WHAT CONTROLS LARGE-SCALE VARIATIONS OF DEEP CONVECTION?

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Suppose deep convective clouds represent the free buoyant ascent of low-level air, once it is brought to its level of free convection (LFC) by a low-level "activation" process energetic enough to overcome the convecive inhibition energy (CIN). Large-scale enhancements of convection should therefore be attributed to processes such as *low-level* adiabatic dynamic lifting; enhanced frequency of occurrence of strong activation mechanisms such as gust fronts; enhancements of the warmest, moistest boundary-layer air; and decreased mixing of dry air into updrafts. Quantifying these bulk sensitivities - which involve (intercorrelated) subgridscale distributions as well as mean values - seems necessary for accurate, physically-based parameterization.

Differences of the right sign are present in tropical aircraft data and enhanced-suppressed sounding composite differences, but actual values of CIN, LFC, etc. are quite delicately sensitive to unresolvable details of entrainment, precipitation, parcel definition, and vertical resolution. Still, a model would presumably develop its own internally-adjusted (and tunable) economy of inhibition & activation, and would perhaps yield better simulations for having appropriate sensitivity to low-level inversions. Its climatology needs only be wrong by the amount of these delicate quantities, making their smallness a blessing in disguise.

1. INTRODUCTION

In-situ studies of convection invariably find that convective clouds represent the free buoyant ascent of potentially-buoyant low-level air. In essentially every convective storm, a gust front or other energetic low-level circulation is found, lifting near-surface air to its level of free convection in a conditionally unstable environment. However, global modelers faced with the problem of parameterizing convection seem to balk at turning control of convection over to the statistics of such local phenomena, which would in turn have to be parameterized. Instead, a "supply-side" mentality has prevailed, perhaps because of its greater perceived elegance. Supply-side (or "equilibrium-control," Mapes 1997) parameterizations assume that convection efficiently consumes something it requires (e.g. moisture, or convective available energy) at the rate at which it is supplied by model-resolved flows. Such parameterizations tacitly assume that low-level triggering disturbances such as gust fronts are so ubiquitous that their availability never limits the rate at which convective mass flux leaves the boundary layer for its journey into the free troposphere (this rate is Ooyama's 1971 "dispatcher function").

There is just one problem. Although large-scale models can *resolve* the divergent flows that supply regional hot spots of convection with their water or energy, these flows are incorrectly represented, because the model doesn't know where to put the convection in the first place. This problem is especially severe in the tropics, where divergent circulations, other than the diabatic circulation driven by the convection itself, are extremely weak. This paper briefly examines convective initiation processes in the tropics, and considers prospects for "demand-side" (or "activation-control", Mapes 1997) representations of convection in global GCMs.

For all the decades of tedious hard work, convective parameterization seems to be perpetually in its infancy. In this spirit of naive beginnership, section 2 inquires about the causes of the substantial observed variations of convection during the latest tropical field campaign, TOGA COARE. In this inquiry, we do not content ourselves with the facile diagnostic observation that large-scale flows converge moisture or "unstable air" into convecting regions at low levels. Rather, we suppose that convective variations must ultimately be related to the buoyancy dynamics of convective clouds.

We shall see that convective initiation is a delicate business, involving precisions in the vertical direction and in thermodynamic variables that may exceed those of our models and measurements. At first, this might seem to suggest that there is no hope for parameterization. But perhaps this smallness of convective activation control variables is really a blessing in disguise.

Surely a model with an activation-controlled ("demand-side") parameterization of deep convection will have a different economy of CIN than nature, just as different "supply-side" parameterizations lead to different, but internally consistent, economies of convective available potential energy (CAPE) in present GCMs. In the latter case, however, different CAPE climatologies correspond to temperature biases through the depth of the troposphere that imply significant biases in midlatitude jets, etc. These errors are in addition to the other climatological errors that these parameterizations make directly in the tropical precipitation and circulation fields. In contrast, a model with its own, internally consistent, inhibition climatology needs only differ from nature by this much smaller amount, and may give considerably better simulations of precipitation because of its convection's (appropriately tuned) sensitivity to low-level thermodynamic variables.

