Radiative–convective equilibrium in a three-dimensional cloud-ensemble model

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SUMMARY
A knowledge of radiative convective interactions is key to an understanding of the tropical climate. In an attempt to address this a cloud-resolving model has been run to a radiative–convective equilibrium state in three dimensions. The model includes a three-phase bulk microphysical scheme and a fully interactive two-stream broadband radiative-transfer scheme for both the infrared and solar radiation. The simulation is performed using a fixed sea surface temperature, and cyclic lateral boundary conditions. No ‘large-scale’ convergence, mean wind shear or background vorticity was imposed.

The total integration lasted 70 days, and a statistical equilibrium state was reached at all heights after 30 days of simulation in all model variables. It is seen that some variables, such as vertical mass flux, adjust quickly to their equilibrium values while others, such as column-integrated water amount, domain-mean temperature and convective available potential energy (CAPE) display variation on a longer 30-day time-scale. The equilibrium state had a column-integrated vapour amount of 42.3 kg m\(^{-2}\), a mean temperature of 258.7 K and a pseudo-adiabatic CAPE value of 1900 J kg\(^{-1}\). The equilibrium-state statistics are consistent with tropical observations.

The convection does not remain randomly distributed but instead becomes organized, aligning in a band structure associated with high moisture values in the boundary layer. This organization seems to result from interactions between radiation, convection and surface fluxes. The surface-flux feedback is due to higher boundary-layer winds, associated with convection, increasing surface fluxes of moisture locally. Horizontally inhomogeneous radiation can act to make clouds longer lasting and also increase convergence into cloudy region. Replacing the wind-sensitive surface-flux calculation with a linear relaxation to surface values appeared to largely destroy this organization, as did the use of an imposed horizontally uniform radiative-heating rate.

KEYWORDS: Cloud-resolving model Convection Tropical climate

1. INTRODUCTION

An external forcing of the earth’s climate, a change in the solar irradiance for example, will produce a corresponding temperature change in that system. This change will be dependent on the feedbacks operating owing to changing cloud characteristics and atmospheric water vapour. To calculate the magnitude of these feedbacks in a convecting atmosphere it will be necessary to understand the moistening and warming processes of convection—not only in a transient sense, considering only the development or lifespan of a single cloud, but in the longer term, over the life cycle of many clouds in a state of radiative–convective equilibrium.

The issue of cloud and water-vapour climate feedbacks has mostly to date been dealt with using observational studies (e.g. Ramanathan et al. 1989; Raval and Ramanathan 1989) or general-circulation models (GCMs) (Cess et al. 1990; Senior and Mitchell 1993; Wetherald and Manabe 1988, among others). One-dimensional idealized radiative-convective models have also played a role in the understanding of climate processes (e.g. Manabe and Strickler 1964; Emanuel 1991; Rennó et al. 1994). An advantage of using models to investigate the tropical climate is the possibility of conducting a controlled experiment in which the system’s external forcing is predetermined and the time-scales for the various fields to reach their respective equilibrium states can be analysed in detail.

With increasing computing resources, a new tool for these studies has recently become viable for these investigations; namely the cloud-resolving model (CRM) that explicitly represents the processes of convection. However, today’s computers are still not sufficient to model the entire globe with convective dynamics resolved, and so it is necessary to deal with the interaction with so-called ‘large-scale’ features of the atmospheric circulation that

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are not contained within the modelled domain. The principal way in which this has been done is by way of ‘scale separation’ in which processes larger and slower than a certain selected space- and time-scale are considered as a ‘large-scale’ forcing for convection. This assumes firstly that there exists some intrinsic domain-scale which is large enough to contain an ensemble of cumulus activity, but is nevertheless small when compared to the large-scale flow, and secondly that these large-scale terms vary slowly enough in space to be represented as averages over the CRM domain. Assuming that such a separation is possible, Grabowski et al. (1996b) gave a careful discussion on how a CRM can be forced in this way and proceeded to use evolving data from the GATE* field study as convective forcing in two-dimensional numerical investigations. The assumed scale separation is also fundamental to cumulus parametrization, and it might be argued that some of the successes of the parametrization effort provide indirect evidence in favour of that underlying assumption.

However, the precise scale at which to separate convection from large-scale processes is often unclear, and it is arguable that such a division is sometimes simply not possible. An obvious example is midlatitude mesoscale convective systems, which are often of a size that is a significant fraction of the radius of deformation and thus not small in comparison with baroclinic weather systems (Gray et al. 1998). Even for tropical convection, if the large scale is defined as a GCM grid box or the domain of a CRM integration, the ‘ensemble’ may consist only of a single organized convective system. This problem was recognized by Xu and Randall (1996) who used mean values from ensembles of three CRM integrations in their study of GATE convection. A major difficulty in defining a scale separation is a lack of knowledge of the intrinsic space and time-scales of the convective ensemble. At the very least, it is known that these will depend upon the organization of the convection, which will depend upon the large-scale environmental factors such as vertical wind shear.

Given this uncertainty it seems prudent, in a first attempt to study cloud–radiative interactions, to avoid the problems of scale separation by considering a simpler physical situation where an imposed large-scale forcing is not present. As a numerical modelling problem this situation is well posed, and allows the intrinsic space- and time-scales of the convection to be identified, provided the parameters of the simulation, such as domain size and run length, are sufficient to accommodate them. This last point is important, as the scales may be unexpectedly large. For example, in two-dimensional simulations, convection becomes organized on the scale of the radius of deformation unless restricted by the size of the domain. This behaviour is predicted by Bretherton (1987), and has been demonstrated in simulations on large domains (up to five times the radius of deformation) (Craig 1995, 1996b). It is not claimed that this simple situation of convection forced only by surface fluxes and radiative cooling is the most important form of tropical convection; only that it represents a physically relevant problem that is consistently posed and, as will be shown, computationally feasible.

