1 of 2

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INTERACTIONS BETWEEN A TROPICAL MIXED BOUNDARY LAYER
AND CUMULUS CONVECTION IN A RADIATIVE-CONVECTIVE MODEL

A Thesis in

Meteorology

by

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ABSTRACT

A radiative-convective model, combining previously developed cumulus, stable cloud and radiation parameterizations with a boundary layer scheme, was developed in the current study. The cloud model (Frank and Cohen, 1985) was modified to incorporate the effects of both small and large clouds. The boundary layer model was adapted from a mixed layer model (Albrecht et al., 1979), was only slightly modified to couple it with the more sophisticated cloud model.

The model was tested for a variety of imposed divergence profiles, which simulate the regions of the tropical ocean from approximately the Intertropical Convergence Zone (ITCZ) to the subtropical high region. The sounding used to initialize the model for most of the runs is from the trade wind region of ATEX. For each experiment, the model was run with a timestep of 300 seconds for a period of 7 days.

The model produced an inversion height of approximately 950 mb for each region. As the subsidence increased, typical of conditions towards the subtropical high region, the boundary layer became moister and warmer, and the periods of deep convection became weaker and less frequent.

Various cloud spectra were also introduced into the model to test for sensitivity to the number of cloud types

represented. The model produced similar results whether one cloud or many clouds were used. However, the runs with more than one cloud generally displayed resultant inversion heights with more stable oscillations than the equivalent runs with only one cloud type. With the exception of higher mixing rates resulting in smaller clouds, and lower mixing rates resulting in larger clouds, the specification of initial updraft vertical velocity and mixing rate made little difference to the final inversion height and boundary layer thermodynamic values.

Finally, sensitivity tests were done for varying sea surface temperatures and surface wind velocities. The model responded well to the changes. As sea surface temperature increased, the inversion height increased, the boundary layer became warmer and the moisture stayed approximately the same. As surface wind velocity increased, the inversion height decreased, and the boundary layer temperature and moisture decreased and increased respectively.

A FGGE sounding from a region closer to the equator was tested and the results were very similar, with an inversion height near 960 mb. Therefore, the model is thought to be physically representative of the physical area described above.

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LIST OF SYMBOLS AND CONSTANTS

α :	$(F_{SV})_0 / (F_{SV})_{B-} = -.1$
c:	condensation
CAPE:	convective available potential energy
c_p :	specific heat at constant pressure = 1004 [J/(deg*kg)]
C_T :	dimensionless transfer coefficient = 1.15E-3
δ :	dimensionless constant used in computing $(F_{SV}) = .608$
dq/dP :	lapse rate of q immediately above the inversion
dS/dP :	lapse rate of S immediately above the inversion
D:	divergence [1/s]
Δq :	inversion jump of mixing ratio
Δs :	inversion jump of dry static energy
Δs_v :	inversion jump of moist static energy
ΔT_v :	inversion jump of virtual temperature
e:	evaporation
$(F_q)_0$:	flux of q at the surface
$(F_q)_{B-}$:	flux of q at the top of the boundary layer
$(F_s)_0$:	flux of S at the surface
$(F_s)_{B-}$:	flux of S at the top of the boundary layer
$(F_{sv})_0$:	flux of S_v at the surface
$(F_{sv})_{B-}$:	flux of S_v at the top of the boundary layer
K:	diffusion constant = .5
L:	latent heat of condensation = 2.5E-6 [J/(deg*Kg)]
LCL:	lifting condensation level
P_B :	pressure difference between the surface and the inversion
q:	mixing ratio [g/g]

q_0 : mixing ratio of sea surface
 q_M : boundary layer value of mixing ratio
 Q_R : diabatic heating of radiation
 ρ : density [kg/m^3]
 s : dry static energy [J/Kg]
 s_0 : sea surface S
 s_M : boundary layer value of S
 s_v : moist static energy [J/Kg]
 T_R : reference temperature used in computing (F_{sv}) = 288K
 v_0 : surface wind speed [m/s]
 ω : vertical velocity [Pa/s]
 ω^* : convective mass flux

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"Whatever Lola wants, Lola gets."

Chapter 1

INTRODUCTION

Modeling has been a part of the study of climate for almost one hundred years. In 1896, Arrhenius published an article hypothesizing a two degrees Celsius warming induced by a doubling of CO_2 , using a simple energy budget model. Modeling grew slowly in popularity until the development of the computer. At this point, numerical models rapidly became more popular. The very first global climate simulation experiment using all the equations of motion was conducted and analyzed by Smagorinsky et. al. (1965). The 1960's also saw the development of one-dimensional radiative-convective models. These models were used to study possible variations in the climate. Even now, although the ability to model the atmosphere in two and three dimensions is very good, the one-dimensional model is still popular. This popularity results from its simple, yet relatively accurate calculations. These calculations can be performed numerous times for a wide variety of initial and/or boundary conditions. The two and three-dimensional models lack some of this flexibility and speed. However, there are also a few disadvantages to using one-dimensional models. Obviously, one of the disadvantages of one-dimensional models is that they are location-specific. That is, the model is only good in one grid box. Essentially,

although one-dimensional models are used with imposed lateral boundary conditions, using a one-dimensional global climate model involves assuming that conditions around the globe are homogeneous, which is obviously not true. These one-dimensional climate models are used primarily to study equilibrium temperature and moisture profiles. This equilibrium structure depends on a balance between parameterized radiative and convective processes only. The parameterizations used for radiation and convection are generally developed in one dimension and refined and tested using a one-dimensional radiative-convective model before they are adapted to more sophisticated three-dimensional, dynamic models.

Historically, one-dimensional radiative-convective models used in climate studies generally used elementary convection schemes and relatively detailed radiation schemes and therefore were used primarily to study greenhouse warming and climate change (Kiehl, 1991). This, however, is no longer the only use for such models, and as the one-dimensional models are used to study more complex, localized situations, more sophisticated convective parameterizations are being introduced.

Cumulus parameterization schemes generally fall into one of two groups. One of these groups includes convective adjustment schemes while the other includes cloud model schemes (Anthes, 1977). The convective adjustment schemes

are the simplest in that, when the environment becomes convectively unstable, they merely adjust the lapse rate, thereby conserving the column integrals of dry static energy (s) and mixing ratio (q) (Kiehl, 1991). As is obvious by the names, a dry adiabatic adjustment always adjusts to a dry adiabatic lapse rate, and a moist adiabatic adjustment adjusts to a moist adiabatic lapse rate. Manabe and Strickler (1964) used an adjustment scheme such as this in their general circulation model. There are some obvious problems with such a scheme, such as physically unrealistic instantaneousness of its grid box heating and the complete decoupling of convection from boundary layer processes and horizontal transport of moisture. Efforts have been made, though, to at least account for some of these problems (e.g. Betts and Miller, 1986).

The other group of parameterization schemes, the cloud model schemes, are more complex, but not necessarily more accurate. Their accuracy depends very much on the accuracy of the included physical processes. This dependence can pose problems because processes such as entrainment and detrainment in clouds, among others, are not yet fully understood or documented. This type of cumulus parameterization scheme can be seen in early work by Arakawa and Schubert (1974), Ooyama (1971), Kuo (1974) and others. Using the cloud model results, calculated estimates of heating/cooling and moistening/drying are applied to the

larger scale equations in the host radiative-convective model to satisfy a closure hypothesis. The closure is usually an assumed equilibrium state between convection and the large-scale circulation. Traditionally, cloud model schemes have focused only on deep, precipitating clouds. Cho (1975) points out that deep cloud and shallow cloud activities appeared to be controlled by separate mechanisms, making the development of a scheme to account for both slightly more challenging. The deep cloud parameterization has been improved upon by Anthes (1977), Frank and Cohen (1985, 1987), as will be discussed later. In this thesis, shallow and deep convection will be parameterized using a single scheme.

Whether a cloud model scheme or a convective adjustment scheme is used, an inversion-capped boundary layer structure is generally assumed. This structure can vary from very simplistic to highly detailed. The boundary layer is the layer of the atmosphere, generally contained in the lowest two kilometers, that interacts with the surface. The boundary layer's structure is dominated by surface-based, turbulent processes. Kiehl (1991) describes it as "the physical link between the surface and the free atmosphere". Some planetary boundary layer schemes include such detail as accounting for various surface types, as well as a variety of thermal structures. However, the more general bulk approach is required when the planetary boundary layer

cannot be adequately resolved with the model vertical resolution.

A bulk model of the boundary layer commonly assumes constant dry static energy (s) and mixing ratio (q), and is referred to as a well-mixed boundary layer. This approach assumes that the turbulent fluxes are sufficient to "mix" the air causing the air beneath the capping inversion to have relatively constant values of s and q . Some of the earliest mixed layer models are by Ball (1960), Lilly (1968), Betts (1973), Carson (1973), and Tennekes (1973). Previously mentioned cumulus parameterization studies that used such a mixed layer model of the boundary layer include those of Arakawa and Schubert (1974), Sarachik (1974), and Ogura and Cho (1974).

The interaction between the cloud layer and the boundary layer has been studied extensively (Sarachik, 1974; Ogura and Cho, 1974; Esbensen, 1975; Betts, 1976a). Betts (1976a) coupled a mixed layer model with the cumulus parameterization scheme described in Betts (1975) to help illustrate cloud layer subcloud boundary layer interaction. This closure argument is based on relating the cloud base mass flux and mass flux gradients to subcloud-layer parameters, such as subcloud-layer depth and surface flux parameterizations (Betts, 1976a).

