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Atmospheric Moist Convection

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Abstract Various forms of atmospheric moist convection are reviewed through a consideration of three prevalent regimes: stratocumulus, trade-wind, and deep, precipitating, maritime convection. These regimes are chosen because they are structural components of the general circulation of the atmosphere and because they highlight distinguishing features of this polymorphous phenomena. In particular we emphasize the ways in which varied forms of moist convection communicates with remote parts of the flow through mechanisms other than the rearrangement of fluid parcels. These include radiative, gravity wave and or microphysical (precipitation) processes. For each regime basic aspects of its phenomenology are presented along with theoretical frameworks which have arisen to help rationalize the phenomenology. Throughout we emphasize that the increased capacity for numerical simulation and increasingly refined remote sensing capabilities position the community well for major advances in the coming years.

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1 INTRODUCTION

Moist (atmospheric) convection, manifest as clouds, both engenders and celebrates the irreversibility of atmospheric motions. It expresses itself with a wonderfully rich phenomenology spanning a fantastic range of scales, and is involved intricately in many of the central problems in meteorology and climate science. Severe weather, including hurricanes, flash-floods, electrical storms, and tornadoes, invariably involves deep precipitating moist convection. Both today's climate and its susceptibility to human perturbations are thought to depend in crucial, yet subtle, ways on the behavior of all forms of moist convection. Despite a concrete appreciation of the role of moist convection in a wide variety of pressing problems, our understanding of this basic process remains amorphous, in part because unlike its dry counterpart moist convection is not one or two, but many things. Despite this remarkable complexity one gets the impression that at least some of the mysteries of moist convection are beginning to succumb to the ever increasing power of numerical simulation and remote sensing—an impression we attempt to develop when possible.

The basic forms of moist convection around which we develop our main ideas are those that one might expect to find in a thermally direct circulation such as sketched in Fig. 1. This type of figure is often thought of in terms of the Earth's Hadley cell, but also can be used to describe the Walker Cell, a zonal overturning circulation in the equatorial Pacific. Here deep precipitating cumulus towers, crowned by anvils of ice, mediate rising motion near the equator; stratocumulus veil the cold subtropical ocean; in between trade-cumulus deepen the atmospheric boundary layer, enhancing surface evaporation and fueling, in part, the overturning circulation. After introducing some basic concepts and terminology in section two, our discussion in sections 3-5 is structured around the basic types of convection, stratocumulus, trade-wind cumulus and cumulonimbus, featured in Fig. 1. Dividing the problem in this way helps highlight the different ways in which moist convection expresses its irreversibility, which in the end might help explain why moist convection is many things. This approach also encourages us to think more about the statistics of convecting layers in their entirety, rather than the details of individual clouds (especially their microphysical characteristics), or their transient behavior, both of which happen to be responsible for some of the more remarkable phenomena that motivate interest in the subject in the first place—but on a topic of this breadth, such compromises seem unavoidable.

2 ATMOSPHERIC THERMODYNAMICS

The language of moist convection is the idiosyncratic language of moist (atmospheric) thermodynamics, where almost everything, from density, to entropy, to moisture content, finds expression as a temperature. Although elementary, and well covered in many texts (e.g., Iribarne and Godson, 1973; Emanuel, 1994; Bohren and Albrecht, 1998), we review the main ideas here for the benefit of those with less background in the atmospheric sciences. For purposes of orientation, we begin by first reviewing dry atmospheric thermodynamics wherein the air is taken to be a perfect gas with a fixed composition.

2.1 Dry Air

The state of the dry system is determined by any two state variables, for instance temperature, T, and pressure, p. However for a hydrostatic atmosphere $\partial_z p = -\rho g$ (where the density ρ is a function of state and g is the gravitational acceleration) the spatial distribution of T determines p as a function of z. Consequently, the state of the hydrostatic system can be completely determined by the distribution of T in space and time, which we denote by (x, y, z, t) with z pointing upwards. For most applications instead of T, it proves useful to describe the state of the system in terms of the potential temperature, θ defined as

$$\theta \equiv T \left(\frac{\pi}{p} \right)^{R/c_p} \tag{1}$$

where π is a specified reference pressure (typically taken as 1000 hPa) $R = 287.15$ J kg^{-1} K^{-1} is the gas constant and $c_p = 1005$ J kg^{-1} K^{-1} the isobaric specific heat of the working fluid, which in this case is "dry air," the name commonly given to an ideal mixture of N_2, O_2 and Ar (Argon) with number fractions of 0.7809, 0.2095 and 0.00934 respectively. Physically θ is the temperature air at some initial temperature T and pressure p would have if it were isentropically brought to a reference state pressure π; consequently it does not vary with pressure (and hence z) for isentropic displacement of fluid parcels—which is one of its chief virtues. Because entropy differences from an arbitrary reference state vary as $c_p \ln \theta$, isopleths of θ can be identified with isentropes of the system.

Vertical accelerations in the atmosphere can be associated with imbalances in gravitational and the hydrostatic component of the pressure forces, i.e., buoyancy perturbations, which appear as the first term on the rhs of the vertical momentum equation:

$$\left(\partial_t + \mathbf{u} \cdot \nabla \right) w = -g \frac{\rho'}{\rho} - \partial_z p' + \nu \nabla^2 w, \tag{2}$$

where here primes denote deviations from a hydrostatic reference state which in what follows is denoted by subscript 0, ν is the kinematic viscosity, and $\mathbf{u} \equiv \{u, v, w\}$ is the velocity vector. The relationship between density and the state of the system is given by the equation of state, which for a dry atmosphere is the ideal gas law, $p = \rho R T$, the linearization of which yields

$$\frac{\rho'}{\rho_0} \approx -\frac{T'}{T_0} + \frac{p'}{p_0} \approx -\frac{\theta'}{\theta_0} + \left(\frac{c_v}{c_p} \right) \frac{p'}{p_0}, \tag{3}$$

where c_v is the isometric specific heat of the working fluid. For most scales of motion in the troposphere, $\rho' << \rho_0$. Because it is also possible to pick a basic state such that pressure perturbations contribute negligibly to density variations, the buoyancy can be related to the state of the system as

$$b = g \frac{\rho'}{\rho} \approx g \frac{\rho'}{\rho_0} \approx -g \frac{\theta'}{\theta_0}. \tag{4}$$

Hence buoyancy is effectively proportional to entropy variations.

In the non-diffusive, inviscid limit it is straightforward to show that for a horizontally homogeneous fluid at rest, at any given point the fluid is stable, neutral or unstable to infinitesimal perturbations according to whether $\partial_z \theta$ is less than, equal to, or greater than zero. In response to such an instability the fluid convects with the purpose of rearranging fluid parcels so that θ is non-decreasing. In practice these motions induce filamentation of fluid elements which then lays the basis for molecular dissipation of both temperature (buoyancy) and velocity perturbations, thus convection tends to drive the fluid to the neutrally stratified state of $\partial_z\theta = 0$, which under suitably chosen constraints is also the state of maximum entropy (Verkley and Gerkema, 2004). In such a state $\Gamma_d \equiv \partial_z T = -g/c_p$ which is exactly the lapse rate in temperature required so that the reduction in the specific enthalpy (c_pT) with height equals the increase in the potential energy (gz). This is called the dry-adiabatic lapse rate, and is precisely what is to be expected for isentropic, vertical displacements of dry fluid parcels.

2.2 Moist (Warm) Air

To describe the state of a moist atmosphere requires some measure of the water within a control volume. A common choice is, q_t, the total water specific humidity (defined as the mass fraction of H_2O in the system). The mass fraction of dry air follows as the remainder, $q_d = 1 - q_t$. Although the partitioning of water mass among its vapor q_v, and condensate q_c forms (which can be liquid, denoted q_l, or ice, denoted q_i) is a strong function of temperature (and hence pressure, equivalently altitude), q_t does not change following reversible fluid displacements. The presence of a variable constituent in the moist system implies that partial derivatives of the working fluid (the gas constant, specific heats, *etc.*) will vary with the composition of the fluid. Both this and phase changes complicate the development of the thermodynamics of moist air. Because the propensity of ice toward non-equilibrium introduces particular complications and because many of the interesting aspects of moist convecting atmospheres are apparent in the absence of ice, most of our ensuing development is for convection in the absence of ice processes.

Similar to the dry system, instead of temperature it is useful to use thermodynamic coordinates, which like q_t (and θ for the dry system) are invariant following reversible rearrangements of fluid parcels. Typically the choice of a coordinate carrying information about temperature is based on a moist generalization of θ. Because this generalization must specify the disposition of the state of the water mass in the reference state two choices arise naturally: (i) a reference state for which $p = \pi$ and in which all the water is in the vapor state; and (ii) a reference state in which $p = \pi$ and in which all the water is in the liquid state. Temperatures obtained by isentropically moving to these reference states are called the liquid water potential temperature, θ_l and equivalent potential temperatures, θ_e, respectively. By neglecting differences among the specific heats for dry air, water vapor and liquid water, and differences between the gas constants for vapor vapor, it is straightforward to show that

$$\theta_l \approx \theta \exp\left(\frac{-q_l L_v}{c_{p,d}T}\right), \quad \text{and} \quad \theta_e \approx \theta \exp\left(\frac{+q_v L_v}{c_{p,d}T}\right) \tag{5}$$

Although more accurate expressions can be derived by accounting for compositional effects on c_p and R (e.g., Emanuel, 1994), the above expressions more clearly expresses the dominant physical processes at play and are sufficient for most purposes. From Eq. 5, θ_l is readily interpreted as an evaporation temperature, which reduces to θ in the absence of condensate. In saturated conditions the difference between θ_l and θ simply expresses the enthalpy of vaporization released through the formation of any condensation, which is responsible for constancy of θ_l. Similarly, θ_e can be interpreted as a condensation temperature. It too is invariant to changes in phase, which can be seen by noting that an increase in θ during a reversible change in phase of water is offset by a decrease in q_v (which must be equal to q_s the saturation specific humidity) in saturated conditions. A $\{\theta_l, q_t\}$ representation of the state space provides a more orthogonal basis than a $\{\theta_e, q_t\}$ representation because typically $q_l \ll q_t$. On the other hand, θ_e has the advantage of being insensitive to changes in the amount of condensate present, which motivates its use in studies of precipitating moist convection. Additionally, by replacing q_v by q_s in the definition for θ_e above, one can construct a state variable called the saturated equivalent potential temperature, $\theta_{e,s}$, whose chief virtue is that it is independent of the moisture content of the atmosphere yet is invariant following reversible displacements of saturated fluid parcels.