This paper describes such an experiment in which a simple radiative–convective equilibrium is achieved for an ensemble of cumulus clouds over a uniform fixed sea surface temperature (SST) surface in the complete absence of any large-scale dynamical forcing. This means that the convection is driven only by radiative cooling of the atmosphere and fluxes of heat and moisture from the surface, and that no mean wind, vertical wind shear, mean ascent or moisture forcings are imposed. Since there is evidence that background rotation can generate large-scale structure (Vallis et al. 1997) the experiment also excludes the Coriolis effect. Although the physical problem is quite simple, the intrinsic space- and time-scales of the convecting atmosphere are not well known. The equilibrium obtained

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will be statistical in nature, that is, there will be fluctuations such as those associated with individual convective events, so due consideration must be given to ensuring that the averaged properties of the equilibrium state are representative of those that would occur in a region of real atmosphere of sufficient size that the fluctuations are small. Therefore, considerable attention is given to ensuring that the results will be independent of the model configuration, in particular the domain size and horizontal resolution.

Unlike the work to be described here, most long-term simulations of convective ensembles have been based on a two-dimensional (2D) vertical slice. In an early numerical study Asai and Nakasuji (1977, 1982) examined the scale and organization of convection in a quasi-steady state using a model with highly simplified microphysics. Nakajima and Matsuno (1988) operated a more complex 2D cloud model to a quasi-steady state using fixed imposed radiative cooling to drive convection. More recently, Held et al. (1993), Sui et al. (1994) (referred to hereafter as H93 and S94, respectively) and Randall et al. (1994) have studied radiative-convective equilibrium states, using 2D cloud-resolving models, this time with interactive radiation schemes. The experiment of S94 was repeated with a similar set-up by Grabowski et al. (1996a) using a different 2D cloud model with the diurnal cycle added to the short-wave radiation parametrization. Unlike the experiment of S94, Grabowski et al. (1996a) obtained a very moist and warm equilibrium climate, with column-integrated water amounts substantially larger than monthly mean observations. Very recently Xu and Randall (1997) have attempted to investigate the possible reasons for differences between S94 and Grabowski et al. (1996a) by integrating their own CRM to equilibrium with a similar set-up to the previous two studies, but with various methods of imposing large-scale forcing. They found that different large-scale forcing can result in an equilibrium total water-vapour amount that is similar to that in both the previous studies. However, the contrasting equilibrium states could also be attributable to disparities in the model codes, and as discussed by Emanuel (1991, 1994), it seems likely that the equilibrium profiles are also sensitive to the parametrization of microphysical processes.

Tao and Soong (1986) and Tao et al. (1987) compared 2D and 3D experiments with a cloud-ensemble model and reported that many convective statistics were similar (possibly as a result of the 2D nature of the forcing). However, it is questionable whether two dimensions can correctly represent the interactions that exist between convection and its environment for anything other than essentially two-dimensional phenomena. Even the commonly quoted example of a 2D phenomena, the squall line, can involve complex 3D dynamical motions and produces anlils that are three-dimensional. Bretherton (1987) discussed the nature of subsidence associated with a heat source in two- and three-dimensional geometry, and it was shown that 3D geometry restricts the subsidence to the locality of the heat source, much more strongly than do the effects of dissipation and the Coriolis effect. This suggests it may be necessary to utilize a 3D cloud model to correctly represent the interactions between clouds. Both Robe and Emanuel (1996) and Islam et al. (1993) have operated 3D cloud models to equilibrium states, but in both cases fixed imposed radiative cooling and simple microphysical schemes were used.

The model naturally needs to have a horizontal resolution that at least marginally resolves deep convective cores. If it is assumed that cloud vertical velocities are largely determined by local buoyancy, then a resolution that is too coarse will result in clouds with excessive entrainment and presumably inaccurate vertical water transport. In addition to being able to model a single cloud, the model should be able to model the long-term behaviour of an ensemble of clouds and their associated subsidence. To accomplish this it is important that the domain size used is large enough, since if the domain size is too small the subsidence resulting from convection will be larger than it would naturally be, as it is artificially restricted. Thus the dry-adiabatic warming of the environment will
also be greater than usual, perhaps even large enough to push the environment into a
convectively stable state. Thus convection becomes "sporadic" with significant pauses
existing between convective events, while the atmospheric radiative cooling and surface
fluxes restore conditional instability. Robe and Emanuel (1996) reported that averaging
over time in their three-dimensional experiments yielded identical results to using a larger
domain. However, even if this is true for measurements of cloud hydrometeors, the same
radiative-convective equilibrium cannot be expected. Because of the nonlinear dependence
of the radiation field on atmospheric composition, the mean environmental heating rates
of a domain that is oscillating between states of over-cloudy and over-dry will not be
the same as a domain with a continuously realistic cloud field (e.g. Zhang et al. 1994).
In the study of H93, for example, rainfall time series seem to indicate that the domain
contained at most one major convective event at any particular time in the equilibrium
state, interspersed with periods with little or no convective activity, possibly indicating an
undersized domain. Likewise, the equilibrium climate of Grabowski et al. (1996a), who
used a similar-sized domain, had an upper-tropospheric cloud fraction of almost 100%
throughout their experiment.

Because of these considerations, this study employs a 3D cloud-resolving model,
with an interactive radiation scheme and three-phase microphysics. The model needs a
resolution that at least marginally resolves convection and a domain size sufficiently large to
represent the subsidence surrounding convection correctly. A cloud-resolving model, with
all the above components and necessary set-up, has now been run to a radiative-convective
equilibrium state. This paper will describe the model used and all its components briefly in
section 2. The set-up used for the experiment is outlined in section 3, which then attempts
to justify the choices made for factors including horizontal resolution and domain size.
Section 4 details the equilibrium state, and the time-scales on which different variables
approach their respective final states. The structure and organization of the convection is
also examined.

2. MODEL DESCRIPTION

This section aims to describe briefly the cloud model used for this experiment, along
with the microphysical, radiation and surface-flux schemes that have been added.

(a) Dynamical model

The cloud model is based on the anelastic quasi-Boussinesq approximation, for which
the main equations are given in Shutts and Gray (1994) (hereafter SG94). The model
uses liquid-water temperature for its thermodynamic quantity, which is more accurately
conserved for parcel ascent in deep convection than other alternatives such as the liquid-
water potential temperature (Wilhelmsen 1977; Shutts 1991).

