

## EQUILIBRIUM VS. ACTIVATION CONTROL OF LARGE-SCALE VARIATIONS OF TROPICAL DEEP CONVECTION

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### Abstract

What processes control large-scale variations of deep convection (LSVDC) in the tropics? Here ‘large-scale’ is taken to mean any coherent variations, in either space or time, comprised of statistical populations of separate convective cloud systems. This essay highlights the distinction between processes which supply moisture or available energy over the depth of the convecting layer (*equilibrium control*), versus inhibition and initiation processes at low levels (*activation control*), as hypotheses for explaining LSVDC.

Conceptual separations of the LSVDC problem are reviewed. Scale separation, though rigorous, is artificial, since net heating makes deep convective clouds multiscale, or spectrally red. Moist-dry, or diabatic-adiabatic, separation is more useful. An ill-posed hybrid separation - ‘the interaction of moist convection with large scales’ - has spawned confusion. Correlations between deep convection and its own large-scale components (suggestively labeled ‘forcing’) have been misinterpreted as evidence for equilibrium control. This externalization of large-scale vertical velocity also encourages overinterpretation of a fictitious ‘compensating subsidence’ term.

Published evidence for equilibrium control theories is critically re-examined. The deep-cloud quasi-equilibrium observations of Arakawa and Schubert would hold for arbitrarily determined variations of convection, because stratified fluid dynamics efficiently redistributes localized heating, not because convection is controlled by slow deep large-scale ‘forcing.’ A more sensitive test indicates that most tropical LSVDC are not forced by preexisting deep upward motions.

Activation-control processes can operate on the large space scales and long time scales that define LSVDC. Modulation of convection by easterly waves and upper-tropospheric troughs, and the climatological distribution of convective cloudiness, are examined as examples. Lower boundary flux enhancements and deep lifting exert both equilibrium and activation influences on convection. The hypothesis that activation control may prevail on all scales short of globally-averaged climate is difficult to refute.

Systematic study of convective cloud-ensemble sensitivities is badly needed. Unfortunately, current cloud ensemble modeling strategies, based on unnatural equilibrium-control assumptions, render the results merely diagnostic.

To represent activation controls, large-scale models need multiple levels, and some prognostic representation of the subgrid-scale inhomogeneity of low-level fields.

## 1. Introduction

Precipitating deep convection is a structural element of the global climate system, linking the incoming solar radiation stream, which is primarily absorbed by the earth's surface, to the outgoing longwave radiation stream, which is largely emitted by the atmosphere. The mean thermodynamic state of the troposphere is determined by a bulk radiative-convective equilibrium, in which net radiative cooling is balanced by latent heat release, primarily in deep convective clouds (e.g. Riehl 1954, p364). In this climatic balance, convection is obedient and responsive to the rather more inflexible radiative forcing, as evidenced by the fact that the mean stratification is nearer to a moist adiabatic than to a radiative-equilibrium profile (Emanuel 1994, p476).<sup>1</sup>

On sub-global scales, however, the intensity of tropical deep convection varies on a wide range of space and time scales that are large with respect to individual clouds and even with respect to mesoscale convective systems. It is these *large-scale variations of deep convection* (LSVDC) that we seek to understand. Large-scale gradients of latent heating (as indicated by precipitation, moisture convergence, 200 mb divergence, upward motion, high cloudiness, outgoing longwave radiation, and other largely redundant indices, here lumped under the term "deep convection") drive deep, large-scale circulations in the vertical plane. These circulations include the climatological Hadley and Walker cells and a wide variety of transient overturnings. Fluid dynamical models simulate observed tropical circulations ranging from mesoscale to synoptic to planetary-scale flows quite well, given heating distributions resembling observed convection fields. But to what extent, and how, do these circulations in turn determine the large-scale structure of the convection field? The answer to this question, sometimes called the "closure problem" in cumulus parameterization literature, is the missing link in our understanding of the moist circulations of the tropical troposphere.<sup>2</sup>

The first step in tackling such a question must be a clear definition of terms. Unfortunately, this subject has suffered from a great deal of confusion. For example, some terminology commonly used to discuss the problem ('the interaction of moist convection with large scales', section 2) contains a fundamental conceptual flaw. Briefly, deep moist convection *is* a large-scale phenomenon, even when it takes the form of spatially small cumulonimbus, because precipitation is a positive definite quantity.

This chapter begins, in section 2, with the foundation issue of how to separate the problem of LSVDC into parts for analysis. Scale separation and moist-dry separation are carefully considered. Section 2 may at first seem something of a digression from the titular theme of deep versus low-level controls on convection. But the 'deep' aspects of the

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1. A similar balance holds for the tropical belt alone, to the extent that heat flux convergence is small compared to radiative cooling and latent heating within the region.

2. This discussion focuses on *dynamically induced* variations of deep convection, as opposed to lower boundary induced variations, because the latter are more ambiguous in terms of our titular distinction (see section 6). Furthermore, we focus on convection within ~15 degrees of the equator, whose heating effects are not highly trapped in a local region by geostrophic adjustment processes.

influential equilibrium-control school of thought are hidden within a deep/large-scale/slowly-varying complex of assumptions that must first be dissected.

Section 3 highlights the distinction between inhibitory processes at low levels (*activation control*), and available energy-generating processes operating over the depth of the convecting layer (*equilibrium control*), as competing hypotheses for explaining LSVDC. These two strains of thought have run through convective meteorology since its inception, encapsulated in the very concept of conditional instability, as measured by its two main energy indices: convective inhibition (CIN) or negative area on a thermodynamic chart, and convective available potential energy (CAPE), or positive area. We reserve some flexibility in how exactly to define these quantities.

Section 4 is devoted to a critique of published evidence for equilibrium control theories. Briefly, these theories postulate that dynamical LSVDC are caused by slow, deep, large-scale circulations differentially supplying available energy (or moisture). Convective cloud populations are envisioned as efficient consumers of this differential supply, whence the convection field inherits its nonuniformity. In this sense, equilibrium-control theories could perhaps be termed “supply-side” theories. The evidence for equilibrium control is found to be weak, mainly because insensitive tests have been used. More sensitive tests indicate that equilibrium control is not prevalent in the tropics.

Section 5 summarizes, and advocates wider consideration of, activation-control ideas, which suppose that deep convection frequency is determined by initiation processes, and inhibition thereof, even on quite large scales. A familiar example involves the trade inversion at ~800 mb over the subtropical and tropical oceans (e.g. Riehl 1954 and refs), which largely prevents the formation of deep convection over vast regions of the globe where conditional instability prevails. When a lower-tropospheric disturbance breaks the inversion, deep convection can form rapidly without deep or upper-level forcing. Evidence for the activation-control viewpoint is illustrated for familiar large-scale phenomena such as African easterly waves, an upper-tropospheric trough, and the climatological spatial distribution of convection.

Section 6 discusses ambiguous cases for the equilibrium vs. activation control distinction. For some situations, such as deep lifting ahead of a midlatitude upper-tropospheric short wave, or over positive sea surface temperature anomalies, the two kinds of theories both predict enhanced convection. The equilibrium vs. activation control distinction probably makes an important quantitative difference, but not an easily-observable qualitative difference for convection predictions in these situations. Such situations are therefore not sensitive or discriminating tests of the two competing hypotheses. What is needed is a careful analysis of processes that change CAPE and CIN independently, or at least in variable proportion.

## **2. The interaction of convection with other**

The phenomenon of deep convection is by now well documented (see e.g. Houze 1993, Emanuel 1994). Free buoyant ascent of air from near the surface occurs quite intermit-

tently, usually in cells or bubbles, often organized on the mesoscale, typically along gust fronts. Copious rain precipitates from the ascending air. Convective cells (with aspect ratio of order unity) detrain great masses of cloudy air, creating subsequent and adjacent decks of precipitating stratiform cloud in the middle and upper troposphere, with larger aspect ratios and slower ascent rates. The convective and stratiform precipitation together are described by the general term mesoscale convective system (MCS).

The complementary set of processes with which convection can meaningfully be said to interact must be defined carefully, lest we find ourselves discussing the interaction of a phenomenon with (a smoothed version of) itself. Two main lines of separation suggest themselves: scale separation (section 2.1), and moist-dry, or diabatic-adiabatic, separation (section 2.2). We argue that the latter is more fruitful for scientific understanding. Scale separation appears to align with engineering problems associated with numerical modeling, but is very artificial in terms of the physics of convection. In addition, differences between mathematical and vernacular concepts of ‘scale’ are dangerously subtle. The greatest source of confusion in the field has been the mixing of these two ways of separating the problem (the ill-posed ‘moist-convective - large-scale’ separation, section 2.3).

## 2.1. SCALE SEPARATION

Scale separation is mathematically well-founded, and has been successfully used for decades in the context of dry turbulence. One approach is to choose a reference size, and define as ‘large-scale’ the variation among averages over regions of that size. Residual variation (deviations from the averaged or smoothed values) is called small-scale. Some common reference scales include the size of a natural phenomenon; a deformation radius; the size of an observing system such as a rawinsonde array; or the size of a grid-box in a numerical model. Spectral decomposition, of which Fourier analysis is the most familiar example, can also serve as a source of unambiguous definitions of scale, at least in one dimension. These methods can be used also in the time domain.

Unfortunately, the word ‘scale’ also has intuitive or vernacular meanings that are subtly different from the mathematically precise meanings. For example, autocorrelation falloff is sometimes taken to indicate scale. Are the subtropical high pressure - trade wind systems, with their fabled steadiness and large autocorrelation spanning many thousands of kilometers, inherently larger in scale than the circulations of the equatorial region, where autocorrelation falls off more rapidly with distance? Here the confusion is between *large-scale* and *the absence of small scales*. Since correlation is covariance divided by total variance, simply adding small-scale noise to a field will make autocorrelation fall off faster with distance, although the large scales are unaffected.

Perhaps the most pervasive confusion is between *scale* and *physical dimension*. For example, what is the scale of a precipitating convective cloud 10 km wide? It may be tempting to say 10 km, and thence to dismiss such clouds as small-scale, but that would be most misleading. The net precipitation (or heating) in such a cloud makes a contribution to integrals over any and all reference scales which encompass the cloud. Equiva-

lently, in the Fourier decomposition of a positive-only point heating, all scales are equally present, including zero wavenumber (largest possible scale). *Every precipitating cloud has a large-scale essence*, along with its decorative halo (in Fourier space) of truly small scales, because precipitation is positive definite.

For example, consider the flow in an infinite pool of shallow water after the introduction of a volume source, such as a stone thrown in the pool<sup>1</sup>? Neglecting nonlinearities (such as the turbulent splash, which could be mitigated by a gentler introduction of the stone), the response consists of an expanding ring, well described by linear wave dynamics, that alters the height of the surface to account for the presence of the stone's volume. The expanding *size* of the response would seem, in the vernacular sense, to constitute an 'up-scale' growth of the response with time. But the linear equations governing free shallow water flow permit no energy to change scales. All scales, including the largest scales, are present from the moment of the stone's entry. The subsequent growth of the affected area is simply the phase evolution of the red spectrum of scales excited by the stone. Throwing a stone into a large pond is in part a large-scale event, whatever the physical size of the stone.

We see, then, that the word *scale* must be used, and read, very carefully. In this light, consider the opening sentence of Arakawa and Schubert (1974, hereafter AS74):

“The many individual cumulus clouds which occur in a large-scale atmospheric disturbance have space and time scales much smaller than the disturbance itself. Because of this scale separation, it may be possible to predict the time change of the large-scale disturbance by describing not each of the many individual clouds, but only their collective influence. This is the goal of cumulus parameterization.”

Here the large-spatial-scale aspects of precipitating clouds are denied, or rather defined away, given over to the mysterious entity called a large-scale disturbance. But at the same time, this disturbance is presumed to have a time scale that is slow relative to that of convection. The possibility of a fast, large-scale component of deep convective flows is implicitly neglected. Furthermore, it is implied that clouds cannot be treated collectively without denying the large-scale part of their nature.

The lens of numerical modeling, through which AS74 viewed the problem of LSVDC, encourages a scale-oriented analysis. Every model has a grid scale, and the modeler's premise seems to be that phenomena which are well resolved will be well represented, so long as sub-gridscale eddy effects are properly accounted for. If one is optimistic about the prospects for success, then perhaps it seems reasonable to take all large-scale variables, including vertical velocity, as given or known in the design of sub-grid-scale parameterizations. These scale separation ideas of the early 1970s spread beyond the engineering problem of cumulus parameterization. The GARP Atlantic Tropical

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1. The linear problem of localized heating in a stratified fluid, as by a precipitating cloud, is identical.