The closure assumption assumed in the current study is the one developed by Albrecht et. al (1979), which assumes

that cloud base and the height of the capping inversion (top of the boundary layer) coincide. Seguin and Garstang (1976) studied the strong coupling between the cloud layer and the subcloud layer. They hypothesized that deep convection stabilizes and dilutes the boundary layer, thereby lessening the effect of the surface turbulent fluxes and those at the top of the boundary layer on the boundary layer structure. These turbulent fluxes are not, however, overpowered when the convection is small and non-precipitating. Instead, there seems to be a strong relationship between the parameterized fluxes, the inversion strength, and the cloud structure. Betts (1976b) also found that strong convection associated with heavy downpours transforms a well-mixed boundary layer into a boundary layer with a more stable structure. In other words, the convection acts to stabilize the layer (Garstang and Betts, 1974).

Generally, convective downdrafts act to cool the boundary layer. They may either dry or moisten the boundary layer depending on the existing conditions. On the other hand, in the wake of convection, warm, dry air from aloft is often entrained into the mixed layer. These two processes of convective downdrafts and entrainment tend to oppose each other. However, Betts found that the layer beneath the clouds usually cools and dries, an indication that both processes are important.

Betts (1976b) modeled separately the effects of

small, non-precipitating clouds and large, precipitating clouds with downdrafts. Betts also showed that this coupling of a mixed layer with a shallow cumulus layer based on a cumulus mass flux parameterization provided satisfactory results. In 1977, Ogura et. al. developed a similar parameterization, but it proved to be too sensitive to various externally specified parameters, such as surface fluxes and large-scale divergence.

Convective clouds tend to exist in clusters, as they are generally organized by large-scale disturbances (Chang, 1970). A cloud spectrum is necessary if the various characteristics of an ensemble of clouds are to be accurately accounted for.

Research has been conducted to study the properties of cloud ensembles and the using of spectra of clouds in cumulus parameterizations. LeMone and Zipser (1980) studied the relationship between vertical velocity, mass flux, cloud diameter, and intensity. Nitta (1975) conducted an observational study of the distribution of cloud mass flux. Yanai et. al. (1973) observationally demonstrated the existence of both shallow and deep clouds in the intertropical convergence zone (ITCZ). They also showed that the heating due to between-cloud subsidence is very important to the energy budget there. The Arakawa parameterization theory referenced in Nitta (1975) allows determination of the bulk properties of clouds as well as

the cloud mass flux and other properties of each cloud type. Frank and Cohen (1985) found that the heating and moistening profiles from cloud model ensembles differ from the profiles produced from single clouds.

In the current study, cloud ensembles are assumed, with the distinguishing feature of the different cloud types within the ensemble usually being the varying entrainment rates. The higher entrainment rates tend to inhibit cloud growth. The cloud model simply models each cloud's effects independently and then a weighted-average of the heating and moistening profiles, based on number of clouds and percentage of area covered, is computed and applied to the model. It has been observed that clouds tend to be distributed bimodally, with maxima for clouds with tops around 200-300 mb and again for clouds with tops below 600 mb (Cho, 1975). Ideally, not only would a cumulus parameterization scheme account for these large and small clouds, it would also accurately account for ensembles of clouds containing any specified combination of each type.

The current study consists of a set of model runs, which are designed to explore the coupling of a boundary layer model with improved cumulus parameterization schemes. A one-dimensional radiative-convective model, which incorporates a radiation parameterization from the Canadian Climate Centre (Morcrette, 1991), an adapted version of the Frank-Cohen cloud model (Frank and Cohen, 1985, 1987), and a

boundary layer model (Albrecht et al., 1979), is used in these runs.

The model was developed to study the effects of increased subsidence on the inversion height and the thermodynamic structure of the boundary layer. The model will also be tested with several cloud spectra, with varying mixing rates and initial updraft velocities, to determine whether specification of cloud type is important in one-dimensional modeling. Lastly, the sea surface temperature and the surface wind velocity will be varied to study the change in the inversion height and the boundary layer structure.

Chapter 2

MODEL DESCRIPTION

Base Model

The radiative-convective model used in the current study will be used to study the effects of increased subsidence on inversion height and boundary layer variables. Furthermore, cloud spectra will be introduced into the cloud model to test the effects of various cloud types on the inversion height and structure of the boundary layer. Finally, these variables will be studied as the sea surface temperature and the surface wind speed vary.

The model uses a 300 second timestep and the primitive equations in pressure coordinates, with a constant surface pressure of 1006 millibars (mb) and top level of 100 mb. There are currently thirty-six levels in the model and the layers are 50 mb thick above 800 mb and 10 mb thick below 800 mb, with the exception of the bottom layer, which is 6 mb thick. The levels are indexed from the top of the atmosphere to the surface. The increased resolution of the model at low levels increases the accuracy of predicting the inversion height. Furthermore, since the modeled small clouds are only 10 to 20 mb deep, this increased resolution makes it possible to include these clouds in the parameterization. The pressure level of the capping inversion that defines the top of the boundary layer is

predicted by the boundary layer scheme. Thus, it changes with time and is not carried specifically as a model level.

The constant, externally-specified divergence profile forces the large-scale vertical velocity to vanish ($\omega=0.0$ Pa/s) at the top and bottom levels. The vertical velocity used to calculate the large-scale vertical advection is defined as $\delta\omega=-D*\delta p$, in units of [Pa/s], where D is the large-scale divergence [s^{-1}].

The bulk of the model is written in terms of dry static energy (s) and mixing ratio (q). The large-scale tendency equations take into account the effects of horizontal and vertical advection, vertical diffusion, the diabatic heating of radiation, and the diabatic heating and moistening of convection. The tendency equation for q is similar to that for s ; however, in the q -equation, the effects of radiation are omitted. The basic form of the equation is shown in terms of s here.

$$\frac{\delta s}{\delta t} + \nabla \cdot \mathbf{v} s + \frac{\omega \delta s}{\delta p} + K \frac{\delta^2 \omega}{\delta p^2} = -L(c-e) + (Q)_R \quad (2.1)$$

In this equation, c is the effect of condensation, e is the effect of evaporation, K is the diffusion coefficient (.5), and $(Q)_R$ is the diabatic heating of radiation.

By applying the continuity equation, this equation can be written in flux form:

$$\frac{\partial s}{\partial t} + \bar{v} \nabla s + \frac{\partial (\omega s)}{\partial p} + K \frac{\partial^2 \omega}{\partial p^2} - L(c - e) + (Q) \quad (2.2)$$

Horizontal gradients of s and q can be externally specified, but are neglected in this study. Thus, this equation reduces to the following equation.

$$\frac{\partial s}{\partial t} = -\bar{v} \nabla s - \frac{\partial (\omega s)}{\partial p} - K \frac{\partial^2 \omega}{\partial p^2} - L(c - e) + (Q) \quad (2.3)$$

The resulting equation in terms of q is:

$$\frac{\partial q}{\partial t} = -\bar{v} \nabla q - \frac{\partial (\omega q)}{\partial p} - K \frac{\partial^2 \omega}{\partial p^2} - L(c - e) \quad (2.4)$$

The large-scale horizontal divergence of s and q is calculated in flux form using a simple pressure-weighted finite differencing method, and both the large-scale vertical advection and the vertical diffusion terms are computed using simple centered finite differencing approximations. The convective tendency terms are computed in a subroutine named CLOUD, and the radiative effects are calculated in a subroutine named RADIATN.

Equations 2.3 and 2.4 are only applied at levels above the inversion, because values of s and q below the inversion are predicted only by the boundary layer scheme (Albrecht et al., 1979), which incorporates the effects of cloud heating

and moistening and radiational heating. This scheme will be discussed in more detail later.

The excess moisture generated in the model is simply converted into liquid water and "rained out", leaving saturated air at that level. The effect of the latent heating associated with this conversion process is calculated and applied to the atmosphere accordingly.

Boundary Layer Model

The boundary layer model to be coupled with the cloud model is essentially that described in Albrecht et al., 1979. It has been extracted from the layered model described in that paper. The main features of the model will be explained here. The diagram in Appendix B shows the main features of the mixed layer.

The boundary layer is assumed to be well-mixed in s and q . Although the superadiabatic layer immediately above the surface has been neglected, it has been shown in previous studies that the modeled structure is realistic to a first approximation (Betts, 1976a). The transition layer lies just above the mixed boundary layer and just below the overlying cloud layer. This infinitesimally thin layer, the inversion level, has discontinuities in s , s_v , and q at that level, where s_v is the virtual dry static energy defined as $s_v = c_p T + g z + L q$, where L is the latent heat of

condensation, g is the acceleration of gravity, and c_p is the specific heat at constant pressure.

The inversion strength, measured by Δs and Δq , is computed by assuming that there is a constant specified lapse rate between the inversion and the model level immediately above the inversion. The reason for this assumption will be described in more detail in Chapter 3. The initial lapse rates are $ds/dP = -.65 \text{ J/kg*Pa}$ and $dq/dP = 3.0 \times 10^{-6} \text{ g/g*Pa}$. If the inversion rises above a level, the lapse rates immediately above the inversion are set equal to the existing lapse rates in that layer. These lapse rates are slowly relaxed back to the constant lapse rates defined above. This relaxation is done in an attempt to keep the lapse rates of s and q constant immediately above the inversion without introducing large shocks into the model. This is also explained further in Chapter 3. However, if the inversion drops below a model level, then the values of s and q at the level which no longer is included in the boundary layer, are defined as those values which are necessary to keep the inversion strength constant for that timestep. The lapse rates of s and q between this level and the inversion are also held constant across the inversion, but the lapse rates continue to relax toward the constant lapse rates, as specified above.