The subgrid model utilizes a first-order closure based on the Smagorinsky–Lilly
model (Mason 1989) that is identical to that of SG94 with the exception of the moist
Richardson-number specification. This is instead calculated from the pointwise flux Richard-
son number \( R_f \), which is in turn calculated from the subgrid buoyancy flux that would
occur by exchanging a small fraction of air by volume (assumed here to be 0.01) between
the layer in question, and the adjacent layer above. This method is detailed in MacVe
and Mason (1990). The dissipation length-scale in the subgrid model is calculated from a
basic length-scale \( \lambda_0 \) as in equation (11) of SG94. The basic mixing length \( \lambda_0 \) is defined
here to be equal to the product \( C_s H_m \), where \( H_m \) is the mixed-layer depth (assumed to be
around 500 m) and \( C_s \) is a constant taken as 0.22. Reducing \( C_s \) to 0.15 in a sensitivity test
had little effect on the simulations.
The time-step used in the model is variable to enable efficient use of computing resources whilst ensuring the Courant–Friedrichs–Lewy and diffusive stability criteria are continuously met. The advection scheme used is a total variance diminishing (TVD) scheme based on Leonard (1991), which has the property of being able to advect sharp gradients without overshoots and is less diffusive than the alternative TVD scheme of van Leer (1974). The advection scheme is applied to all prognostic variables.

A Newtonian damping layer applied at the model top damps all prognostic variables to their horizontal mean values at a rate $\alpha$ given by

$$\alpha = \alpha_0 \left\{ \exp \left( \frac{z - z_0}{H_D} \right) - 1 \right\},$$

where $\alpha_0 (=0.0005)$ is the damping coefficient, $z_0$ is 15 km, and $H_D$ is the damping height scale, set to 3 km. The present model configuration implies that damping is applied to the three models layers above 16.3 km which proved to be sufficient to prevent significant gravity-wave reflection at the domain top.

(b) Microphysical scheme

For these experiments an early version of the three-phase ice scheme of Swann (1994) and Brown and Swann (1997) is used. In this scheme, cloud water is assumed to form immediately in the presence of supersaturated air and subsaturation likewise evaporates any cloud present. As a result of this assumption, water vapour and cloud are combined into one variable, the total-water mixing ratio. The scheme therefore integrates prognostic equations for the mass mixing ratios of five microphysical variables: total-water mixing ratio, raindrops, ice crystals, snow crystals and graupel. Parametrizations for 38 difference mechanisms for exchanging mass between the various categories are represented. Cloud water and cloud ice have zero fall speeds.

(c) Radiation scheme

The radiation scheme that was interfaced to the cloud model is based on the two-stream plane-parallel approximation and is detailed in Edwards and Slingo (1996), and compared with observations in Taylor et al. (1996). The scheme has been successfully tested against line-by-line radiation schemes, with heating rates comparing well throughout the troposphere.

The scheme calculates heating rates for both the long-wave and short-wave spectra at each model grid point, and additionally for ten extra layers of equal mass that divide the region between the model top and the top of the atmosphere. The long-wave spectrum is divided into six spectral bands and the short-wave into nine bands. The calculation of heating rates is based on the amounts of water vapour, CO$_2$, ozone, and all hydrometeor mixing ratios, with the effective radii of all water substances based on their mass mixing ratios (Petch 1998). The asymmetry correction for the ice crystals of Petch (1998) is also applied. The profiles of McClatchey et al. (1972) are used to specify temperature and water vapour for the levels above the model domain, in addition to the CO$_2$ and ozone distributions.

The radiation scheme is called every 120 time-steps—equivalent to an average of every 16 minutes. This was chosen to be shorter than a cumulus life-time in an attempt to eliminate 'ghost clouds' in the forcing, but is longer than the interval employed in the simulations of Xu and Randall (1995a) where there was significant movement of the clouds due to an imposed background horizontal wind. In order to observe the intrinsic variability of the convective ensemble no diurnal cycle is currently included. If the diurnal cycle were
simply averaged over a day a small solar zenith angle results, giving continuously long and undesirable path lengths. Instead the solar constant is halved to 685 W m\(^{-2}\) and the solar zenith angle is set to 51.7\(^\circ\).

\(d\) Surface-flux parametrization

Surface fluxes are calculated according to similarity theory (MacVean, personal communication) that obviates the need for an imposed minimum surface wind speed that is necessary when bulk aerodynamic formulae are used. Equations (3.23), (3.24), (3.29) and (3.33) of Garratt (1992) are used to calculate the friction velocity and a turbulent potential-temperature scale in terms of \(z/L\), where \(L\) is the Monin–Obukhov length. The constants used in the similarity functions of (3.29) and (3.33) are \(\gamma_1 = 19.3\), \(\gamma_2 = 12.0\), \(\beta_{1M} = 4.8\) and \(\beta_{1H} = 7.8\). The surface roughness length \(z_0\) adopted for both heat and moisture is 0.01 m. The increases in fluxes with wind speed compare reasonably well to TOGA-COARE* observations made by Jabouille et al. (1996).

3. Experimental set-up

\(a\) Number of horizontal grid points

As discussed in the introduction, it is important to have a domain sufficiently large that it does not constrain the behaviour of the convective ensemble. It was argued that this is achieved when convection is occurring continuously at some location in the domain. Since the minimum area that can be occupied by the convection is a single grid point, the number of grid points occupied by the subsidence must be at least \(1/\sigma\), where \(\sigma\) is the fractional area of the convective motions.