Compositional effects on the gas constant of the moist fluid can not be neglected in so far as they contribute to density perturbations, where such effects are leading order. In particular, the gas constant R, which appears in the ideal gas law $p = \rho R_d T$ for the moist fluid, varies with the composition of the fluid, according to

$$R = R_d q_d + R_v q_v = R_d(1 + q_v R_v/R_d - q_t). \tag{6}$$

Because meteorologists prefer to work with the dry air gas constant, R_d, it has become customary to define a virtual, or effective temperature, $T_v = T(1 + q_v(R_v/R_d) - q_t)$ which carries the compositional dependence of R in the moist system. Physically, T_v (or analogously $\theta_v \equiv T_v(\pi/p)^{R_d/c_{p,d}}$) is the temperature required of dry air to have the same density as moist air. Defined in this fashion it follows that whereas for the dry system $\rho'/\rho_0 \approx -\theta'/\theta_0$, for the moist system $\rho'/\rho_0 \approx -\theta_v'/\theta_{v,0}$. However, unlike dry systems, for which density perturbations can be linearly related to entropy perturbations, in the moist system at best one can define a piecewise linear relationship between buoyancy and state variables, depending on whether or not air is saturated. For instance, for a horizontally homogeneous system whose state is described by θ_l, q_t

$$b = -g\frac{\rho'}{\rho_0} = g\begin{cases} \alpha_u(\theta_l'/\theta_0) + \beta_u q_t' & q_t < q_s \\ \alpha_s(\theta_l'/\theta_0) + \beta_s q_t' & \text{otherwise.} \end{cases} \tag{7}$$

The partial derivatives $(\alpha_u, \alpha_s, \beta_u, \beta_s)$ are functions of state, which (with the help of the Clausius-Clapeyron equation) can be determined analytically . For shallow flows they are often approximated as constant, i.e., for $\theta_l = 288K$ and $q_t = 10$ g kg^{-1},

$$\alpha_u = 1.06, \quad \beta_u = 0.608 \quad \alpha_s = 0.49, \quad \beta_s = 3.3.$$

Compositional effects on fluid density account for the non-zero value of β_u, but the difference between the saturated and unsaturated value of the coefficients primarily encapsulate the effects of phase changes. For instance, $\alpha_s < \alpha_u$ implies that in a saturated fluid a positive perturbation in θ_l projects less strongly on to temperature (and hence density) than for an unsaturated fluid, as it is partially offset by evaporation which cools the fluid. In contrast, because q_t perturbations induce phase changes in saturated fluids they project strongly onto density through temperature (rather than just compositional) variations when the fluid is saturated. Although the compositional effect is usually thought to be small, over the tropical oceans it is not unusual for roughly half of the buoyancy of thermals in the unsaturated marine boundary layer to be attributable to this effect.

In general, the effect of phase changes, embodied in the discontinuous nature of the partial derivatives α and β (but also in other partial derivatives of the system, for instance temperature changes along isentropes) is the origin of many fascinating aspects of moist (atmospheric) convection. Given $\{\theta_l, q_t\}$, whether or not a fluid parcel is saturated depends only on pressure, which (because hydrostatic pressure variations dominate) varies principally with the vertical coordinate. Thus the basic behavior of fluid parcels (as measured by their partial derivatives) depends on their position. It is almost as if the fluid is magically transformed into another form once it crosses a certain threshold. This can be thought of as a Cinderella effect, where instead of pumpkins turning into carriages and back again at a certain hour, fluid parcels change their qualitative behavior at a certain altitude. But more magical still, the witching hour (location) of each pumpkin (fluid parcel) varies with its state. In summary, moist convection can in many instances be thought of as a two fluid problem, where one fluid (unsaturated air) can transform itself into another (saturated air) simply through a vertical displacement. This possibility greatly augments the basic anisotropy due to gravity, which begets convection of all kinds to begin with.

A striking implication of the non-linearity, or two-fluid behavior, embodied in Eq. (7) is the concept of buoyancy reversal, wherein when dry air is mixed with saturated air of greater density it attains densities greater than those of its individual components. A common geometry for this phenomenon is illustrated in the left panel of Fig. 2. Roughly speaking buoyancy reversal occurs when there exists some mixing fraction χ such that $\rho_*(\chi) < \rho_-$, where subscript "*" and "-" denote the state of the mixture, and the underlying air respectively. This situation can be thought to occur when the temperature difference between the upper and lower layer is not sufficient to offset evaporative cooling which will accompany any mixing between the two layers. At this level of approximation this occurs whenever $\theta_{e+} < \theta_{e-}$. The inclusion of compositional effects in the equation of state results in a somewhat more stringent criterion:

$$\kappa \equiv 1 + \frac{\theta_{l+} - \theta_{l,-}}{(L/c_p)(q_{t+} - q_{t-})} = \frac{\theta_{e+} - \theta_{e,-}}{(L/c_p)(q_{t+} - q_{t-})} > \frac{c_p\theta_0}{L_v\alpha_s} \equiv \kappa_* \tag{8}$$

which says that whenever κ exceeds some threshold value ($\kappa_* \approx 0.2$) then some mixtures will support buoyancy reversal. How strong this effect is depends on two non-dimensional numbers, κ and χ_s, the latter being the mixing fraction of the just saturated mixture which measures the sub-saturation of the upper layer relative to the condensate available for evaporation in the lower layer. Buoyancy reversal has no counter part in dry convection, although to the extent it is realized only under dissipative processes it bears some similarity to double-diffusion. Situations in which buoyancy reversal is to be expected are commonly encountered in the atmosphere, but its role in regulating the structure of moist convecting atmospheres remains controversial, a point we shall return to.

The Cinderella effect also greatly enriches conceptions of stability. With respect to infinitesimal perturbations, $\partial_z b$ still demarcates the condition of neutral stability; however, from (7) it is apparent that this condition implies different relationships between $\partial_z\theta_l$ and $\partial_z q_t$ based on whether or not the fluid is saturated, which in turn depends on the amplitude of the displacement of fluid parcels. Thus the stability of the atmosphere can look different for finite amplitude, as opposed to infinitesimal, fluid displacements. Even in the isentropic limit the stability of the fluid column has to account for the energy available to global rearrangements of fluid parcels.

To illustrate these types of effects consider Fig. 3. Here we describe the state of the atmosphere using the solid lines in $(q_t, -\ln p)$ space (left panel), and in $(T, -\ln p)$ space (right panel) — where in the discussion that follows, z and $-\ln p$ are used interchangeably. The environmental temperature is chosen to decrease uniformly with height at the environmental lapse rate, Γ_e. The moisture field is chosen to decay exponentially such that the environment is everywhere sub-saturated. In this state the atmosphere is stable to small displacements because $\Gamma_e < \Gamma_u$, where $\Gamma_u \equiv (\partial_z T)_{\theta_l, q_t, u}$ is the temperature lapse rate for isentropic, unsaturated, vertical displacements. Because the corrections to c_p are small for an unsaturated fluid $\Gamma_u \approx \Gamma_d \approx 10$ K km^{-1}.

A starting point for addressing the global stability of the atmospheric profile is the analysis of the energetics of test parcels displaced isentropically from a specified level. Such a process is labeled by subscript $''*''$ and illustrated by the dotted/dashed lines in the thermodynamic diagrams of Fig. 3, drawn for a test parcel lifted from the surface. By definition T_* initially decreases at the rate Γ_u and q_{t*} will remain constant. For a linear decrease in T_* with altitude q_{s*} decreases exponentially, rapidly attaining the value of q_{t*}. The pressure, p_c at which they are first equal is called the lifting condensation level (or saturation pressure level), above which the test parcel becomes saturated. Further vertical displacements result in a reduced rate of temperature decline with height, according to: $\Gamma_s \equiv (\partial_z T_v)_{\theta_l, q_t, s} = -g/c_*$ where

$$c_* \approx c_{p,d} \left(\frac{1 + \frac{L_v^2 q_s}{c_{p,d} R_v T^2}}{1 + \frac{L_v q_s}{R_d T}} \right)$$

is an effective heat capacity; c_* is a strong function of temperature, ranging from $2.5 c_{p,d}$ at high temperatures and asymptoting to $c_{p,d}$ as temperature decreases. Consequently Γ_s can be as small as 4 K km^{-1} in the lower troposphere, and asymptotes to Γ_d aloft. In this example $\Gamma_u < \Gamma_e < \Gamma_s$ at the warmer temperatures of the lower troposphere, but $\Gamma_u < \Gamma_s < \Gamma_e$ aloft. Consequently, during the initial part of its saturated ascent the test parcel warms relative to the environment, while aloft it cools relative to the environment. Moreover, we have chosen Γ_e and q_t such that the warming above the lifting condensation level is sufficient for the parcel to become warmer than the environment (at what is called the level of free convection, p_f). Aloft as the parcels cools more rapidly with further ascent, it once again attains the environmental temperature at the level of neutral buoyancy, p_n.

The distinction between the saturated and unsaturated lapse rates in the lower troposphere permits meta-stable profiles (or sub-critical instabilities), as implied by Fig. 3. For these situations the atmosphere is stable to infinitesimal displacements, but unstable to larger displacements. This is a common situation in the atmosphere. Profiles for which $\Gamma_u < \Gamma_s < \Gamma_e$ somewhere, are often called conditionally unstable; although this terminology is something of a misnomer (cf., Sherwood, 2000) because even for finite displacements, whether or not any instability is realized depends on sufficient moisture being present for test parcels to take advantage of the favorable thermal structure of the atmosphere. For example, by sufficiently reducing the available water vapor in the profile in Fig. 3 the sounding would remain conditionally unstable, but the displacement of any test parcel, finite amplitude or otherwise, would be capable of extracting energy from the mean state. Thus the convective instability is better measured in terms of the work done by a test parcel in moving from its initial position p_1 (in the above example the surface) to some final pressure level, p_2 (in the above example, p_n) as simply

$$W(p_1, p_2) = R_d \int_{p_1}^{p_2} (T_{v*} - T_v) d \ln p. \tag{9}$$

With respect to Fig. 3 $W(p_s, p_f)$ measures the negative area on the thermodynamic diagram, and is called the convective inhibition (sometimes abbreviated CIN), while $W(p_f, p_n)$ measures the positive area (similar to what is sketched in Fig. 3 but incorporating compositional, or virtual, effects as expressed by T_v) and is sometimes called the convective available potential energy, or CAPE. The presence of CIN facilitates the accumulation of CAPE, which when tapped often results in strong transient convection associated with severe weather.

Although, W is frequently used as a measure of the energy available for convection it has three major shortcomings: (i) it is sensitive to the starting and ending points of the test parcel; (ii) it neglects the work demanded by continuity, i.e., the compensating motions of the environment; (iii) it neglects the effects of irreversible processes (ranging from precipitation to mixing) in determining T_{v*}. The first two points can be addressed by using a generalized measure of CAPE (e.g., GCAPE Randall and Wang, 1992) as follows: Let $W_{i,j}$ be a discrete version of W such that i and j index the starting and ending pressures p_i and p_j respectively. Then $\tilde{W} = \sum_{i,j} W_{ij} P_{ij}$ defines the energy of some permutation of the system as defined by the permutation matrix[1] P_{ij}. In an n-layer system there are $n!$ possible permutations, so if we index by k the energetic cost (reward) of the kth permutation, then generalized CAPE, or GCAPE is simply given by $\max\{\tilde{W}_k; k = 1, .., n!\} \geq 0$, where the lower bound is set by the trivial perturbation (i.e., no change). Although the generalized CAPE suffers from neither the starting point sensitivity nor the neglect of compensating motions implicit in the definition of W it does implicitly assume (through the definition of T_{v*} in Eq. (9)) that rearrangements are isentropic. Both CAPE and GCAPE are limited in that they attempt to characterize with a single number the stability characteristics of the atmosphere, although the latter provides a more general framework to look at the energetics of a family of displacements (and hence things like CIN).

In constructing a language for our subsequent discussion we have, for the most part, focused on moist convection as an isentropic process. The buoyancy reversal argument is an irreversible argument; however, it is irreversible in a way which is common to most other forms of convection, namely via local mixing of fluid parcels. But as we shall see, one of the more fascinating aspects of moist convection is the varying ways in which it expresses its irreversibility. The formation of condensate not only allows for strong, local interactions with radiant energy at both terrestrial and solar wavelengths, it also leads to the formation of precipitation which transports enthalpy across fluid trajectories. Both are long-range interactions for which there is no analog in dry convection.

[1] A permutation matrix of an n-element set is an $n \times n$ matrix with one element of each row and one element of each column being unity and the remaining elements being zero.