Horizontal advection terms are neglected in the boundary layer, so only the divergence of the turbulent

fluxes in the boundary layer, convection, and radiation affect s_m and q_m , which are the mixed layer values of s and q respectively.

$$\frac{ds_H}{dt} = \frac{-g[(F_s)_s - (F_s)_0]}{\rho_s} + \left(\frac{\partial s_H}{\partial t}\right)_{RAD} + \left(\frac{\partial s_H}{\partial t}\right)_{CONV} \quad (2.5)$$

$$\frac{dq_H}{dt} = \frac{-g[(F_q)_s - (F_q)_0]}{\rho_s} + \left(\frac{\partial q_H}{\partial t}\right)_{CONV} \quad (2.6)$$

The surface fluxes are calculated by the bulk aerodynamic method, and α -theory, which states that the flux of s_v at the top of the boundary layer is a constant fraction of the surface flux of s_v , (α) (equation 2.8), is used to calculate the fluxes at the top of the boundary layer, as shown below.

$$\frac{(F_s)_s}{\Delta s} = \frac{(F_{sv})_s}{\Delta s_v} = \frac{(F_q)_s}{\Delta q} \quad (2.7)$$

$$(F_{sv})_s = \alpha (F_{sv})_0 \quad (2.8)$$

$$(F_s)_0 = \rho_0 C_T V_0 (s_0 - s_H) \quad (2.9)$$

$$(F_q)_0 = \rho_0 C_T V_0 (q_0 - q_H) \quad (2.10)$$

$$F_{sv} = F_s + \delta C_p T_R F_q \quad (2.11)$$

By assuming that the discontinuities in s and q are always at the same height as each other, the fluxes at the top of the mixed layer can be calculated using the proportionality equations shown above (equation 2.7). Equations 2.9 and 2.10 are the bulk aerodynamic calculations of the surface fluxes of s and q respectively, where ρ_0 is the density of the surface air, C_T is the transfer coefficient (1.15×10^{-3}), V_0 is the surface wind speed, and s_0 and q_0 are the sea surface values of dry static energy and mixing ratio. Equation 2.11 shows the relationship of the flux of s_v , (F_{sv}) , to the fluxes of s and q , (F_s, F_q) , where δ is a constant value (.608), C_p is the specific heat at constant pressure (1004. J/(deg*kg)), and T_R is a reference temperature (288K).

The inversion height, or pressure difference between the surface and the inversion, (P_B) is predicted from the turbulent entrainment across the inversion (ω_{INV}), the convective fluxes across the inversion (ω_{CONV}), and the large-scale vertical velocity (ω_B). The large-scale vertical velocity is, as defined earlier, $\omega_B = -D \cdot P_B$, where D is the large-scale divergence [s⁻¹], and ω_B is the large-scale vertical velocity [Pa/s].

$$\frac{dP_B}{dt} = \left(\frac{\partial P_B}{\partial t}\right)_{INV} + \left(\frac{\partial P_B}{\partial t}\right)_{CONV} + \left(\frac{\partial P_B}{\partial t}\right)_{LARGESCALE} \quad (2.12)$$

$$\left(\frac{\partial P_B}{\partial t}\right)_{INV} = \frac{-g(F_B)_B}{\Delta s} \quad (2.13)$$

$$\left(\frac{\partial P_B}{\partial t}\right)_{CONV} = \frac{g(F_B)_B}{\Delta s} \quad (2.14)$$

$$\left(\frac{\partial P_B}{\partial t}\right)_{LARGESCALE} = -DP_B \quad (2.15)$$

Equation 2.13 shows the effect of the turbulent entrainment across the inversion on the rate of rise of the inversion. As the flux of s at the top of the boundary layer grows more negative, the inversion will rise more quickly, since the air just above the inversion becomes entrained into the boundary layer faster. The rate of rise of the inversion is also affected by the convective fluxes across the inversion. Equation 2.14 shows that as the convective flux of s , $(F_s)_B$, increases, the rate of rise of the inversion increases. Also, increasing the strength of the inversion (Δs) decreases the rate of rise of the inversion height. Lastly, equation 2.15 explains how the rate of rise of the inversion increases as the divergence decreases and this rate is also directly proportional to the depth of the boundary layer (P_B) . Each of these effects essentially works to change the amount of mass in the boundary layer, thereby forcing the inversion to rise (increased mass in the boundary layer) or fall (decreased mass in the boundary

layer).

In general, the surface flux of s warms the boundary layer, while the downward flux of s at the top of the inversion warms the boundary layer and cools the small layer above the inversion. This small layer above the inversion is also warmed by large-scale and convection-induced subsidence. Therefore, an equilibrium is reached. The surface flux of q moistens the boundary layer, and the upward flux of q at the top of the boundary layer tends to moisten the small layer above the inversion. The large-scale subsidence, and sometimes the convection-induced subsidence, tends to dry out this small layer above the inversion, and again an equilibrium is reached.

The convective mass flux, w' , is defined as the product of the fractional area of updrafts, updraft velocity, and updraft density. The model closure assumption requires that enough convective mass flux be generated to keep the inversion approximately at the lifting condensation level (LCL) by generating an equivalent compensating subsidence mass flux between the clouds. This closure assumption is that the lifted parcels become saturated at the top of the mixed layer, ensuring that cloud base is at the top of the mixed layer. If the mass flux is positive, then it follows that a cloud must exist in order to satisfy the closure assumption. Thus, the added fluxes of s and q in the boundary layer which are needed to maintain the height of

the inversion at cloud base, or the LCL, is the convective mass flux. These added fluxes cause a cloud to be forced at every timestep that the LCL would otherwise fall below the inversion. This mass flux is then used to scale the heating and moistening profile from the cloud model (Frank and Cohen, 1985). The cloud model produces heating and moistening profiles per unit cloud base mass flux.

Cloud Model

The cloud model has been adapted from a version of the Frank and Cohen (1985) cloud model, which simulates deep, precipitating clouds. The primary difference between the version used here and the published version is the inclusion of small, non-precipitating clouds with no downdrafts.

The model is written in pressure coordinates, and indexed from the surface to the top of the atmosphere. The program setup requires that all of the model levels be included in the cloud model. Currently, this cloud model is run with the same resolution as the main program.

The basic cloud model consists of a steady-state plume convective updraft with a steady state plume convective downdraft, and a constant lateral mixing rate for each cloud type. The lateral entrainment and detrainment rates are computed from the buoyancy sorting program of Kain and Fritsch (1990). Surface air, from the well-mixed boundary

layer, feeds the updraft, and cloud base is assumed to be at the LCL, or inversion height. Equivalent potential temperature is conserved for mixing processes, and T and q are then computed using Bolton's formula (1980).

Each cloud is assigned an updraft velocity at cloud base. At each level, the non-hydrostatic vertical velocity is calculated, and the top level of the cloud is defined as the highest level above the inversion with positive updraft velocity. All of the remaining updraft mass flux is detrained at this level. This version of the cloud model forces all clouds to detrain no lower than the second model level above the inversion (at least 10 mb above the inversion), because clouds smaller than this would form highly unrealistic heating and moistening profiles due to lack of vertical resolution of the model. Furthermore, this type of model does not take into account the mechanisms responsible for formation of the very small clouds, which have a vertical extent of less than 20 mb.

After completion of the updraft calculations, the downdraft is initiated at the highest level where there is rain. The cloud model uses a RWATER subroutine, taken from Kain and Fritsch (1990), which computes rain when positive vertical velocity and cloud water coexist. Only 60% of the cloud water participates in the conversion to rainwater, and no rain is allowed if the cloud top is lower than 900 mb, which keeps unrealistic downdrafts from forming in small,

supposedly non-precipitating clouds. In this study, however, an automatic conversion factor is used. As its name suggest, this method automatically converts all cloud water above a certain threshold, currently .001 g/g, into rain.

In a mass flux cloud model, convective heating can be expressed in terms of several cloud processes: descending/ascending motion, which compensates for the net convective mass flux, detrainment mixing from the updraft and downdraft, and evaporation of detrained cloud water. The compensating vertical motion usually dominates in the middle of the cloud, and the detrainment mixing effects usually dominate near the top and bottom of the cloud. The moistening is determined from similar processes: cloud-induced subsidence drying, detrainment of water vapor from updrafts and downdrafts, and evaporation of detrained cloud water not converted into rain. Once again, the compensating subsidence drying effects tend to dominate in the cloud's middle layers and the detrainment moistening dominates near the cloud's top and bottom.

The cloud model is designed to compute profiles for a spectrum of clouds. The distinguishing features of the cloud types are mixing rates, initial updraft velocities, and their relative cloud base mass fluxes. The heating/moistening profiles are calculated separately for each type of cloud and then a single, weighted-average

profile is computed based on the specified mass flux percentage of each cloud type present.