From mass continuity, \(\sigma\) is related to the mean convective vertical velocity \(w_c\) and the mean vertical velocity in the surrounding subsidence region \(w_s\) by

\[
\sigma w_c + (1 - \sigma) w_s = 0.
\]  

If \(\sigma \ll 1\), then

\[
\sigma \approx \frac{-w_s}{w_c}.
\]

If the domain is sufficiently large enough to avoid large fluctuations in domain-averaged quantities, the subsidence velocity \(w_s\) will balance the vertically integrated radiative cooling \(\langle Q \rangle\) in the absence of large-scale convergence, hence

\[
w_s = \frac{-\langle Q_{\text{rad}} \rangle}{\rho c_p H \frac{\partial \theta}{\partial z}},
\]

where \(H\) is the depth of the convecting layer, \(\rho\) is air density, \(c_p\) is the specific heat of air at constant pressure, and \(\partial \theta / \partial z\) is the gradient of potential temperature. Assuming \(\langle Q \rangle = 125\) W m\(^{-2}\), \(\rho H = 9000\) kg m\(^{-2}\) and \(\partial \theta / \partial z = 3\) K km\(^{-1}\) gives an estimate for \(w_s\) of 46 cm s\(^{-1}\). The cumulus velocity is assumed to scale with the convective available potential energy (CAPE),

\[
w_c^2 = \mu \text{ CAPE}.
\]

The constant represents the coefficient of dissipation of mechanical energy, assumed to take the value of 16 (Rennó and Ingersoll 1996). With CAPE = 2000 J kg\(^{-1}\), a value of \(w_c = 11\) m s\(^{-1}\) is obtained.

* Tropical Ocean/Global Atmosphere Coupled Ocean–Atmosphere Response Experiment.
Combining the estimates from (3), (4) and (5) suggests that the number of grid points required to maintain continuous convection is approximately \( \sigma^{-1} = 2400 \), indicating that a 50 by 50 grid is required in a three-dimensional experiment. It is worth noting that the number of grid points required, and hence the expense of the simulations, is the same in two and three dimensions. To verify that a domain size of 50 by 50 grid points is adequate, time series of the maximum and minimum vertical velocity within the domain were analysed (not shown), which showed that convective updraughts and downdraughts are occurring continuously throughout the integration, using the common threshold velocity of ±1 m s\(^{-1}\) to identify convective cores (Lemone and Zipser 1980; Jorgensen and Lemone 1989; Lucas et al. 1994; Robe and Emanuel 1996).

While the estimates given above appear to depend in a rather arbitrary way on characteristics of the convecting atmosphere, such as CAPE, they are consistent with a much more general analysis of quasi-equilibrium convection by Craig (1996a) based upon the heat engine arguments of Rennó and Ingersoll (1996), which led to a relationship for \( \sigma \) of the form

\[
\sigma = c(\Delta h)^{-3/2} F,
\]

where \( c \) is a constant, \( \Delta h \) is the enthalpy difference between the surface and tropopause, and \( F \) is the large-scale forcing (the sum of radiative cooling and cooling by dynamically driven ascent). It was argued that these two parameters represent externally imposed constraints on the convective ensemble, implying that the convective fraction is also externally determined. A related result has also been obtained by Emanuel and Bister (1996). An important consequence of (6) is the linear dependence of \( \sigma \) on the external forcing. This has also been observed in previous numerical studies including Asai and Nakasuji (1977, 1982), Gregory and Miller (1989) and Robe and Emanuel (1996). This indicates that the estimated required domain size will be proportionately smaller when a large-scale forcing is imposed.

(b) Resolution and boundary conditions

Given an estimate for the number of grid points, knowledge of the size of a typical convective core is required to determine the domain size. A brief review of observational deep-convection studies made by Tompkins (1997) indicates that to marginally resolve deep convective cores a horizontal resolution of about 2 km is needed. Thus for these experiments a grid of 50 grid points in each direction is used, giving a horizontal domain size of 100 km by 100 km.

In the vertical a stretched grid consisting of 31 levels is used to enable more levels to be placed in and near the boundary layer. The lowest model layer is located at 40 m with the grid spacing varying from 125 m near the surface to 1500 m at the top of the domain at 20 km. The lateral boundary conditions are periodic. Since the complexities of interaction with large-scale processes are to be neglected, there is no imposed convergence into the domain, no forced uplift, no imposed wind shear and no rotation. This does not imply, however, that the imposed boundary conditions are necessarily unrealistic. A fixed SST of 300 K was the mean measured in the equatorial mid-Pacific ocean during the intensive observation period of TOGA-COARE (Vincent 1994). The absence of large-scale mass and moisture forcing in this experiment is also consistent with the mean moisture convergence for certain mid-Pacific regions, which can have approximately zero monthly-mean values at certain times during the year (Dodd and James 1996). Of course, these monthly-mean observations do not imply zero moisture convergence on shorter synoptic time-scales, but analysis of the adjustment time-scales present in this model should allow predictions as to how the model will react to synoptic-scale events.
4. EQUILIBRIUM STATE

(a) Approach to equilibrium

The approach to equilibrium is charted in Fig. 1 which shows time series of the domain-averaged temperature, column-integrated water-vapour content ($W$) and the rain rate. Examination of the $W$ and temperature graphs reveals a long time-scale for equilibrium to be reached. The long time-scale in temperature is not obvious in this experiment since the mean temperature of the initial state was very similar to that of the final equilibrium state, with the variation in temperature limited to less than 1 K. This is because the experiment was initialized from the end of a preliminary investigation using fixed radiative cooling but also with an SST of 300 K. In other experiments a long adjustment time-scale in the temperature field is more apparent. When exponential curves are fitted to the data, the time constant for the approach to equilibrium is around 20 days, of similar order to the adjustment time-scales of all the earlier 2D investigations of H93, S94 and Grabowski et al. (1996a). Although of a similar order of magnitude, this time-scale is somewhat longer.
than that of S94, perhaps because of the absence of imposed convergence which acts as an additional forcing.

- In addition to this long time-scale, shorter time-scales are also in evidence. Firstly, oscillations of the period of hours associated with the cumulus time-scale are apparent. There are also short-time oscillations of time-scales up to five days or so, which could possibly be associated with a 'cumulus cluster' time-scale. Variability on a variety of time-scales was also seen in the 2D simulations of Randall et al. (1994), but it is unclear whether that variability is related to the 3D results presented here.