2. TOGA COARE CONVECTION VARIATIONS: DISTURBED-DRY

TOGA COARE was a 4-month program of intensive observation in the equatorial western Pacific region. Convective cloudiness in the COARE observational domain exhibited large variations from hour to hour, day to day, week to week, and even month to month (e.g., Chen et al 1996). Many of these variations were "large-scale," meaning that they consisted of systematic behaviour among a statistically significant ensemble of convective clouds, behaving in a coherent fashion. Of course, there is also a lot of variability in rainfall which is "mesoscale," or due simply to the peculiar structure or behavior of a single mesoscale storm such as a squall line (or two or three). Mesoscale storm dynamics will not be discussed here.

If dozens or hundreds of buoyant convective clouds sharing common environments behave coherently, then the ensemble of radiosondes and aircraft data sampling those environments should be able to sample the reasons why. Here we consider two subsets of the COARE soundings data, based on the satellite-observed infrared (~cloud-top) temperature in the (~50 km,~3h) vicinity of the sounding launch site and time. The "disturbed" set, defined by cloud top T < 210K, contains 324 soundings; the "dry" set, with T>285K, contains 392.

Figure 1 shows the virtual temperature difference between the disturbed and dry mean soundings. As is typical of such disturbed-dry composites (several reviewed in Mapes 1997), the disturbed conditions are ~0.5C cooler in the lower troposphere, and warmer aloft, then cooler again at tropopause levels. The virtual temperature correction is significant, but the sense of the temperature difference (dotted) is preserved in the virtual temperature difference (solid). The greatest difference is in the boundary layer, as the disturbed soundings sampled a lot of convective outflow ("cold pool") air (e.g. Addis et al. 1984). Of course, the difference between temperature and virtual temperature also indicates that the dry composite sounding is less humid, which may have more direct impacts on convective cloud viability, through mixing processes (see e.g. Fig. 3 below).

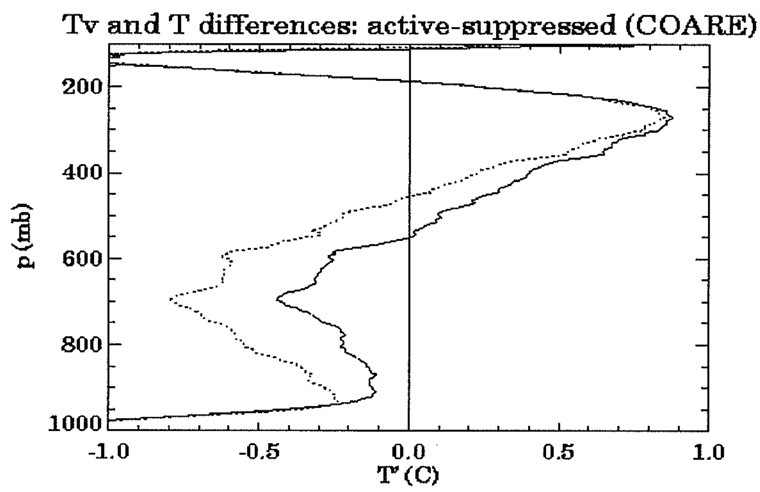

Figure 1. Disturbed minus dry composite virtual temperature (solid) and temperature (dotted) differences. Categories are based on satellite-observed cloud top temperature criteria (text).

Are these 0.1 - 1C temperature differences too small to be relevant to the cloudiness difference between disturbed and dry conditions? Let us see how they impact the buoyancy experienced by parcels lifted from the boundary layer.

3. INSTABILITY DIAGNOSIS OF DISTURBED-DRY SOUNDINGS

3.1 Defining the parcel

The first uncertainty in parcel instability diagnosis is what to assume about the parcel. Low-level air properties vary strongly in the horizontal and vertical, both because of airmass history (recent downdraft outflow vs. "recovered" air) and because of boundary-layer circulations such as rolls. Only the warmest and moistest air is capable of buoyant ascent to the high altitudes at which deep convective cloud tops are observed. This air is only available in limited quantities, but convection preferentially feeds from the high end of the distribution of boundary-layer equivalent potential temperature θ_e (Weckwerth et al., 1996). Assumptions about mixing and precipitation in such parcel calculations are simply ad-hoc. Ultimately, cloud-model guidance is needed here, but unfortunately model studies rarely focus on convective initiation and development sensitivities except in the context of severe storms (e.g. Crook 1996).