Experiment (GATE) convection observation program was designed around the concept of scales and their interactions (Betts 1974b).

The peculiar importance of deep convection to large-scale dynamics comes from the fact that each precipitating cloud contains large-scale latent heating. The peculiar difficulty springs from the fact that this large-scale heating is partly subject to small-scale control. The central scientific challenge in problems of dynamically determined LSVDC<sup>1</sup> is not to formulate the properties of ensembles of small-scale convective eddies *in terms of a known large-scale average vertical velocity*, as assumed by AS74. Rather, we need to understand how known large-scale *state* variables, in concert with small-scale fluctuations that cannot be explicitly resolved, produce the observed amount and type of precipitating convection.

## 2.2. MOIST-DRY SEPARATION

The occurrence and form of convective clouds is intimately linked, by mass continuity, to motions in the surrounding atmosphere. To the extent that the motions of the environment occur in unsaturated air, they are invisible in visual and radar observations. But to that same extent, they are described to high accuracy by equations with very little inherent uncertainty. This suggests the utility of a moist/dry, or diabatic/adiabatic, separation of the problem [e.g. Ooyama 1971, Raymond 1983, Mapes and Houze 1995, Mapes (this volume)]. The dynamics of moist convection are quite complex, owing not only to saturation and phase change effects but also to complications involving the microphysics of condensed water particles. But at least these ‘moist’ motions are vigorous and remotely observable. Here is the hard nucleus of the problem, where attention should be focused. The corresponding ‘dry’ motions are hard to observe, but are highly constrained by inescapable, simple equations.

The usefulness of the view of ‘moist’ convective heating interacting with ‘dry’ dynamics was called into question in the introduction of Emanuel et al (1994, ENB):

Riehl and Malkus (1958)...showed that latent heat release in tall cumulonimbus clouds is an important energy source for large-scale tropical circulations...[their analysis] embodied a view of tropical dynamics that persists today, i.e. a view in which convection is regarded as a heat source for an otherwise dry circulation. We shall argue that this ‘externalization’ of convective heating has had a large and unfortunate effect on thinking about the interaction of moist convection with large-scale flows...

Criticism of the final phrase is the subject of the next section. As for the utility of the dry-moist separation, ENB’s pessimism seems misguided.

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1. For climate questions, in contrast, the eddy fluxes of microphysical species are extremely important for simulations of cloud and water vapor for radiative transfer calculations.

ENB offer instead a view of convection as acting to reduce the (still positive) ‘effective static stability’ or ‘gross moist stability’ felt by large-scale dynamical motions (Gill 1982, Neelin and Held 1987, ENB). These ideas can be used to alleviate the discrepancy between the observed slow phase speeds of dynamical convective variations, such as the intraseasonal or Madden-Julian (1994) oscillation, and the much faster phase speeds of dry waves with the same vertical structure. The reasoning is that latent heating in convection partly balances the adiabatic cooling in regions of upward motion, so the restoring force (buoyancy) felt by waves is reduced, and hence wave speeds are reduced.

But if moisture did indeed modify large-scale wave dynamics in this way, then a separated theory cast in terms of convective heating interacting with dry dynamics would still give the right answers. We might compare this with a physicist’s explanation for why tinfoil is opaque: the light goes right through the foil, traveling as in free space, but the electrons in the metal, jiggled by the light’s fluctuating electric field, emit light of precisely canceling phase. A separated view is not wrong, it would just be inconveniently complex if the moist processes were indeed utterly obedient to the dry. Certainly there is evidence of wave signatures in cloudiness data (e.g. Takayabu 1994), but the relationships are not crisp enough to indicate such obedience. Furthermore, other lines of theoretical reasoning can also be tuned to predict reduced wave speeds.

In summary, the ‘externalization’ of convective heating decried by ENB is actually a framework general enough to encompass any particular closed theory of condensation-modified wave dynamics. To abandon moist-dry separation in favor of one particular version of moist dynamics therefore seems unwarranted. In fact, it may be argued that moist-dry separation has never been taken to its full theoretical fruition, mainly because the relevant sensitivities of convective cloud ensembles to environmental conditions have never been properly assessed.

### 2.3. CONFUSION OF MOIST-DRY AND SCALE SEPARATIONS

The mixing of different types of separations has been a source of considerable confusion. For example, if the *diabatic* upward motion in spatially small precipitating cumulonimbi is implicitly considered *small-scale*, then the inevitable correlations between convection and large-scale vertical motion are misinterpreted as ‘control’ of the former by the latter (Fig. 1, section 2.3.1). When *large-scale* upward motion is tacitly reinterpreted as *adiabatic* (or environmental) upward motion, a fictitious term, popularly described as ‘compensating subsidence,’ is mistakenly elevated to the status of physical reality (section 2.3.2). The effects of these two misconceptions cancel each other in artificially-forced cloud ensemble models with reflecting or periodic boundary conditions (section 2.3.3).

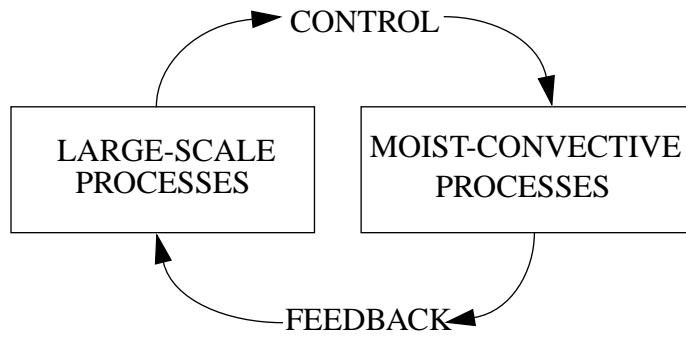


Figure 1. Schematic of the interaction between large-scale and moist-convective processes. Adapted from Arakawa (1993).

Figure 1 is the first figure, on the second page, of the 1993 AMS monograph on the representation of convection in numerical models (Arakawa 1993), so can fairly be taken as influential enough to warrant criticism. Two non-overlapping boxes, labeled “moist-convective processes” and “large-scale processes,” are connected by arrows labeled “control” and “feedback.” Similar diagrams, with more detail, may be found in Schubert (1974), who credits Betts (1974b) for the feedback-control terminology, and in Hack et al. (1984).

Suppose we ask, which box contains net precipitation, the (for some purposes) most important process in deep convection? If the words in the right hand box are covered up, then the diagram shows large-scale processes interacting with the other, nonoverlapping box. This appears to be a scale separation, so as discussed in section 2.1, the net precipitation (by definition of ‘net’) is in the large-scale processes box. But if we cover the words in the left hand box, the diagram now seems to be describing a moist-other, hence moist-dry, separation. Here the net precipitation lies in the *moist-convective* box. In short, although the two boxes are drawn in a nonoverlapping manner, *large-scale processes and moist-convective processes are not mutually exclusive: precipitation is both.*

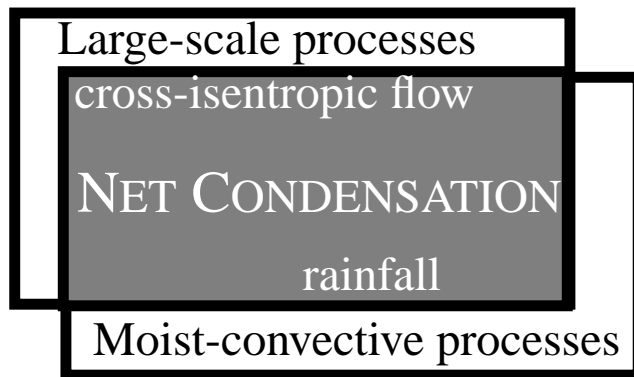


Figure 2. Suggested overlapping version of Figure 1.



To summarize, a more appropriate diagram of overlapping large-scale and moist-convective processes in a tropical convecting region might look like Figure 2. Figure 3 shows the two ways, discussed above, of breaking the problem into mutually exclusive parts: scale separation, and moist-dry separation.

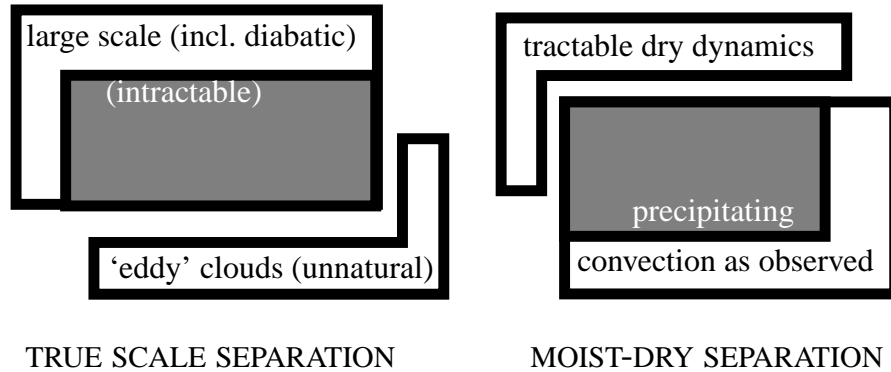


Figure 3. The two ways of separating the dynamical LSVDC problem into mutually exclusive, interacting parts.

One additional distinction is worth making in the context of the moist-dry separation: between divergent and rotational dry dynamics. Although this distinction is formally untidy, qualitatively one can associate rotational dynamics with horizontal wind circulations as seen on synoptic charts, and divergent dynamics with vertical circulations. Deep convection is linked directly with deep divergent flows. Although rotational circulations have slow characteristic time scales, e.g. they persist from day to day on weather maps, we shall see in section 4.2.2 that dry divergent flow has a fast time scale, even on large spatial scales.

### 2.3.1. Interpretation of large-scale / moist-convective correlations

The form that precipitation takes in the *large-scale processes* box is a net condensation, observable with large-scale data as both an apparent heat source (flow across smoothed isentropes), and an apparent moisture sink (flow across smoothed specific humidity isopleths). Isentropes, and to a lesser but considerable extent specific humidity surfaces, are nearly flat, level, and unchanging in the tropics. As a result, the cross-isentropic flow is, within observational errors, well captured in diagnoses of vertical velocity, whether in isentropic, pressure, or height coordinates. Any of these vertical velocities, either at a midtropospheric level or integrated over height, whether plain or multiplied by the vertical gradients of entropy or moisture to yield an apparent large-scale advection, contains the large-scale signal of net precipitation in deep convection. In the *moist-convective processes* box, precipitation appears more explicitly: it might be measured by radar, or by a network of rain gauges.

In the diagnostic community, radar and rawinsonde-array precipitation estimates appear to be about equally weighted in the search for a consensus rainfall estimate,

sometimes leaving the ostensibly most direct estimates from raingauges as outliers. Here the inevitable temporal correlation between large-scale deep upward motion (or moisture convergence) and rainfall is properly viewed as a test of our various techniques for measuring or inferring area-integrated precipitation.

The problem comes when these same sorts of temporal correlations are interpreted as evidence for an *a priori* presumption about the causal links responsible for LSVDC. In particular, when large-scale / moist-convective correlations are interpreted according to the flawed conceptual model of Fig. 1, one is led to the erroneous conclusion that large-scale processes (deep upward motion in particular) ‘control’ LSVDC. Confusion over this issue is illustrated by Wang and Randall’s (1994) replacement, in the context of such an analysis, of the term *large-scale processes* with *nonconvective processes*, despite their recognition that area-averaged deep upward motion in a convection-containing GATE rawinsonde array was the dominant source of variability in this ‘nonconvective’ term. This substitution not only ignores, but actually obscures the semantic clue to, the nonexclusivity of the moist-convective and large-scale categories.

Discussing the ‘interaction’ of area-averaged upward motion with ensembles of precipitating convective clouds is akin to discussing the interaction of traffic flow with the ensemble of moving cars. Since traffic consists mostly of cars, the two are highly correlated, but does it follow that traffic flow ‘controls’ car-ensemble flux? Would focusing on the correlation between traffic and car flow facilitate the discovery of useful dynamics, say that southbound left-turning trucks dispatched by a certain company were limiting traffic (and car) flow at rush hour?

### 2.3.2. *Do clouds modify their environment through ‘compensating subsidence?’*

The latent heat released in precipitating clouds is exported (low static energy air flows in in the lower troposphere, while high static energy air flows out aloft), to gently warm large distant regions of the atmosphere. The form of the warming caused by convective heating is a non-steady expanding pattern of downwelling wavefronts in the stratified environment (Bretherton and Smolarkiewicz 1989, Nicholls et al 1992, Bretherton 1993, Mapes 1997). Unfortunately, the terms ‘compensating subsidence’ and ‘cloud-induced subsidence,’ which might appear to describe this response, were taken long ago to describe a mathematical term that arose as a definitional artifact in literature on the interaction of convection with large scales (Ooyama 1971, Yanai et al 1973, AS74). This section examines the conceptual roots of this term, in the tacit ‘externalization’ of large-scale vertical velocity from convection.

