Experiments with Lilly's Cloud-Topped Mixed Layer Model

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ABSTRACT

Lilly's model of a horizontally homogeneous cloud-topped mixed layer is studied. The model is closed by taking a weighted (weighting factor or entrainment parameter $k$) average of Lilly's maximum and minimum entrainment cases. The dependence of steady-state solutions on large-scale divergence, sea surface temperature and entrainment parameter $k$ is investigated. By numerical integration the response of the mixed layer to a diurnally varying radiative flux is investigated. Significant variations in the state of the mixed layer and in the convective fluxes are found.

1. Introduction

One of the difficult remaining problems for the modeling of large-scale atmospheric motions is to parametrically include into the heat, moisture and momentum equations the vertical transport and source/sink terms which result from small-scale convective motions. This problem is of particular importance in tropical and subtropical latitudes.

The advent of the meteorological satellite has greatly increased our understanding of the global distribution of convective regimes. For example, Fig. 1, prepared by Miller and Feddes (1971), shows a 4-year average (1967–70) of cloud brightness for January and July.

![Image of cloud brightness maps for January and July](image-url)

**Fig. 1.** Four-year average (1967–70) cloud brightness for January and July (from Miller and Feddes, 1971).
One striking feature of Fig. 1 is the bright band of cloudiness between 5° and 10°N associated with the ITCZ. This time-averaged band of cloudiness is the result of westward propagating clusters of deep cumulonimbus clouds and their associated cirrus cloud shields (Chang, 1970). Another striking feature of Fig. 1 is the existence of large areas of low-level stratocumulus convection in the strong subsidence regions to the east of the subtropical highs. A typical instantaneous view of the stratocumulus regime is shown in Fig. 2. It is composed mostly of closed convective cells (Agee and Dowell, 1974) which are generally confined to the lowest kilometer.

A schematic cross section (see also Riehl et al., 1951; Malkus, 1958; Neiburger et al., 1961) from San Francisco to the Hawaiian Islands to the Marshall Islands is shown in Fig. 3. This cross section runs approximately along the trajectories of low-level air parcels. Three regimes are shown. The stratocumulus regime and the trade cumulus regime make up the broad descending branch of the Hadley circulation. The cumulonimbus regime, or ITCZ, makes up the narrow ascending branch of the Hadley circulation.

In the steady-state stratocumulus regime, clouds exist in the face of strong large-scale subsidence through the clouds themselves. Air parcels crossing into the cloud must jump to a higher moisture content and to a lower potential temperature. The jump to higher moisture content is accomplished by a near discontinuity in the upward turbulent flux of total water. The jump to lower potential temperature is accomplished by near discontinuities in the radiative flux and in the turbulent fluxes of liquid water (causing cooling by evaporation) and potential temperature (Lilly, 1968).

In the trade cumulus regime shallow, primarily non-precipitating cumulus exist in the face of large-scale subsidence. Just below the trade inversion this subsidence, in combination with the large vertical gradients of potential temperature and mixing ratio, causes large warming and drying effects. These effects are opposed primarily by the cooling and moistening caused by the detrainment of water substance from clouds which lose buoyancy just below the trade inversion. This picture is consistent with the recent BOMEX and ATEX data (Holland and Rasmussen, 1973; Nitta and Esbensen, 1974; Nitta, 1975; Augstein et al., 1973), and with the predictions of Betts (1973).

The tendency of the subtropical atmosphere to divide itself into layers means that air parcels in the downward branch of the Hadley circulation change their potential temperature and moisture content not only gradually when descending through a layer but also in sudden jumps when crossing from one layer to another.

A general circulation model capable of realistic simulation of the Hadley circulation should incorporate the recent advances in cumulus parameterization theory (e.g., Ooyama, 1971; Arakawa and Schubert, 1974) in combination with a unified mixed layer theory, i.e., a mixed layer theory capable of handling both situations where cloud base lies below the top of the mixed layer and situations where cloud base lies above the top of the mixed layer. Such an approach is presently being undertaken for the NCAR GCM by Deardorff (1972, 1976) and for the UCLA GCM by Randall and Arakawa (see Arakawa et al., 1974, and Arakawa, 1974).