Figure 2 shows histograms of θ_e in the 950-995 mb layers of the disturbed (solid) and dry (dotted) sounding sets. The distributions overlap considerably, and some high-θ_e air is available in all conditions (see also Kingsmill and Houze, 1997, who used higher-accuracy aircraft data). Given that fact, the greater frequency of occurrence of downdraft air in disturbed conditions is not necessarily a negative factor for convection, even though it makes the mean θ_e lower. Gust fronts capable of lifting air through its CIN to its LFC are certainly more prevalent during disturbed conditions, although the efficacy of these gust fronts depends on the *virtual* temperature difference across them, not the θ_e difference *per se*. To fix ideas, we set aside delicate sounding humidity accuracy issues, and take the mean boundary-layer air in undisturbed conditions as our reference parcel (θ_e = 352K).

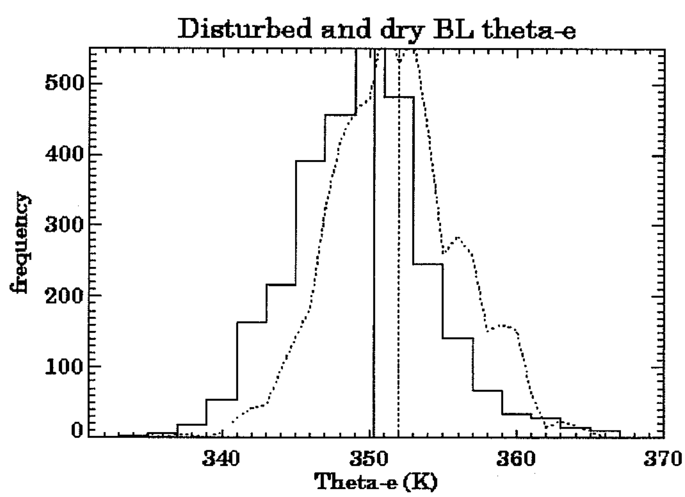

Figure 2. Histograms of boundary-layer theta-e in disturbed (solid) and dry (dot) conditions Means are indicated by vertical lines.

Figure 3a shows the buoyancy of a mixture of the air from the lowest 50 mb of the "dry" category mean sounding, as a function of pressure. The buoyancy is expressed as a difference in "density temperature," an equivalent temperature which takes into account the density effects not only of water vapor, but also of liquid water (Emanuel, 1994). In this case, precipitation is assumed to limit the parcel's liquid water content to a maximum of 3 g/kg, and mixing with environmental air is considered through continuous entrainment, with rates [0,5,10,15,...] %/100mb. The ordinary, nonentraining virtual temperature difference (parcel-env) is also shown, in a dotted line. For 0 entrainment rate, the parcel experiences some negative buoyancy (CIN = -10.2 J/kg) below its LFC at 835 mb, then positive buoyancy above (CAPE = 953 J/kg). For this dry composite sounding, entrainment is particularly devastating to parcel buoyancy: for rates above 20%/100mb, the parcel never attains any positive buoyancy (with freezing neglected).