It is possible to see a high degree of correlation between the moisture and temperature fields on all time-scales. In contrast to the W and temperature plots, the rain rate adjusts very quickly to an equilibrium state and all the variability is on the short 'cumulus' time-scale, although obviously the amplitude of this dominant variability will be a function of domain size. The three other microphysical variables of snow, ice and graupel also quickly adjust to equilibrium status on the same time-scale as the rainfall, and no continued accumulation of any species occurs which could have a serious impact on the equilibrium status when using an interactive radiation scheme. The correlation between the water and temperature fields and the mechanisms that determine the various time-scales of adjustment are examined further in Tompkins and Craig (1998b).

It is not possible to see from Fig. 1 if the model climate has reached equilibrium at all heights and so Fig. 2 shows the time variation in the vertical of the anomalies of dry-static energy and relative humidity with respect to their mean values over the last 40 days of the experiment. The domain-average relative humidity is calculated from the domain-average temperature and moisture profiles (rather than the average of the relative humidity of each grid point). It appears that the troposphere has indeed reached equilibrium at all heights by day 30 of the integration. The stratospheric part of the model domain also reaches equilibrium within the integration period, on a slower time-scale than that operating in the troposphere.

If the quantities are averaged between day 30 and 70, then the equilibrium state has a domain-average temperature of 258.7 K, pseudo-adiabatic and reversible convective available potential-energy values (CAPEp and CAPEv) of 1900 and 1100 J kg⁻¹, respectively, a column-integrated water-vapour content of 42.3 kg m⁻¹, and a rain rate of 4.6 kg m⁻² d⁻¹. The domain-mean CAPE values are calculated from the domain-mean sounding rather than the average CAPE value of each column. CAPE is calculated using the bottom model level (at 40 m) for the starting point of the parcel ascent and taking virtual-temperature effects and the heat capacity of water into account. The adjustment of the CAPE is shown in Fig. 3. It is seen that although CAPE shows large variability on the short time-scales, there is a perhaps surprisingly substantial adjustment on the longer 30-day time-scale too.

The average rain rate is almost exactly equal to the average surface water flux of 133 W m⁻², with the residual following the trend in W. These values can be compared to the control run of Lau et al. (1993) (which gave more details concerning the water budget of S94 and will be referred to as L93) that utilized a surface temperature of 301 K, one degree warmer. The domain-averaged temperature is very similar to the 257.6 K of L93. On the other hand the equilibrium atmosphere of L93 is substantially more moist (W = 51.3 kg m⁻²). Stephens (1990) showed that observations suggest a strong dependence of W on the surface temperature, and so this difference is perhaps partly due to the warmer surface temperature used in L93, and also because L93 included a large-scale convergence moisture forcing which is absent from this experiment. The limited-domain investigations of Randall et al. (1994) produced a similar equilibrium value for W of 45 kg m⁻². Grabowski et al. (1996a) on the other hand obtained an equilibrium W of 70 kg m⁻²—a much higher value.

The equilibrium value for W here appears to be in line with mid-Pacific observations.
Figure 2. Time–height sections of (a) dry-static energy perturbation and (b) relative humidity with respect to the equilibrium mean values.

For example, analysis of satellite data by Prabhakara et al. (1985) and sonde data by Raval et al. (1994) would indicate a seasonal average total-column water content in the 40 to 45 kg m$^{-2}$ range, not in disagreement with the instantaneous sonde data reported by Weaver et al. (1994). In particular, Stephens (1990) analysed the dependence of $W$ on surface temperature over the oceans and for an SST of 300 K gave an annual mean $W$ of approximately $41 \pm 4$ kg m$^{-2}$. Gaffen et al. (1992) used sonde data to measure this relationship and found results in close agreement with those of Stephens (1990).

(b) Cloud structure and organization in equilibrium state

To reveal the nature of a deep convective event in this model, a 2D slice through a typical convective core occurring on day 70 of the experiment is shown in Fig. 4, with
Figure 3. Time series of convective available potential energy (CAPE) calculated for a parcel lifted pseudo-adiabatically (CAPE$_p$) and reversibly (CAPE$_r$) from the bottom model level at 40 m.

Figure 4. Vertical section through a deep convective event occurring on day 70 of the integration. Mass mixing-ratio contours are plotted for ice + liquid cloud (solid contours), rain + graupel (short dashed contours) and snow (long dashed contours). The mass mixing-ratio values plotted for each category are 0.005, 0.05 and 0.5 g kg$^{-1}$, respectively.
contours of ice + liquid cloud, graupel + rain and snow mass mixing ratios. This cell was associated with core vertical velocities approaching a maximum of 12 m s$^{-1}$. In the later stages of a deep convective event much of the liquid cloud content at lower levels has been swept out by falling rain and graupel or has evaporated, leaving relatively small amounts. An example of shallow convection is also present at $x = 80$ km. Additionally, in the anvil region snow mixing ratios are an order of magnitude greater than those of cloud ice, since the microphysical scheme quickly converts ice to snow. This could mean that the vertical distribution of upper-tropospheric convective moistening by this microphysical model is more homogeneous than in nature, since snow has an appreciable fall velocity relative to ice cloud. The snow anvil associated with this event is approximately 25 km in diameter at this particular time. Since the ice has a fall speed of zero the contours of ice give a good indication of the entrainment levels of deep convection.

The structure of the convection in the equilibrium state does not occur in a completely random fashion but in fact adopts a degree of self-organization despite the absence of imposed vertical shear. After approximately four days of apparently random convection, the cells organize themselves into a band structure and the model mean winds develop a very weak vertical shear of around 0.15 m s$^{-1}$ km$^{-1}$ (Fig. 5) which appears to be a steady feature of the equilibrium state.