As a step in this direction we will study one part of the total problem. In particular, we will study the stratocumulus model developed by Lilly (1968). In Section 2 we review Lilly's model. In Section 3 we investigate its steady-state solutions, and in Section 4 we investigate some of its time-dependent solutions.

2. Review of Lilly's model

In the absence of condensation we can regard the dry static energy $s = c_p T + g z$ and the water vapor mixing
ratio $q$ as conservative quantities. Thus, in a non-saturated mixed layer $s$ and $q$ are constant with height up to the top of the mixed layer $p_B$, and the lifting condensation level $p_C$ lies above $p_B$.

When water changes phase we can regard the moist static energy $h = c_s T + g z + L q$ as a conservative quantity. In addition, if the condensed water is not precipitated but is carried along with the air, we can regard the sum of the water vapor mixing ratio $q$ and the liquid water mixing ratio $l$ as a conservative quantity. Thus, in a saturated non-precipitating mixed layer $h$ and $q + l$ are constant with height up to the top of the mixed layer $p_B$, and the lifting condensation level $p_C$ lies below $p_B$. If $h$ and $q + l$ are constant with height in the mixed layer, it follows that the same is true for $s - L l$. The temperature and moisture fields of the mixed layer are predicted if any two of the three quantities $h$, $q + l$ and $s - L l$ are predicted. We have chosen to predict $h$ and $q + l$.

Assuming no variations in the horizontal direction and denoting the mixed layer values of $h$ and $q + l$ by $h_M$ and $(q + l)_M$, the mixed layer budget equations are

$$\frac{\partial h_M}{\partial t} = \frac{g}{p_0 - p_B} [(F_h)_0 - (F_h)_B],$$

$$\frac{\partial (q + l)_M}{\partial t} = \frac{g}{p_0 - p_B} [(F_{q + l})_0 - (F_{q + l})_B],$$

where $p_0$ is the surface pressure and $(F_h)_B$ and $(F_{q + l})_B$ are the turbulent fluxes of $h$ and $q + l$ just below $p_B$. Above $p_B$ the turbulent fluxes jump to zero. The surface turbulent fluxes of $h$ and $q + l$ are given by

$$(F_h)_0 = \rho_0 C_T V_0 [h_0 - h_M],$$

$$(F_{q + l})_0 = \rho_0 C_T V_0 [q_0^* - (q + l)_M],$$

where $\rho_0$ is the surface air density, $V_0$ the surface wind speed, $C_T$ the transfer coefficient, $h_0^*$ the surface saturation static energy, and $q_0^*$ the surface saturation mixing ratio.

Application of the budget equations for $h$ and $q + l$ in the infinitesimally thin layer at the cloud top leads to

$$\frac{1}{g} \left( \frac{\partial \rho_B}{\partial t} - \omega_B \right) \Delta h + (F_h)_B = \Delta F_B,$$

$$\frac{1}{g} \left( \frac{\partial \rho_B}{\partial t} - \omega_B \right) \Delta (q + l) + (F_{q + l})_B = 0,$$

where $\omega_B$ is the large-scale vertical $\rho$-velocity at $p_B$, $\Delta h$ and $\Delta (q + l)$ are the jumps in $h$ and $q + l$ across $p_B$ and $\Delta F_B$ is the jump in the radiative flux across $p_B$. The jump of a function is defined as the value of the function at a pressure slightly lower than $p_B$ minus the value at a pressure slightly higher than $p_B$; e.g., $\Delta h$ is defined as $h(p_B) - h_M$.

The cloud base pressure $p_C$ is approximately given in terms of the mixed layer total water content $(q + l)_M$ and the saturation mixing ratio of the mixed layer air just above the surface $q^*(p_0^*)$ as

$$\frac{p_0 - p_C}{p_0} = \frac{q^*(p_0^*) - (q + l)_M}{b} = \frac{(1 + \gamma)[q_0^* - (q + l)_M] - (\gamma/L)[h_0^* - h_M]}{b},$$

where $b$ is a dimensionless constant and $\gamma$ is defined in (11).