3 STRATOCUMULUS CONVECTION

Stratocumulus is a low-lying, characteristically stratiform cloud type, usually exhibiting evidence of underlying cellular structure. Because moist convection is often identified with cumuliform clouds, stratocumulus is easily omitted from the canon of moist convection. Because it highlights what is arguably a defining aspect of moist convection in the earth's atmosphere, namely its capacity to efficiently interact with radiant streams of energy, such omissions are a mistake. From a climatological perspective regions where stratocumulus prevail are most evident where there exists great thermal contrast between the overlying free atmosphere and the underlying surface: for instance over the upwelling regions of the subtropical oceans, which the right side of Fig. 1 is meant to reflect; but also in the storm tracks, where in the latter they are most prevalent during periods dominated by anticyclonic circulations; and at the poles.

The tendency of stratocumulus to be most evident in subsiding maritime environments, where the thermal contrast between the surface and free atmosphere can be very pronounced, has been recognized for nearly a century. Recent studies have more systematically explored these relationships be examining the correlation between stratocumulus cloud incidence and the degree of thermal contrast between the surface and the free atmosphere, with the latter being measured by the lower tropospheric stability, defined as difference between θ at 700 hPa and its value at the surface. Nearly two-thirds of the inter-annual variability in the low cloud amount in the subtropical regions can be explained by variability in the lower tropospheric stability (Klein and Hartmann, 1993). On shorter timescales the variability is less, as is the fraction of the variance it can explain. Additionally as one moves out of the heart of the stratocumulus regimes, the lower tropospheric stability becomes a less important indicator of low cloud amount, and the role of cold advection (as measured by the the wind speed times the surface temperature gradient in the direction of the mean wind) becomes more important (e.g., Klein, 1997). The suggestion of this analysis is that cloud amount strongly reflects the upstream conditions, indicative of memory in the system. These ideas are amplified by recent studies using satellite data (e.g., Pincus et al., 1997) which indicate that cloud amounts best correlate with conditions 24 hours upstream of a given observation. In what follows we elaborate on this statistical overview of stratocumulus and its importance by presenting an overview of important aspects of its phenomenology, and the theoretical framework that has arisen to help rationalize this phenomenology. A more comprehensive overview is presented by (Moeng, 1998).

3.1 Phenomenological Overview

Because it is relatively easy to sample experimentally, and because of its importance to the radiative balance of the planet as a whole, stratocumulus may be one of the best sampled, best understood, and yet least recognized, forms of moist convection. The latter is the result of a common misconception wherein moist convection is synonymous with cumuliform clouds for which updrafts are saturated, and downdrafts are unsaturated. In contrast, the upper parts of both the up and downdrafts in stratocumulus are saturated, with the basic geometry illustrated by Fig. 4. Here a relatively shallow, cool and moist thermal boundary layer is capped by a much warmer and drier subsiding atmosphere. Quantities which are invariant under displacements along isentropes (e.g., θ_l and q_t in the profiles superimposed on the the figure) transition very sharply between their boundary layer and free tropospheric values. In the boundary layer they tend to be well mixed as a result of convective turbulence generated by infrared radiative cooling at the top of the clouds. Unlike the dry convective boundary layer, which develops from the solar heating of the land surface every day, the infrared driving of stratocumulus favors the night. This nocturnal proclivity is especially evident from five days of ship-board measurements made in regions of the south-east pacific known to favor stratocumulus convection (Bretherton et al., 2004).

The basic processes thought to compete in determining the thermodynamic state of stratocumulus forced in this way are also illustrated in Fig. 4. Here the radiatively driven convective turbulence impinges on the cloud top interface, entraining warm and dry air from aloft. From the point of view of the mass budget, the entrainment deepening of the boundary layer works against a gently subsiding large-scale environment. The radiative cooling competes with entrainment warming and surface heat fluxes, while entrainment drying acts to offset the moistening effect of surface fluxes.

Because the degree to which adiabatic invariants of the flow are well mixed with height in the stratocumulus boundary layers is a foundation of many theoretical descriptions, it is important to establish the degree to which this idealization has merit. For the flight whose data is incorporated into Fig. 4 this has been evaluated in several ways: by comparing the extent to which the mean cloud base, as measured by an upward looking lidar, agrees with that predicted by the saturation pressure, p_s of the state variables measured *in situ* below cloud base (e.g., Feingold and Morley, 2003); by comparing the degree to which q_l increases with height at the rate one would expect for a layer in which q_t and θ_l are constant (illustrated by the dash-dot line in Fig. 4); by examining the extent to which rms vertical velocity maximize in the flow interior and decrease to near zero toward the boundaries (Stevens et al., 2003). By all of these measures this limit has merit.

Many studies similar to that discussed above has helped establish the canonical view of stratocumulus as a well-mixed, radiatively driven, non-precipitating thermal boundary. Naturally more recent work has begun to probe the limits of such idealizations, focusing in turn on when and how the well-mixed structure of the layer breaks down, how other driving forces affect the evolution of the layer, and the extent to which drizzle helps balance the water and heat budget. Despite the degree to which well mixedness serves as a useful organizing principle for understanding stratocumulus, departures from this state are not hard to find. Very early studies (e.g., Vernon, 1936) documented the degree to which the diurnal variation of stratocumulus in coastal regions is accompanied by vertical stratification in state variables, with θ_l increasing and q_t, decreasing with height. Aircraft measurements above the North-Sea, (Nicholls, 1984) show that such situations are also common over the ocean, and tend to be accompanied by a two layer structure wherein a warmer drier cloud layer is separated from a cooler moister surface layer by a stably stratified transition layer which can be of variable depth. Such structures tend to be more evident during the day, and in deeper boundary layers, and have come to be called decoupled boundary layers. For such layers radiative cooling from the cloud top can help maintain an elevated mixed layer which continues to dry by entrainment, but the lack of turbulent penetration to the surface cuts the cloud off from its compensating moisture supply.

The development of this decoupled structure has been hypothesized to be the first step in the dessication of the cloud layer. Its causes have been attributed to a variety of processes including the stabilizing effect of precipitation (Paluch and Lenschow, 1991; Wang, 1993; Wang and Wang, 1994) or solar radiation which is preferentially absorbed in the cloud layer (Bougeault, 1985; Turton and Nicholls, 1987; Duynkerke, 1989; Ciesielski et al., 2001).

Relatively little is known about precipitation from stratocumulus; although often idealized as non-precipitating (e.g., Fig. 4), over the years a variety of circumstantial evidence has accumulated to suggest that at times precipitation rates can be dynamically significant (e.g., Nicholls, 1984; Austin et al., 1995; Bretherton et al., 1995). A drizzle flux of 1 mm d^{-1} will warm a layer by approximately 30 W m^{-2}, which although somewhat smaller than the nocturnal flux divergences across the boundary layer, is commensurate with the diurnally averaged radiative cooling rates. Thus it has been hypothesized that sufficient drizzle may be capable of offsetting the radiative driving of the layer, both directly and indirectly (by depleting the cloud), leading to the desiccation of the cloud (Paluch and Lenschow, 1991). Air and ship-borne remote sensing (vanZanten et al., 2004; Comstock et al., 2004) has made possible more comprehensive surveys of precipitation in stratocumulus layers, which suggest that precipitation can be prevalent, with surface precipitation rates being most pronounced at night. Showers with rain rates of 1 cm d^{-1} are not uncommon, and precipitation rates averaged over large areas being sustained at values near 1 mm d^{-1} for many hours were observed approximately a third of the time, suggesting that drizzle may play a pivotal role in limiting stratocumulus depth. Yet more intriguing are recent observations (Stevens et al., 2004) which show that drizzle seems to be connected to the emergence of pockets, and perhaps even broader regions, of open cellular convection embedded in otherwise more overcast cloud regions—a striking example of which is shown in Fig. 5.

One of the more vexing problems pertaining to stratocumulus convection, is how and why the stratiform cloud layer breaks up into more broken or scattered convection, as is characterized by the downstream transition in low-level cloudiness in Fig. 1. One idea is that as the stratocumulus layer advects over warmer water the thermal contrast between the boundary layer is reduced while the moisture contrast is enhanced, leading to an increased potential for buoyancy reversal, (e.g., as measured by κ in Eq. 8). In the buoyancy reversal regime mixing with the free troposphere is argued to induce further mixing promoting the destruction of the cloud. This idea, first stated in the 1960s (Kraus, 1963; Lilly, 1968) is best known as the cloud top entrainment instability (CTEI) hypothesis, a term which was coined after it was refined to account for compositional effects on buoyancy (Deardorff, 1980; Randall, 1980). Early tests showed that regions of stratiform clouds were prevalent in conditions where the CTEI hypothesis would predict their demise (Kuo and Schubert, 1988; Albrecht, 1991), leading to refined arguments and more stringent criteria for cloud dissolution (MacVean and Mason, 1990; Siems and Bretherton, 1992; Duynkerke, 1993). Although none of these measures has proven to provide a compelling ordering of the data, the more stringent criteria tend to perform better. More recent observational data (De Roode and Duynkerke, 1997), and simulations (Lewellen and Lewellen, 1998; Moeng, 2000) lead to renewed interest in the original CTEI formulation, but analysis of yet more recent measurements for which $\kappa > \kappa_*$ is accompanied by a thickening of the cloud seemed to indicate that at the very least CTEI is not a sufficient conditions for cloud desiccation (Stevens et al., 2003).

In the past decade theoretical work (Krueger et al., 1995; Bretherton and Wyant, 1997; Stevens, 2000; Lewellen and Lewellen, 2002) has turned away from CTEI and begun to focus on a broader accounting of the energetics, hypothesizing instead that as non-radiative forcings begin to dominate the energetics of the stratocumulus-topped boundary layer the cloud can begin to entrain sufficiently large amounts of air to negate the radiative cooling, thereby requiring work to be done to mix the entrained air below cloud base. The concept of the cloud layer needing to do work on the sub-cloud layer in order to maintain a well mixed layer (elegantly illustrated by Schubert et al., 1979) arises from the two-fluid nature of moist convection embodied in (7). In such a situation the circulation should transition from an overcast stratocumulus-like circulation, to a more cumulus-like (unsaturated downdrafts) circulation. Consistent with the observations a variety of factors can influence this change, ranging from the state of the cloud top interface, to precipitation, solar radiation, enhanced cloud base warming by longwave radiative fluxes, and enhanced sensible and latent heat fluxes, some of which (surface latent heat fluxes and cloud base radiative warming) are particularly prevalent for deeper boundary layers.

3.2 Theoretical Perspectives

The theoretical foundation of our understanding of stratocumulus is Lilly's 1968 mixed layer theory. By integrating over the boundary layer (e.g., from $z = 0$ to h in Fig. 4), and by assuming a well mixed vertical structure, it is possible to describe the evolution in terms of a system of three ordinary differential equations:

$$\frac{D}{Dt}h = W + E \tag{10}$$

$$\frac{D}{Dt}\hat{\theta}_l = -\Delta F + V(\theta_{l,0} - \hat{\theta}_l) + E(\theta_{l,+} - \hat{\theta}_l) \tag{11}$$

$$\frac{D}{Dt}\hat{q}_t = -\Delta R + V(q_{t,0} - \hat{q}_t) + E(q_{t,+} - \hat{q}_t) \tag{12}$$

where hats denote the vertical average of variables (i.e., $\hat{\theta}_l = 288.96$ K and $\hat{q}_t = 8.91$ in Fig. 4.), and subscripts 0, 1 denote values of state variables at the surface and just above the boundary layer respectively. The substantial derivatives D/Dt denote a change following the mean horizontal flow, while ΔF and ΔR are sources of θ_l and q_t respectively, V is a surface exchange velocity used to parameterize the surface fluxes, similarly E parameterizes the entrainment fluxes. With V determined by a surface-exchange law, all that remains to close the theory is to determine E. Just how to do this has been, and remains a topic of great interest.