Figure 6 shows a series of four snapshots of the moisture in the lowest model on day 60 of the integration separated by 45 minutes; the band structure is clearly seen running diagonally across the domain, orientated roughly parallel to the weak mean vertical wind shear. Although the band is most often found with this orientation it propagates across the domain and rotates, and often lies perpendicular to the mean shear. In other orientations the band sometimes loses its coherency. While the existence of the band and the mean wind shear is likely to have a physical interpretation, its common orientation of 45° could be a result of numerical effects. On the other hand, the fact that there exists only one band in the domain may indicate that the domain size is artificially restricting this organization and could also be the cause of this preferred orientation. The horizontal velocity vectors
at the lowest model level are also shown, to highlight regions of convergent and divergent flow. The contours indicating the positions of deep convective cells reveal that the deep convective activity is completely confined to the band of high moisture values at the 6 km model level. This moist band corresponds to the highest moist static energy ($h$) of this bottom model layer (not shown) and so the convection is occurring in the region where boundary-layer parcels are most energetic. The moist-band air is slightly cooler than that in the adjacent 'dry band' and so it seems that in this experiment it is moisture, rather than temperature, that determines the location of convective activity.

The propagation of the cells within the moist band is apparent. Animations have revealed that new convective cells are initiated as a result of the interaction of downdraughts from older cells with the boundary layer, at or just within the boundaries of the spreading cold-pool air. The speed of propagation varies considerably, with the convective activity sometimes almost stationary and at other times propagating right across the domain in less than two hours.

(i) Surface-flux organization. In Fig. 6 the square box highlights a dry area that has been caused by earlier downdraught activity resulting from a particularly vigorous cumulus event. Examining this part of the domain at subsequent times reveals the fast recovery time of the boundary layer in the vicinity of convection. In Fig. 6(a) the mixing-ratio value of this highlighted cold pool is similar to that of the dry band. However, since surface winds here are higher the latent-heat flux is also higher, and 135 minutes later (Fig. 6(d)) the cold-pool water-vapour content has increased by around 3 g kg$^{-1}$.

This wind sensitivity of the surface fluxes suggests a mechanism for the localization of the convection. The local time rate-of-change of the moist-static energy of the boundary layer away from the direct influence of convective downdraughts (i.e. not within cold-pool areas) can be simply written in terms of the surface fluxes, entrainment of air from above the boundary layer and radiative cooling, with horizontal advection ignored:

$$\frac{\partial h}{\partial t} = C_D |V_{bl}|(h_{bl} - h_0) + w_e(h_{bl} - h_m) + c_p Q_{rad}, \quad (7)$$

where $w_e$ is the vertical velocity above the boundary layer. The subscripts $bl$, 0 and m indicate quantities at the model bottom level, the sea surface and above the boundary layer, respectively. The surface fluxes have been written in terms of the bulk aerodynamic formula, with $C_D$ representing a constant drag coefficient, and $V$ is the wind velocity. The time rate-of-change of $h$ can be ignored to estimate the boundary-layer quasi-equilibrium value as in the ideas of Emanuel et al. (1994), Raymond (1995), and Emanuel (1995). Within the dry band, away from convection, surface wind speeds $V$ are much smaller, indicating that the local boundary-layer equilibrium value of $h$ is also lower. Thus convection continues to be less likely to occur in this part of the domain. This mechanism could localize convection, but the presence of the background mean wind is necessary to 'smear' the high moisture values into the band structure observed.

(ii) Radiation organization. It has been suggested previously that radiation may interact with convection by (i) large-scale destabilization of the environment by infrared (IR) emission to space (in conjunction with surface fluxes) (Dudhia 1989); (ii) by destabilizing cloud layers due to cloud-base (top) warming (cooling) in the IR spectrum (Webster and Stephens 1980); and (iii) by differential horizontal heating rates between cloudy and clear-sky areas, causing a secondary circulation with convergence into cloudy regions (Gray and Jacobson 1977). However, horizontally uniform large-scale IR cooling can not produce localization of convection. Interactions between radiation and convection have
Figure 6. Water-vapour mixing ratio in the lowest model layer taken at (a) day 60 of the experiment and (b) 45, (c) 90 and (d) 135 minutes later. Horizontal velocity vectors are overplotted (the maximum horizontal velocity in all four pictures does not exceed 7 m s\(^{-1}\)). The contours show vertical velocity at 6 km exceeding 1 m s\(^{-1}\) to indicate the areas of deep convective activity within the domain. The square box highlights a dry ‘patch’ within the moist band caused by the downdraught of an earlier deep convective event.

Previously been investigated in cloud-resolving models by Xu and Randall (1995b), Tao et al. (1996) and Dharssi et al. (1997), and all three studies indicated that in-cloud vertical differential heating rates (mechanism (ii)) had a bigger effect on precipitation rates than clear–cloudy horizontal differences (iii). In contrast, Craig (1995, 1996b) showed mechanism (iii) to be important in axisymmetric simulations of warm-core convective systems (tropical cyclones and polar lows). However, these studies all used a 2D CRM and therefore were unable to examine the organization of convection that may occur with horizontally inhomogeneous interactive radiation in three dimensions.

(iii) Numerical experiments. To investigate this further two further shorter model runs were performed. In the first, all features of the CRM were identical to the control run except that the surface fluxes of heat and moisture are calculated as a Newtonian relaxation towards the sea surface state, with the rate constant based on the bulk aerodynamic formulae:

\[
F_{Rl} = \frac{C_D V_*}{c_p} (s_{bl} - s_0)
\]  

(8)
Figure 7. Water-vapour mixing ratio in the bottom model layer for: (a) day 70 of the control run using interactive radiation + wind-sensitive surface fluxes, (b) after 4 days simulation with interactive radiation + wind-insensitive surface fluxes and (c) after 4 days simulation using non-interactive radiation and wind-sensitive surface fluxes. Runs (b) and (c) were initialized using the organized state shown in (a). The domain is repeated twice in each direction for clarity. The colour scale is as for Figure 6.

\[ F_{qv} = C_D V_s (q_{v1} - q_{v0}). \]  

(9)

Here \( F \) is the surface flux of the subscript quantity, where \( T_L \) is the liquid-water temperature (the model's thermodynamic variable), and \( q_v \) is the mixing ratio of water vapour. The surface flux drag coefficient, \( C_D \), is taken as \( 1.1 \times 10^{-3} \), \( V_s \) is the surface scale velocity assumed to be 10 m s\(^{-1}\) (to give similar mean surface fluxes to the control run) and \( s \) represents the dry-static energy. The second run used the original calculation for surface fluxes but replaced the interactive radiative scheme with horizontally uniform non-interactive radiative-cooling rates taken from the control run.