Radiation Model

The detailed radiation scheme used in the radiative-convective model is from Morcrette (1991). The radiation code includes a diurnal cycle, as well as a seasonal cycle, and no significant changes have been made to the code. It is run as Morcrette described and the basic features are outlined here. It is known as the EC3 radiation scheme, which was developed from high resolution spectral models and then simplified by making some assumptions which added computational efficiency at the same time. The clear-sky longwave fluxes are calculated using an emissivity method with a temperature and pressure dependent representation of the absorption. Clouds are assumed to be gray bodies and have a longwave emissivity which depends on the cloud water path. The shortwave fluxes are calculated using a photon path distribution method. A delta-Eddington approximation is used to compute the scattering, which is separated from the effects of absorption. The transmission functions are Pade approximations. The parameters for the cloud shortwave radiation include optical thickness and single-scattering albedo. Because stratiform clouds tend to contain smaller droplets than do cumulus and cirrus clouds, the cloud types

are taken into account by including the effective radius of the cloud particles and the liquid water path in the cloud in the calculations of optical thickness. The input variables are temperature and mixing ratio. The only output directly used in this version of the base model is the total heating profile.

In summary, this program provides a detailed radiation scheme and accounts for both deep and shallow cumulus clouds, through specification of a spectrum of clouds. The program is designed to combine a number of previously used and tested ideas to provide a realistic framework with which to study the equilibrium state of the atmosphere.

Chapter 3

METHODOLOGY

The current study seeks to better understand and improve upon the solutions to various problems associated with cloud parameterizations in radiative-convective models. The primary foci are as follows:

- 1) differences between large and small clouds,
- 2) method of calculating the inversion strength, and
- 3) determination of cloud spectrum.

Differences between Large (Precipitating) and Small (Non-precipitating) Clouds

The Frank-Cohen cloud model (1985) was designed for use in the Frank and Cohen (1987) cumulus parameterization, which only accounts for large, precipitating clouds with tops above 800 mb. Although, this type of cloud parameterization is useful, ideally a cumulus parameterization should also include small, non-precipitating clouds. However, several difficulties arise in trying to do this. Two of the problems in differentiating small and large clouds are of interest in the current study, although they were not both completely resolved in the course of this research. These are:

- 1) method of determining cloud existence and/or size,
and
- 2) method of calculating precipitation.

One of the primary purposes of this study was to adapt the Frank-Cohen (1985) single-mode cloud model so that it could be used to simulate both large and small clouds simultaneously, through the specification of a spectrum of clouds. However, the above problems needed to be addressed in order for this adaptation to be possible.

The first of these problems is that large clouds are normally assumed to form where the convective available potential energy (CAPE) is above a certain threshold ($CAPE > 0$, in the Frank/Cohen, 1985, model). The clouds in that version of the model are fed from the level below 700 mb with the highest value of CAPE, and this level is subsequently defined as the cloud base. However, in the atmosphere, small, non-precipitating clouds form even when the CAPE values are negative. These clouds are simply an overshooting of the boundary layer turbulent updrafts, and they form when the lifting condensation level (LCL) is at or below the inversion height. No positive buoyancy is necessary in order for these clouds to form, as there is almost always a slight overshooting from the boundary layer. The negative buoyancy associated with negative CAPE values simply inhibits the clouds from reaching the level of free

convection (LFC) and forming the large clouds.

This method of using CAPE to determine cloud existence has been modified in the current version of the model. Because small clouds are now incorporated, CAPE is no longer a consideration in the formation of the clouds. The updrafts are fed from the boundary layer air and the cloud base is assumed to be at the LCL. If the inversion is at the same height as or above the LCL, then there is assumed to be a cloud. As in the Frank/Cohen model (1985), the non-hydrostatic vertical velocity at each level is calculated and the top of the cloud is defined to be the last level in the cloud with upward motion.

If the direction of vertical motion is downward at the first or second model level above cloud base, then the cloud is forced to continue to the next model level, thereby forcing the top level of the cloud to be at least two model levels above the LCL. This is required for adequate vertical resolution of the shallow clouds. Although the CAPE no longer plays a part in the decision of whether a cloud should form, CAPE still affects the height of clouds that reach the LFC. Sometimes, however, the integrated value of CAPE is large, but the bottom levels of the cloud are stable enough to inhibit large-cloud formation. In this situation, only small clouds form.

Another dissimilarity between small and large clouds is in parameterizing precipitation. In the Frank-Cohen model

(1985), there is assumed to be precipitation when upward motion coexists at the same level as liquid water. However, the small clouds, which are, by definition, non-precipitating, may still satisfy these two conditions, and, therefore, the model sometimes unrealistically predicts precipitation even for clouds with cloud tops under 900 mb.

In order to ensure that precipitation does not form in these non-precipitating clouds, this method for computing the precipitation, which was adopted from the Kain and Fritsch (1990) model, was replaced with an automatic conversion factor. Precipitation is assumed to form at a given level when the moisture content exceeds a preset threshold value, which is currently set to .001 g/g. Any moisture in excess of this value is automatically converted to rainwater, and if necessary, snow. This method prevents the small clouds from unrealistically forming rain or snow just because upward motion and liquid water coexist.

Method of Calculating Inversion Strength

The inversion strength at the top of the boundary layer is a highly unstable variable. Albrecht et al. (1979) predict the jumps in s , q , and s_v at the top of the inversion by enforcing a constant lapse rate in their highly simplified cloud layer. Therefore, they extrapolate this constant lapse rate to the inversion level to get the

magnitude of the inversion strength. However, more complex cloud models do not have constant lapse rates in the cloud layer, but predict the values of S , q , and S_v at each level. Therefore, estimating the inversion strength becomes a somewhat more complicated matter.

Several problems are inherent in estimating this inversion strength. It is physically unrealistic for the discontinuities in s and s_v to become negative; therefore, a scheme for calculating these jumps must be formulated with this condition strictly enforced. Another consideration is the continuity of the inversion strength when the inversion passes a model level. It is highly unrealistic to see a discontinuity in the magnitude of the inversion strength when the inversion level passes by a non-physical model level. Infinite vertical resolution would solve this problem, but as this level of detail is highly uneconomical, avoidance of this problem is desirable. Ideally, a method for determining the inversion strengths will be immune to these model levels. Several methods of calculating these jumps were tested, and the procedure outlined below was implemented.

Enforcing a constant lapse rate in s and q in the small layer immediately above the inversion usually prevents Δs from becoming negative, and is somewhat physically realistic with the model's increased resolution below 800 mb. However, when the inversion level passes a model level, the

temperature and moisture just above the inversion experience jump discontinuities. Therefore, when the inversion falls below a level, the lapse rates of s and q are defined empirically as $-.65 \text{ (J/kg)/Pa}$ and $3.0 \times 10^{-6} \text{ (g/g)/Pa}$, respectively. However, when the inversion rises above a level, the lapse rates in the layer immediately above the inversion are redefined as the existing lapse rates in that layer. Then, the lapse rates in the layer above the inversion are gradually adjusted toward the specified constant lapse rates at a rate of 10% per time step until the inversion passes another level. This method prevents shocks from being introduced into the system.

Determination of Cloud Spectra

Because in nature, various types and sizes of clouds coexist, a cloud model spectrum or ensemble should be more realistic than a single type of cloud. Cloud spectra distinguish between cloud types in a variety of ways. Although it is commonly accepted that an ensemble of cloud types is preferable to the single cloud type in a radiative-convective model, the method of specifying the cloud types' differences is still under debate. In the current study, the specified parameters are: number of cloud types, the mixing rate of each cloud type, and the initial updraft velocity of each cloud type.

When clouds have large mixing, or entrainment, rates, which allows environmental air to mix rapidly with cloud air, the clouds tend to be small. Likewise, small mixing rates tend to produce larger clouds. Therefore, specifying the mixing rate has a big impact on whether the cloud will be a small, non-precipitating cloud or a large, precipitating one. However, other environmental factors affect a cloud's growth. For example, a large value of CAPE can help a cloud to overcome large mixing rates by encouraging the growth that the large mixing rate inhibits. Also, even with a very small mixing rate and high net CAPE, a layer of strong negative buoyancy just above cloud base can stunt a cloud's growth by preventing it from reaching the LFC.

Another important factor in cloud type specification is the initial updraft velocity. Large clouds generally have larger initial updraft velocities at cloud base, and small clouds generally have smaller initial updraft velocities, though this is variable. A large value of updraft velocity can overcome the layer of negative buoyancy that tends to exist just above the cloud base. Therefore, the mixing rates, CAPE, lid, and initial updraft velocities work together to determine cloud size. The initial updraft velocities may be a random distribution around some typical mean value. In this case, use of a random number generator would be helpful in determining these velocities. Over the

tropical oceans and with the types of clouds modeled here, an average updraft velocity value at cloud base is on the order of 1.0-1.5 m/s (e.g. Emmitt, 1978). These values vary, but are generally confined to between .5 m/s and 2.0 m/s. This relationship, though, is still imperfect and, therefore, there is room for improvement on the exact relationship between the updraft vertical velocity and cloud height.

In the experiments, several combinations of mixing rates, initial updraft velocities, and number of cloud types are examined. These experiments will attempt to improve the understanding of these relationships discussed above.

Chapter 4

RESULTS

Experiment Design

The current study tests whether the model can produce realistic results in the range of its assumed validity, which extends from the tradewind regions to the edges of the Intertropical Convergence Zone (ITCZ) regions of the tropics. Therefore, several sensitivity studies, as well as experiments involving varied divergence profiles and cloud spectra, are conducted. The model is tested with two soundings. The first data set with which the model is initialized approximates a trade wind sounding used by Albrecht et al. (1979). This Atlantic Trade-Wind Experiment (ATEX) sounding (sounding 1, see Appendix A) is from the period of February 7-12, 1969 (figures 4.0a,b). The second sounding, from the ITCZ regime, will be discussed later.