A starting point for evaluating E is the energetics of stratocumulus topped boundary layer. Such an approach takes advantage of a feature of the mixed layer theory in that the assumption on the vertical structure allows one to consistently couple the energetics of the flow (as measured by the buoyancy flux, $\mathcal{B} \equiv \overline{w'b'}$,) to the evolution of the mean state. This becomes evident by noting that the necessary condition for a mixed layer to remain well mixed is for the sum of the diabatic and turbulent fluxes of a state variable to be linear. That is, for a horizontally homogeneous flow

$$\partial_t \bar{\theta}_l = \partial_z \left(\overline{w'\theta_l'} + F \right), \tag{13}$$

then in order for $\partial_t \partial_z \theta_l$ to vanish, $\left(\overline{w'\theta_l'} + F \right)$ must be a linear function of z. Such a situation is referred to as a *quasi*-steady state. Given a knowledge of the diabatic forcings F and turbulent fluxes at the flow boundaries (which are given if E and V are known), quasi-stationarity determines the structure of $\overline{w'\theta_l'}$ in the flow interior. Because similar relations constrain $\overline{w'q_t'}$ the resultant profiles of the turbulent fluxes of q_t and θ_l can be used to determine $\mathcal{B}$, given (7).

Over the years a variety of approaches have been used to determine E as a functional of $\mathcal{B}$. Most initial work made closure assumptions for E in analogy to the dry convective boundary layer, for instance by choosing E such that $\mathcal{B}_{\min}/\mathcal{B}_{\max}$ is some fixed ratio, or so that the integral of $\mathcal{B}$ over its negative area is some fixed fraction of its integral over its positive area (Schubert, 1976; Kraus and Schaller, 1978),. Inspired in part by large-eddy simulation it has gradually come to be realized that constraints on the net buoyancy flux poorly bound the energetics of the flow, and that a more useful approach is to fix E by requiring that ratio of $\mathcal{B}$ to its value for a hypothetical, non-entraining, flow be fixed (Manins and Turner, 1978; Stage and Businger, 1981; Lewellen and Lewellen, 1998; Lock, 1998; vanZanten et al., 1999). Exactly how best to count the energetics is complicated by uncertainty about how to best account for diabatic fluxes which occur in the entrainment zone (e.g., Moeng et al., 1999; Moeng and Stevens, 1999), as well as how to relate fluxes of conserved variables to $\mathcal{B}$ at partially saturated interfaces such as cloud top (Lilly, 2002; Randall and Schubert, 2004). Many recent closure hypotheses for E have been calibrated using large-eddy simulation, but the ability of this approach remains controversial, in part because entrainment rates from LES can vary substantially from model to model and as a function of the representation of the cloud top interface (Stevens, 2002; Stevens et al., 2003). Despite significant uncertainty, a variety of recent work (ranging from observations of pronounced drizzle, to direct measurements of entrainment) seems consistent with the idea that entrainment in stratocumulus topped mixed layers is less efficient than predicted by many of the early closures.

Notwithstanding some vexing problems, and ready complications, idealized stratocumulus flows is arguably the form of moist convection most analogous to common forms of dry convection. For instance, the simplest stratocumulus flow could consist of a saturated fluid bounded above and below and driven by fixed buoyancy fluxes. Such a problem is isomorphic to the Rayleigh-Bénard problem for dry convection. Beginning with this problem, gradual steps toward more realistic stratocumulus layers could be made by studying the dynamics of partially saturated layers, in which case the non-dimensional cloud base height and the relative driving of q_t and θ_l fluxes to the flow enter as two non-dimensional numbers. In the limit of the saturated fluid consisting of infinitely small and numerous drops one could pose the problem in a way that is tractable for direct numerical simulation. By prescribing the boundary forcings this approach avoids dealing with the critical issue of entrainment, nonetheless it can provide interesting and important insight into the dynamics of stratocumulus layers. Despite its simplicity and relevance, very little research of this kind has taken place. It would seem to be a natural entry point for those from the broader fluids community who are looking to bring their expertise and creativity to meteorological flows.

4 SHALLOW CUMULUS CONVECTION

Shallow cumulus convection is ubiquitous. It spans large expanses of the world ocean, and is common over land during the day in periods when fair-weather prevails. From a popular perspective it is the archetypical form of moist atmospheric convection, although as compared to other forms of convection it is relatively infrequently studied. Its diminutive size and tendency not to be associated with precipitation renders it less visible than its shallower, albeit more stratiform counterpart on the one hand, and its deeper and more copiously precipitating counterpart on the other. In Fig. 1 it is represented by the middle portion of the circulation as isolated cumulus drafts confined to the lower troposphere. In this context it is often referred to trade-wind cumulus convection and is seen as a structural component of the thermal boundary layer in the trades. If the various modes of convection were drafted in proportion to their frequency of occurrence, the shallow regime would occupy most of the figure.

The motivation for studying shallow convection stems from an early recognition of its role in maintaining the structure of the lower troposphere in the trades and hence the intensity of large-scale circulations in the tropics (Riehl et al., 1951). The trade-wind structure is most significantly marked by the presence of a "trade-wind inversion" that is a zone of increasing temperatures, usually found somewhere between one and two kilometers, which marks a usually sharp transition between a relatively cool and moist turbulent boundary layer from the warm dry overlying free-troposphere. This inversion tends to be somewhat more diffuse and elevated than its counterpart which caps the stratocumulus topped boundary layer. It is ubiquitous in the trades (von Ficker, 1936; Neiburger et al., 1961) sub similar structures are sometimes evident between periods of precipitating convection in the deep tropics (Johnson and Lin, 1997). Studies have long argued that the maintenance of this inversion structure is accomplished through the action of shallow cumulus, which moisten and cool the subsiding free atmosphere (e.g., Nitta and Esbensen, 1974), thereby accumulating the large amounts of latent heat in the lower troposphere necessary to feed regions of deep precipitating convection and drive associated large-scale circulations (e.g., Riehl et al., 1951; Tiedtke et al., 1988). In what follows we briefly discuss the phenomenology and theory of shallow cumulus convection in the trades, with an eye towards recent developments. Excellent, and more comprehensive, reviews on this topic are given by Betts (1997) and Siebesma (1998)

4.1 Phenomenological Overview

The mean structure of the atmosphere in regions where shallow convection prevails is illustrated by the cartoon in Fig. 6. Many elements of the mean structure were described on the basis of aircraft observations quite early, most notably in a remarkable body of work by Malkus (1954, 1956, 1958). The mean structure in the cartoon is derived from large-eddy simulation, but it well illustrates many generic features of the trade-wind boundary layer: (i) A well mixed subcloud layer extending to cloud base; (ii) a transition layer of 100-300 m in depth through which cloud fraction increases and q_t decreases rapidly with height; (iii) a conditionally unstable cloud layer spanned by cumuliform clouds whose area coverage is typically less than 10%; (iv) an inversion layer of several hundred meters in extent which caps the clouds and separates the thermodynamics boundary layer from the overlying free atmosphere. This structure is also illustrated from the 915 Mhz reflectivity signal from an upward pointing radar in Fig. 7. The tendency for clouds to be confined to a relatively shallow layer (less than 2km) and their short lifetime is thought to inhibit the formation of precipitation, and for the most part *shallow* cumulus convection has come to be synonymous with *non-precipitating* cumulus convection (Betts, 1973). Unlike other forms of moist convection, for which important aspects of their description requires energy to be transported across flow streamlines, either by falling hydrometeors or by photons, in its ideal form shallow cumulus convection acts irreversibly only through local mixing with its environment; in this sense it is similar to dry convection. It differs from both dry convection and stratocumulus convection primarily in its inherent asymmetry of up- and down-ward motions. Upward currents are saturated and extract energy from the flow, downward currents tend to be unsaturated and must do work on the flow—this "up-moist, down-dry" asymmetry has long been thought to be why cumulus clouds tend to have small area fractions (Bjerknes, 1938).

Over the ocean the depth of the sub-cloud layer is relatively constant; cloud base is usually found between 600-800m above sea level (Betts and Albrecht, 1987). Its turbulent structure is remarkably similar to that found in the dry CBL: the flux of buoyancy normalized by its surface value decreases from unity at the surface to a value near $-1/5$ at the top of the sub-cloud layer; and fluctuations of the vertical velocity tend to be well scaled by a convective velocity scale given by the depth h of the layer and the magnitude of the surface buoyancy flux following (Deardorff, 1970). Such relationships are evident both in observations and simulations (Nicholls and LeMone, 1980; Stevens et al., 2001; Siebesma et al., 2003). That said there are indications that momentum and moisture fluxes are enhanced under patches or lines of active convection, and of cloud "roots" extending 100 m or more into the top of the sub-cloud layer (LeMone and Pennell, 1976; Nicholls and LeMone, 1980; Ötles and Young, 1996). Such features are indicative of underlying support for developing cumulus elements, and are evident in cross-sections of the turbulent circulation derived from LES, but their implication for modeling and theoretical work remains unclear. As for the patterns of organization of the cloud field, while the role of mean wind in developing boundary layer rolls and cloud streets has been the focus of many studies, relatively little attention has been devoted to the mechanisms for clustering on scales of 50 km or so (Malkus, 1958; LeMone and Pennell, 1976). Although usually idealized as non-precipitating there is ample evidence of precipitation from shallow convection (Short and Nakamura, 2000) raising the question as to whether this often neglected process might play a critical role in the organization, and other elements of shallow cumulus convection (see also, Jensen et al., 2000).

The role of the transition layer in regulating trade-wind convection remains controversial. Typically it varies from 100-300 m in depth and is most evident by a decrease in q_v by more than 1 g/kg, and to a lesser extent a slightly enhanced lapse rate of θ. In a recent study of soundings from the Eastern tropical Pacific, transition layers were evident roughly 45% of the time (Yin and Albrecht, 2000). Although spatial variations on the cloud-separation scale (order kms) tend to wash out its structure in the mean profiles in Fig. 6, it can be associated with region over which the average cloud fraction is increasing. A more marked structure would be evident in individual vertical profiles from the large-eddy simulation. Its ubiquity has led to speculation about its dynamic role (e.g., Yin and Albrecht, 2000; Stevens et al., 2001). However, the incorporation of such a layer into theories of trade-wind cumulus has been spotty, indicative of the lack of consensus in this respect

Past studies of non-precipitating cumulus convection tend to focus predominantly on the structure of clouds (as test parcels or families of test parcels), and to a lesser extent on the structure of the cloud layer. This distinction is somewhat analogous to the differences between CAPE and GCAPE, where the latter views clouds as just one part of a global rearrangement of the fluid. An excellent, and still current, review of the structure of individual clouds is given by (Blyth, 1993). Some consensus characteristics of clouds summarized at that time, include: (i) well defined clouds boundaries; (ii) the tendency for updrafts to be concentrated within the cloudy envelope and down-drafts to predominate around the edges of the cloud, especially near cloud top; (iii) similar cloud and environmental lapse rates; (iv) heterogeneous cloud properties indicative of active mixing between the clouds and their environment. To this could be added the tendency for turbulence to be concentrated in the clouds and their near fields. Blyth's review primarily summarized information gleaned from decades of *in situ* cloud measurements. In the decade since, our understanding has advanced most rapidly as a result of improved remote sensing capabilities and large-eddy simulation. Fig. 8 from Kollias et al. (2001) shows the tendency for updrafts to penetrate into the stably stratified air capping the cloud layer, with maximum vertical velocities greater that 5 m s^{-1} and downdrafts of nearly -3 m s^{-1} crowning cloud top. Regions of active mixing, in this case measured by the spreading of the Doppler spectrum, tend to locate near cloud edges.

