applied.

7. Conclusions

Given that the tropical moisture budget and the whole nature of the large-scale tropical circulation will be dependent on the organization of convection, it is important to establish what factors control this. Previous studies have mostly focused on convective organization by large vertical wind shears, which can organize the convection into mesoscale structures such as squall lines or mesoscale cloud systems, or by large-scale SST gradients. Here we attempt to ascertain if other mechanisms can organize convection in the absence of these two factors, and if so, whether these mechanisms also modify organization by wind shear or SST gradients.

In order to approach this problem, a large domain 3D experiment was conducted that extended over a region of 1024 km by 64 km, using a cloud-resolving model. Using a model allows the effects of wind shear and SST gradients to be isolated by applying horizontally uniform boundary conditions. The simulation lasted for a total of 15 days, and it was found that after an initial period of about 2 days, during which the convection was randomly distributed, organization began to establish itself. The organization consisted of convective clusters of a dimension of around 200 km, consisting of many convective events that propagated throughout the clusters. The clusters themselves also propagated through the domain at minimal velocities of approximately 0.5 m s\(^{-1}\). In between the clusters, the atmosphere remains largely free of convection. These regions were found to be drier than average, especially in the lower troposphere, where the dry perturbation was also accompanied by a positive temperature perturbation, consistent with a horizontally homogeneous virtual temperature field.

It is interesting to hypothesize the structure this organization would take in a domain of larger horizontal extent. Since the boundary conditions used here are cyclic, the organization produced is essentially a periodic two-dimensional banded structure (one can imagine infinitely tessellating the domain in each direction). Although two-dimensional structures such as squall lines do occur readily in the tropical atmosphere, they are most often associated with a reasonable vertical shear, and the 2D structure could simply be an artifact of the limited third spatial dimension. The 64 km third axis was of large enough horizontal extent to contain the subsidence associated with a deep convective event, and it is presumed therefore that the 200-km cluster scale was not artificially affected by the short size of this axis, although without further sensitivity tests, this possibility cannot be ruled out. On a much larger domain, the convective organization could maybe take the form similar to closed-cellular convection as observed in the planetary boundary layer on a smaller scale. In this scenario, deep convection would be present in circular clusters with a radius on the order of 100 km, separated by clear sky subsidence regions. That said, during the Tropical Ocean Global Atmosphere Comprehensive Ocean–Atmosphere Response Experiment, Rickenbach and Rutledge (1998) found that MSCs were organized linearly in the vast majority of cases, and the disorganized nonlinear MSCs were actually quite rare.

The mesoscale organization of convection was found to be the result of feedback with the water vapor field. Convective activity moistens the local atmosphere, which makes it more favorable for future convection. In order to show that water vapor plays an active role in determining the location of convection, and that its distribution was not simply a passive artifact of convection organized by other mechanisms, experiments were conducted in which vapor perturbations were applied in the free troposphere. These proved that water vapor does indeed control the location of convection. They also indicated that convection is much more sensitive to relative changes in the lower troposphere, most probably through the role of convective scale downdrafts that inject the dry perturbation directly into the boundary layer, although the coarse resolution demands that this conclusion be made with caution.

Examining the distribution of CAPE, it was seen that the peak CAPE values are found on the boundaries of cold pool air of previous convective events. The large values of CAPE are the result of higher than average boundary layer moisture, the origin of which is undetermined. The indication is that the action of cold pools in the initiation of new convective cells could be more thermodynamic than dynamic in nature in cases of limited vertical wind shear. The specific role that cold pools play in the initiation of convection and its organization are the subject of a companion paper (Tompkins 2001b, submitted to *J. Atmos. Sci.*).

Further experiments were conducted to see if the feedback still operates and possibly modifies the organization due to underlying SST gradients. An idealized SST hot spot experiment was conducted, where an underlying SST gradient was imposed for 5 days to establish a large-scale circulation, which was then reversed to introduce a surrogate SST hot spot, and to allow the atmospheric dynamical response to be examined. During the first 5 days the atmosphere over the cooler SSTs dries out relative to the convectively active areas as one would expect. It found that this dry atmosphere prevents the spontaneous eruption of convection over the new SST warm anomaly after the SST reversal, and instead the convection must propagate into the area at the speed at which it can moisten the atmosphere. This has significant implications for our understanding of the relationship between SST, convection, and the large-scale flow in the Tropics, investigated further by Tompkins (2001a).

Finally it was shown that, although very strong wind
shears prevent the feedback by strongly mixing water vapor, it is still important in the cases of weak wind shears.

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