One additional equation is needed to close the system. Lilly has argued that the turbulent energy balance sets maximum and minimum bounds on the entrainment. The maximum and minimum entrainment conditions proposed by Lilly are respectively

$$\frac{1}{p_0 - p_B} \int_{p_B}^{p_0} F_{*} dp = 0 \quad \text{but} \quad F_{*} \neq 0 \text{ somewhere,}$$

$$F_{*} \mid_{p_B} = 0 \quad \text{but} \quad \frac{1}{p_0 - p_B} \int_{p_B}^{p_0} F_{*} dp > 0,$$

where $F_{*}$ is the flux of virtual dry static energy and is given in terms of the fluxes of moist static energy and total water as

$$F_{*} = \begin{cases} \beta F_h - \epsilon LF_{q + l}, & p_B < p < p_C \\ F_h - (1 - \epsilon \delta) LF_{q + l}, & p_C < p \leq p_0. \end{cases}$$

The dimensionless constants appearing in (10) are defined as

$$\beta = \frac{1 + \gamma(\delta + 1)}{1 + \gamma}, \quad \epsilon = \frac{L \left( \frac{\partial q^*}{\partial T} \right)_p}{c_p T}, \quad \delta = 0.608,$$

where $F_{*}$ is proportional to the generation of turbulent kinetic energy by buoyancy, the maximum entrainment condition (8) states that regions of positive generation of turbulent kinetic energy are balanced by regions of negative generation, the net generation for the layer thus vanishing. The minimum entrainment condition (9) states that there is no region of negative generation of turbulent kinetic energy and that the net generation for the layer is positive. Thus, all energy dissipation occurs in the positive region. In this paper we shall use a weighted average of (8) and (9), which can be
written\footnote{The factor 1/2 in (12) is somewhat arbitrary and is included so that (12) will reduce to the customary closure assumption in the non-saturated case (see explanation at the end of this section). Omission of this factor simply results in a revised interpretation of the parameter $k$.}

\[
\frac{k}{\rho_0 - \rho_B} \int_{\rho_B}^{\rho_0} F_{*} d\rho + \frac{1}{2} (1 - k) (F_{*})_{\text{min}} = 0. \quad (12)
\]

Since $0 \leq k \leq 1$, a positive $(F_{*})_{\text{min}}$ in (12) would require a negative vertically averaged $F_{*}$. This is clearly impossible. Thus, the solutions of (12) satisfy

\[
\frac{1}{\rho_0 - \rho_B} \int_{\rho_B}^{\rho_0} F_{*} d\rho \geq 0 \quad \text{and} \quad (F_{*})_{\text{min}} \leq 0. \quad (13)
\]

The fluxes of $h$ and $q+l$ are linear in $\rho$ in the mixed layer, so that

\[
F_{h} = \left( \frac{\rho - \rho_B}{\rho_0 - \rho_B} \right) (F_{h})_0 + \left( \frac{\rho - \rho_B}{\rho_0 - \rho_B} \right) (F_{h})_B, \quad (14)
\]

\[
F_{q+l} = \left( \frac{\rho - \rho_B}{\rho_0 - \rho_B} \right) (F_{q+l})_0 + \left( \frac{\rho - \rho_B}{\rho_0 - \rho_B} \right) (F_{q+l})_B. \quad (15)
\]

Using (10), (14) and (15), the entrainment relation (12) can be written

\[
\left\{ \frac{1 - \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right)^2}{\beta} + \left[ 1 - \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right)^2 \right] \right\} (F_{h})_B - \left\{ 1 - \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right)^2 \right\} + (1 - \epsilon_0) \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right) L (F_{q+l})_B
\]

\[
+ \left\{ \frac{1 - \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right)^2}{\beta} + \left[ 1 - \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right)^2 \right] \right\} (F_{h})_0 - \left\{ 1 - \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right)^2 \right\} + (1 - \epsilon_0) \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right) L (F_{q+l})_0
\]

\[
= \frac{1 - k}{k} \left[ \frac{\left( \frac{\rho_0 - \rho_B}{\rho_0 - \rho_B} \right) (F_{h})_0 + \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right) (F_{h})_B}{\beta} \right] - \epsilon L \left[ \frac{\left( \frac{\rho_0 - \rho_B}{\rho_0 - \rho_B} \right) (F_{q+l})_0 + \left( \frac{\rho_0 - \rho_C}{\rho_0 - \rho_B} \right) (F_{q+l})_B}{\beta} \right]
\]

\[
= 0. \quad (16)
\]