The control run for the experiments will be called run 3, which is described in Table 1. The experiments are divided into three sections: divergence profile experiments, cloud spectra experiments, and sensitivity experiments. The divergence profile experiments are conducted by varying the divergence profiles to induce a range of subsidence profiles from weak lifting to strong subsidence (runs 1-6). The cloud spectra experiments involve changing the specified

Figure 4.0a. ATEX trade wind region sounding from February 4-12, 1969, (sounding 1) and FGGE average sounding from an area near the equator (sounding 2) of dry static energy [s].

Figure 4.0b. ATREX sounding from February 4-12, 1969, (sounding 1) and FGGE sounding (sounding 2) of mixing ratio (q).

cloud spectrum by varying the number of cloud types, mixing rate of each cloud type, and initial vertical velocity of each cloud type (runs 7-12). Finally, the sensitivity experiments involve varying the sea surface temperature, surface wind speed, and initial sounding (runs 13-17).

Divergence Profile Experiments

The divergence/omega profiles chosen for these experiments represent different regions, between approximately 10 N and 25 N and between 10 S and 25 S, of the tropical and sub-tropical oceans. The sub-tropical high regime is represented by very strong subsidence (run 5), the trade wind regime is represented by moderate to weak subsidence (runs 2,3,4), and the intertropical convergence zone is represented by upward motion or no vertical motion (runs 1,6). These six runs are made with varying divergence profiles, while the other variables remain as they were in the control run (Table 1).

Five of the divergence profiles used in the runs consist of a layer of divergence extending from the surface to 800 mb, and a layer of convergence extending from 300 mb to the top of the modeled atmosphere, 100 mb. The sixth run has the same structure, but with divergence at the top of the atmosphere and convergence at the surface, to induce

Table 1. Control run description.

sea surface temperature	298 K
divergence at lower levels/ divergence at upper levels [s^{-1}]	$3.3 \times 10^{-6}/$ -3.0×10^{-6}
number of clouds	2
mixing rates of clouds [mb^{-1}]	.002, .01
initial updraft vertical velocities [m/s]	1., 1.
surface wind speed [m/s]	5.0

lifting (Table 2). While the vertical structure is the same in each run, the magnitude of the divergence varies (figures 4.0c,d).

The type of convection formed depends on which regime it is in, and therefore, it depends on the vertical velocity profile. In the subtropical high regime, observations show stratocumulus dominating the atmosphere. The trade wind regime is dominated by fair weather trade cumulus, which are shallow cumulus clouds with tops penetrating into a high level inversion (near 800 mb) known as the trade inversion, and cumulus congestus. Closer to the equator, in the Intertropical Convergence Zone (ITCZ), the atmosphere is dominated by strong convection, or thunderstorms. This model is able to resolve the strong convection and trade

Figure 4.0c. The divergence profiles for each of the six runs with varying divergence.

Figure 4.0d. The corresponding omega profiles for each of the six runs with varying divergence.

Table 2. Description of runs 1-6 with varying divergence. All other variables are as described in Table 1.

run number	value of divergence at lower levels [s ⁻¹]	value of divergence at upper levels [s ⁻¹]
run 1	0.0	0.0
run 2	1.12×10^{-6}	-1.0×10^{-6}
run 3	2.23×10^{-6}	-2.0×10^{-6}
run 4	3.3×10^{-6}	-3.0×10^{-6}
run 5	4.3×10^{-6}	-3.9×10^{-6}
run 6	-0.6×10^{-6}	0.54×10^{-6}

cumulus reasonably well. It cannot, however, model stratocumulus because of the restrictions imposed by its closure assumption, namely that the LCL and inversion height must stay together through the forcing of small trade cumulus. For stratocumulus regimes, this assumption must be relaxed. This expansion is reserved for future research.

Throughout the model runs with varying divergence, the inversion height stays near 950 mb, as is commonly observed over the tropical ocean, and always exhibits oscillatory behavior. Observational data over the tropical Western

Pacific ocean during the TOGA Pilot cruise of the RV Wecoma (Young et al., 1992) shows oscillatory behavior in the values of boundary layer temperature and moisture due to the effects of convection (figures 4.0e,f). Despite the subtleness of interregional differences in the boundary layer behavior, reasonable interregional variations in convection can be found in the following experiments.

First, in the run with no vertical motion, which is typical of regions near the equator and the ITCZ, the inversion oscillates around 950 mb (figure 4.1a). The entire run is dominated by hours long periods of deep convection with strong downdrafts, which imply heavy rain, interspersed with hours-long periods of alternating deep and shallow convection (figures 4.1b,c). Because the downdrafts produced by deep convection are associated with rain, this alternating of clouds with downdrafts and clouds without downdrafts implies intermittent rain. During the periods of strong, continuous heavy rain, the boundary layer dry static energy decreases and the boundary layer moisture increases. These deep clouds stabilize the atmosphere by causing a net heating in the atmosphere, thereby increasing the static stability, and by warming the boundary layer, thereby decreasing the convective instability. This stabilizing causes the inversion to sink. In contrast, during the periods of more intermittent heavy rain, the shallow convection causes the atmosphere to destabilize by causing a

Figure 4.0e. Observational data from TOGA pilot cruise of boundary layer temperature for a period of 7 days gathered over the tropical ocean (Young et al., 1992).

Figure 4.0f. As in figure 4.0e, but for boundary layer moisture (q).

Figure 4.1a. For run 1, the time series of inversion pressure for one week.

Figure 4.1b. For run 1, with no divergence, the time series of cloud tops for one week for the cloud with a low mixing rate ($.002 \text{ mb}^{-1}$).

Figure 4.1c. For run 1, the time series of cloud tops for one week for the cloud with a high mixing rate ($.01 \text{ mb}^{-1}$).

net cooling in the deep layer and by cooling the boundary layer, making the atmosphere more convectively unstable. By warming the atmosphere, the deep convection stabilizes the atmosphere and only small convection is allowed to form. Then, the shallow convection cools the layer right above cloud base and causes deep convection. These two effects reach an equilibrium state in time. Ideally, the cloud spectra will enable the atmosphere to reach an equilibrium state without having to cycle between periods of small and deep convection. However, this is reserved for further study in this ongoing research project.

During the period following the fourth day of this run, the inversion height behavior alternates between periods of very small oscillations and periods of no oscillations. The oscillatory periods are associated with the extended periods of deep convection, where the wake from the downdrafts causes inversion lowering, followed by rainless periods long enough for boundary layer recovery. The non-oscillatory periods are associated with the periods of alternating deep and shallow convection. During this period, the downdrafts have little lasting effect on the inversion, resulting in nearly constant inversion height.

The boundary layer remains very dry and cool during this run because of the effects of convective downdrafts (figure 4.1d,e). The boundary layer stays approximately two degrees cooler than the sea surface and the mixing ratio

Figure 4.1d. For run 1, the time series of boundary layer dry static energy (s) and the dry static energy at the top of the inversion. The difference between these two values is one measure of the inversion strength. Sea surface s is 299192 J/kg.

Figure 4.1e. For run 1, the time series of boundary layer mixing ratio (q) and the mixing ratio at the top of the inversion.

stays at around 13.8 g/kg. Thus, surface fluxes of sensible and latent heat are large. The inversion strength is approximately one degree and is therefore fairly weak, which allows for the deep convection. These convective effects generally agree with the observations of Young et al., (1992).

The inversion height oscillates more with subsidence (run 2) than it did without any imposed vertical motion, yet it still oscillates around 950 mb (figure 4.2a). As the subsidence increases, the periods of no convection increase in both number and duration (figures 4.2b,c). The deep convection is followed by these periods of no convection, during which the boundary layer recovers from the downdraft cooling associated with the deep convection. The same pattern of oscillations is generally followed as in the case of no divergence: The periods of deep convection stabilize the boundary layer and force the boundary layer top down by causing dry static energy to decrease and mixing ratio to increase, which lowers the LCL and, therefore, the inversion. When the deep convection stops, the atmosphere destabilizes, or becomes drier and warmer increasing the chance that deep convection will form. The atmosphere then becomes unstable enough to form thunderstorms, and the deep convection begins again. Thus, the entire cycle is repeated. This process is seen throughout the experiments and will therefore only be referred to from now on as the

Figure 4.2a. As in figure 4.1a, but for run 2.

Figure 4.2b. As in figure 4.1b, but for run 2.

Figure 4.2c. As in figure 4.1c, but for run 2.

stabilizing/destabilizing effect of the convection.

In this run with weak subsidence, the boundary layer is slightly warmer and moister than in the run with no subsidence (figures 4.2d,e). Because of the reduced convection, the boundary layer temperature is only about .5-1 degree cooler than the sea surface temperature, and the boundary layer mixing ratio is between 14 and 15 g/kg. This moistening will be explained later.

When the subsidence increases to a value typical of the trade wind region (run 3), the inversion oscillates around 950 mb (figure 4.3a), but with smaller, more regular oscillations, and with longer periods of clear skies between the periods of deep convection (figures 4.3b,c). The small clouds cool the area right above cloud base, thereby causing deeper convection; whereas the deep convection warms the area right above cloud base and the clouds are thus suppressed. The periods of clear skies correspond, once again, with destabilization following the stabilizing effect of the downdrafts. The inversion strength increases with the increased subsidence. The boundary layer temperature is warmer (less than .5 degrees cooler than sea surface temperature) and moister (15.2 g/kg) than in the previous runs (figures 4.3d,e). Otherwise, the model produces similar results and shows evidence of the type of convection characteristic of the trade wind region, namely fairly long periods of small or no convection alternating with periods

Figure 4.2d. As in figure 4.1d, but for run 2.