Another view of the structure of the cloud layer is provided by the distribution of state variables $\{\theta_l, q_t\}$ at a fixed pressure level within the cloud layer as shown in Fig. 9, (Neggers et al., 2002). For this pressure level the temperature at which $q_t = q_s$ is represented by the dash-dot line, and the line of neutral buoyancy whose slope changes in the saturated, versus unsaturated region of state space (cf., 7), is indicated by the dashed line. This division separates the plane into four quadrants, positively buoyant saturated air on the upper right, positively buoyant unsaturated air on the lower right, negatively buoyant saturated air on the upper left, and negatively buoyant unsaturated air on the lower left. The symbols also separate the air into strong updrafts, strong downdrafts and environmental air. The tendency of updrafts to be relatively rare, saturated, and positively buoyant, while downdrafts are subsaturated and negatively buoyant, is clearly evident (although the symbols separating up and downdrafts are difficult to make out). The strong drafts almost appear to be drawn from a distinct population relative to the more Gaussian spread of the background flow, which we take as being further indicative of the two-fluid nature of the problem as per Section 2.2. The degree to which updrafts are moister and cooler (in the sense of θ_l) than the environment also emphasizes their role in the transport of heat and moisture through the layer. This latter property helps motivate the atomistic view of clouds emphasized by the mass flux methodology discussed below.

The necessity of considering the cloud field as an ensemble of drafts is emphasized in Fig. 6. Here, increased gradients of q_l within the inversion layer, and the very small values of cloud fraction at this height are indicative of the role of relatively rare and strong plumes in penetrating into the stably stratified environment. The decreasing cloud coverage with height suggests that most clouds penetrate through the depth of the cloud layer for a relatively short period of their lifetime, if ever. Attempts to quantify this distribution find that the probability, p, of finding a cloud of scale ℓ (associated with the square root of cloud area projected to a horizontal plane) is a power law of the form $p(\ell) = a\ell^b$ with $-2 < b < -5/3$ up until a scale break at some size ℓ_c. Clouds with size $\ell > \ell_c$ being relatively rare (Plank, 1969; Benner and Curry, 1998; Neggers et al., 2003). This scaling implies order ℓ_c clouds dominate the area coverage, and although not obvious (but related to the tendency of larger clouds to have more vigorous mean updrafts) these clouds are even more dominant from the point of view of energetics. Thus the scale break is a controlling parameter of the cloud size distribution. What determines this has been a matter of speculation, but the depth of the cloud and sub-cloud layer are obvious candidates (Neggers et al., 2003). Simulations have also argued for a scaling of cloud perimeter area versus volume implying a fractal dimension of 7/3 for cloud surface area, consistent with observational analyses of area, perimeter length, scalings from observations (Siebesma and Jonker, 2000). All of which point to the cloud field being rich in scales, an idea in accord with even the most casual observation of the untrained eye.

4.2 Theoretical Perspectives

Theoretical work has tended to focus on models of the cloud elements as a function of the mean state. Given a mean field ϕ (usually taken to be either θ_l or q_t), its evolution due to moist convection can be written as

$$\partial_t \phi|_{\text{clouds}} = -\partial_z F^\phi \tag{14}$$

where F^ϕ is a turbulent flux largely identified with the clouds. From this point of view the job of a cloud model is to specify the flux. Much recent work has adopted the "mass flux" perspective where

$$F^\phi = \frac{M}{\rho}(\phi^c - \phi). \tag{15}$$

where M is the mass flux carried by the clouds, ρ is the density, and $\phi^c - \phi$ measures the difference between cloud averaged values of ϕ (denoted by superscript c) and their environmental values (see Tiedtke, 1989; Siebesma, 1998, for an application of these ideas to non-precipitating shallow convection). In the sense of Fig. 9 this is as if the flux is entirely carried by a single plume whose properties are identical to those obtained by conditioning the average over just cloudy, or perhaps buoyant and cloudy, parcels. Based on such an averaging procedure one can derive equations for M and ϕ^c :

$$\partial_z M = (\epsilon - \delta) M \tag{16}$$

$$\partial_z \phi^c = -\epsilon(\phi^c - \phi). \tag{17}$$

Here two reciprocal length scales, ϵ and δ, determine the change of M with height. Physically they are interpreted as the rate at which the cloud ensemble is diluted through the incorporation of environmental air (cf., Eq. 17) and diminished through the detrainment of cloud mass. Although (16)-(17) make ready reference to physical processes, such as entrainment and detrainment, the specification of ϵ and δ is equivalent to specifying the non-dimensional profile of M and ϕ^c (Bellon and Stevens, 2004).

Closure of the above system requires a specification of $\{\epsilon, \delta\}$, boundary conditions which determine M and ϕ^c at cloud base, and some means for determining the support of F^ϕ, i.e., the depth of the cloud layer. The model (16)-(17) differs from a similarity plume through the inclusion of δ which arises when one associates M with an ensemble of drafts. Recently, however, the similarity plume approach has been re-emphasized, but for a distribution of plumes, each of which satisfies (17), chosen to span the distribution of cloudy elements (Neggers et al., 2002; Cheinet, 2004). In such an approach the number of cloudy plumes and their respective updraft velocities, typically modeled using a simplified momentum equation (e.g., Simpson and Wiggert, 1969; Siebesma et al., 2003) implicitly determines the vertical profile of M. Both the bulk and multi-parcel approaches share a common need to model ϵ. Similar to the entrainment problem for stratocumulus this is a central and long standing problem in the modeling of cumulus clouds, for which many ideas (Raymond and Blyth, 1986; Siebesma and Cuijpers, 1995; Siebesma, 1998; Cheinet, 2004) have been proposed. Although there is an emerging consensus that cumulus drafts entrain more rapidly than similarity arguments would seem to indicate (largely based on an analysis of large-eddy simulation, e.g., Siebesma and Cuijpers, 1995), little consensus has developed around any specific approach. Simulations have also been used to evaluate closure assumptions for M and ϕ^c at cloud base. Typically, ϕ^c is associated with the statistics of the sub-cloud layer, while if the atmosphere is determined to be convecting M is determined in a variety of ways, most typically through a constraint on the sub-cloud layer (Albrecht et al., 1979; Tiedtke, 1989; Grant, 2001). Here however, there does appear to be an emerging consensus that the cloud base value of M scales with w_*, as initially argued by Nicholls and LeMone (1980)

With most research focusing on a description of the cloud layer, relatively little work has addressed the broader question of how clouds help maintain the observed boundary layer structure. Numerical experiments (McCaa and Bretherton, 2004) suggest that cumulus induced deepening of the layer is crucial in mediating the observed transition between regions of more stratiform clouds and the more broken trades. However, most models of cumulus convection neglect any explicit consideration of the interaction of the cumulus layer with the overlying free atmosphere, although this (in the end) will determine the statistics of the cloud layer. Betts and Ridgway (1989) proposed a simple steady-state theory of the trade-wind boundary layer which only incorporates the cumulus dynamics in so far as they determine the mean structure of the cloud layer, energetically such a model is constrained by the heat balance of the sub-cloud layer. A two layer model developed by Albrecht et al. (1979) attempts to more formally incorporate cloud processes into its dynamics, however in addition to being sensitive to parameters whose values are difficult to constrain (Bretherton, 1993) it was recently pointed out that this framework rests on assumptions about the cloud layer structure which are inconsistent with the assumed model of mixing (Bellon and Stevens, 2004). These frameworks are also difficult to close because for partially saturated layers relationships similar to (7) depend more fully on the full joint pdf (Bechtold and Siebesma, 1998). Although some recent work has attempted to unify the representation of cloud and sub-cloud processes (Lappen and Randall, 2001; Golaz et al., 2002a,b; Cheinet, 2003, 2004), these models require a vertically resolved representation of the trade-wind layer. Overall efforts to understand trade-wind cumulus convection remain hampered by the lack of a compelling representation of the bulk structure of the trade-wind layer that is capable of consistently incorporating more sophisticated statements of cloud processes, as for instance represented by the models discussed above (Bellon and Stevens, 2004).

Another approach to studying trade wind clouds has been to study their similarity structure, in analogy with the study of dry convective boundary layers (Grant and Brown, 1999; Brown and Grant, 2000). The evolution of shallow cumulus as air advects downstream over warmer water in the trades has much in common with the growth of the dry convective boundary layer. Recently (Stevens, 2004) proposed a framework for studying trade-wind cumuli wherein the cloud layer grows into a layer with a fixed lapse rate of θ_v in the free troposphere, an exponentially decaying initial moisture profile, and a sea-surface temperature which is continually adjusted to maintain a constant surface buoyancy flux. Even with a paucity of additional parameters (the moisture scale height, decay scale and initial value) this problem is difficult to constrain by similarity approaches. From the point of view of large-eddy simulation one finds that initially the boundary layer depth increases with time, t, following the square-root law $h \sim t^{1/2}$ as would be expected from similarity arguments for the similar problem in dry convection. But with the onset of moist convection we find that $h \sim t$. Because the flux of condensed water into the inversion layer can be expected to depend linearly on cloud depth, such a model is consistent with the longstanding view of trade-cumulus maintaining the depth of the trade-inversion by evaporating cloud water into it. Results from this analysis also show that the proportionality scaling between cloud base mass flux and w_* proposed by (Nicholls and LeMone, 1980) well represents the variability in the data.

5 DEEP PRECIPITATING CUMULUS CONVECTION

Amongst students of convection, atmospheric moist convection is often identified with deep precipitating cumulus convection (cumulonimbus) as sketched over the warm water segment of Fig. 1. Because deep cumulus convection is so strongly associated with precipitation the statistics of the latter often serve as a surrogate for the former. From this perspective we have learned that deep precipitating cumulus convection is a relatively rare phenomena. Early work hypothesized that roughly 2000 hot towers (i.e., non-dilute cumulonimbus clouds, which would span between 0.1 and 0.5% of the available area) are all that are necessary on a daily basis to satisfy the energy balance of the equatorial trough (low-pressure) zone. Satellite studies suggest that actual precipitation zones tend to be more extensive but still highly textured in space and time. Statistically speaking they correlate well with oceanic regions where sea-surface temperatures are warmer than 27-28° C, surface winds are convergent, and where a bulk measure of the atmospheric relative humidity is high (Bretherton et al., 2004). Budget studies show that the diabatic processes associated with cloud processes in these regions tends to warm and dry the troposphere with most of the drying being concentrated below the freezing level (roughly 5 km) and most of the warming being evident at higher levels (Yanai et al., 1973). These patterns of heating and moistening are consistent with two dominant modes of convection, one shallow and non-precipitating, another deep and precipitating. Because the impact of such diabatic processes feeds back strongly on adiabatic circulations (Reed and Recker, 1971; Emanuel et al., 1994; Stevens et al., 1997), it is more difficult to separate deep moist convection from its environment, as is commonly done for non-precipitating convection. Such difficulties are reinforced by noting that although deep convection tends to locate in flows whose adiabatic component is upwelling, these same budget studies suggest that the net mass flux in ascending branches of large-scale circulations is less than the upward mass fluxes within the cumulus clouds themselves. Given that the area-fraction of deep convection is relatively small, this suggests that most of the air in the ascent regions of the tropics is actually descending.