Since $F_{*}$ is linear in pressure in the subcloud layer and in the cloud layer but is discontinuous across cloud base $\rho_C$, the minimum $F_{*}$ may occur at the top of the cloud layer $\rho_B$, the bottom of the cloud layer $\rho_C$, the top of the subcloud layer $\rho_C^T$, or the bottom of the subcloud layer $\rho_B$. These four possibilities for the minimum $F_{*}$ are reflected in the four rows within the large braces of (16).

The purpose of introducing the entrainment condition is to close the system by relating the fluxes at some level above the surface to the fluxes at the surface. Once this is done it becomes possible to compute all the fluxes at all the levels. Only then do we know where the minimum $F_{*}$ occurs. Thus, (16) has a somewhat implicit form. This is of no consequence in steady-state solutions since the steady-state minimum $F_{*}$ always falls at $\rho_C^T$. It is of consequence in time-dependent solutions and is discussed in Section 4.

A special case of (16) occurs when the minimum $F_{*}$ falls at $\rho_C^T$ and $\rho_C \to \rho_B$, i.e., the mixed layer becomes non-saturated. Then, using (10), Eq. (16) reduces to

\[(F_{*})_B = -k (F_{*})_0, \] which is the customary closure assumption in the non-saturated case.

3. Steady-state solutions

In the steady state the fluxes of $h$ and $q+l$ are constant with height in the mixed layer and are equal to their surface values. Eqs. (1) through (6) can be combined to give

\[
\frac{\omega_B}{\rho_0 C_T V_0} h^* + h^* - \frac{\Delta F_R}{\rho_0 C_T V_0} = \frac{\omega_B}{\rho_0 C_T V_0} + 1, \quad (17)
\]

\[
\frac{\omega_B}{\rho_0 C_T V_0} q^* + q^* = \frac{\omega_B}{\rho_0 C_T V_0} + 1. \quad (18)
\]

Cloud base is given by (7). The entrainment relation
can be written

\[
\frac{k}{\rho_0 - \rho_B} \left[ (F_{se})_0 (p_0 - p_B) + (F_{se})_0 (p_0 - p_c) \right] = \frac{1}{1 + \gamma} (1 - k) (F_{se})_0 = 0. \tag{19}
\]

The minimum \( F_{se} \) occurs in the subcloud layer because Lilly has shown that in the steady state

\[
(F_{se})_0 - (F_{se})_0 = \frac{1 - e^{(\delta + 1)}}{1 + \gamma} \rho_0 C_T V_0 L = \frac{\rho_0 - p_c}{\rho_0} > 0. \tag{20}
\]

Let us assume that the large-scale subsidence field and the moist static energy and water vapor mixing ratio above the mixed layer are given by

\[
\omega_B = D (\rho_0 - \rho_B), \tag{21}
\]

\[
h(\rho_B) = h_0 - \frac{\partial h}{\partial \rho} (\rho_0 - \rho_B), \tag{22}
\]

\[
q(\rho_B) = q_0 - \frac{\partial q}{\partial \rho} (\rho_0 - \rho_B), \tag{23}
\]

where \( h_0, \partial h/\partial \rho, q_0, \partial q/\partial \rho, \) and the large-scale divergence \( D \) are known constants. We can now compute steady-state solutions using the following numerical values:

\[
\rho_0 C_T V_0 = 0.0129 \text{ kg m}^{-2} \text{ s}^{-1}, \quad \Delta F_R = 65.65 \text{ W m}^{-2},
\]

\[
h_0^* = \begin{cases} 310.62 \text{ kJ kg}^{-1} \\ 315.90 \text{ kJ kg}^{-1} \end{cases},
\]