Ooyama (1971) wrote down conservation equations for variables describing the large-scale environment of convective clouds in the tropics. An overbar was used to denote these *environmental* average variables. He then derived terms for these equations expressing the effects of an upward mass flux in an implicit population of embedded precipitating convective clouds, assumed to be of negligible physical size.

Under this small-size assumption, area-averaged and environmental values of *state* variables are approximately equal, since in-cloud and out-of-cloud values aren’t radically different. However, this is not true of vertical velocity, because in-cloud vertical mass

flux is important no matter how small the clouds are assumed to be. Here Ooyama introduced an interesting exception to his notation. For all other variables, an overbar indicated *environmental values*. But for vertical velocity, an overbar was used to indicate a *total area average*, including the convective cloud mass flux. This large-scale vertical velocity  $\bar{\omega}$  in pressure coordinates (equal to -gravity times vertical mass flux) is then  $\bar{\omega} = \omega_e + \omega_c$ , where *for  $\omega$  only* the subscript e is used to denote the environmental value (eq. 22 of Ooyama 1971). The in-cloud mass flux is represented by  $\omega_c$ . Then the vertical advection of an arbitrary environmental quantity  $\bar{\alpha}$  becomes

$$\omega_e \frac{\partial}{\partial p} \bar{\alpha} \equiv \bar{\omega} \frac{\partial}{\partial p} \bar{\alpha} - \omega_c \frac{\partial}{\partial p} \bar{\alpha} \quad (1)$$

This permits the generic conservation equation for an environmental variable  $\bar{\alpha}$ , with source  $S_\alpha$ , to be rewritten in the form:

$$\frac{\partial}{\partial t} \bar{\alpha} + \bar{V} \cdot (\nabla_h \bar{\alpha}) + \bar{\omega} \frac{\partial}{\partial p} \bar{\alpha} = \omega_c \frac{\partial}{\partial p} \bar{\alpha} + S_\alpha \quad (2)$$

Because of the notational irregularity, the left hand side now looks just like an ordinary expansion of the material derivative, while a new term has appeared on the right. Ooyama called this term:

...a virtual source of  $\bar{\alpha}$  when the large-scale budget is formulated in terms of the “mean” vertical velocity  $\bar{\omega}$ .

He noted that other forms for writing the equation were more “convenient for interpretation of physical processes.”

If the distinction between *large-scale* and *adiabatic* (environmental) vertical velocity is ignored, this ‘virtual’ term on the RHS of (2) is mistakenly elevated to the status of physical reality. Later authors seem to have associated this term with a more literal image of advection by compensating subsidence in the environment between convective clouds. For example, Arakawa and Schubert (1974) stated:

...cumulus clouds modify the environment through the cumulus-induced subsidence,  $-M_c$ , in the environment.

In their context, ‘environment’ explicitly meant *local* environment, while the word ‘induce’ is defined in the Random House dictionary as “...to bring about, produce, or cause.” Yet no equation governing vertical velocity has been considered in the derivation. How did a diagnostic mathematical statement come to be associated with a causal verbal statement? The answer lies in interpretive assumptions about  $\bar{\omega}$ .

It is an interesting foray into scientific culture to consider how the word “subsidence” became attached to this term. Algebraic forms with negative signs appearing

before the cloud mass flux seem to have been carefully arranged. In AS74, the “ $-M_c$ ” quoted in the excerpt above actually appears with an additional negation, as  $+M_c \frac{\partial}{\partial z} \bar{\alpha}$ , when substituted into the final equations (their 29-30) of form (2). In Yanai et al (1973), pressure coordinates were used for the vertical gradient, but the symbolic substitution  $M_c = -\omega_c$  rendered the term  $-M_c \frac{\partial}{\partial p} \bar{\alpha}$ . The sense that this term reflects physical action by clouds was perhaps bolstered by the fact that both these papers arrived at the term only after detailed derivations involving equations for ensembles of model clouds.

To fix ideas, consider the case of the thermodynamic equation ( $\alpha$  becomes potential temperature  $\theta$ ) for a region containing strong tropical convection. Since local change, horizontal advection, and environmental  $\theta$  sources (mainly radiation, but also evaporation of detrained condensate) are all considerably smaller, the dominant balance in (2) for this case is:

$$\bar{\omega} \frac{\partial}{\partial p} \bar{\theta} \approx \omega_c \frac{\partial}{\partial p} \bar{\theta} \quad (3)$$

The large-scale apparent advection term on the left and the term on the right are both positive for this case, with  $\bar{\omega} \sim \omega_c < 0$ , indicating upward motion. Why would one associate the word “subsidence” with the term on the right? It is *the upward convective mass flux itself*, converted to isentropic coordinates. Of course, we know on physical grounds that if air goes up in clouds, there must be compensating subsidence somewhere. But equations (2)-(3) say nothing about that.

Origins of the ‘compensating subsidence’ image can be glimpsed in Ooyama’s (1971) unnumbered equation between his (28) and (29). He notes that if one sets  $\bar{\omega} = 0$  then the thermodynamic equation becomes (neglecting horizontal advection and  $S_\alpha$ )

$$\left. \frac{\partial}{\partial t} \right|_p (\bar{\theta}) = \omega_c \frac{\partial}{\partial p} \bar{\theta} \quad (4)$$

where the subscript p refers to local change at constant pressure. Now (4) does describe clouds warming their environment through local compensating subsidence. Converting

to isentropic coordinates clarifies the statement:  $\left. \frac{\partial}{\partial t} \right|_p (\theta) = -\left( \frac{\partial \theta}{\partial p} \right) \left. \frac{\partial}{\partial t} \right|_\theta (p)$  so (4) becomes

$$\left. \frac{\partial}{\partial t} \right|_\theta (p) = -\omega_c > 0 \quad (5)$$

Surfaces of constant  $\theta$  descend (increase their pressure) at a rate equal and opposite to the upward mass flux in clouds. But this instantaneous relationship of isentropic

descent to cloud mass flux is not contained in the thermodynamic equation in form (2). Rather, it is a consequence of introducing an additional, very strong, unjustifiable vertical equation of motion:  $\overline{\omega} = 0$ .

More generally, *the image of local cloud-induced subsidence arises from the conceptual externalization of  $\overline{\omega}$*  - that is, from the large-scale - moist-convection separation that is the subject of this section. AS74 seems to be based on the fundamental presumption that because  $\overline{\omega}$  is ‘large-scale,’ it is a known quantity in the context of a GCM, as if its fluctuations were entirely determined by the part of the atmosphere outside the convection-containing grid box in question. Under this presumption, the observed fact that  $\omega_c = \overline{\omega} - \omega_c$  remains small might appear to be a powerful constraint on  $\omega_c$ , a firm foundation for the parameterization of  $\omega_c$  in terms of  $\overline{\omega}$ . This presumption is at the root of their quasi-equilibrium hypothesis, and of equilibrium-control theories generally.

However,  $\overline{\omega}$  averaged over any limited area of the atmosphere has an internally determined (diabatic) part that, in the convecting tropics, is much larger than the externally determined (adiabatic) part. Physically, when upwelling and downwelling wavefronts, launched by recent convective heating changes in the area, cross the area’s boundary, then  $\overline{\omega}$  changes. The only way  $\overline{\omega}$  could be externally controlled is if the external dynamics were a special and elaborate function of the internal. For example, the external dynamics might have incoming wavefronts that are mirror images of the outgoing wavefronts just inside the area’s boundary. In this case the words “externally determined” would hardly seem to apply. In fact, the case (4)-(5) where local compensating subsidence really exists ( $\overline{\omega} = 0$ ) can only be realized with such a ‘reflecting’ boundary surrounding a convecting region, or equivalently with periodic boundary conditions.

### 2.3.3. *Comments on cloud ensemble modeling methodology*

The growing field of cloud ensemble modeling (CEM) has embraced, for reasons of computational convenience, the problematical notions of large-scale ‘forcing’ and ‘compensating’ subsidence. In a typical CEM experiment, area-averaged deep upward motion<sup>1</sup>, as measured by a tropical rawinsonde array, is imposed uniformly on a computational domain capable of resolving and representing moist convection. The ensuing adiabatic cooling destabilizes the domain rapidly, so convection soon breaks out. Because the lateral boundary conditions are artificially closed (typically, periodic), the subsiding motions driven by the resulting latent heating gradients, which in the atmosphere would expand rapidly to affect a very large region, are trapped within the domain.

In such an experiment, *large-scale* upward motion is quietly reinterpreted as *adiabatic* upward motion. But the effects of this conceptual error are disguised or compensated by the unrealistic closed boundary conditions. A state of statistical balance prevails between the artificially adiabatic upward motion and the artificially trapped subsidence. The fact that these models maintain a statistical state that is grossly similar to nature (e.g., the temperature stays ‘near’ a moist adiabat) appears to validate the overarching

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1. In many cases, large-scale vertical *advection* is specified, in a forcing term that also includes small horizontal advection and observed local change terms. In this case, model-predicted vertical gradients are not used..

conceptual framework. For example, CEM experiments are the basis of Arakawa's (1993) statements that "Cumulus activity is rather strongly modulated by large-scale processes...In conclusion, cumulus activity is basically parameterizable in terms of large-scale processes." These statements contain the false premise that cumulus activity (meaning precipitating convection) is *small-scale*, and more specifically that observed large-scale vertical motion is a 'forcing' for, rather than an essential part of, precipitating moist convection.

We shall see in section 4 that the output from CEM experiments is in fact tellingly different from nature: the real tropical troposphere does not cool throughout its depth a couple of hours before and during outbreaks of deep convection. Indeed, from an a priori dynamical standpoint, there simply aren't any recognized dynamical phenomena in the tropical troposphere that could cause the kind of intense, adiabatic forced lifting used in CEM experiments.

CEM experiments are like puppet shows. To the extent that the puppetmaker and puppetmaster have observed nature, there is a certain realism to the behavior of the puppets, but the sinews of cause and effect are all wrong. Is this a relevant framework for evaluating the sensitivities of convective cloud populations to external parameters, such as SST changes or radiative processes?

### 3. Equilibrium vs. activation controls on convection

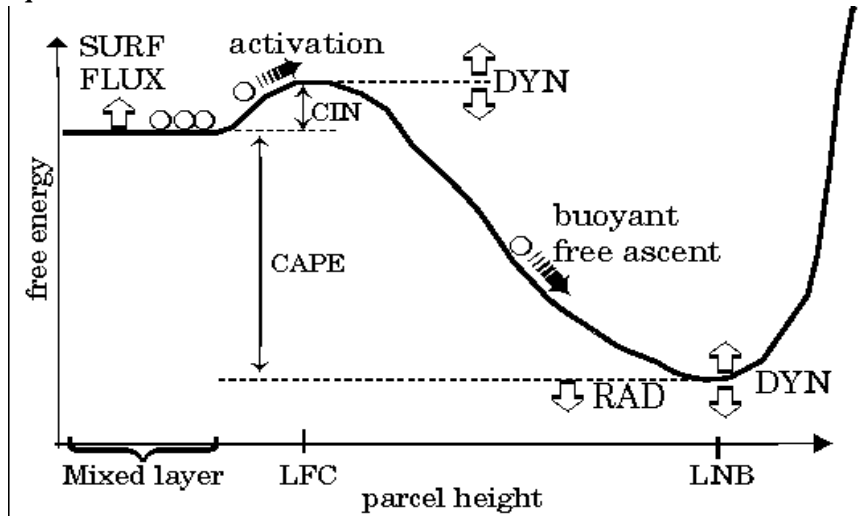


Figure 4. A free-energy depiction of a parcel model for convection. The height of a parcel is shown in the horizontal, with the level of free convection (LFC) indicated. The cumulative integral of work done in lifting the parcel is indicated with the heavy line, as a function of the parcel's height. A parcel of boundary-layer air (circle) requires activation to overcome the energy barrier (CIN) and liberate the CAPE. Dynamical processes (DYN) vary CAPE and CIN, while surface flux (SURF FLUX) raises the platform on the left and radiation (RAD) lowers the platform on the right.