Both runs were initialized using the model state at day 70 of the control run with the band structure and weak vertical wind shear already present. The water vapour from the bottom model level is shown in Fig. 7 for day 4 of each of these short experiments along with day 70 of the control run for comparison. The image is repeated twice in each direction to make any organization present easily apparent. In the run with wind insensitive fluxes (Fig. 7(b)) the coherent structures are larger scale than in the case of fixed radiation (Fig. 7(c)) but it is seen that remarkably, the strongly banded organization that was such a robust feature of the control run, is destroyed in both cases. The vertical wind shear did not seem to decrease significantly in either case during the short runs. It thus appears that radiation could be interacting via dynamics with surface fluxes, by increasing convergence into convective areas. Whether the influence of radiation is via horizontal or vertical heating-rate gradients is beyond the scope of these experiments and further work must be conducted to determine this. However, it is definitely the case that both cloud-altered radiative forcing and wind-sensitive fluxes are required to produce spontaneous convective organization.

(c) Vertical structure of the equilibrium state

To examine the vertical structure of the equilibrium atmosphere, a tephigram of the domain-averaged temperature and vapour profiles for the last 20 days of integration is shown in Fig. 8. Three features worthy of note are apparent:

- It can be seen that the model has a resolved boundary layer that is dry-adiabatic and almost constant in \( q_v \). This resolved boundary layer appears to be approximately
400 m in depth, similar to, but slightly shallower than observations. This characteristic was absent from the $q_v$ profiles of L93 and H93.

- The temperature profile is roughly moist-adiabatic, although conditional instability is in evidence.
- Enhanced values of $q_v$ are seen in the 200 to 250 hPa layer, suggesting that the majority of deep convective clouds detrain at this level, as seen for example in Fig. 4. This is below the temperature minimum, situated at the 100 hPa level. This behaviour of convection detraining substantially below the conventionally defined tropopause height has been noted both in observations and in column-model studies (Forster et al. 1997; Highwood and Hoskins 1998).

To focus on the vertical moisture structure more closely, the average equilibrium relative humidity with respect to water and ice as a function of height is shown in Fig. 9. The distribution of relative humidity (RH) as a function of height is not unreasonable when compared to observed soundings (for example Liu et al. 1991; Sun and Lindzen 1993), although the lower stratosphere appears too moist with an RH of 95% with respect to ice. This was found to be partly attributable to an inaccuracy that was discovered in the calculation for the saturation mixing ratio in the model. Calculations have shown that the correction of this error reduces this peak to around 40% and 80% RH with respect to water and ice. More interesting is the bulge between 9 and 13 km which was identified as the possible region of detrainment from the tephigram.

The bulge in RH is also consistent with the cloud fraction in the model shown in Fig. 10. If the sum for all the hydrometeors' mixing ratios (liquid cloud, rain, snow, ice and graupel) exceeds 0.005 g kg$^{-1}$ then it is identified as a 'cloudy' grid point. This threshold value was also used previously by S94 to identify cloudy grid boxes. Two peaks in the cloud fraction are seen at the 930 and 250 hPa levels, corresponding to shallow convective activity and the detrainment level of deep convection. The anvil cloud detrains
Figure 9. Domain-average relative humidity with respect to both liquid water (solid line) and ice (dashed) as a function of height.

Figure 10. Fractional cloud coverage as a function of height for the last 20 days of simulation.
at the same height as the RH bulge and covers about 6% of the model domain. Larger anvil coverage is not necessarily expected since the clouds do not experience large vertical wind shears in this experiment. The values shown are reasonably consistent with the numerical experiments of Robe and Emanuel (1996) except below 950 hPa where the cloud fraction here is higher. This is mainly due to the fact that rain was taken into account in the determination of a cloudy grid box, resulting in non-zero cloud fractions in the lowest model levels.

Figure 11 examines the radiative-heating rates of this equilibrium state, showing the profiles for the clear and cloudy regions, with the cloud regions defined using the microphysical criteria as above. The heating rates compare well with those in S94. The difference between the two regions is not as distinct as one might expect owing to the inclusion in the cloud statistics of many ‘marginal’ cloud columns containing small but nevertheless significant cloud amounts. The clear-sky heating rates are very representative of the domain-average rates due to the low cloud fractions, and unlike previous 2D studies which had much higher cloud fractions. The domain-average net radiative-cooling rate that is also shown in Fig. 11 drops off sharply above the 200 mb level, becoming zero at around 150 mb. Since, over long time-scales in an equilibrium state, the convective vertical mass flux is constrained to balance radiative cooling (ignoring large-scale effects which are anyway absent from this experiment), this radiative profile is consistent with the picture of clouds detraining at 200 mb with only the occasional event penetrating to higher altitudes. The average top-of-atmosphere long-wave flux is 287 W m⁻² in the equilibrium state, agreeing well with the observations of Raval et al. (1994), with a short-wave albedo value close to 18%. The term ‘top-of-atmosphere’ is applied to the top radiation model level at 0 hPa.

Attention has been focused recently on the structure that cumulus convection in the tropics is expected to take. Renno and Ingersoll (1996) gave a theory that treated the atmosphere as a heat engine and which gave a prediction for the fractional area covered by
convection ($\sigma_c$) of 0.05%, and Craig (1996a) extended this analysis giving a similar estimate for $\sigma_c$. Emanuel (1994) used a different scaling argument to give an estimate of $\sigma_c = 0.3\%$. In the equilibrium state the convective fraction (defined as the fraction of vertical columns with at least one convective grid point with an absolute vertical velocity exceeding 1 m s$^{-1}$) is 1.7%. Examining $\sigma_c$ as a function of height for updraught, downdraught and total convection fractional area (Fig. 12) it is seen that total $\sigma_c$ lies between 0.25% and 0.55% for the whole of the free troposphere. These figures are in close agreement with the numerical experiments of Robe and Emanuel (1996). The downdraught convection-fraction curve has two distinct peaks at 12 km and at 2 km, the latter representing the downdraught that develops below the convective turret as the precipitation evaporates when falling through unsaturated air. It is these downdraughts that directly affect the boundary layer. However, it is perhaps surprising to see also a relatively large downdraught fraction at the 12 km level. These downdraughts are formed from the outflow of deep convective events. The air detrained from the clouds contains ice and snow which evaporates as the air leaving the region of ascent mixes with environmental unsaturated air, forming these high-altitude downdraughts that normally remain until around 8 km in altitude. These downdraughts, along with slower but wider mesoscale downdraughts, are an essential feature in the way that deep convection moistens the tropical troposphere above the boundary layer (Sun and Lindzen 1993). High-altitude oceanic convective observations are rare and so it is difficult to validate the structure of these upper-level downdraughts. For example, none of the observational studies of LeMone and Zipser (1980), Jorgensen and LeMone (1989) or Lucas et al. (1994) reported statistics above 8 km altitude.