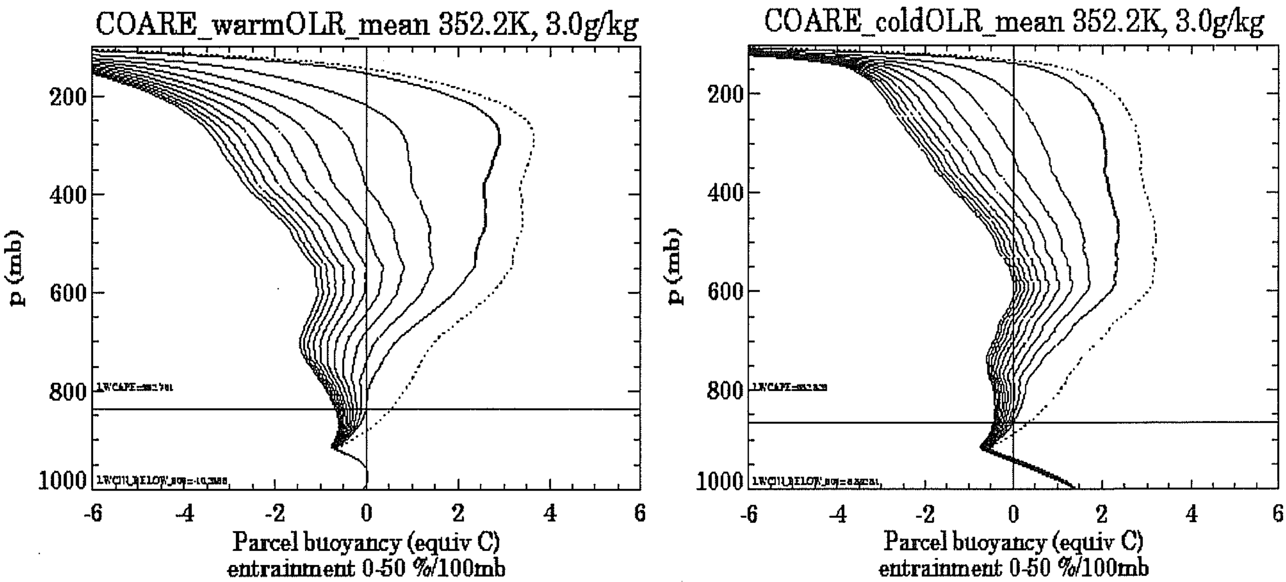

Figure 3. Buoyancy profiles of a parcel of "dry" composite mean boundary-layer air, through the "dry" mean sounding (left; CAPE = 953 J/kg, CIN = -10.3 J/kg, LFC = 835mb) and the "disturbed" mean sounding (right; CAPE = 883 J/kg, CIN = -5.9 J/kg, LFC = 865mb). Values quoted are for non-entraining parcels, but buoyancy for entrainment rates 0-50 %/100mb are also shown. Precipitation is assumed to maintain liquid water content <3 g/kg. A nearly-pseudoadiabatic virtual buoyancy (without liquid water loading, but with the 3g/kg liquid heat capacity included) is shown in dotted lines. Ice processes are neglected.

3.2 Disturbed conditions: less CIN, humid aloft, with gust fronts

Figure 3b shows the buoyancy of the same parcel, in the mean sounding characterizing disturbed conditions. The undilute parcel's buoyancy is greater below the 530 mb level, and less above, as could be deduced from Fig. 1. The decrease in upper-tropospheric buoyancy makes a slightly greater contribution to CAPE, which is now 885 J/kg, or ~10% less. Negative area or CIN, however, differs by over 40%, now -5.9 J/kg. Bouyancy is much less affected by entrainment in this moist environment: every entrainment rate shown permits some positive buoyancy at 600 mb.

Furthermore, this parcel experiences positive buoyancy exceeding the CIN in the cooler mean mixed layer of the disturbed conditions, below the ~940 mb level. This somewhat unusual convention of using a parcel with different virtual temperature than the mixed-layer sounding through which it ascends serves as a simple indicator that typical gust fronts (cold outflows) present in disturbed conditions can perhaps possess sufficient energy to lift adjacent air through 6 J/kg of CIN.

3.3 Is 5-10 J/kg of convective inhibition worth noticing?

Given the existence of theories which neglect CAPE values of hundreds to thousands of J/kg, perhaps a number like 10 J/kg cannot possibly be of large-scale or climatic importance. Suppose we convert this number to a vertical kinetic energy that boundary-layer air would need in order to coast its way through this several hundred meter layer in which it is negatively buoyant (Fig. 3a). If $w^2/2 = 10$ J/kg, this requires $w = 20^{1/2} = 4.5$ m/s. Aircraft observations (Fig. 4) suggest that the plentiful existence of gust fronts of the required strength cannot be taken for granted, even in large gridboxes hundreds of km on a side.