Figure 4 shows a schematic illustration of the state of conditional instability toward deep convection that prevails across much of the tropics. The altitude of a convecting parcel (indicated by a ball) is shown on the horizontal axis, while the vertical dimension indicates a system “free energy.” The heavy curve is an energy surface along which the process of deep parcel ascent occurs. This depiction mimics free energy diagrams used in chemistry to discuss reaction rates for energetically favorable, but inhibited, chemical reactions. The convective inhibition (CIN, analogous to *free energy of activation* in chemistry) is depicted as a barrier, or distance uphill, while the available energy that is released by spontaneous buoyant parcel ascent from the Level of Free Convection (LFC) to the Level of Neutral Buoyancy (LNB) -the CAPE- is shown as a downhill plunge. Fig. 4 is simply a parcel model of buoyant convection, as discussed in any elementary meteorology text, turned on its side.

The convective mass flux can be considered a “reaction rate,” the rate at which parcels of air from near the earth’s surface cross the activation energy barrier at the LFC and ascend to the upper troposphere, in a given grid-box or large-scale patch of the atmosphere. Is this rate controlled by changes in the amount of the downhill plunge (as a large-scale function of space and time)? Or is it controlled by the rate at which parcels are lifted over the (space and time variable) activation energy barrier by intense small-scale lifting processes (whose frequency of occurrence also varies on large scales)? The former scenario will hereafter be referred to as *equilibrium control*, while the latter is termed *activation control*.

The answer to this question is scale-dependent. On the mesoscale, activation control clearly prevails: the spatial organization of convection into arcs or lines, as well as temporal variations of mass flux and precipitation over mesoscale regions, are related to the existence and vigor of gust fronts, sea and land breezes, dryline convergence, etc. On the other hand, on the largest space scale (the global integral) and climatic time scales, convection responds to radiative destabilization quite continuously, maintaining a state of near radiative-convective equilibrium, without any apparent role for inhibition.

Somewhere between the mesoscale and global climate scales, then, lies the crossover between activation control and equilibrium control. At scales larger and longer than the crossover, activation can be taken for granted and mass flux variations are controlled by a deep, bulk destabilization rate. At smaller scales, activation frequency plays the leading role in determining when and where the climatically necessary convection will occur. In midlatitudes, this crossover scale is the Rossby deformation radius (Ooyama 1982, Frank 1983). Equatorward of about 10-15 degrees, however (where much of the world’s convection occurs), the deformation radius is essentially infinite in the zonal direction and this simple reasoning breaks down.

We wish to inquire, with regard to observed tropical deep convection fluctuations on various space and time scales - from synoptic to planetary, from diurnal to climatological - does equilibrium or activation control prevail on this scale? With what observables, and what observations, can we make this distinction? We begin with a critical reconsideration of existing evidence that has been offered in support of equilibrium control.

#### 4. Equilibrium control theories on sub-global scales: a critique

The defining characteristic of equilibrium-control thinking, referring to Figure 4, is that processes modulating the degree of deep convective available energy are assumed to control LSVDC. Any theory or parameterization in which deep convection amount is determined by a quantity integrated over the depth of the troposphere (such as CAPE, its time rate of change, or column-integrated moisture convergence) will be considered an equilibrium-control theory. In practice, however, the *deep control* assumption is often buried in a large/slow/deep complex of space and time scale assumptions, sometimes encapsulated in phrases such as *slowly-varying large-scale disturbances*, with the *deep* assumption implicit. After a brief, subjective listing of historical currents relevant to the rise of the equilibrium-control school of thought (section 4.1), section 4.2 reviews published evidence for equilibrium control of tropical LSVDC.

##### 4.1. RELEVANT HISTORICAL CURRENTS

- Temperature in the tropical troposphere varies little, by midlatitude standards ( $\sim 1\text{K}$  standard deviation). Because this is close to the accuracy of temperature sensors, it has been tempting to hope that, despite the buoyant nature of convection, with  $o(1\text{K})$  buoyancies, the source of its variability is not a  $<1\text{K}$  signal in the stratification profile.
- Sounding indices used successfully by midlatitude convection forecasters fail when applied in the tropics. Instability indices of regional *mean* conditions, as obtained by unbiased radiosonde sampling, vary oppositely with convection amount over the tropical ocean, because of surface convective outflows. Forecasters have been more successful looking at upstream disturbances in the wind field.
- Deep, large-scale upward motion, as measured by rawinsonde arrays, invariably accompanies convection. *Surface* convergence is observed prior to convective out-breaks in some cases (e.g. in easterly waves, see Figs. 11-12).
- Early cloud models had difficulty simulating convection. Absurd perturbations were necessary to initiate convection, which tended to thrive better with imposed domain-wide forced vertical motion.
- Most of the disturbance energy in the tropical troposphere is captured by a first baroclinic mode structure in the vertical, so it is tempting to try to close large-scale moist tropical circulations in a two-level, or one-internal-mode, framework.
- Quasi-geostrophic theory was enormously successful at explaining large-scale weather phenomena in the midlatitudes, and cultivated a generalized view of the world in which circulations of large horizontal scale are assumed to be balanced, deterministic, and slowly evolving, governing (in the aggregate) small-scale or vertical circulations.

##### 4.2. PUBLISHED EVIDENCE FOR EQUILIBRIUM-CONTROL THEORIES

The most elaborate development of an equilibrium-control theory was put forth in the AS74 paper describing the design of a cumulus parameterization. This paper postulated that mass flux in deep cumulus clouds varies in response to deep large-scale destabiliza-



tion processes. Formally, the AS74 parameterization was intended for global atmospheric models, so the functional definition of large and small scales is gridscale and subgridscale. However, in AS74's diagnostic evidence for quasi-equilibrium, discussed in section 4.2.1, the Marshall Islands rawinsonde array (~500 x 1000 km in dimension) operationally defines large scale. More importantly, AS74 argued that there is a separation between cumulus-ensemble and large-scale destabilization *time* scales (section 4.2.2). 'Single-column' tests of cumulus parameterizations may appear to constitute de facto evidence for the validity of equilibrium-control thinking (section 4.2.3).

Evidence related to the shift of the equilibrium state with forcing (Betts 1974) is discussed in section 4.2.4. This sensitive test suggests that equilibrium control is not prevalent in the tropics.

#### 4.2.1. $d(\text{CAPE})/dt$ is "small"

AS74 defined, as a particular measure of convective available energy, the cloud work function  $A$ .  $A$  is the integral over height of the buoyancy experienced by a continuously entraining parcel of air from a mixed layer near the surface ascending to its highest level of neutral buoyancy. AS74 proposed that a quasi-equilibrium for work function prevails, in which large-scale generation terms are statistically balanced by convective consumption terms. When time derivatives are represented with an overdot, and subscripts LS and C represent large-scale 'generation' and convective 'consumption', the quasi-equilibrium hypothesis may be written simply as

$$\dot{A} = \dot{A}_{LS} + \dot{A}_C \sim 0 \quad (6)$$

AS74 examined six-hourly data from the Marshall Islands rawinsonde array in the tropical western Pacific to obtain estimates of  $\dot{A}_{LS}$  and  $\dot{A}$ . Figure 5a is a reproduction of AS74's Fig. 13a, also reprinted by ENB and in the recent textbook by Emanuel (1994, p481), as "striking observational evidence supporting the quasi-equilibrium hypothesis."

The data points in Fig. 5a show that the observed time changes of  $A$  (in 6-hourly data), are always smaller than they would be if  $\dot{A}_{LS}$  acted in the absence of  $\dot{A}_C$  (a scenario represented by the dashed line of unit slope). But how relevant is that comparison?

The large-scale generation term  $\dot{A}_{LS}$  consists of processes which tend to change the vertically integrated buoyancy of an entraining ascending parcel. This includes both processes which alter the ambient density profile and those which change the parcel's density as a function of height. The dominant contribution to  $\dot{A}_{LS}$  when it is large (during periods of strong convection) is cooling of the ambient troposphere by large-scale (area averaged) upward motion, *taken to occur adiabatically*. This dominant 'generation' term,

$\bar{\omega} \frac{\partial}{\partial p} \bar{s}$ , (where  $s = C_p T + gz$  is dry static energy) is nearly balanced by convective 'consumption' of  $A$ , whose dominant term is the 'compensating subsidence' term discussed

in section 2.3.2, given by  $\omega_c \frac{\partial}{\partial p} \bar{s}$  [see Eq. (4)]. Since  $\bar{\omega} = \omega_e + \omega_c$  by definition, and environmental vertical mass flux  $\omega_e$  remains small even during convective outbreaks (large density anomalies would be created if it did not), these generation and consumption terms remain nearly equal.

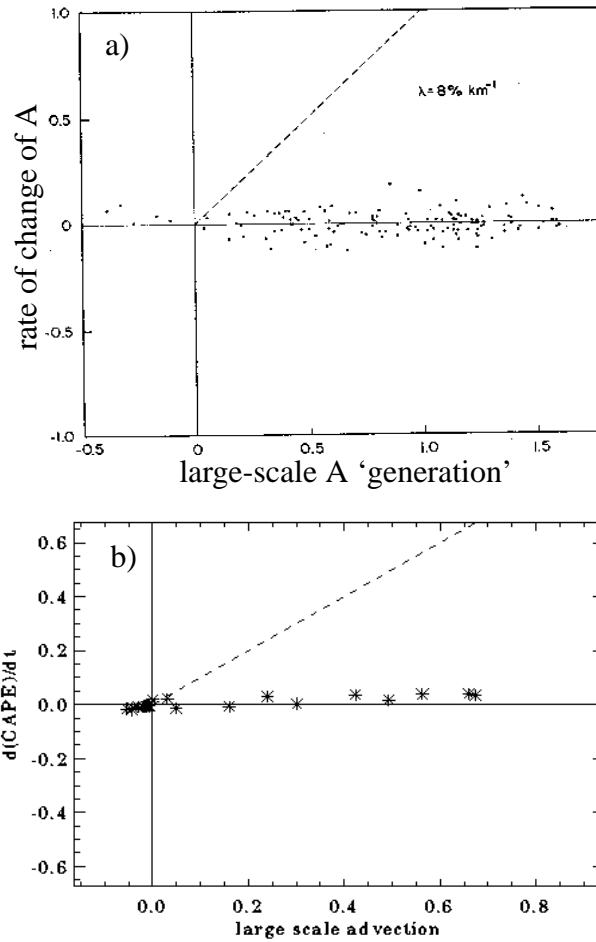


Figure 5. Tests of the quasi-equilibrium hypothesis, for a) Marshall Islands rawinsonde array data (AS74 Fig. 13) and b) imposed-heating model output (Mapes 1997 Fig. 15).

AS74 interpreted this balance to indicate that cloud mass flux  $\omega_c$  responds rapidly and obediently to variations in  $\bar{\omega}$ . This interpretation was apparently based on a tacit assumption that, because  $\bar{\omega}$  is “large-scale,” it is controlled by larger, external forces. Though not explicitly argued, this assumption is apparent in their choice of the word

“forcing” to describe a quantity dominated by  $\bar{\omega} \frac{\partial}{\partial p} \bar{s}$ . Within this interpretation, the results of Fig. 5 do indeed seem to suggest that quasi-equilibrium is a powerful statement about convection, and a strong basis for a parameterization of convective clouds in large-scale models.

Unfortunately, the apparently remarkable balance that keeps  $\dot{A}$  ‘small,’ meaning small with respect to the parts into which it has been divided, says more about how unnatural that division is than it does about the physics of convection (section 2.3).

To illustrate the trivial nature of the correlation displayed in Fig. 5a, Fig. 5b shows a similar plot from Mapes (1997), constructed using synthetic data from a simple linear dry primitive-equation model. In the model, a heating process scaled to resemble a typical mesoscale convective system is specified to fluctuate within a synthetic tropical radiosonde array similar in dimensions to the Marshall Islands array. Quasi-equilibrium holds even for this case, in which the convective heating variation is arbitrarily specified, completely unresponsive to the large-scale flow.<sup>1</sup> Although this synthetic convection “consumes” CAPE (it heats the troposphere), the reduction in CAPE is not observed locally, because the fast, large-scale part of the convection’s diabatic circulation quickly redistributes the heating over very large regions. This experiment indicates that *Fig. 5a says nothing about what causes rawinsonde array-scale LSVDC, or about the relationship of convection to any externally driven, pre-existing, or underlying large-scale motions.*

4.2.2. *AS74’s timescale separation between large-scale ‘forcing’ and cloud ensembles*  
 AS74 stated that quasi-equilibrium holds because there is a clear separation between the time scale characterizing large-scale processes and moist-convective processes. The characteristic adjustment time of convection (which they labeled  $\tau_{\text{ADJ}}$  and estimated at  $10^3 - 10^4$  seconds) was postulated to be an order of magnitude smaller than the time scale characterizing the large scales ( $\tau_{\text{LS}}$ , which they asserted to be “typically”  $10^5$  s or greater). This section argues that quasi-equilibrium holds to observed accuracy if *either* the convection *or* the large-scale flow adjusts on a time scale that is faster than the temporal resolution of the observations. We shall see that both  $\tau_{\text{LS}}$  and  $\tau_{\text{ADJ}}$  are “fast.” This makes it much more difficult to distinguish the direction of cause and effect using observations, but section 4.2.4 offers a method for doing so.