Figure 4.2e. As in figure 4.1e, but for run 2.

Figure 4.3a. As in figure 4.1a, but for run 3.

Figure 4.3b. As in figure 4.1b, but for run 3.

Figure 4.3c. As in figure 4.1c, but for run 3.

Figure 4.3d. As in figure 4.1d, but for run 3.

Figure 4.3e. As in figure 4.1e, but for run 3.

of deep convection.

The inversion in the next run (run 4) lowers only slightly to a level near 955 mb (figure 4.4a), and the small oscillations resemble those seen in the run with slightly less subsidence (run 3). With this greater subsidence, deep convection becomes even less frequent and much weaker (figures 4.4b,c). The inversion is only slightly stronger, and the boundary layer is moister and warmer than in the lower subsidence runs. The boundary layer temperature approaches the sea surface temperature and the boundary layer mixing ratio is approximately 16-16.5 g/kg (figures 4.4d,e).

When the subsidence becomes large enough to completely suppress cumulus convection as in the sub-tropical high region (run 5), a region generally dominated by a strong inversion and stratocumulus clouds, the modeled inversion height falls to the surface (figure 4.5a). In order to form stratocumulus, the LCL and the inversion height must decouple and clouds must be allowed to form in the top layer of the mixed layer. Because the model's closure assumption does not allow these stratocumulus clouds to exist, the model continues to force small trade cumulus and suppress all deep convection (figures 4.5b,c). Therefore, the trade cumulus updrafts are unrealistically venting mass out of the boundary layer. This mass is not being replaced by downdraft air, since the trade cumulus downdrafts are non-

Figure 4.4a. As in figure 4.1a, but for run 4.

Figure 4.4b. As in figure 4.1b, but for run 4.

Figure 4.4c. As in figure 4.1c, but for run 4.

Figure 4.4d. As in figure 4.1d, but for run 4.

Figure 4.4e. As in figure 4.1e, but for run 4.

Figure 4.5a. As in figure 4.1a, but for run 5.

Figure 4.5b. As in figure 4.1b, but for run 5.

Figure 4.5c. As in figure 4.1c, but for run 5.

existent. Thus, the inversion sinks to the sea surface as all the air is pumped out of the boundary layer. The boundary layer temperature becomes warmer than the sea surface temperature, thereby cutting off the surface heat flux, but the continuing surface flux of moisture causes the boundary layer to moisten too quickly (figures 4.5d,e). Thus, the LCL sinks, and the model forms fog and shuts down.

When the model is applied to a region with weak upward vertical motion (run 6), typical of a weak ITCZ, the inversion still oscillates around 950 mb (figure 4.6a), and experiences the stabilizing/destabilizing effects of convection (figures 4.6b,c). The boundary layer remains very cool and dry with a weak inversion (figures 4.6d,e). As in the zero subsidence case, the region is dominated by rainy periods followed by periods with intermittent rain. This behavior is typical of this regime where strong thunderstorms frequently form.

When the model was run for strong upward motion, the model crashed because the inversion strength became too small. However, the model should work under such conditions, and, therefore, this problem will be addressed in future work, which is already underway. Therefore, the model gives the most accurate results in the regions poleward of the Intertropical Convergence Zone and equatorward of the subtropical high.

From these runs, it is apparent that as subsidence

Figure 4.5d. As in figure 4.1d, but for run 5.

Figure 4.5e. As in figure 4.1e, but for run 5.

Figure 4.6a. As in figure 4.1a, but for run 6.

Figure 4.6b. As in figure 4.1b, but for run 6.

Figure 4.6c. As in figure 4.1c, but for run 6.

Figure 4.6d. As in figure 4.1d, but for run 6.

Figure 4.6e. As in figure 4.1e, but for run 6.

increases, the inversion strength increases and the boundary layer becomes warmer and moister, with longer periods of no convection and weaker periods of deep convection (figures 4.6f-h). As seen in equation 2.7, the downward flux of s at the top of the boundary layer is directly proportional to the inversion strength. Therefore, the increased inversion strength results in an increased flux of s through the inversion into the boundary layer. Because there is a surface heat flux from the ocean surface as well, the flux convergence of s causes a warmer boundary layer in response to the inversion strength increasing. In contrast, the upward flux of q at the top of the inversion is inversely proportional to the strength of the inversion (eqn. 2.6), so as the inversion increases, the flux of q out of the boundary layer decreases. This process allows the boundary layer to moisten. The inversion height, however, stays near 950 mb during each run with subsidence characteristic of the regions that the model is intended to represent (runs 1-4,6).

The final soundings of s and q are displayed (figures 4.6i,j) for three of the six soundings (runs 2,3,4). As subsidence decreases (run 2), the static stability through most of the troposphere increases since there is relatively less subsidence warming of the lower levels, which in turn allows more deep convection, producing warming of the upper levels. The less stable lapse rate produced in the run with

2 of 2

Figure 4.6f. Composite graph of the time series of the inversion pressure for each of the six runs with varying divergence.

Figure 4.6g. Composite graph of the time series of the boundary layer dry static energy for each of the six runs with varying divergence.

Figure 4.6h. Composite graph of the time series of the boundary layer moisture for each of the six runs with varying divergence.

Figure 4.61. Soundings of s at the end of runs 2, 3, and 4.

Figure 4.6j. As in 4.6i, but for q.

more subsidence (run 4) is caused by strong subsidence warming in the lower troposphere, which prevents deep convection and therefore leads to radiative cooling of the upper troposphere. Lastly, the soundings of q are drier and more variable in the vertical with increased subsidence. The results presented in this section are from runs with no horizontal advection of s and q . The addition of this horizontal advection will decrease the magnitude of the net effect of the radiation and the convection on the upper levels of the atmosphere, resulting in a more realistic sounding, and will be the subject of future research.

Cloud Spectrum Experiments

The externally-specified cloud spectrum influences the inversion height by encouraging or discouraging growth of various types of cloud. The results shown in figures 4.1-4.6 are from runs with two cloud types. One of the cloud types has a high mixing rate ($.01 \text{ mb}^{-1}$) and the other has a low mixing ratio ($.002 \text{ mb}^{-1}$). They are initialized with the same updraft vertical velocity (1.0 m/s). However, when this spectrum varies, it stands to reason that the resultant inversion height and boundary layer properties can vary as well. Because the equilibrium of the atmosphere is highly dependent on the forms of convection, and the convective heating and moistening effects are very strong in tropical

regions, changing the clouds could change the entire result of a climate model.

This section investigates the degree of this sensitivity to details of the specified cloud spectrum. Using the same subsidence from the run illustrated in figures 4.3a-e, which is representative of the trade wind region, the cloud spectrum is varied to study the effects of this spectrum specification on the atmosphere. Table 3 lists the spectra to be tested with this subsidence. For each run, each cloud type's updraft is equally weighted in the gridbox.

When only one of the previously used cloud types is used, the inversion height behaves differently. The cloud type with a lower mixing rate ($.002 \text{ mb}^{-1}$) causes more oscillations in inversion height (figure 4.7a) than the run with two cloud types experiences, but the oscillations still center around 950 mb. These oscillations are caused by the stabilizing/destabilizing effect of the convection. Because only one size of large clouds form and the small clouds are not there to damp out the effects of this deep convection, the downdraft cooling affects the atmosphere more radically than in the run with both types of clouds (figure 4.7b). Periods of intermittent strong convection alternate with periods of no convection at all.

When a single cloud type with a higher mixing ratio ($.01 \text{ mb}^{-1}$) is used, the inversion height is still centered

Figure 4.7a. Time series of the inversion pressure for the run with only one cloud type with mixing rate of $.002 \text{ mb}^{-1}$, for run 7.

Figure 4.7b. Time series of cloud tops for the run with only one cloud type with mixing rate of $.002 \text{ mb}^{-1}$, for run 7.

Table 3. The spectra to be tested in the cloud model in runs 7-12.

run #	number of clouds types	mixing rates [mb ⁻¹]	initial updraft velocities [m/s]
7	1	.002	1.0
8	1	.01	1.0
9	3	.002	.5, 1.0, 1.5
10	3	.002, .005, .01	1.0
11	5	.002, .004, .005, .008, .01	1.0
12	5	.002	.5, 1.0, 1.5, 2.0 , 2.5

around 950 mb (figure 4.8a) and smaller clouds form (figure 4.8b). This result is expected because the higher mixing rate generally stunts the clouds' growth. These periods of deep convection, in contrast to those resulting from the clouds with smaller mixing rate, are fairly steady, although they still alternate with periods of no convection. This steadiness of rainfall within deep convective periods

Figure 4.8a. Same as in 4.7a, but for only one cloud type with mixing rate of $.01 \text{ mb}^{-1}$, for run 8.

Figure 4.8b Same as in 4.7b, but for only one cloud type with mixing rate of $.01 \text{ mb}^{-1}$, for run 8.

seems to damp the inversion height's oscillations, which are more regular.

Of the two runs with only one cloud type, the one with the lower mixing rate results in a cooler boundary layer because of the associated stronger downdrafts. The moisture, however, is similar in both runs (figures 4.8c,d). The run with both cloud types more closely resembles the single-cloud type run with the lower mixing rate ($.002 \text{ mb}^{-1}$). This resemblance results, in part, from the dominating effect of the deep convection in the cloud model, as in the tropical atmosphere.