\[
q_0^* = \begin{cases} 9.27 \times 10^{-3} \\ 1.058 \times 10^{-3} \end{cases}, \quad h_0 = 313.95 \text{ kJ kg}^{-1},
\]

\[
\frac{\partial h}{\partial \rho} = - 0.25 \text{ kJ kg}^{-1} \text{ kPa}^{-1},
\]

\[
q_0 = 3.3 \times 10^{-1}, \quad \frac{\partial q}{\partial \rho} = 0.043 \times 10^{-2} \text{ kPa}^{-1},
\]

\[
\rho_0 = 102.0 \text{ kPa}
\]

\[
D = \begin{cases} 3.5 \times 10^{-6} \text{ s}^{-1} \\ 5.0 \times 10^{-6} \text{ s}^{-1} \end{cases}, \quad \beta = 0.532, \quad \gamma = 1.340,
\]

\[
\epsilon = 0.114, \quad b = 0.0356.
\]

The lower values of \( h_0^* \) and \( q_0^* \) result from a sea surface temperature of 13°C, the higher values from a temperature of 15°C. Although \( \Delta F_R \) is not independent of mixed layer variables, for simplicity we shall treat it as so.

The method we have used to compute steady-state solutions is as follows. Choose a value of \( \rho_B \) and compute \( h_M \) and \( (q + l)_M \) from (17) and (18) using the auxiliary relations (21) through (23). Then compute \( \rho_B \) from (7), \( F_{se} \) and \( F_{se} \) from (3) and (4), \( (F_{se})_0 \) and \( (F_{se})_0 \) from (10), and \( k \) from (19). This procedure has resulted in an acceptable solution if \( 0 \leq k \leq 1 \) and \( \rho_B \leq \rho_c \leq \rho_0 \), and can easily be repeated for many choices of \( \rho_B \).

The case \( h_0^* = 310.62 \text{ kJ kg}^{-1}, \quad q_0^* = 9.27 \times 10^{-3}, \quad D = 5 \times 10^{-6} \text{ s}^{-1} \) and \( k = 0.20 \) is shown in Fig. 4. As required by (18), \( (q + l)_M \) is a weighted average of \( q(\rho_{se}) \) and \( q_0^* \). As is required by (17), \( h_M \) is a weighted average of \( h(\rho_{se}) \) and \( h_0^* - \Delta F_R / (\rho_0 C_T V_0) \). For the numerical values given above, \( \Delta F_R / (\rho_0 C_T V_0) = 5.089 \text{ kJ kg}^{-1} \), and it is the discontinuity in the radiative flux which allows \( h_M < h_0^* \) and hence a positive \( F_{se} \).

When compared from the viewpoint of convective fluxes, the similarities and differences between strato-cumulus and trade cumulus are evident. This is illustrated in Fig. 5. Fig. 5a shows vertical profiles of convective fluxes for the steady-state stratum-cumulus case shown in Fig. 4. \( LF_{se+1} \) exceeds \( F_{se} \), causing \( F_{se-1} \) to be negative. Fig. 5b shows the observed vertical profiles of the trade cumulus convective fluxes for five undisturbed days during BOMEX. These profiles were computed by Betts (1975) from the heat and moisture budget analysis of Nitta and Esbensen (1974) and the radiation budget analysis of Cox (1973). The profiles are 5-day averages. Instantaneous profiles often show a thinner trade inversion and larger vertical gradients of the fluxes in the trade inversion layer. The trade
cumulus situation has fluxes approximately 3-4 times as strong over a layer 3-4 times as deep as the strato-
cumulus situation. Except near its top, the trade cumulus subcloud layer has $F_{v-l1}$ positive, i.e., it 
has positive surface sensible heat flux and positive Bowen ratio. These subcloud layer differences 
are consistent with the role of radiation in each regime. In the strato-
cumulus regime the radiative cooling is confined to a 
thin layer at cloud top. In the trade cumulus regime 
radiative cooling extends throughout the cloud layer 
and subcloud layer, the subcloud layer radiative cooling 
being balanced by a vertical convergence of the 
convective flux of $(s - L1)$.