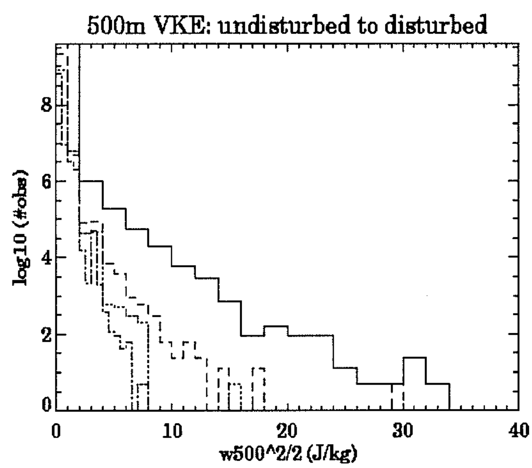

Figure 4. Histograms of estimated vertical kinetic energy at boundary-layer top, derived from 1 Hz aircraft observations. Vertical velocities from 100-1000m altitudes have been scaled by the factor 500m/altitude to yield an estimate of 500 m vertical velocity, which was then converted to kinetic energy. Data are from NOAA WP-3D aircraft during COARE flights on dates 921128H (dash-dot), 921126H (dot), 930209I (dash), and 930206I (solid), typically 4-5 hours of data each.

4. SUBGRIDSCALE FLUCTUATIONS: THE BAD NEWS

The numbers, such as CIN (Colby 1984, Crook 1996), involved in these convective inhibition and initiation arguments are dreadfully small. Not small enough to be physically unimportant, as discussed above, but too small to resolve in low-vertical-resolution models, or even to measure accurately with rawinsondes. Furthermore, sub-gridbox-scale fluctuations of boundary-layer θ_e, low-level virtual temperature, and gust-front vertical velocity are large and undountedly correlated. For example, consider the COARE aircraft observations of Fig. 4. On 15 December, the NCAR Electra L-188 aircraft was departing a decaying mesoscale convective system (MCS). It descended from 700 mb to >1000mb while traversing ~10-75 km southeast of the front edge of this system, then ascended through the same levels while continuing in the same direction. The left panel shows temperature and mixing ratio differences between these two soundings. The ~1C cooler temperatures and ~1g/kg higher humidity at 850mb and 700mb in the sounding nearer the convection are indicative of upward displacements, as in the crest of a gravity wave. These density gradients of ~1C / 100km must be highly transient.

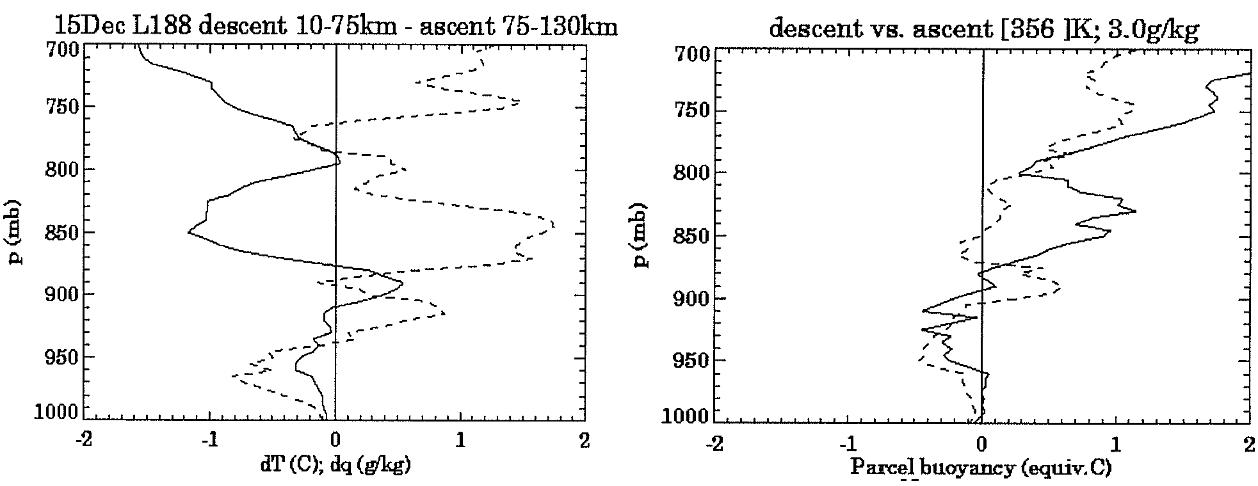

Figure 5. COARE electra observations

The buoyancy profiles experienced by a parcel of the mean boundary-layer (950-1000mb) air, in each of the two soundings, is plotted in the right hand panel. The fate of buoyant convective clouds would be considerably different in these two environments. However, it is doubtful whether either environment even existed in a single vertical column, and would exist long enough for a cloud to feel it! Clearly the gravity-wave and convective cloud fields are intimately linked, along with boundary-layer fields. Systematically correlated subgridscale fluctuations are as large or larger than the disturbed-dry differences identified in Figs 1-3. See also Weckwerth et al., 1996.