Next, a spectrum is used that incorporates three cloud types. As seen in Table 3, the cloud types have the same mixing rate, which at $.002 \text{ mb}^{-1}$ is a relatively low value, but the clouds are initialized with different initial vertical velocities that vary from $.5 \text{ m/s}$ to 1.5 m/s . The inversion still oscillates around 950 mb , but the oscillations are more irregular in this run than in the one with two cloud types (figure 4.9a). The resulting cloud heights vary only slightly, if at all (figure 4.9b). Similar to the run with two cloud types, there are periods of deep convection alternating with periods of no convection. However, rainfall within these periods of deep convection is intermittent in the run with three cloud types. There are fewer clouds than in the run with only one cloud type of the same mixing rate; however, the clouds in

Figure 4.8c. Time series of the boundary layer s and the s at the top of the boundary layer for each run with one cloud type and the control run with two cloud types (runs 7, 8, and 3).

Figure 4.8d. Time series of the boundary layer q and the q at the top of the boundary layer for each run with one cloud type and the control run with two cloud types (runs 7, 8, and 3).

Figure 4.9a. As in figure 4.7a, but for run 9, with three cloud types.

Figure 4.9b. As in figure 4.7b, but for run 9, with three cloud types. Only one of the cloud types was shown since they were identical.

each run behave very similarly.

The three cloud types in the next run have varying mixing rates and identical initial updraft velocities. Again, the inversion height stays near 950 mb (figure 4.10a). The cloud heights vary predictably with the mixing rate: the largest mixing rate produces the smallest clouds and the smallest rate produces the largest clouds (figures 4.10b-d). These clouds produce steady rainfall, alternating with periods of shallow or no convection, and affect the boundary layer through the stabilizing/ destabilizing effect of the convection, and seem to damp the oscillations of the inversion height. The boundary layer temperature in both runs with three cloud types (runs 9 and 10) approaches the sea surface temperature, but still remains almost .5 degrees cooler than ocean. The boundary layer moisture also stays around 15.3 g/kg during both of these runs. These boundary layer properties are similar to those seen in the control run with two cloud types (figures 4.10e,f).

Increasing the number of cloud types to five has little effect on the overall height of the inversion. The clouds in the first experiment with five types have varying mixing rates with constant initial updraft velocities (Table 3). The inversion height remains near 950 mb with some irregular behavior toward the end of the run (figure 4.11a). Again, the clouds with the lowest mixing rates are slightly taller than those with the highest (figures 4.11b-f). The periods

Figure 4.10a. As in figure 4.7a, but for run 10, with three cloud types.

Figure 4.10b. As in figure 4.7b, but for run 10, with the cloud type with mixing rate of $.002 \text{ mb}^{-1}$.

Figure 4.10c. As in figure 4.7b, but for run 10, with the cloud type with mixing rate of $.005 \text{ mb}^{-1}$.

Figure 4.10d. As in figure 4.7b, but for run 10, with the cloud type with mixing rate of $.01 \text{ mb}^{-1}$.

Figure 4.10e. As in figure 4.8c, but for the two runs with three cloud types and the control run with two cloud types (runs 9, 10, and 3).

Figure 4.10f. As in figure 4.8d, but for the two runs with three cloud types and the control run with two cloud types (runs 9, 10, and 3).

Figure 4.11a As in figure 4.7a, but for run 11, with five cloud types.

Figure 4.11b As in figure 4.7b, but for run 11, with five cloud types, for the cloud with mixing rate of .002 mb^{-1} .

Figure 4.11c As in figure 4.7b, but for run 11, with five cloud types, for the cloud with mixing rate of .004 mb^{-1} .

Figure 4.11d As in figure 4.7b, but for run 11, with five cloud types, for the cloud with mixing rate of $.005 \text{ mb}^{-1}$.

Figure 4.11e. As in figure 4.7b, but for run 11, with five cloud types, for the cloud with mixing rate of $.008 \text{ mb}^{-1}$.

Figure 4.11f. As in figure 4.7b, but for run 11, with five cloud types, for the cloud with mixing rate of $.01 \text{ mb}^{-1}$.

of deep convection yield steady rain and alternate with short periods of shallow or no convection. The small clouds at the end of the run account for the sudden rise in inversion height. If the run were continued, the inversion would most probably be forced down again by deep convection.

The last run in this series is with five cloud types of varying initial updraft velocities (between .5 m/s and 2.5 m/s) and constant mixing rates. The inversion is once again oscillating around 950 mb (figure 4.12a). The oscillations are fairly irregular, and they are caused by the same processes of alternating deep and shallow convection as in the other runs. The varying initial updraft velocities have little if any effect on the cloud height (figure 4.12b). Therefore, the clouds will all have similar effects on the environment. Interestingly, the rain during convective periods is more intermittent without long periods of sustained deep convection.

The boundary layer temperature and moisture from the runs with five cloud types and the control run with two cloud types are very similar, implying that specification of a cloud spectrum, although important, is not crucial to the final values of boundary layer temperature, moisture, and inversion height (figures 4.12c,d). There are, however, differences in the oscillations of the inversion height. The runs with an enhanced cloud spectrum seem to have much more regular oscillations than the runs with only one cloud

Figure 4.12a. As in figure 4.7a, but for run 12, with five cloud types.

Figure 4.12b. As in figure 4.7b, but for run 12, with five cloud types, with mixing rate of .002 mb^{-1} . Only one time series is shown since the cloud types are identical.

Figure 4.12c. As in figure 4.8c, but for the two runs with five cloud types and the control run with two cloud types (runs 11, 12, and 3).

Figure 4.12d. As in figure 4.8d, but for the two runs with five cloud types and the control run with two cloud types (runs 11, 12, and 3).

type. Therefore, it is concluded that cloud spectra are important in modeling the atmosphere. Varying the vertical velocity with clouds of the same mixing ratio seems to have little effect. These clouds are deep enough to be fairly insensitive to small changes in initial updraft vertical velocity. Likewise, the results of the model are fairly insensitive to mixing rate changes. The cloud size, however, does respond to changing the mixing rates. The lower mixing rates produce larger clouds and the higher mixing rates produce smaller clouds. Primarily, the model depends more on whether a cloud spectrum exists that incorporates both small and large clouds, as opposed to exactly how these cloud types are specified. The model behaves more realistically with more than one cloud type than with just one cloud type.

Sensitivity Experiments

The surface wind speed and sea surface temperature affect the physics of the boundary layer (eqn 2.8) by changing the surface fluxes of dry static energy and moisture. As the boundary layer cools and moistens from the deep convection described earlier, the surface heat flux increases and the surface moisture flux decreases, causing the boundary layer to recover. Therefore, if the constants used to compute the surface fluxes change, the results of

the model will be different. Several sensitivity studies were conducted using the same subsidence profile as in the previous set of experiments (run 3). The two variables to be tested are: surface wind speed and sea surface temperature. A different sounding (sounding 2) is used as input to the model to test the robustness of the model. The sensitivity studies use the same variables as in the control run, with the exceptions of those described in Table 4.

Table 4. Experiment descriptions for the sensitivity studies.

run number	SST	surface wind speed [m/s]
run 13	298 K	3.0
run 14	298 K	10.0
run 15	296 K	5.0
run 16	300 K	5.0

The increased surface wind speed causes a drop in the inversion height to between 970 and 980 mb (figure 4.13a). The resultant boundary layer temperature oscillates between sea surface temperature and approximately .5 degrees cooler than sea surface temperature (figure 4.13b). The boundary layer also becomes very moist as a result of this increased

Figure 4.13a. Time series of inversion pressure with varying surface wind speed.

Figure 4.13b. Time series of boundary layer s for runs with varying surface wind speed.

surface flux of moisture (figure 4.13c). As the moisture increases, the LCL lowers and, therefore, the inversion is lower. There are also larger clouds in this run than in the control run. The increased moisture and temperature of the boundary layer air, which feeds the clouds' updrafts will naturally lead to more buoyant air and the clouds will be taller.

Lowering the surface wind speed to 3.0 m/s has, naturally, the opposite effect. The surface fluxes are decreased, causing a drier, cooler boundary layer than in runs with higher surface wind velocities (figures 4.13b,c). The inversion, therefore, oscillates at a higher level, namely between 930 and 940 mb (figure 4.13a). The clouds are smaller and more sporadic, with long periods of no deep convection.

The sea surface temperature has a somewhat similar effect on the resultant inversion height. Lowering the sea surface temperature from 25C to 23C lowers the inversion height to an average value of approximately 975 mb (figure 4.14a) because the boundary layer cools, as a result of the decreased surface heat flux, but the moisture does not change. Thus, the LCL, and therefore the inversion, lowers. The boundary layer temperature and moisture fluxes are limited by the reduced sea surface temperature, but the boundary layer temperature still stays within .5 degrees of the sea surface temperature (figure 4.14b). The moisture

Figure 4.13c. Time series of boundary layer q for the runs with varying surface wind speed.

Figure 4.14a. Same as in 4.13a, but for varying sea surface temperature.

Figure 4.14b. Same as in 4.13b, but for varying sea surface temperature.

stays approximately between 15 g/kg and 16 g/kg (figure 4.14c). The clouds are taller and have periods of both extended rain and intermittent rain along with the periods of no convection.