One major difference between the stratocumulus and 
trade cumulus regimes is the change of $h$ across cloud 
top. The air overlaying the stratocumulus deck has an 
h typically 4 kJ kg$^{-1}$ higher than the mixed layer air. 
The air overlaying the inversion above the trade cumulus 
has an $h$ typically 4 kJ kg$^{-1}$ lower than the cloud layer 
air. In the trade cumulus regime there is a large con-
vergence of the convective flux of $h$ at cloud top. In the 
steady state this is used to raise the $h$ of the mean 
subsiding air to its higher cloud layer value. Although 
radiation can play a role due to sharp changes in water 
vapor across the inversion, it is not essential for balance. 
The situation in the stratocumulus regime is different. 
The mean subsiding air must jump to a lower value 
of $h$, but the convergence of the convective flux of $h$ 
has just the opposite effect. It is through the discon-
tinuity in radiative flux that it is possible to maintain 
a steady state with $F_{v-l1} > 0$ and $\Delta h > 0$.

The dependence of the steady-state solutions on sea 
surface temperature, large-scale divergence and en-
trainment parameter $k$ is shown in Fig. 6. Many of the 
features of Fig. 6 can be understood from a considera-
tion of the minimum entrainment case. In this case, 
substitution of (17) and (18) into (19), with the aid of 
(3), (4), (10) and (21), yields

$$p_o - p_B = \frac{g \Delta F_{R}}{D(s_v(p_B) - s_{o}^*)},$$

where $s_v(p_B)$ is the virtual dry static energy just above 
p$_B$, and $s_{o}^*$ is the saturation virtual dry static energy at 
the ocean surface temperature. This is not an explicit 
relation for $p_B$. However, if $[s_v(p_B) - s_{o}^*]$ varies slowly 
with $p_B$ (the typical case) we can regard it as explicit. 
Then, for fixed sea surface temperature, the depth of 
the mixed layer is just proportional to $\Delta F_{R}$ and in-
versely proportional to $D$. If $\Delta F_{R}$ is fixed, a decrease 
in $D$ simply causes the height of the top of the mixed 
layer to increase until an identical subsidence rate is 
found. If $h(p_B)$ and $q(p_B)$ are slowly varying functions 
of $p_B$, the same subsidence rate implies the same $h_M$ 
and $(q + l)_M$ by (17) and (18), the same surface fluxes 
by (3) and (4), and the same cloud base by (7). All 
these features are evident in Fig. 6.

**Fig. 5.** Steady-state vertical profiles of stratocumulus convective fluxes (a) for the same case as shown in Fig. 4, and observed vertical profiles of trade cumulus convective fluxes (b) for five undisturbed days during BOMEX (after Betts, 1975).

For fixed $\Delta F_{R}$ and fixed $D$, an increase in sea surface 
temperature results in increases of $p_B$, $\rho_C$, $h_M$, $(q + l)_M$, 
$F_h$ and $F_{v-l1}$. While there is some dependence of the 
steady-state results on $k$, this dependence is not large, 
especially if atmospheric values of $k$ lie between 
approximately 0.1 and 0.3.

4. Time-dependent solutions

In this section we show some results of numerical 
integrations of Lilly's model, which consists of Eqs. (1)
through (7) and (16). As it stands this system is not convenient for numerical integration because (5) and (6) are both prognostic equations for $p_B$. In order that (5) and (6) predict $p_B$ in a consistent manner it is required that

$$\frac{L\Delta(q+I)}{\Delta h}(F_{h})_{B} - L(q+I)_{B} = \frac{L\Delta(q+I)}{\Delta h} \Delta F_{R}. \tag{24}$$

Eq. (5) can be written

$$\frac{\partial p_B}{\partial t} = \omega_{B} + \frac{(F_{h})_{B} - \Delta F_{R}}{\Delta h}. \tag{25}$$

We can now regard our system as composed of the five diagnostic equations [(3), (4), (7), (24) and (16)] and the three prognostic equations [(1), (2) and (25)].
The actual sequence of calculations involved in going from one step to another is as follows:

(i) Compute the surface fluxes \((F_h)_0\) and \((F_{q+1})_0\) from (3) and (4).
(ii) Compute the cloud base \(p_c\) from (7).
(iii) Solve (24) and (16) for \((F_h)_B\) and \(L(F_{q+1})_B\).
(iv) Compute the tendencies of \(h_{v}\) and \((q+l)_M\) from (1) and (2), and the tendency of \(p_B\) from (25).