Furthermore, surface fluxes in the absence of any moist convection (deep or shallow) can completely destroy the small convective inhibition in minutes to hours (e.g. Raymond 1995). Presumably this is why shallow convection is so ubiquitous over tropical oceans. What role do inhibition and triggering processes play in determining the distinction between shallow (nonprecipitating) and deep (precipitating) convection? Mixing with dry air aloft is very important to parcel buoyancy

(Fig. 3). How does it depend on "parcel" size, and on the *local* humidity aloft (which may differ systematically from area-mean values). Is the *energy* of activation the only relevant variable, or does mesoscale organization matter too (e.g. through an *entropy* of activation)?

5. MUST CONVECTIVE INHIBITION BE TAKEN SERIOUSLY ON LARGE SCALES?

All in all, an activation-control descrition of convection appears daunting, and it is tempting to keep trying to tune supply-side parameterizations rather than trying to represent the fickle, fluctuating elements of convective demand. On the other hand, if this is the physics of convection, perhaps we need to replicate it in parameterizations, at least in spirit, since full detail is impossible. Cumulus parameterizations for mesoscale models are designed with some consideration of the low-level activation processes necessary for convective development, and these models seem to predict rainfall patterns with some accuracy. Is this success necessarily due to higher model resolution?

A simple parameterization of activation control of convection might require carrying another prognostic variable, representing low-level variance or noise. This variable might be thought of as crudely representing the widths of the statistical distributions of θ_e (Fig. 2), of the vigor of gust fronts (Fig. 4), and of the amplitude of the field of gravity-waves that modulate low-level buoyancy (as in Fig. 5), all lumped together. As a result, this low-level noise variable would be unverifiable; it would have to be viewed simply as an internal or control variable. Convection would tend to increase this subgridscale noise, so it would function as a positive feedback and as a mechanism for persistence - "when it rains, it pours." Land and sea breezes could be represented as local enhancements of this variance. On the other hand, convective downdrafts also act to decrease the *mean* boundary-layer θ_e to the point where even a high-variance environment cannot support convection, so this approach need not give rise to runaway gridpoint storms. Variance should decay with time, on a timescale of hours (representing surface fluxes, friction, vertical gravity wave propagation).

The main value of a low-level control scheme for deep convection would be its sensitivity to low-level processes, including inversions, such as the trade inversion, and low-level dynamical lifting in clear air, such as that performed by easterly-wave secondary circulations. Current models seem to have their convection occurring too broadly and blandly, i.e. they fail to adequately simulate the frequent "inhibitedness" of deep convection. Double ITCZs and too-weak large-scale wind fields are common, and these problems are exacerbated in coupled models (R. Seager, pers comm). Surely making the convection sensitive to the ubiquitous trade inversion, and to dryness aloft (through mixing), would improve matters.

Ideally, shallow and deep convection should be handled by the same scheme. Some form of mixing, whether entrainment or stochastic mixing (e.g. Raymond and Blyth 1992), must be included to give a range of clouds, sensitive to humidity aloft. The association with the boundary layer is so tight that perhaps the whole problem is best considered together (e.g. Qian et al 1997).

Cloud-resolving models could be useful for quantifying and formulating the details. I suspect that eventually such models will validate the common observation that low-level inhibition is the valve controlling deep convection variations. At present, however, lateral boundary condition issues, both numerical and conceptual, are preventing these models from being used even to quantify the obvious. Current work by the author aims to develop a modeling strategy to address these issues.

Perhaps inhibition processes can be grafted onto current GCM parameterization schemes, and retuning can achieve some relevant balance of sensitivites of convection to low-level inhibition and to moisture or available-energy supplies. Certainly this would be better than a scheme with no particular low-level sensitivities at all.

6. ACKNOWLEDGEMENTS

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