Although, in terms of fractional area, convective downdraughts are comparable to updraughts, their mass transport is considerably smaller since the core velocities are significantly lower (Fig. 13). At the top of the boundary layer the ratio of the downdraught to the updraught mass flux is approximately 0.05, rising to a maximum of 0.2 at around 2.5 km. Since the reduction in downdraught mass flux below 2 km is mostly due to mass continuity as the outflow spreads horizontally at the surface, this latter value of 0.2 is more
Figure 13. Vertical mass flux of convective updraughts and downdraughts (defined using the 1 m s$^{-1}$ vertical velocity criterion), and the net convective mass flux for the mean state.

representative of the value that would be used in parametrization schemes that utilize this ratio as part of a closure (e.g. Raymond 1995). In comparison to the observation studies of LeMone and Zipser (1980), Jorgensen and LeMone (1989) and Lucas et al. (1994) the ratio of the downdraught to updraught mass flux is relatively small, which may be attributable to the upright nature of the convective cores in our non-sheared environment, in contrast to many of the observed convective events which were part of organized convective systems (Lucas et al. 1996). Cheng (1989a,b) for example, proposes that increasing the tilt of deep convective updraughts will increase convective-scale downdraught activity, and found little downdraught activity in cases of scattered convection in his diagnostic analysis of GATE tropical convection observations.

5. CONCLUSIONS

In order to examine the role of cloud and water-vapour feedbacks and the water-vapour cycle in the tropical climate, a cloud-resolving model with a fully interactive radiation scheme has been integrated to an equilibrium state. The experiment was conducted over a tropical ocean of fixed specified properties in a three-dimensional box that was sufficiently large to represent the subsidence associated with convection in this experimental set-up and to allow an ensemble of convective cells to exist continuously in the final equilibrium state. The lateral boundary conditions were cyclic and no ‘large-scale’ forcing terms were specified, so that the convection is forced only by radiative cooling of the atmosphere and fluxes of heat and moisture from the surface. Thus this experiment represents a simple system in which all the processes such as convective heating, mixing, water transport, radiative heating and surface fluxes are explicitly represented, allowing a two-way interaction between them. The simplicity of this system will, it is hoped, lead to a greater understanding of the interactions between this self-contained set of processes.

The aim of this paper has been to describe the model and the experimental set-up in some detail, to illustrate that a robust equilibrium state has been achieved and to describe
the structure of the equilibrium state, highlighting features of the structure and organization of the convective cells themselves.

In the approach to equilibrium, two time-scales were clearly apparent, a short cumulus time-scale of a few hours, associated with the life cycle of individual cumulus elements and a longer time-scale of around 30 days. These time-scales were also observed in earlier two-dimensional radiative-convective equilibrium studies also using cloud-resolving models. This longer time-scale could result from the radiative subsidence time-scale in the tropical clear-sky regions. Further experiments to investigate the mechanisms that determine the various time-scales will be published separately (Tompkins and Craig 1998b).

The convection was shown not to be completely random in the equilibrium state, but had instead evolved some degree of self-organization. This was in the form of a ‘band of convection’ associated with a moist channel in the boundary layer. The convection propagates throughout this moist channel most probably via the action of downdraughts lifting boundary-layer air and increasing surface fluxes. It would appear that the self-organization of convection occurs, at least in part, because of the interaction of higher winds associated with convective horizontal mass convergence/divergence and the wind-sensitive surface fluxes. In the areas away from convection, the surface wind speeds are lower and thus the equilibrium moist-static energy value for the boundary layer is also lower, prohibiting further convection. It was also seen, however, that radiation also acts to localize convection, possibly by increasing convergence into cloudy regions, enhancing the surface-flux effect. Replacing the wind-sensitive surface-flux scheme with Newtonian relaxation terms for heat and moisture fluxes, or imposing a horizontally constant radiative-heating rate instead of the interactive scheme, destroyed the convective organization after four days. This interaction between surface fluxes, radiation and convective dynamics is obviously complex and requires further investigation to be fully understood.

An equilibrium was achieved at all heights of the model atmosphere including the lower stratosphere, and in all model quantities including water vapour and all microphysical quantities. The equilibrium state had a column-integrated vapour amount of 42.3 kg m\(^{-2}\), a mean temperature of 257.8 K and a pseudo-adiabatic CAPE value of 1900 J kg\(^{-1}\). Despite the apparently simple experimental set-up, the profile of relative humidity, column-integrated humidity, and temperature profiles are consistent with mid-Pacific observations.

The agreement of water-vapour statistics with observations in this study is significant, since prediction of equilibrium water-vapour profiles is actually very difficult for models to achieve, and thus it gives confidence that this model represents a tool with which much can be learnt about the sensitivities of the tropical water cycle to various external forcings. With this knowledge, two companion papers will examine further aspects of the CRM’s equilibrium state. The first will examine the sensitivity of the equilibrium status to perturbations in SST, allowing the feedbacks of cloud and water vapour to be calculated (Tompkins and Craig 1998a). The second study will focus on the mechanisms for both the short and long time-scales involved in the achievement of equilibrium status for the various model fields, which will hopefully be of relevance to the issue of cumulus parametrization (Tompkins and Craig 1998b).

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