This procedure is straightforward except for step (iii). When beginning this step we regard \(p_B\), \(p_c\), \(\Delta h_{S}\), \(\Delta (q+l)\), \((F_h)_0\), \(L(F_{q+1})_B\) and \(\Delta F_B\) as known. Then the consistency relation (24) and the entrainment relation (16) are regarded as two equations in the unknowns \((F_h)_B\) and \(L(F_{q+1})_B\). These equations take the form

\[
\begin{bmatrix}
a_{11} & a_{12} \\
a_{21} & a_{22}
\end{bmatrix}
\begin{bmatrix}
(F_h)_B \\
L(F_{q+1})_B
\end{bmatrix}
= \begin{bmatrix}
b_1 \\
b_2
\end{bmatrix},
\]

(26)

However, because of the form of (16) the coefficients \(a_{12}\) and \(a_{22}\) and the constant \(b_2\) are unknown until \((F_h)_B\) and \(L(F_{q+1})_B\) are known, i.e., until the location of the minimum \(F_{ve}\) is known. Our procedure is to assume that the minimum \(F_{ve}\) occurs at \(p_B\), solve (26) for \((F_h)_B\) and \(L(F_{q+1})_B\), then using (10), (14) and (15) check to see if the minimum \(F_{ve}\) actually occurs at \(p_B\). This procedure is repeated for \(p_c\), \(p_c^2\) and \(p_0\). If one and only one of the four possibilities does not lead to a contradiction, we have found the unique solution. It is possible that no solution exists or that more than one solution exists. In the numerical integrations reported here we have not encountered a problem with existence or uniqueness. However, the special case \(k=0\) requires extra care. As can be seen from (16), if \(k=0\) and the minimum \(F_{ve}\) at \(p_0\) case is attempted, both \(a_{12}\) and \(a_{22}\) vanish, causing the determinant of the coefficients in (26) to vanish. In other words, the minimum entrainment condition applied at \(p_0\) does not close the theory in the time-dependent case. Another interpretation of this difficulty is that there is no guarantee that the minimum entrainment requirement \((F_{ve})_0=0\) will be consistent with (3) and (4).

Perhaps the most interesting time-dependent behavior of the model is that resulting from diurnal variations in radiation. We thus proceed to a brief discussion of the diurnal variation of \(\Delta F_B\).

The radiation calculation performed by Lilly is based on considering the cloud to be a blackbody. Lilly estimates that the blackbody assumption is valid if the cloud layer is approximately 125–150 m thick. If this condition holds, the net upward radiative energy flux off the cloud top consists of the upward blackbody flux at the cloud top temperature minus the downward longwave flux from the atmosphere above, minus the solar radiation absorbed by the cloud.

In the time-dependent problem the radiation calculations and mixed layer calculations should be coupled. For example, as \(p_B\), \(h_M\) and \((q+l)_M\) change, the cloud top temperature and hence the upward blackbody flux change. In addition, as \(p_0\) changes, the downward longwave flux from the atmosphere above changes.

This closure problem does not occur in the steady state. However, in that case a minimum at \(p_0\) can also be interpreted as a minimum at \(p_c^2\).
Rather than make such coupled calculations we have chosen the simple alternative of making $\Delta F_B$ a specified function of time, which takes the form

$$\Delta F_B = 90.00 - 69.77 \max\left\{ 0.202 + 0.779 \cos\left(\frac{2\pi t_d}{24}\right) \right\} \quad [\text{W m}^{-2}], \quad (27)$$

where $t_d$ is the time of day measured in hours from local noon. The first term is the net upward longwave flux, while the second term is the solar radiation absorbed by the cloud. At 0500 ($t_d = -7$) and 1900 ($t_d = 7$) the absorbed solar radiation is zero. Such sunrise and sunset hours are found at $33^\circ$N in mid-July.