Resounding confirmation that convective cloud ensembles have a response time of  $\sim 10^4$  s has come from cloud ensemble models (CEMs, discussed in section 2.3.3). Xu and Arakawa (1992) subjected a periodic 2-dimensional cloud-resolving model domain to a time-varying imposed domain-average upward motion, with a deep profile through

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1. Boundary-layer changes were entirely neglected in construction of Fig. 5b, while Fig. 5a implicitly included them in  $\dot{A}$ , but not in  $\dot{A}_{\text{LS}}$  (footnote 12 of AS74). Analogous inclusion of boundary-layer humidity fluctuations would increase the vertical scatter of points in Fig. 5b, but would not change the conclusions.

the troposphere and a 27 hour period. The rainfall produced by model clouds lags the forcing by 2-3 hours, with a slightly more dramatic delay in convective development when strong wind shear is present (the case shown in Fig. 6). This response time is consistent with AS74's estimate of  $10^3 - 10^4$  seconds discussed above, and is indeed much less than the 27 hour forcing period (the  $10^5$  seconds mentioned by AS74).

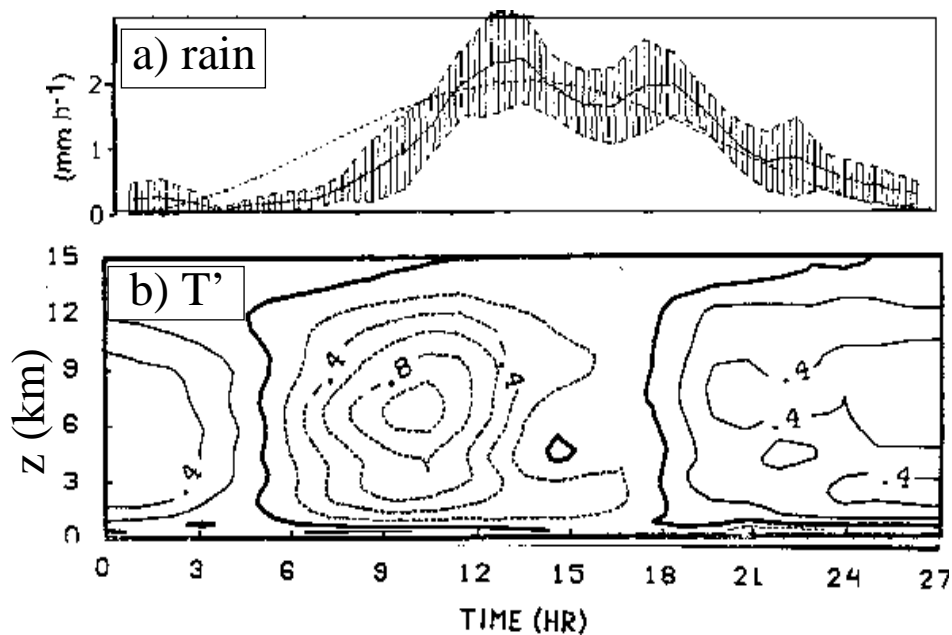


Figure 6. Results from the CEM experiments of Xu and Arakawa (1992). Upper panel: time series of cyclical forcing by deep adiabatic forced uplift (dotted line), and domain-averaged rainfall (ensemble range from several cycles indicated with hatching, mean is solid line). Bottom panel: temperature deviation cycle as a function of height (from Xu 1991).

What about the quasi-equilibrium exhibited in Fig. 5b? Here it is useful to look closer at the synthetic radiosonde-array data from the Mapes (1997) experiment with imposed MCS-like heating in a linear hydrostatic atmospheric model. Figure 7 shows time series of imposed heating, and of large-scale (radiosonde array-integrated) vertical motion for two reasonable radiosonde array sizes. The large-scale vertical motion is seen to respond to convective heating with a characteristic delay time of just 2-3 hours, comparable to the delay of clouds to large-scale forcing seen in Fig. 6. This delay time is the time for  $\sim 50$  m/s gravity waves (this speed is set by the vertical wavelength) to cross the array. Here is the *fast, large-scale* part of the story that was implicitly excluded by AS74's claim that both the space and time scales of convection are smaller than those of 'large-scale disturbances' (section 2.1).

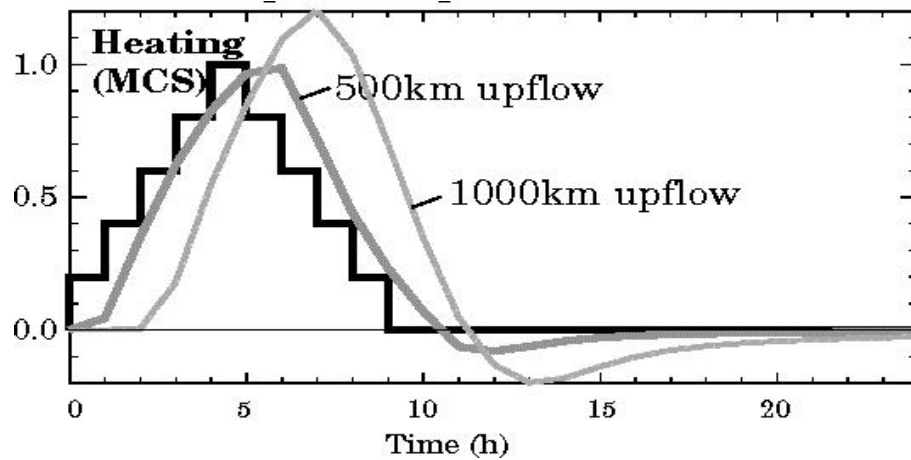


Figure 7. Time series of imposed MCS-like heating (heavy piecewise constant line) in the Mapes (1997) model, and area-integrated upward motion as sampled by synthetic low-latitude rawinsonde arrays of ~500 and ~1000 km diameter (heavier and lighter gray lines, respectively).

In summary, the reason that quasi-equilibrium prevails in the imposed-heating model (Fig. 5b) is that large-scale vertical motion responds quickly to convection. Quasi-equilibrium holds trivially at the scales represented by radiosonde networks, because gravity wave processes are efficient at redistributing convective heating, and not necessarily because convection is obedient to large-scale forcing. We note again that the tropical atmosphere largely lacks dynamical processes capable of causing strong large-scale *adiabatic* lifting (forcing) like that utilized in CEM experiments.

#### 4.2.3. *Semi-prognostic and single-column tests of convection schemes*

Cumulus parameterization schemes are sometimes subjected to ‘single-column’ tests. In these tests, area-averaged vertical velocity, vertical advection, or moisture convergence, as functions of time and height, are passed to a scheme that ‘predicts’ rainfall rates and heating and drying profiles. When the inputs are taken from rawinsonde-array observations, the outputs reproduce well the rainfall, heating, and drying diagnosed from those same observations. Such results are interpreted as evidence for the validity of the parameterization and, by extension, for the underlying equilibrium-control hypothesis. Section 2.3.1 discussed the logical flaws in this thinking. Here we merely reiterate that correlations between cross-isentropic (upward) motion and rainfall arise because both are redundant indices of net condensation, not because the former controls the latter.

#### 4.2.4. *Le Chatelier’s principle: the shift of the equilibrium state with forcing*

We have seen that the observed state of quasi-equilibrium characterizing tropical LSVDC holds because both large-scale vertical motion and convective cloud ensembles have response times short compared to the resolution of observations. How can we assess

whether large-scale upward motion drives convection in LSVDC, as assumed in equilibrium-control theories, or the other way round?

The 19th century chemist Le Chatelier proposed that the state of a continually forced equilibrium-controlled chemical reaction shifts slightly away from equilibrium, in the direction of the forcing. This intuitively appealing principle allows precision of measurement of the mean state to be substituted, where the time resolution of observations is inadequate to resolve cause and effect relationships.

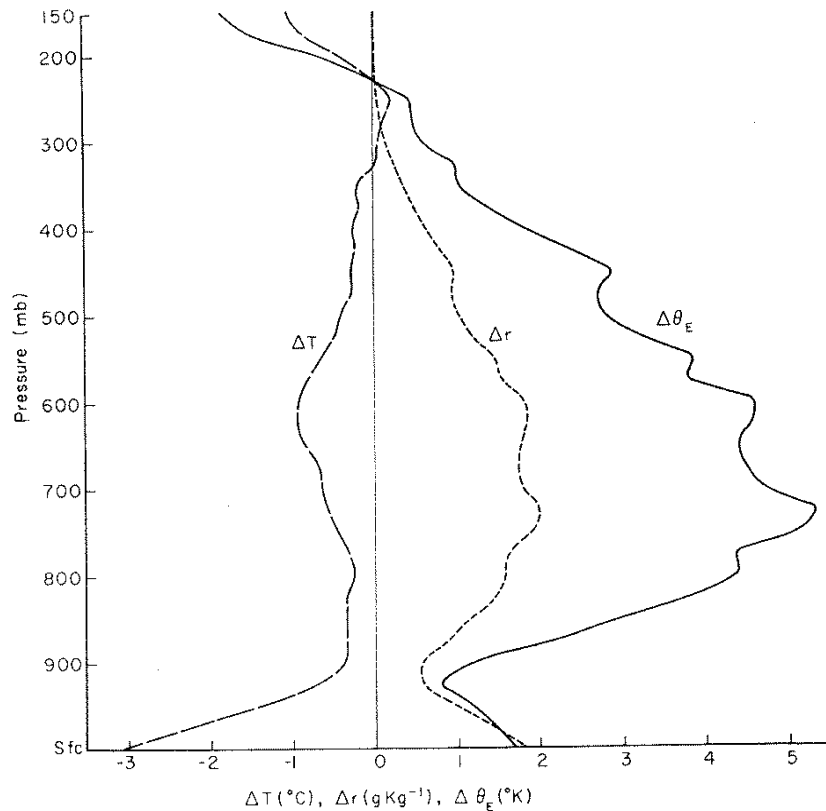


Figure 8. Betts (1974)'s disturbed minus dry day differences in composite temperature, mixing ratio  $r$ , and equivalent potential temperature, from ~300 Venezuelan soundings.

Betts (1974) divided Venezuelan rainy-season soundings into disturbed and dry categories, and examined the difference between the disturbed and dry means (Fig. 8). The temperature, in particular, was observed to be cooler during disturbed periods throughout the troposphere. Betts interpreted these results as evidence for AS74's contemporary quasi-equilibrium idea:

These trends are in the direction of the forcing: for example, large-scale mean ascent will tend to cool and moisten the atmosphere in dis-

turbed conditions. The convective transports are a response to this forcing, and in the main produce the opposite effect - a warming and drying. The mean atmospheric state which we observe represents some balance between these opposing processes. However, as one might expect, this “balanced state” shifts in the direction of the forcing. The concept of a balanced state while convection is in progress is closely related to the quasi-equilibrium hypothesis...

Betts's (1974) method constitutes perhaps the most distinguishing test of the equilibrium-control hypothesis ever published. Similar cooling of the troposphere during disturbed periods also characterizes forced cloud ensemble model results (Fig. 6). When the latent heating in precipitating clouds lags a forced deep uplift of the troposphere (Fig. 6a), the atmosphere tends to be cooler during rainy periods (Fig. 6b).

There is also an energetic relevance to this question of the temperature in convecting regions of disturbances, as summarized by ENB:

Suppose that the convection lags the forcing (vertical velocity) by a small amount. The convective heating is slightly displaced into the cold phase of the wave, leading to a negative correlation of heating and temperature, and thus to decay of the wave...In general, waves will experience Moist Convective Damping (MCD).

Rainy periods are also more humid, both in Fig. 8 and in forced cloud ensemble model experiments. However, interpretation of this humidity signal is ambiguous, since both adiabatic uplift of unsaturated environmental air and detrainment of saturated air and liquid water by cumulus clouds could contribute to humidification. The somewhat convoluted nature of scale-separated thinking is evident in Betts's comments above. Moistening during disturbed conditions is interpreted as a failure of convective clouds to dry the atmosphere fast enough. The tendency of convective clouds to moisten their environment has been defined away, given over to a large-scale mean ascent that is presumed to be external, a ‘forcing.’