The "hot-tower" perspective of many early studies emphasized what can be thought of as a precipitating version of Fig. 6, wherein the trade-inversion is displaced to the tropopause and isolated "hot-towers" span the depth of the troposphere (12-18km) producing copious amounts of surface precipitation and perhaps a cirroform cloud shield (Riehl et al., 1951). Despite the grip this view maintains on the community's collective imagination, on the basis of early satellite measurements, airborne photogrammetric studies, and *in situ* measurements it has long been apparent that much of the deep convection over oceanic regions is intricately structured with dynamically coherent lines or clusters on scales ranging from hundreds to thousands of kilometers (Zipser, 1969; Nakazawa, 1988). A hierarchy of structures across these scales, and the interplay between the adiabatic and diabatic components of the flow which it hints at, further frustrates attempts to separate deep convection from its larger-scale environment.

The importance and dynamical coherence of meso-scale convective systems (called MCSs) has led researchers to develop a two component classification system for tropical precipitation: convective (hot towers); and stratiform (MCS) (Schumacher and Houze, 2003). This terminology helps to emphasize the vigor of the convection in the convecting regions, but can be misleading because the stratiform regions are often convective, just not cumuliform. Recent studies based on suites of satellite sensors designed to distinguish among the micro- and macrophysical signatures of these different forms of convection suggest that MCSs and hot-towers are each responsible for commensurate amounts of tropical precipitation, roughly 40% each, with the remainder associated with "shallow" systems. The two classes of precipitating convection have distinct spatial statistics: convective regions tend to more frequent, cover less area, and precipitate much more vigorously (by a factor of four or greater) than their stratiform counterparts, all of which is consistent with such regions having a greater convective intensity (Schumacher and Houze, 2003; Nesbitt et al., 2000). Stratiform regions tend to have a more pronounced diurnal cycle (over the ocean), with area-fractions and mean sizes increasing at night (Nesbitt and Zipser, 2003), consistent with them being more influenced by radiative processes (cf., Section 3). Although larger-scale organization appears to be endemic to moist atmospheric convection, it is perhaps the most difficult to ignore for the case of deep convection, even more so for oceanic deep convection where even the prevalence of tropospheric spanning "hot-towers" has recently come into question (Zipser, 2003). For this reason below we attempt to briefly summarize important physical characteristics associated with both the hot-tower and MCS view of tropical convection. Further background is provided in the texts by Cotton and Anthes (1989), Houze (1993), and (Emanuel, 1994). For the case of MCSs we refer the reader to reviews by Redelsperger (1997) and Houze (2004).

5.1 Phenomenological Overview

To the extent that the hot-tower picture of deep moist convection resembles that of trade-wind convection, many of the concepts from the previous section are relevant to deep convection as well. For instance, recent work has begun to focus on the interactions of deep convection with the upper troposphere and lower stratosphere leading to the development of a tropical tropopause layer (indicated roughly by the dashed lines above the regions of deep convection in Fig. 1) which in some sense is thought to be analogous to the entrainment/inversion layer capping trade-wind clouds (Sherwood and Dessler, 2001). Although many analogies exist, the depth of the convecting layer introduces new phenomena including: precipitation, which transports enthalpy across fluid stream lines; ice-microphysics, whose non-equilibrium behavior both influences precipitation formation and interacts strongly with radiant streams of energy; and more disparate timescales for moisture and temperature profiles to adjust to the convection.

Like the shallow convecting layers, layers of deep convective clouds exhibit a spectrum of cloud types. Aircraft and radar studies based on data collected during GATE suggest that the spectrum of clouds is lognormally distributed, whether they be measured in terms of radar reflectivities, updraft area, intensity or mass flux (e.g., López, 1977; LeMone and Zipser, 1980). Notwithstanding the evidence for a broad spectrum of cloud types much theoretical work is based on a categorization of discrete modes. As a result of the early budget studies the emphasis has been on two modes: hot towers and shallow convection, although recent work has sought to emphasize the importance of a mid-level congestus mode thought to account for roughly 20% of the observed precipitation (Johnson et al., 1999; Tung et al., 1999); a point of view rationalized by the presence of three distinguished layers in the troposphere: the trade wind inversion, a freezing-level stable layer, and the tropopause.

There is also evidence that the structure of maritime cumulus convection changes as the depth of the convecting layer becomes deeper. In particular, maximum updraft velocities stop increasing as the clouds extend above 3km, and the ratio of observed liquid water to adiabatic liquid water becomes more approximately constant. During GATE maximum updraft velocities tend to be less than 5 m s^{-1} and hardly ever exceed 10 m s^{-1}. (LeMone and Zipser, 1980). On average the mean diameter of convective elements (roughly 1km) does not appear to vary greatly with height, and like trade cumulus they tend to cover less than 5% of the available areas. Likewise, θ_v perturbations tend to be around 0.5K, irrespective of height, indicative of active mixing (Zipser and LeMone, 1980). As is the case for shallow convecting layers, downdrafts in deep convecting layers tend to be weaker and smaller than the updrafts (cf., Fig. 8). However, precipitation has been found to maintain deep saturated downdrafts which are an important component to the balance of heat, moisture and mass within the convecting layer (Johnson, 1976). Moreover, because shafts of precipitation falling through the updraft can make the cloud collapse, the trajectory of the precipitation and the tilt of the cloud have a strong influence on the convective lifecycle. Because both are determined in part by the profile of the environmental wind, the development of precipitation provides additional pathways for the mean wind to organize convection, this being a particularly rich and active area of research which has little in common with studies of the organization of dry or non-precipitating convection through the interplay between buoyancy and wind shear. These effects of precipitation, and the radiative effects of ice-processes further complicate attempts to measure the stability of the atmosphere based on the displacements of test parcels, e.g., Section 2.2.

One question, which although relevant to shallow convection is more pertinent to hot-towers, is how the unsaturated environment is influenced by the enthalpy of vaporization made available through condensation in the cumulus drafts. Here the usual picture is one of cumulus clouds forcing isopycnal surfaces to adjust to some effective saturated adiabat in the saturated component of the flow. This can be thought of in terms of a redistribution of mass among isopycnal surfaces as shown in Fig. 10. This leads to a tilt in the isopycnal surfaces that induces gravity waves, which in the absence of rotation act to flatten the isopycnals, thereby realigning them with the isobars (Bretherton and Smolarkiewicz, 1989). From this point of view the buoyancy (largely temperature/entropy in the cloud free environment) adjusts on the time-scale of the gravity waves in the cloud-free environment, while the moisture field adjusts on a much slower timescale determined by the rate of horizontal stirring along isentropes and the rate of compensating subsidence. Because the gravity wave speed scales with the depth of the layer, this disparity between timescales is most evident for deep convection.

Deep precipitating cumuliform convection also differs from shallow cumuliform convection in terms of the range and significance of the larger-scale structures into which it so often aggregates. Generically such features are called Mesoscale convective systems which Houze (2004) defines as any *cumulonimbus cloud system that produces a contiguous precipitation area greater than 100km in any direction.* This definition is sufficiently broad to include hurricanes, as well as a wide variety of other agglomerations of convective clouds, and recognizes what (Mapes, 1993) calls the gregarious nature of cumulus convection. Such systems typically share structural, rather than simply morphological attributes. For instance, they are commonly associated with large regions of stratiform precipitation and organized mesoscale circulations, both of which interact with the cumulus convection itself to warrant the nomenclature of *system.* That is the resultant circulation differs from a non-interacting envelope of cumulus clouds in many respects ranging from its longevity, characteristics of propagation, and mechanisms for producing precipitation. A prototypical MCS is the squall-line, which is illustrated by the schematic in Fig. 11. This figure illustrates how convective cells fuel a much larger circulation, which in turn helps organize and maintain the convection itself. Here a circulation develops in which deep, vigorous convective cells are arranged along a line going into the page. As they decay they are swept back into the trailing stratiform graveyard, where through precipitation they help moisten and cool a low-level inflow of low θ_e air, which in turn helps form a pool of cold air near the surface whose leading-edge gust front helps initiate new convection. This couplet of upper-level front-to-rear flow and lower-level inflow is consistent with the response to deep convective heating shown in Fig. 10, (Pandya and Durran, 1996). Of the various forms of mesoscale convective systems, the squall line is perhaps the best understood, with the idea being that the ambient wind shear is critical to the development of such structure (Rotunno et al., 1988). Indeed, almost all anisotropic forms of convective organization are thought to develop their anisotropy from the mean wind, although the details by which different structures are selected remains sketchy. Even more uncertain is the extent to which systems self-aggregate independent of any organizing influence provided by the mean flow in which they are embedded.

5.2 Theoretical Perspectives

Many of the theoretical questions pertaining to precipitating cumuliform convection have counterparts in our discussion of the energetics of test parcels in Section 2.2 and our review of mass flux theories in Section 4. Arakawa (2004) casts this emphasis in terms of what he calls principal and supplementary closures. The former being some constraint on the intensity of convection (as measured by $M_{\rm cb}$ the cloud base mass flux), the latter being a constraint in form, typically in terms of a cloud model which determines the vertical structure of the convection, i.e., $M(z)/M_{\rm cb}$. Unlike for shallow convection, where $M_{\rm cb}$ is often determined by a budget constraint on the sub-cloud layer, or the velocity scale valid for the dry convective sub-cloud layer, for the case of precipitating cumuliform convection the community has largely congealed around *quasi-equilibrium* ideas of Arakawa (1969); Arakawa and Schubert (1974), which constrain M_{cb} based on the temporally evolving energetics of test parcels. In an early implementation of this idea (Arakawa and Schubert, 1974) argue that convection acts to just offset the destabilization of the atmosphere by non-convective processes, where in this case the stability is measured with respect to the energetics of an ensemble of entraining plumes—in essence their test parcels. This idea, and the cloud model underlying it, has been elaborated upon, simplified, and relaxed (e.g., Betts and Miller, 1986; Emanuel, 1991; Moorthi and Suarez, 1992; Zhang and McFarlane, 1995; Raymond, 1997; Pan and Randall, 1998; Raymond et al., 2003), but in one form or another it serves as the conceptual foundation for almost all closures, largely displacing earlier ideas based solely on moisture constraints (see a discussion of these issues by Raymond and Emanuel, 1993). Nonetheless, the question of exactly how best to frame it, and how much the details matter is a question of active research (Emanuel et al., 1994; Stevens et al., 1997; Arakawa, 2004).

A question of contemporary interest is the way in which the humidity of the free atmosphere may regulate convection. The basic idea is that convection into an especially dry atmosphere tends to be more readily diminished by entrainment of very dry air (Mapes and Zuidema, 1996; Parsons et al., 2000). From this perspective the spectrum of cloud depths in a convecting atmosphere, and hence precipitation, may be regulated not only by the thermal structure of the free-troposphere, but also its humidity structure, where the latter is modulated by intrusions of dry air into the convecting zones. Such effects are not evident in many theories of moist convection, and recent work has focused on better incorporating these and other effects such as the regulating role of convective inhibition (Derbyshire et al., 2004; Mapes, 2000). To the extent the equilibrium ideas of Arakawa and colleagues represent mean field theories, another exciting area of recent work has been on the incorporation of stochastic effects (Buizza et al., 1999; Lin and Neelin, 2000; Majda and Khouider, 2002).