With an increased sea surface temperature, the inversion hovers around 930 mb (figure 4.14a), and the boundary layer is warmer, but only slightly moister (figures 4.14b,c). This increase in temperature and moisture results from the increased surface fluxes of s and q . The LCL thus rises, causing the height of the inversion to increase. The clouds are taller than in the control run and experience primarily intermittent rain. The boundary layer temperature stays within approximately 1 degree of the sea surface temperature.

Summarizing, as the surface wind speed increases, the inversion height decreases, the boundary layer temperature decreases, the boundary layer moisture increases, and the cloud height increases. As sea surface temperature lowers, the inversion height lowers, the boundary layer temperature increases, and the boundary layer moisture decreases.

When the model was initialized with a different sounding (sounding 2, see Appendix A) that is more characteristic of the atmosphere near the equator, and run under the same conditions as in the control run, the inversion comes to equilibrium near 960 mb (figure 4.15a). Periods of deep convection and shallow convection still

Figure 4.14c. Same as in 4.13c, but for varying sea surface temperature.

Figure 4.15a. Time series of the inversion pressure for the FGGE sounding (sounding 2).

alternate and these periods affect the inversion height and boundary layer temperature and moisture in the same way as before (figures 4.15b,c). The boundary layer is very moist and very warm reflecting the warmer, moister sounding that was introduced with a warmer sea surface temperature (figures 4.15d,e). The model, though, seems to work for various soundings in the area between the ITCZ and the subtropical high region.

Although all the runs presented in this section were run for 7 days, the model was tested by running the control run (run 3) for 14 days and another run (run 9) for 27 days. The boundary layer depth and did not vary enough to warrant extended runs for the other experiments. Therefore, this extension of the research is left as future work.

Figure 4.15b. Time series of the cloud tops for the clouds with mixing rate of $.002 \text{ mb}^{-1}$ for the FGGE sounding.

Figure 4.15c. Time series of the cloud tops for the clouds with mixing rate of $.01 \text{ mb}^{-1}$ for the FGGE sounding.

Figure 4.15d. Time series of boundary layer s and s at the top of the inversion for the FGGE sounding.

Figure 4.15e. Time series of boundary layer s and s at the top of the inversion for the FCG soundings.

Chapter 5

CONCLUSION/FUTURE WORK

A radiative-convective model which included previously-developed radiation (Morcrette, 1991) and cumulus parameterizations (Frank and Cohen, 1987) coupled with a previously-developed boundary layer scheme (Albrecht et al., 1979) was developed. The cloud model used in the cumulus parameterization was modified to include the effects of small, non-precipitating clouds, and the parameterization closure was altered to use the convective mass flux of the Albrecht, et. al. (1979) planetary boundary layer model for closure. The model was tested between the physical regions of the ITCZ and the subtropical high by varying the divergence/vertical velocity profiles. The inversion height stayed remarkably near 950 mb for each run in this region. The boundary layer temperature and moisture increased as the subsidence increased, because of the increased strength of the mixed layer-capping inversion, which inhibited deep convection.

The model sensitivity to cloud spectrum specification was tested with various cloud spectra as well as with different types of single clouds. With only one cloud type, the inversion height, although, still centered around 950 mb, oscillated considerably. The results improved with a cloud spectrum, but the structure of the boundary layer

remained relatively constant with each run. Varying the initial updraft vertical velocity of the cloud types had no effect on the clouds' final sizes.

Several sensitivity studies were performed by varying the sea surface temperature and the surface wind speed. Changing the sea surface temperature causes predictable changes in the inversion height: Higher sea surface temperatures result in higher inversions and warmer boundary layers. The moisture does not vary significantly with the different sea surface temperatures. As the surface wind speed increases, the inversion height lowers because the boundary layer becomes warmer and moister. Thus, the sensitivity studies show that the model behaves similarly to the atmosphere.

A warmer, moister sounding was also used to initialize the model, with the same divergence profile as in run 3. The inversion stabilized near 960 mb. These results show that the model gives realistic results under various conditions and is, therefore, a valid tool for studying a variety of atmospheric phenomena.

In modeling a complex and nonlinear phenomenon such as an inversion-capped boundary layer coupled with cumulus clouds, much liberty must be taken in making assumptions and parameterizations. Therefore, especially with a one-dimensional model, there will always be room for improvement concerning these assumptions. The primary concern of the

project was to construct a useful base model to be improved upon and perhaps incorporated into a two and/or three dimensional dynamical model. The basic premise of the model is very accurate, as can be seen from the results. It is only fair to point out the inherent problems with a model, so that improving upon it is made slightly more simple. The primary assumptions made in the current study which should be examined further are:

- method of approximating the lapse rates
immediately above the cloud base,
- method of determining cloud existence,
- closure assumption that the inversion must stay with
cloud base under all conditions,
- method of specifying cloud properties in the cloud
spectrum,
- method of determining net mass flux profile with
height, and
- method of partitioning the downdraft air.

The lapse rate approximation for the layer immediately above the inversion is highly variable in this study and depends quite a bit on the previously existing thermodynamic structure in a given layer. This relationship means that a slight error in the cloud calculations could lead to an error in the thermodynamic structure, which would also lead to possibly a bigger error in the calculation of inversion

strength because of the assumed lapse rate. Obviously, this feedback process is undesirable and is avoided as much as possible in the current runs. Only infinite resolution could completely correct this problem, but despite possible shortcomings of this technique, the model results proved robust and fairly stable.

The closure for the boundary layer assumes that enough cloud base mass flux exists to force the inversion to stay with the LCL. However, sometimes, this required cumulus mass flux cannot be realistically achieved, but when the buoyancy condition is checked, the clouds do not reach their second level above the LCL, or, in other words, do not penetrate high enough to form a realistic cumulus. Currently, the model forces these small clouds to form anyway, which potentially causes some unrealistic cloud heating and moistening profiles. One way to reduce possible errors of this section would be to increase the vertical resolution near the inversion. Furthermore, a more accurate way of handling the transition to stratocumulus would improve the realism of the model. Thus, the model would benefit from taking into account the effects of stratocumulus clouds as well as very shallow cumulus clouds. The assumption that the inversion, or cloud base, must stay at the height of the LCL needs to be relaxed for realistic stratocumulus clouds to form.

The use of a cloud spectrum is necessary, but it is not

yet certain what the important variables are to specify. Much work needs to be done with researching the correlations between initial updraft velocity, and mixing rate of each cloud type. There are several other correlations that might be important for this problem, such as correlations with cloud diameter and correlations with gridbox CAPE values.

The problem of the net mass flux profile is also an important one to consider. Because entrainment and detrainment rates differ with type and size of cloud, the corresponding mass flux profiles will also vary. In this model, the clouds assume a relatively constant mass flux profile with height. However, this approximation is slightly inaccurate. Very high or low entrainment or detrainment rates can produce strongly increasing or decreasing mass flux with height. This difference will certainly affect the clouds, as well as the clouds' effect on the atmosphere.

Yet another problem that needs consideration in this model is the partitioning of the downdraft cooling. In models with a large areal extent, such as this one, the downdraft cooling is automatically evenly distributed, and affects the entire grid box. This uniformity is, however, undesirable as it tends to cause convection to switch fully on or off. In nature, the cooling from this downdraft is seen locally and forms the convectively important phenomenon known as a gust front. Thus, the overall effect of the

downdraft cooling may cause some of the oscillatory behavior observed in the model runs. Work on including an embedded cold outflow region scheme in the parameterization is in progress.

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APPENDIX A

Soundings

Sounding 1. (ATEX)

Pressure [mb]	Temperature [C]	Mixing ratio [g/kg]
100	-75.67	.005
150	-69.22	.01
200	-54.63	.04
250	-43.12	.16
300	-32.40	.24
350	-24.37	.31
400	-17.60	.47
450	-11.33	.58
500	-6.47	.87
550	-2.98	.96
600	0.17	1.80
650	3.87	2.00
700	8.87	2.80
750	11.87	3.20
800	14.35	4.30
810	14.76	4.62
820	15.14	5.42
830	15.51	6.22
840	15.88	7.02
850	16.19	7.57
860	16.39	7.75
870	16.60	7.92
880	16.81	8.10
890	17.00	8.43
900	17.18	9.01
910	17.35	9.58

920	17.50	10.16
930	17.70	10.65
940	17.88	11.025
950	18.06	11.30
960	18.69	11.40
970	18.70	12.00
980	19.57	12.00
990	20.43	12.00
1000	21.30	12.00
1006	21.82	12.00

Sounding 2. (FGGE)

Pressure [mb]	Temperature [C]	Mixing Ratio [g/kg]
100	-65.67	.005
150	-58.22	.01
200	-50.63	.04
250	-41.13	.08
300	-30.72	.23
350	-22.66	.52
400	-16.72	.88
450	-10.58	1.37
500	-5.66	1.76
550	-1.37	2.16
600	2.64	2.79
650	6.24	4.01
700	9.99	5.08
750	13.13	5.99
800	14.75	8.24
810	15.04	8.89
820	15.34	9.57

830	15.59	10.17
840	15.97	10.56
850	16.46	10.81
860	16.90	10.93
870	17.35	11.18
880	17.90	11.52
890	18.38	12.05
900	18.88	12.57
910	19.40	13.03
920	19.97	13.55
930	20.48	14.29
940	21.05	14.99
950	21.71	15.57
960	22.93	16.07
970	23.24	16.43
980	24.04	16.65
990	24.89	16.85
1000	25.61	16.92
1006	26.25	17.09

APPENDIX B
Diagram of Boundary Layer

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