What kind of temperature difference between disturbed and dry periods would prevail if activation control, rather than equilibrium control, were operating? A simpler case is one in which convection fluctuations simply have no precursor or underlying temperature signals. Figure 9 shows the (virtual) temperature perturbation in the vicinity of the imposed MCS-like heating in the Mapes (1997) experiments described above. The heating profile (dashed) is top-heavy, as is characteristic of MCSs. The local temperature perturbation (solid) is two-signed, with strong warming aloft and cooling of the lower troposphere. No cold downdraft outflows are included, so the absence of temperature anomalies in the boundary layer is unrealistic. Nevertheless, Fig. 9 can function as an extreme null hypothesis, a contrast to Fig. 6b or Fig. 8.

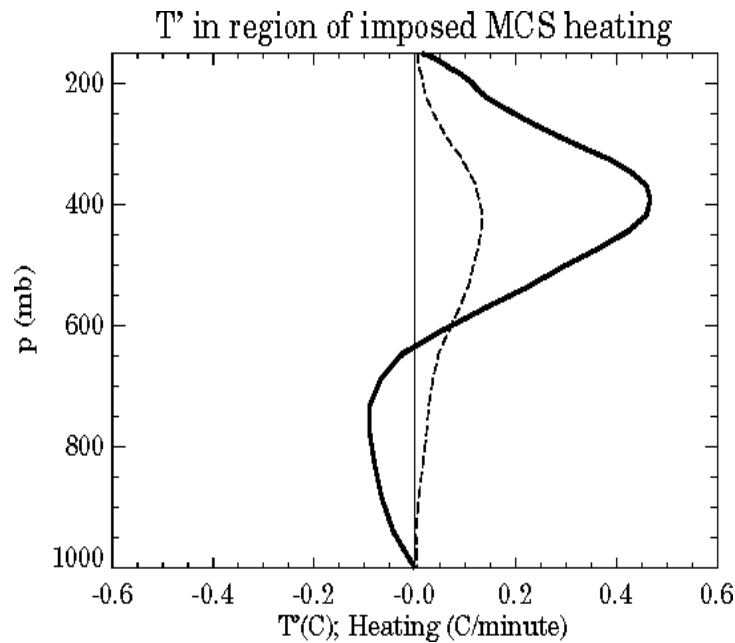


Figure 9. Composite temperature perturbation (solid) in the vicinity of MCS-like heating (profile shown dashed) in the imposed-heating model of Mapes (1997).

A great deal of sounding data has been collected in the tropics since the time of Betts (1974). How has the equilibrium-control supposition fared? Figure 10 shows temperature profiles in disturbed-undisturbed composites from GATE and COARE sounding data. Both these major data sets show cooling of the lower troposphere and warming of the upper troposphere during disturbed periods. Similar results in Australian monsoon data were shown by Mapes and Houze (1992). On a slightly larger scale, correlations between temperature and rawinsonde array-average upward motion have been shown to have similar 2-layer structure, indicating upper warming during convectively active periods (e.g. Kung and Merritt 1974, Stevens et al 1996 and refs therein).

The deep cooling during disturbed periods seen by Betts (Fig. 8) must be considered the exception. The reason seems to be that Venezuelan rainfall is affected by upper-tropospheric troughs, even in summer (Riehl 1977). As discussed in section 6, deep lifting ahead of upper-tropospheric troughs could cause enhanced convection by either equilibrium or activation control mechanisms, so the results in such situations cannot easily be taken as evidence for one hypothesis or the other.



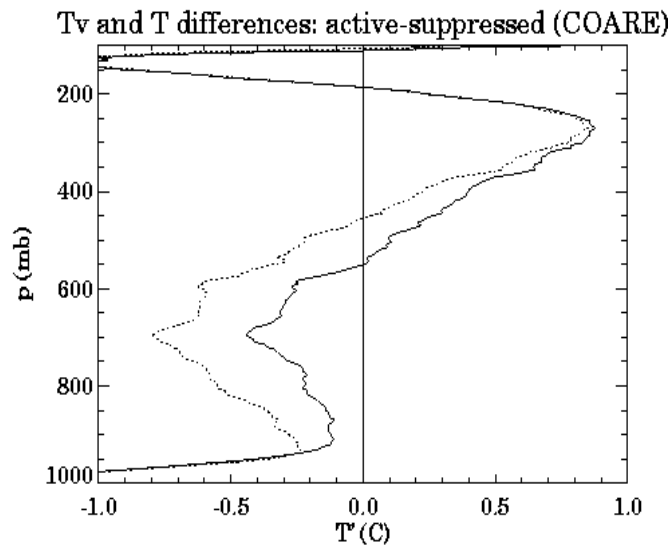
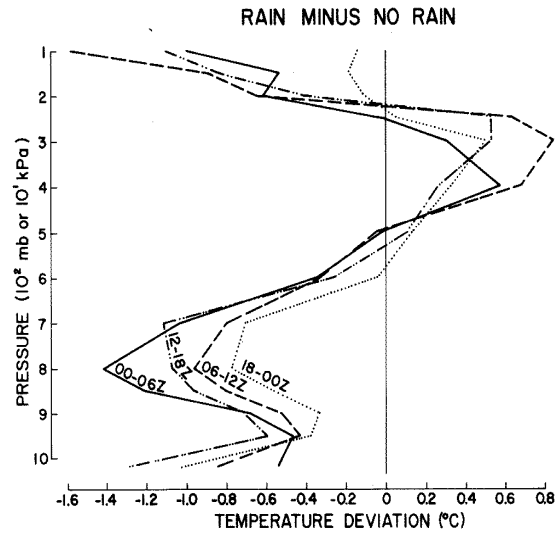


Figure 10. Active minus suppressed composite temperature  $T$  and virtual temperature  $T_v$  profiles. Top panel: GATE  $T$  data categorized based on surface rainfall (Grube 1979). Bottom panel: COARE  $T$  (broken) and  $T_v$  (solid) data, lowest minus highest sextiles of minimum satellite-observed IR temperature within a 55 km square box centered on the sounding site (Mapes 1997b).

In summary, Betts's (1974) test is more definitive than the other published lines of evidence for equilibrium control considered above. When applied across the tropics, however, this test indicates that equilibrium control is not the main mechanism for dynamically determined LSVDC in the tropics.

## 5. Elements of an activation-control theory for LSVDC

Since the evidence does not support equilibrium control theories, perhaps activation control can provide explanations for various types of LSVDC. In this section we consider briefly a few familiar examples, such as African easterly waves (section 5.1), the climatological enhancements of convection observed where concave coastlines focus land breezes at low levels (section 5.2), and events in the case of a shallow upper-troposphere shortwave trough. In African waves, wave modulation of temperature near the LFC is the basis for the hypothesized activation control, while in the latter case it is the frequency of occurrence of sufficiently energetic triggering disturbances that modulates the LSVDC in question. Section 5.4 discusses how the variable coexistence of low-level and deep dynamical waves of cooling excited by convective heating might provide a useful signal for distinguishing equilibrium from activation controls, in the general context of self-organizing tropical convective disturbances.

This section tacitly assumes near-ubiquitous availability of some boundary-layer air with high enough equivalent potential temperature ( $\theta_e$ ) to support deep convection. This assumption is based on the widespread notion that over the tropical oceans, the subcloud layer has a ‘fully recovered’ state, essentially an equilibrium value of  $\theta_e$  to which convective outflow wakes are gradually restored by surface fluxes. For example, estimates of the ‘disturbed’ fraction of the boundary layer, i.e. the fractional area covered by ‘unrecovered’ convective outflow wakes, was about 30% for GATE (Gaynor and Ropelewski 1979). Raymond (1995 and this volume) deduces that over warm oceans, subcloud-layer  $\theta_e$  hovers near a threshold value necessary for the nearly ubiquitous occurrence of moist convection (mainly shallow). Additional criteria then presumably act to determine where the much more sporadic outbreaks of *deep* convection occur.

More discussion of boundary layer variability and its role in LSVDC is in section 6.

### 5.1. AFRICAN EASTERLY WAVES: AN ACTIVATION-CONTROL VIEW

Perhaps the most familiar, strongest example of synoptically-controlled convective variation in the Tropics is African easterly waves (see e.g. Thorncroft and Hoskins 1994 and numerous earlier references therein). The convection observed in the GATE field program was strongly modulated by these waves (e.g. Houze and Betts 1981 and refs). The waves arise from instability of the African easterly jet at about 650 mb. A schematic east-west cross section of an easterly wave train is shown in Fig. 11. Waves are indicated as vorticity perturbations at jet level, with the trough (positive vorticity) in the center of the diagram. The adiabatic clear-air vertical motions that occur as part of the wave-induced secondary circulation (fat arrows) can be rationalized from quasi-geostrophic reasoning. For example, ahead of (west of) the trough, low-level adiabatic ascent occurs, in order to prepare the cool core necessary for simultaneous hydrostatic and geostrophic balance in the oncoming 650 mb trough (e.g. Jenkins, 1995). Scale analysis ( $T' \sim 1K$ , 4 day period) indicates that this vertical motion has peak values of  $\sim 10 - 15$  mb/day. Meanwhile, radiative subsidence is  $30 - 50$  mb/day, while the total observed vertical motion is a deep ascent, exceeding 150 mb/day in the trough, where the convection is most active.

The *adiabatic* vertical velocity is only about a tenth as large as total vertical velocity, which consists mainly of diabatic upward motion in clouds, centered in the trough.

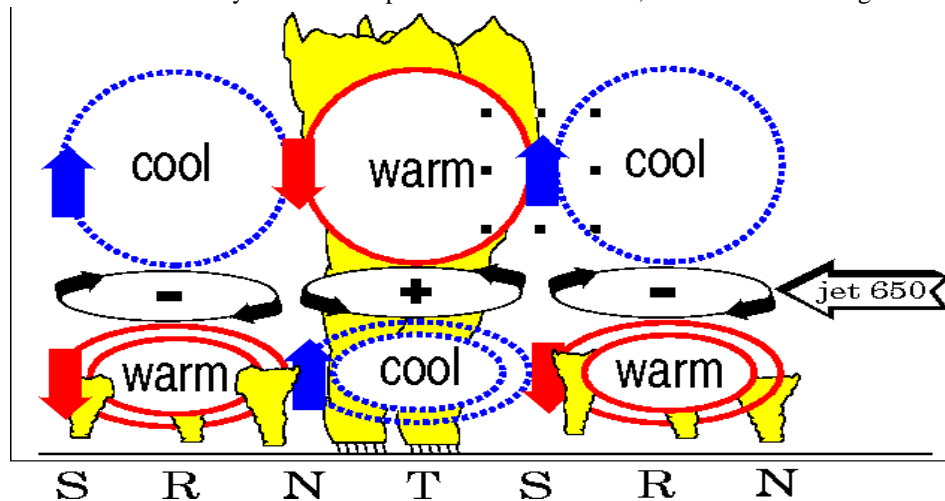


Figure 11. Schematic E-W cross-section through an African easterly wave. Ridges (R), northerlies (N), troughs (T), and southerlies (S) are indicated. *Adiabatic* vertical motions indicated by arrows. Deep convection is most frequent in and just ahead of the trough.

How does such a wave organize convection? An activation-control theory might suggest that the cool core at low levels (near the LFC) reduces CIN, and hence allows deep convective clouds to develop more easily. Enhanced convection would then be predicted in the trough region T, in agreement with observations (e.g. Thompson et al 1979, Houze and Betts 1981 and refs). Furthermore, Fig. 12 (from Grube 1979), shows that cooling near the LFC preceded the onset of convective rainfall in the GATE data.

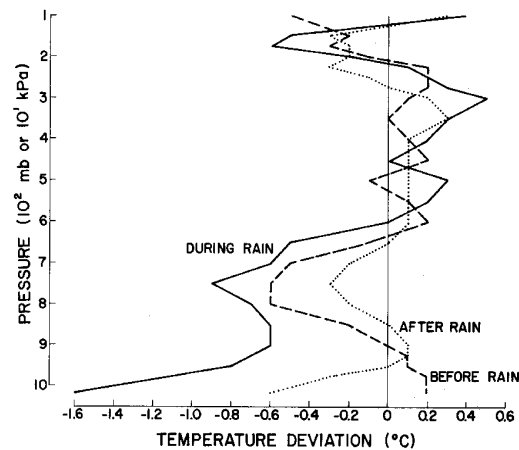


Figure 12. Temperature profiles before, during and after rain, from 3-hourly GATE data (Grube 1979).