From the perspective of simulation, many interesting questions can and have been posed from the point of view of radiative convective equilibrium (e.g., Held et al., 1993; Robe and Emanuel, 1996; Thompkins and Craig, 1998). The geometry of this problem is one of an atmosphere which is being destabilized through a fixed depth H over a surface of fixed temperature. Convection then acts to equilibrate the atmosphere over this depth. The presence of many hidden scales, from moisture scale heights, precipitation and ice formation timescales, etc., and the spatial and temporal range they and their more obvious counterparts embody makes this problem difficult to simulate. Moreover the simplest reasonable configuration of this problem is still a matter of some debate. For instance, to what extent should a realistic representation of radiative processes be included? What makes moist convection fascinating is the ways in which it can be diabatic. In this vein it is not clear that the diabatic tendency of precipitation formation is any more essential than the cloud's diabatic interaction with radiant streams of energy which drive the system in the first place (cf., Held et al., 1993). Although these types of calculations still challenges our computational capacity, they are beginning to be employed to ask fundamental questions. For instance this framework has been used by Robe and Emanuel (2001) to investigate the interplay between the mean shear and convective organization; by (Thompkins, 2001), to investigate the effect of water vapor feedbacks on convection; by (Grabowski, 2003) to investigate the interaction of convection with larger-scale circulations; and by Emanuel and Bister (1996); Pauluis and Held (2002a,b) to test theories of convective scaling. From the point of view of (Pauluis and Held, 2002a) deep convection can fruitfully be thought of as a reversible dehumidifier which in equilibrium is balanced by an irreversible re-humidification of dried air away from the convecting regions. The dehumidification/moistening cycle can be linked to the latent heat transport and limits the amount of energy available for the production of kinetic energy. This effect is sufficiently strong so that the dissipation in the shear zones in the wake of falling hydrometeors is larger than the dissipation of kinetic energy generated by convective accelerations (Pauluis et al., 2000), thus pointing to yet another way in which moist convection expresses its irreversibility.

6 SYNTHESIS

Throughout this review we have attempted to emphasize how moist convection is many, rather than one thing. An emphasis motivated by the question: "*What is moist convection?*" which the author interpreted the request for this review to embody. Such questions are often asked from the perspective of students of dry convection, for whom the corresponding question is somewhat simpler to answer. One of the facets of moist convection which make it such a vast, challenging and difficult subject is the range of regimes it embodies, and the variety of physical processes particular regimes support. We should not, however, loose sight of the fact that beyond curiosity, much interest in moist convection is motivated by the need to accurately represent its effects in large-scale models for use in numerical weather prediction or in climate studies. This raises the question as to whether moist convection must be modeled in many ways, or if some simple set of ideas can provide a unified basis capable of representing its collective effects across a variety of regimes. For the better part of the last half century work has adopted the former approach, with large-scale models often employing distinct, and increasingly complex, models to represent stratocumulus, versus trade-cumulus, versus deep cumulus convection. Some evidence of this is present in this review, where mass flux concepts dominate the discussion of cumuliform convection, and mixed layer theory (consistent with local mixing theories typified by eddy diffusion) dominates the discussion of stratiform convection. Bucking this trend is a renewed interest in simple models whose physical principals allow them to recognize a variety of forcings and naturally switch among regimes, from dry to stratocumulus convection, from shallow to deep convection, from stratocumulus to trade-cumulus. For instance, Lappen and Randall (2001) and Golaz et al.

(2002a) can in some sense be viewed as an attempt to blend mass-flux approaches with more standard turbulence closure theories used for dry convection. Cheinet (2003) explores the unity of mass flux concepts for dry convection, while Cheinet and Teixeira (2003) show that eddy-diffusivity approaches can express behavior usually reserved for mass flux models.

Motivated by a similar desire to represent moist convection with a single set of equations, increased computer power has made it possible to consider embedding two or three-dimensional cloud scale models, whose domains span some subset of the space of a large-scale grid volume in a three dimensional model, as a basis for representing the collective effects of moist convection across regimes. Such an approach has been referred to as cloud resolving convective parameterization (Grabowski and Smolarkiewicz, 1999) and super-parameterization (Randall et al., 2003; Arakawa, 2004). Although conceptually simple, such calculations are computationally intensive. Nonetheless, as they become increasingly common, and begin to capture a wider range of scales within convecting atmospheres, they should prove useful in both bounding our expectations, and (along with the ever increasing power of satellite remote sensing) enriching the phenomenology from which our insights and intuition are drawn, thus brightening the prospects for theoretical advancement.

7 SUMMARY

Unlike dry convection, moist convection is not one, but many things. To illustrate this we reviewed some essential aspects of moist convection, both in terms of its expression within the context of moist atmospheric thermodynamics, and through a consideration of three paradigms: stratocumulus, shallow non-precipitating, and deep precipitating convection. These problems highlight the many ways in which moist convection distinguishes itself, most fundamentally through its non-local interactions with other parts of the flow either through radiative, gravity wave, or microphysical (precipitation) processes. For each regime modern remote sensing is providing wondrous new insights, whether it be with respect to the statistics of precipitating convection and its interaction with the larger-scale environment as seen by satellite, or the structure of shallow cloud circulations, and precipitation, as seen by surface based, or airborne cloud radars. Simulation, and appropriately simply simulation paradigms are also being worked out for the various regimes. Together with long-standing techniques for probing the structure of moist convecting atmospheres these approaches bode well for new leaps in understanding of one of natures more majestic phenomena.

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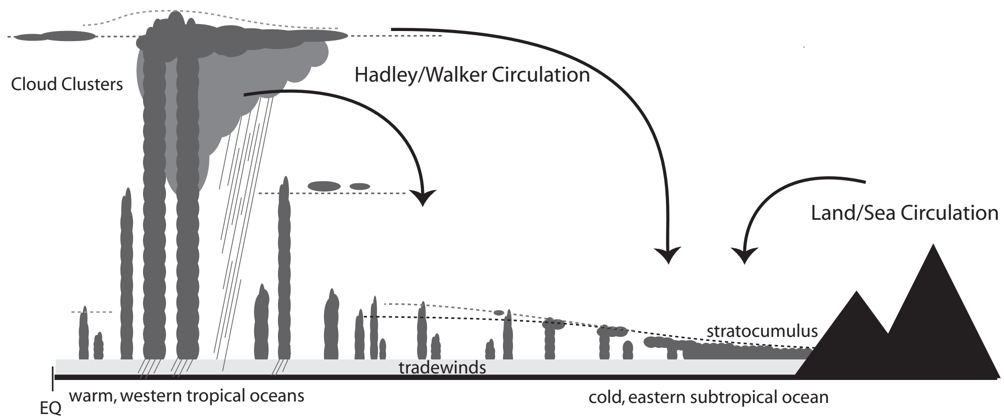

Figure 1: Cloud-Regimes in thermally direct circulations, adapted from (Arakawa, 1975).

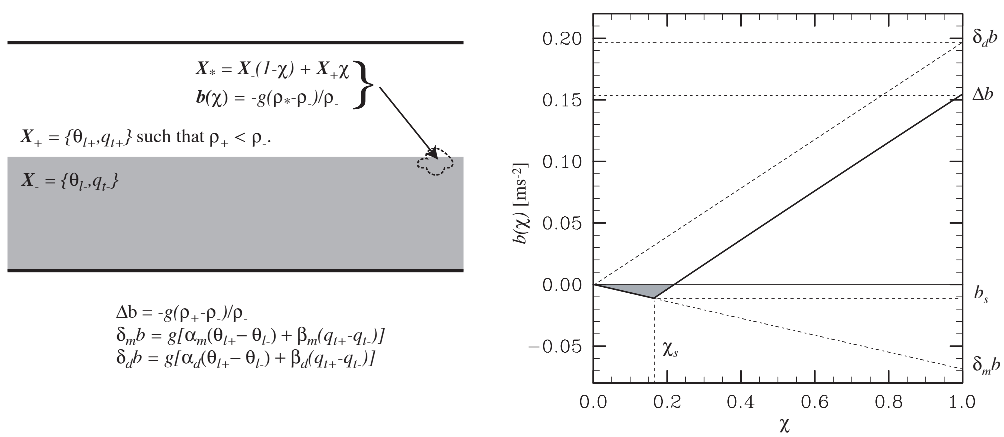

Figure 2: The buoyancy of mixed parcel as a function of the mixing fraction χ at a statically stable interface between dry and saturated fluids.

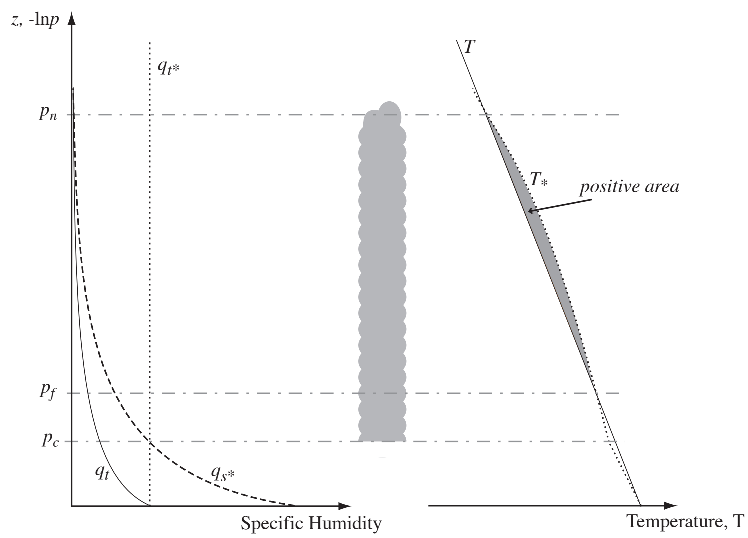

Figure 3: Finite amplitude displacements in moist atmospheres stable to infinitesimal perturbations.

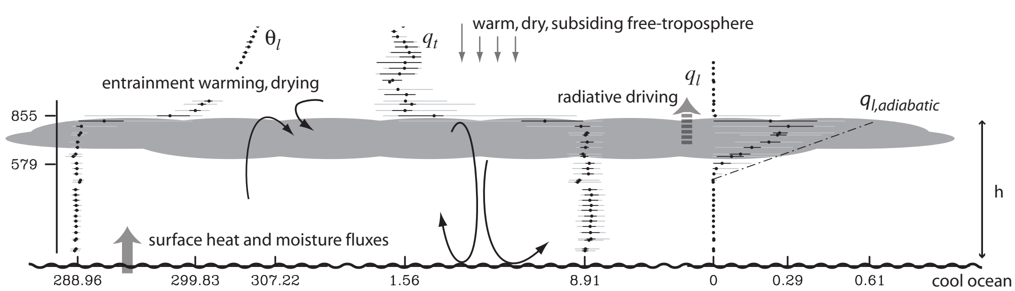

Figure 4: Cartoon of well mixed, non precipitating, stratocumulus layer, overlaid with data from research flight 1 of DYCOMS-II. Plotted are the full range, middle quartile and mean of θ_l, q_t and q_l from all the data over target region binned in 30m intervals. Heights of cloud base and top are indicated as is mixed layer values, and values just above the top of the boundary layer, of various thermodynamic quantities. The adiabatic liquid water content is indicated by the dash-dot line.