An activation-control hypothesis, in which convection is sensitive to the low-level adiabatic displacement (temperature) field rather than the adiabatic vertical velocity per se, also predicts an amplitude-dependent phase for convection enhancement in these waves. Strong waves would have convection-enhancing displacements farther ahead of the trough, while in a weak wave only the largest displacement, right in the trough, could make a difference to convective cloud viability. Here is a testable hypothesis.

It is worth contrasting Fig. 11 with the schematic diagram in the idealized instability study of Thorncroft and Hoskins (1994), which suggested that African easterly waves would enhance deep convection a quarter wavelength ahead of the trough. The reasoning is that since low-level convergence and deep convection are known to be correlated, the convection will occur where the wave-induced low-level convergence is maximum. This reasoning fails to distinguish between the almost unobservably weak (10-15 mb/day), but important, *adiabatic* vertical motions induced in clear air by the wave, and the *total large-scale* (150 mb/day, mostly diabatic, in clouds) vertical velocity. Correlations of convection with the latter have apparently been transferred to the former.

## 5.2. CLIMATOLOGICAL RAINFALL ENHANCEMENTS IN CONCAVE GULFS

Climatological maps of convection from satellite data often show pronounced maxima offshore of landmasses, especially in concave bays or gulfs. For example, Fig. 13 shows 15-year mean Highly Reflective Cloud (HRC) data, indicative of organized deep convection, in the near-equatorial Maritime Continent region. Although HRC has morning-biased diurnal sampling that tends to underestimate land convection, it has relatively high spatial resolution (1 degree).

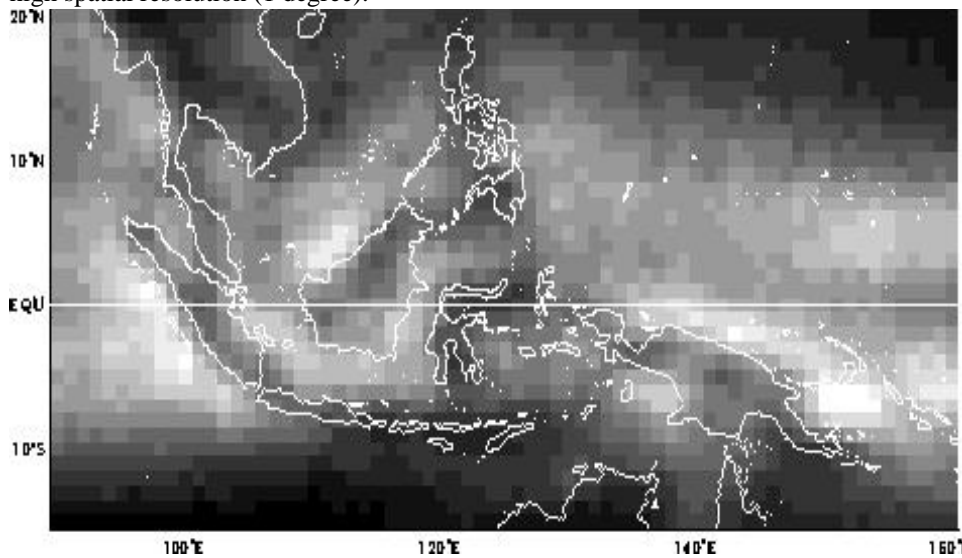


Figure 13. Highly Reflective Cloud (HRC) climatology for the maritime continent region. Values range from 0 (dark) to 6 (white) days per month.

Consider for example the climatological maximum in Batu Bay, off the northwest coast of Borneo, at 4N, 112E. The offshore convection in Batu Bay was intensively studied with radar by Houze et al (1981). A land breeze converging offshore initiated the development of organized convection near midnight many nights. The enhanced frequency of activation at low levels translates directly into a climatological enhancement of rainfall. We may deduce that the deep circulations which supply the moisture and redistribute the heating are simply those circulations driven by the convection itself.

Similar enhancements of convection, presumably also activation-controlled, are found in the Gulf of Panama, the Gulf of Guinea, and essentially everywhere that land meets sea in the convecting latitudes of the tropics. Other features of Fig. 13, such as the equatorial minimum at the east edge of the figure, and ultimately the overall climatological pattern, may also have activation-controlled origins, awaiting discovery by careful investigation.

### 5.3. THE ABSENCE OF CONVECTION IN EQUILIBRIUM-FORCED CASES

It may be more profitable in some cases to ask, not “why did convection occur at location X,” but rather as “why did convection not occur at location Y?”

Consider the adiabatic forcing event of 23-24 January, 1987, over the Gulf of Carpentaria in Australia. A cutoff low from the southern hemisphere upper-tropospheric westerlies drifted over Thursday Island (at latitude 10S). This cyclonic potential vorticity anomaly is apparent in the meridional winds (Fig. 14a), which swung from ~20 m/s southerlies to ~20 m/s northerlies at the 200 mb level at the trough passed over the site. Thermal wind balance entailed a considerable thermal anomaly field - cool core below, warm core above - despite the low latitude. The troposphere from 450 to 200 mb was about 2 degrees cooler than the monthly mean as the cyclone passed over (Fig. 14b).

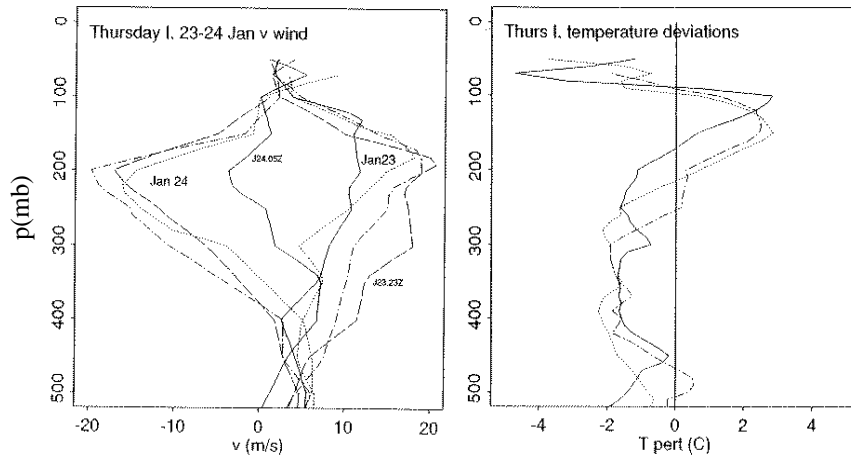


Figure 14. a) meridional wind profiles of 23-24 January 1987, and b) perturbation temperature profiles (relative to a monthly mean sounding) of 23Z 23 January, 05Z 24 January, and 11Z 24 January, at Thursday Island (10S, 142E).

Since the standard deviation of temperature at those levels is about 1 degree, this can be considered quite a strong case of upper-level thermal forcing. However, because of the small horizontal scale  $L$  of the feature, its “Rossby penetration depth”  $NL/f$  was also small, so the adiabatic lifting beneath and ahead of the trough only extended down to  $\sim 500$  mb. The cooling caused by passage of this feature caused an adiabatic generation of CAPE (however defined) of about 400 J/kg, but no convection ensued, despite a moist boundary layer. Strong low-level trade inversions at 850 and 720 mb, with dry air above them, apparently prevented deep convective clouds from forming. In this case the *absence* of convection indicates the predominance of activation control. The upper-level forcing, though quite strong for the tropics, simply didn’t reach far enough down to break the inversion inhibiting convection at low levels. Larger upper troughs, by contrast, do cause outbreaks of tropical convection (e.g., Kiladis and Weickmann 1992).

#### 5.4. THE VERTICAL STRUCTURES OF SELF-EXCITING DYNAMICAL LSVDC

The convecting tropics contains a broad spectrum of dynamically determined LSVDC. Here we consider the general basis for control of convection by convective heating-generated wave motions. The framework is the diabatic-adiabatic, or moist-dry, separation advocated in section 2. Again, boundary-layer recovery subtleties are ignored for purposes of simplifying this discussion. What kinds of adiabatic motions does fluctuating convective heating excite in its stratified environment? Specifically, where and when do the adiabatic vertical displacements, caused by convective heating in one location, most strongly destabilize the atmosphere toward development of additional convection?

The MCS heating-profile measurements of Mapes and Houze (1995) indicate that two vertical ‘modes’ (or more correctly spectral bands, Mapes 1997) are important (Fig. 15). Figure 15a shows the temperature field in an initially resting stratified atmosphere surrounding a MCS-like heating event, after 6 hours of constant heating. A deep single-signed temperature wave has a horizontal phase speed of  $\sim 50$  m/s, while a two-signed temperature signal has a phase speed about half as great. Although the heating is positive-only, some negative temperature perturbations at low levels are apparent near  $r=450$  km. When ensembles of such MCS-like heating events come and go, on a rotating earth, both the 1-signed and 2-signed temperature signals can appear, with either phase, at various times and places (Mapes 1997).

The relative importance of the two modes in determining LSVDC depends on whether equilibrium or activation control prevails in nature. Here is the forefront of our understanding of moist dynamics. The deeper, faster mode contains far more energy (Fig. 15b; recall that geopotential and kinetic energy are equipartitioned for linear gravity waves). However, the second mode has disproportionately large amplitude in terms of temperature at low levels, say near 800 mb, where temperature changes may have a particularly strong effect on subsequent convection (Fig. 12).

The deep mode strongly modulates CAPE, while the two-signed mode affects CAPE (and related integrals over the whole troposphere) much less, but modulates CIN rather strongly. Because these modes travel at different speeds, they cause independent dynamical modulations of CAPE and CIN. These independent modulations might make it pos-

sible to isolate and quantify the degrees of equilibrium and activation control on convection. Of course, relevant CAPE and CIN indices must first be defined, perhaps from cloud modeling results. CIN's values in particular are quite sensitive to its definition (Smith, this volume; Mapes 1997b).

If convection is activation-controlled, then this two-signed temperature mode might actually be more important in shaping the convection field than the one-signed first baroclinic mode. If so, the validity of two-level or one-internal-mode models of the moist tropical troposphere would be questionable, despite the fact that they can represent most of the *energy* in tropical circulations.

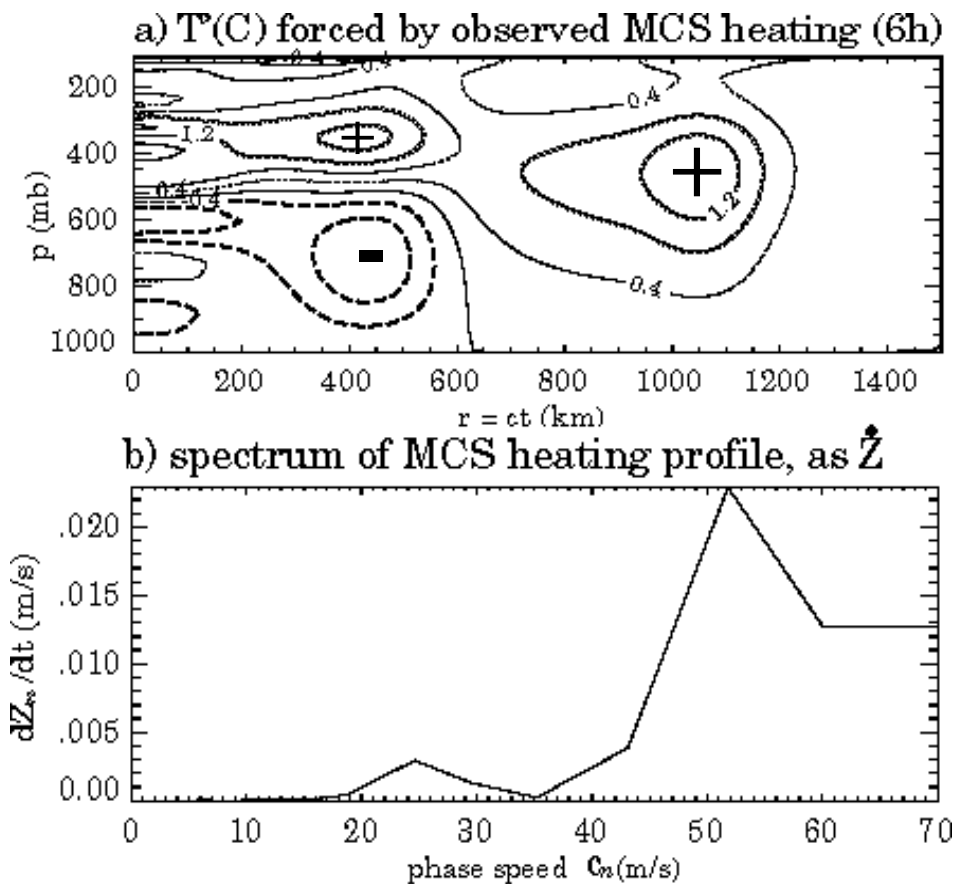


Figure 15. a) the temperature perturbation, as a function of radius and height, induced by an idealized transient (6h) MCS-like heating process in a 140 km circular patch (centered at the origin) of a realistically stratified atmosphere (Mapes and Houze 1995). b) The corresponding spectral coefficient expansion of the MH95 observed MCS heating profile, expressed in terms of geopotential height changes which would occur if the heating went directly into locally changing temperature.

## 6. Ambiguous cases: deep lifting, surface flux enhancements

Two of the most common, and important, processes that cause LSVDC are deep adiabatic lifting, as occurs ahead of upper-tropospheric troughs (section 6.1), and surface flux enhancement, as over a sea surface temperature (SST) anomaly (section 6.2). Unfortunately, qualitative studies of these types of LSVDC cannot yield unambiguous results about whether equilibrium or activation control predominates. In both cases, the “forcing” acts simultaneously as a supply of moisture and instability through a deep layer and a weakening influence on convective inhibition. Note, however, that the equilibrium vs. activation control distinction may have important quantitative implications for our understanding and predictions of these phenomena.

### 6.1. LARGE-SCALE DEEP LIFTING

Deep lifting increases CAPE and decreases CIN simultaneously. For example, Fig. 8 indicates that deep cooling (everywhere below the 320 mb level) accompanied convective enhancements in Venezuela, where upper troughs are common, even in summer (Riehl 1977). Yet it is possible that only the low-level portion of this deep adiabatic lifting is truly necessary for the enhancement of convection (as suggested by Figs. 11-12). Recall also Fig. 14, section 5.3, which indicated that upper-level lifting alone was insufficient to excite a deep convective outbreak.

Situations with strong upper-level forcing have often been used as test cases for evaluating convective parameterizations. For example, Grell (1993) tested various cumulus parameterization assumptions in a mesoscale model study of the 10-11 June 1985 Oklahoma squall line, which occurred ahead of an upper-level shortwave trough, and concluded:

...dynamic control...determines the modulation of the convection by the environment. It is shown that rate of destabilization, as well as instantaneous stability, work well for the dynamic control.

Such a conclusion would presumably not be generalizable to the less strongly forced situations which make up the bulk of the world's LSVDC.<sup>1</sup>

### 6.2. SURFACE FLUX ENHANCEMENTS

Deep convection anomalies can be caused by changes in SST (as in the ENSO phenomenon), or by changes in land-surface conditions (such as changes during the diurnal cycle, or during droughts). Unfortunately, there are many different lines of reasoning that are

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1. Both of the dynamic control assumptions tested by Grell were equilibrium-control assumptions, since they involved integrals over the whole troposphere (of cloud buoyancy and its rate of forced increase). Very few parameterization schemes are truly under activation control (e.g. Gregory and Rowntree 1990), although some schemes have low-level triggering conditions in addition to equilibrium-control closures.



qualitatively consistent with the observation that convection preferentially occurs over warmer, moister surfaces.

In the realm of local mechanisms, one can think of surface heat and moisture flux enhancements tending to increase subcloud-layer  $\theta_e$ . Like deep lifting, however, subcloud-layer  $\theta_e$  enhancement simultaneously increases CAPE and decreases CIN, by raising the level of the platform on the left side of Fig. 4. Strictly speaking, however, it is necessary to discuss the *convergence* of enhanced vertical fluxes across the surface. Here the complexities of the subcloud and shallow moist convection (“trade cumulus”) layers intervene to prevent us from making any easy statements about deep convection. Section 7 contains more discussion about these issues.

Sorting out the equilibrium vs. activation distinction in boundary-forced LSVDC would require a model of how the structure of the whole surface-affected layer of the atmosphere changes, e.g. with a given pattern of SST change. This is a formidable task, as the problem is nonlocal in the horizontal as well as the vertical.

## 7. Discussion and conclusions

Why does deep convection occur when and where it does? On the largest scales convection is inevitable and nearly constant from day to day, like the radiative processes that ultimately drive it. On subplanetary scales, convection inherits spatial structure from land-sea and sea surface temperature (SST) maps. In the time domain, strong outbreaks of convection occur in warm moist airmasses ahead of strong upper-tropospheric troughs. Even crude models, with any of a wide range of assumptions about convection, can get this much roughly right. To go further, one must look more closely at climatological spatial structure (e.g. Fig. 13), or at more subtle aspects of temporal variation. In some cases it may be more discriminating to ask why convection does *not* occur when and where it doesn’t (e.g. Fig. 14, section 5.3).

The ascent of air in deep convective clouds is a spontaneous free buoyant process, which occurs when and where there is conditional instability and a sufficiently vigorous low-level circulation to lift air to its level of free convection (LFC). Of course there are uncertainties in precisely defining instability and LFC, chiefly involving mixing and microphysics. Still, with modern numerical atmospheric modeling capabilities, we could surely get started on mapping the relevant physical regime. What are the quantitative sensitivities of convective cloud ensembles to their environment?

These sensitivities are the final building blocks we lack for solid theories of a wide variety of important large-scale convection-dependent phenomena, and for physically realistic parameterization. They can be expected to be rather delicate: tropical cloud buoyancies are comparable to the error range of many measurement systems, and to the ambiguities of parcel theories of cloud buoyancy. Nonetheless, the observation that deep convection is a buoyant phenomenon is unambiguous. It does not seem profitable to begin by declaring, as equilibrium theories do, that nature’s entire range of variability in buoyancy-related quantities is negligible, compared to some imaginary scenario (e.g.

Fig. 5). Natural buoyancy variability may be beyond our capacity to measure accurately or sample adequately, but it contains important information about cause and effect that is essential if we are to gain a predictive understanding of the atmosphere.

Precipitating convection is a spectrally red phenomenon: the energy generated by the correlation of heating and temperature in a buoyant updraft is not cloud-scale, but broadband. Unfortunately, the study of tropical deep convection has been artificially divided along lines of scale. Mesoscale convection studies zoom in on cloud and storm morphology, deferring questions of existence to a vague ‘large-scale forcing.’ Large-scale studies, based on the false premise that convection is small-scale, define this ‘forcing’ in such a way that it is dominated by (a smoothed version of) the upward motion in the convection itself. In essence, clouds are depicted as puppets of their own diabatic circulations. This depiction is enforced in closed-domain forced cloud ensemble modeling experiments, and codified in supply-side parameterizations. The question of why convection varies in the first place falls entirely between the cracks.

Could it be that LSVDC are caused by large-scale variations in the rate at which air is lifted to its LFC? Adiabatic atmospheric wave dynamics can modulate the height of the LFC, and convective inhibition (CIN), on broad scales. Gust fronts from previous convective outflows trigger new cells, lending a time scale many times longer than a single cloud lifetime to the development of deep convective outbreaks. The hypothesis that deep convection amount is controlled by the low-level processes that govern its initiation is here termed “activation control.” Note that surface warmth and moisture flux anomalies, and deep adiabatic lifting of atmospheric columns, tend to favor convection by activation control, in addition to their oft-cited “equilibrium control” effects (supplying moisture, generating CAPE). The difference between the two control hypotheses is only distinguishable quantitatively in these cases.

Essentially all meteorological phenomena in the tropics that are large enough to contain statistically significant numbers of embedded precipitating cloud systems - including ENSO, monsoons, easterly waves, hurricanes, the Madden-Julian oscillation, the diurnal cycle, etc. - could stand critical reexamination in light of the activation-control hypothesis. Inhibition must be quantified on an energy scale set by the statistical vigor of the boundary-layer eddy motions that trigger deep convection. However, there are several difficulties that must be overcome in order to develop a quantitative activation-control theory.

Simply defining CIN (and CAPE) in a useful way is hard (see e.g. Smith, this volume). First, it is far from clear what air becomes the rising ‘parcel’ in parcel models of deep convection. Assumptions about mixing and microphysical processes strongly affect computations of the parcel’s buoyancy. In the tropics, nearly undilute ascent of air from the lowest 50 mb, with water precipitating out, seems to be necessary for buoyant deep convection to reach tropopause altitudes as is observed. However, deep convection also involves substantial mass transfers among intermediate layers of the troposphere. Second, boundary-layer  $\theta_e$  is horizontally and temporally inhomogeneous on very small scales, especially in regions in which deep convection is underway (e.g. Weckwerth et al. 1996).

Convection draws preferentially from a supply of high- $\theta_e$  air, and creates cold outflow ‘wakes,’ typically with lower  $\theta_e$  (see Addis et al 1984 for some interesting exceptions). The ‘gust fronts’ which bound these wakes are a key mechanism for lifting the high- $\theta_e$  air to its LFC. The *average* thermodynamic properties of these two types of air-masses, which are intimately intermingled on the mesoscale, but not truly mixed, is irrelevant. The ideal situation for deep convection is probably to have both an adequate supply of high- $\theta_e$  air and plenty of gust fronts. Models of activation-controlled LSVDC will need to carry more than just one mean value for temperature and humidity in the boundary-layer of a grid cell.

The role of surface flux anomalies in LSVDC is a particularly challenging issue. Enhancements of surface moisture flux (e.g. by increased wind speed) must cause enhanced moisture fluxes by convective clouds, since the storage capacity of the sub-cloud layer is small (Raymond, 1995 and this volume). However, it is by no means clear that these enhanced fluxes at the surface and the top of the subcloud layer translate straightforwardly into enhancements of *deep* convection. Observational studies (cited by Raymond) show that the surface moisture flux does not tend to converge within the sub-cloud layer, but instead is carried upward in shallow cumuli. This renders dubious ENB’s claim, in connection with the AS74 quasi-equilibrium observations (Fig. 5a above) that “surface fluxes and radiative cooling...generate about  $4000 \text{ J kg}^{-1}$  of available energy each day.” AS74 declined even to estimate the contribution of surface fluxes to available energy generation (their footnote 12).

The role of mesoscale “organization” in determining convection amount remains unclear. For example, coherent boundary-layer rolls, or a linear gust front aligned across ambient wind shear, may be able to lift air to its LFC much more effectively than an equivalent area of spotty boundary-layer thermals of equivalent energy. In the chemistry analogy of Fig. 4, the free energy of activation is composed of two parts: an enthalpy of activation, and an entropy of activation. A reaction with a high entropy of activation requires organization: it might involve, for example, two reactant molecules that must collide with a precise orientation. Even if the required collision energy is small, the reaction might be highly inhibited and slow, because of the low likelihood of organized geometry. Increasing collision *energy* (temperature) might not speed up such a reaction as much as increasing *organization*, say by an electric field that aligns the molecules.

In the case of convection, “organization” might make itself felt through reduced mixing. If clouds can be made to entrain less, say by two-dimensional geometry, or if clouds can preferentially entrain moister-than-average air, then more air can achieve deep buoyant ascent. Alternatively, if the cool phase of gravity waves at the LFC is systematically coordinated with low-level lifting processes, more air can be “activated” than if the same amount of energy were more randomly distributed. In short, predicting the activation rate of deep clouds (Ooyama’s 1971 “dispatcher function”) requires knowledge not only of the subgridscale distributions of  $\theta_e$ , entrainment rate, and gravity wave energy near the LFC, but also of their intercorrelations (see Mapes 1997b for examples and further discussion).

Clearly this situation is hopeless in detail. In practice, one could perhaps collapse the complexity into a single, empirically-tuned subgridscale distribution of “effective convective inhibition,” whose tail of negative values gives an estimate of the fraction of the gridbox that is convecting. Lest this discussion sound too pessimistic, we note that a relatively simple activation-control theory, based on simply-defined CIN alone, can skillfully hindcast convection in the central US (e.g. Colby 1984).

A better understanding of the statistical sensitivities of the convective initiation process is badly needed. The sensitivity of cumulus ensembles which include deep precipitating clouds to humidity, temperature, and wind shear changes in various layers should be carefully examined. Here is an excellent climate-relevant use for cloud-resolving models. Unfortunately, these issues cannot be addressed by model experiments run in equilibrium-controlled ‘puppet show’ modes. Meanwhile, prognostic convective storm model runs are influenced by arbitrary initial conditions throughout the integration.

Many fundamental, model-accessible questions remain unanswered and even unasked. The field is ripe for exploration. Demanding high standards of explanation for LSVDC could ultimately lead us to considerably better tropical, and hence global, predictions on a wide range of space and time scales.

## 8. Acknowledgments

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