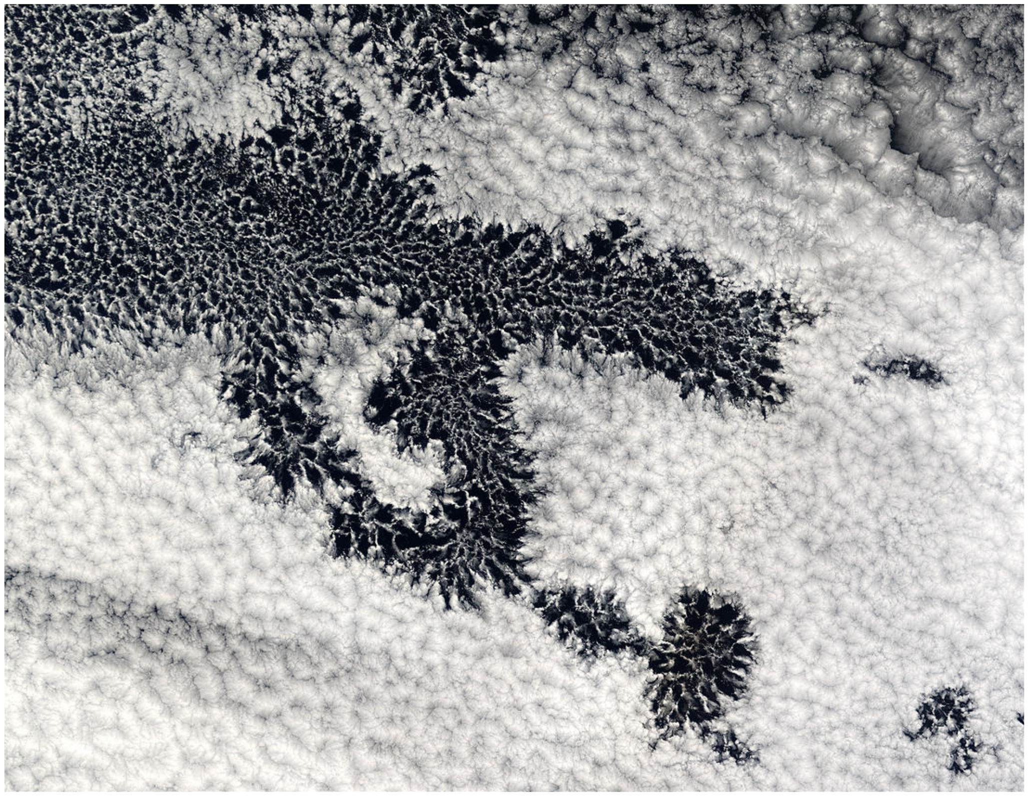

Figure 5: Example of a region of open cellular convection (dark cell interiors, with bright cell walls) embedded in a broader region of closed cellular convection (bright cells with darkened cell walls). Open cellular regions have been hypothesized to be envelopes where drizzle is more prevalent

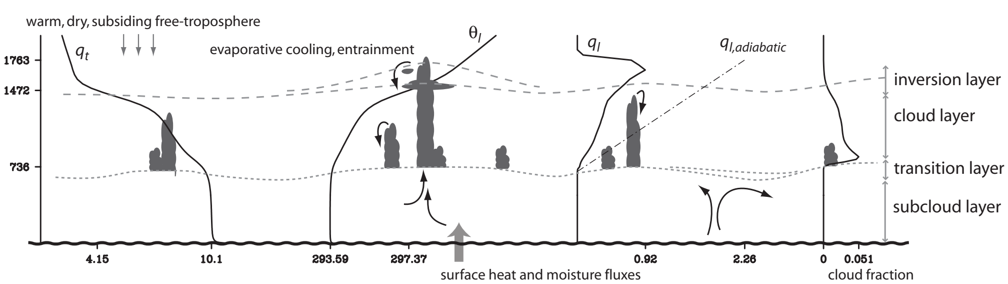

Figure 6: Cartoon of trade-wind boundary layer from Large-Eddy Simulation. Heights of cloud base, level of maximum θ_l gradient (inversion height), and maximum cloud penetration depth are indicated, as are sub-cloud layer, and inversion level values of thermodynamic quantities. Cloud water contents are averaged over cloudy points only, with adiabatic liquid water contents indicated by dash-dot line. Far right panel shows cloud fraction, which maximizes near cloud base at just over 5%.

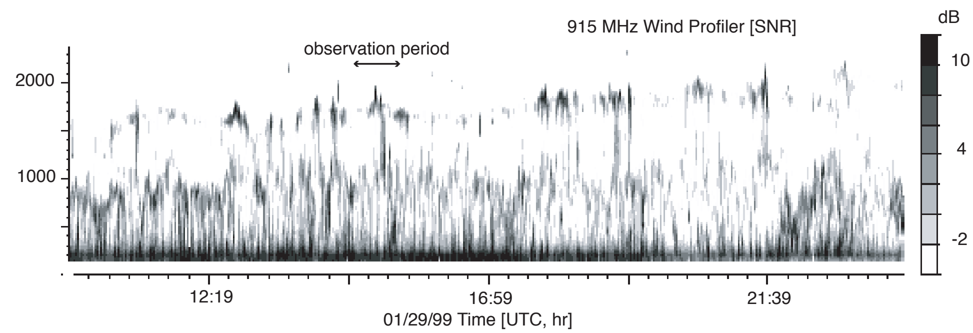

Figure 7: Signal-to-Noise Ratio from a 915 Mhz wind profiler. At this wavelength Bragg scattering from humidity gradients dominate the signal, thus illustrating the turbulent structure of the trade-wind boundary layer by highlighting regions of vigorous mixing associated with sub-cloud thermals and cloud boundaries. Both the trade-inversion and transition layer are evident at this frequency.

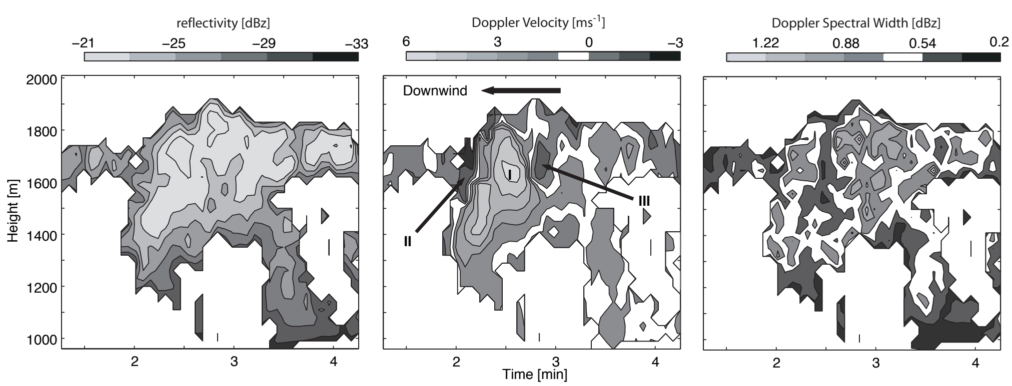

Figure 8: Radar reflectivities (left), Doppler velocities (center) and Doppler spectral width (right) for a cumulus cloud sampled during the observational period marked in Fig. 7. In this figure the updrafts and downdrafts are marked on the central panel. Notice the rather broad spectral widths in the updraft core in the rightmost panel.

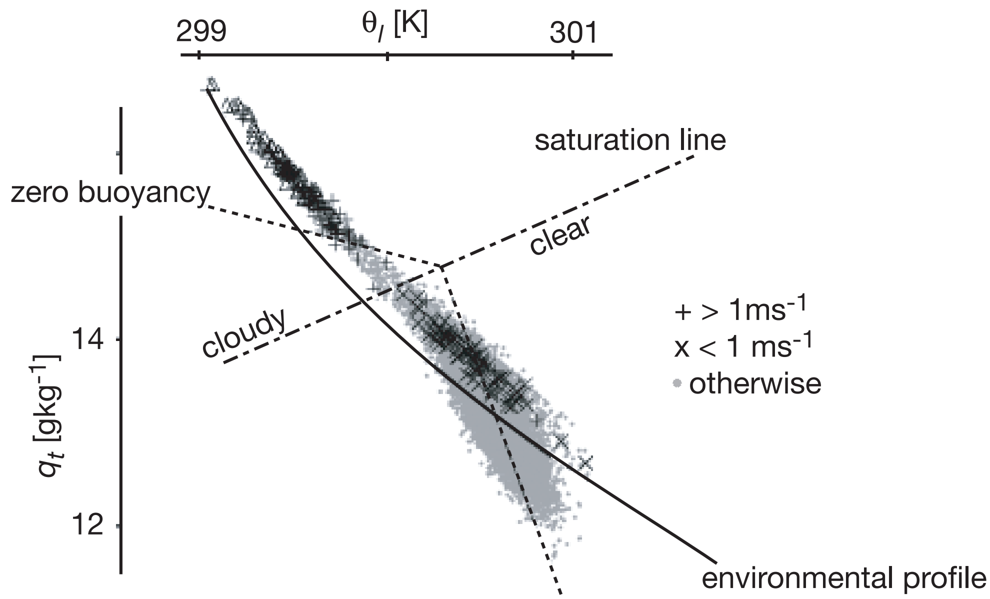

Figure 9: Distribution of state variables on a $q_t - \theta_l$ diagram for a fixed level within the cloud layer as derived from LES and first published by (Neggers et al., 2002). The environmental profile shows the change in the mean state with height, the mean value at the given height is indicated by the intersection of the environmental and zero-buoyancy line. Points above or to the right of the zero buoyancy line are positively buoyant, points below or to the left are negatively buoyant. The kink at the intersection with the saturation line reflects the change in buoyancy for the saturated versus the unsaturated fluid. Note that strong updrafts are rare, cloudy and buoyant.

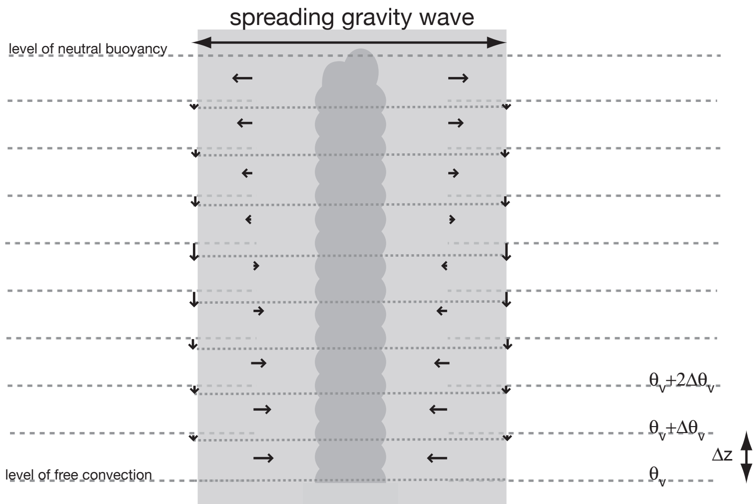

Figure 10: Schematic of adjustment due to cumulus "heating" following (Bretherton and Smolarkiewicz, 1989). Here the transient adjustment to the moist-adiabatic lapse rate of the convecting tower from Fig. 3 is shown to be accomplished be a spreading gravity wave. As the subsidence wave moves away from the region of convection it is associated with a downward displacement of isentropes which equilibrates the buoyancy (θ_v) profile.

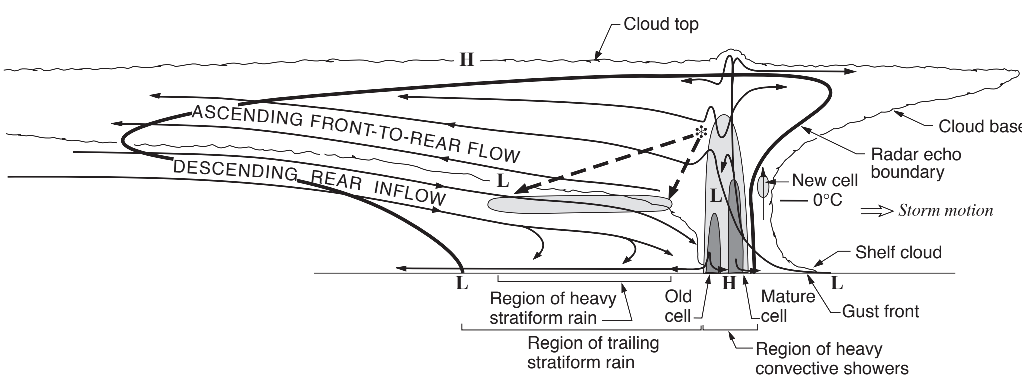

Figure 11: Example of a prototypical mesoscale convective system, a squall line. Taken from (Houze, 1989)