The model was integrated using the classical fourth-order Runge-Kutta scheme with a time step of 10 min. Any reasonable initial condition could be used if several days were allowed for the variables to begin to repeat themselves.

Fig. 7 shows the results of a typical run with a divergence of $5.0 \times 10^{-6}$ s$^{-1}$, a sea surface temperature of $13^\circ$C, and an entrainment parameter $k = 0.20$.

The diurnal cycle can be described as follows. $(F_B)_B$ and $L(F_{eq1})_B$ follow very closely the dip in $\Delta F_B$. Just after sunrise $(F_B)_B$ becomes less than $(F_B)_0$ and $L(F_{eq1})_B$ becomes less than $L(F_{eq1})_0$. This means that just after sunrise $h_M$ and $(q+l)_M$ begin to increase. Since the ocean surface temperature is fixed, the increase in $h_M$ and $(q+l)_M$ is accompanied by a decrease in $(F_B)_0$ and $L(F_{eq1})_0$. Just before 1600 $(F_B)_B$ becomes greater than $(F_B)_0$ and $L(F_{eq1})_B$ becomes greater than $L(F_{eq1})_0$. Then $h_M$ and $(q+l)_M$ increase throughout the evening until just after sunrise the following day. The daytime rise in $h_M$ and $(q+l)_M$ and the daytime fall in $(F_B)_0$ and $L(F_{eq1})_0$ last approximately 10 h, while the corresponding evening and nighttime changes last approximately 14 h. Also shown in Fig. 7a is the variation of $(s-L)_M$, which can also be regarded as a measure of the variation of surface air temperature. The surface air temperature falls slightly from morning to afternoon (about 0.14°C), and this tends to slightly lower cloud base. This effect is greatly augmented by the moistening of the layer, the result being that the cloud base is lowered $\sim 130$ m. Although the cloud top is also lowered, its variation is not as large, the result being that the cloud layer thickens $\sim 70$ m from morning to afternoon. Apparently, the concept of the sun “burning off the stratus” is not valid in the present situation.

We now turn to the question of how well the results of the diurnal computation agree with observation. Neiburger et al. (1961) were apparently the first to notice diurnal variations of the stratocumulus over the open ocean. They obtained their observations off the
California coast and summarized them as follows:

“A surprising feature first revealed by the 1949 data is the apparent diurnal variation of inversion height over the entire area, even 500 miles from shore. This was detected by comparing each observation with the average of the ones taken twelve hours before and after. In 52 of the 68 instances in which such a comparison could be made, the morning (local time) inversion heights were higher than the average of adjacent evening ones.”

This statement is in agreement with the results shown in Fig. 7.

Kraus (1963) studied the twice daily (0300 and 1500 local time) soundings from weather station N (30°N, 140°W) for five July months from 1958 to 1962. After locating, for each sounding, the pressure level at which the temperature has a minimum (we shall call this level $p_a$), Kraus computed the second-order difference

$$F = p_a(0300) - \frac{1}{2}[p_a(1500 \text{ preceding afternoon})$$

$$+ p_a(1500 \text{ following afternoon})],$$

for each 0300 sounding. If the mixed layer is deeper at night than during the day, $p_a(0300)$ will be smaller than the average of the two neighboring soundings, and $F$ will be negative. Out of a total of 136 computations of $F$, Kraus found 40 with $F > 0$ and 96 with $F < 0$ and thus concluded that the mixed layer at weather station N in July tends to be deeper at night than during the day. The results in Fig. 7 are also in agreement with Kraus’ observational study.

5. Conclusions

We have examined both steady-state and time-dependent aspects of Lilly’s stratocumulus model. The steady-state results are not highly dependent on the entrainment parameter $k$ and show a deepening and moistening of the layer as divergence is decreased and sea surface temperature is increased. Numerical integration of the model with diurnally varying radiation off cloud top has revealed significant diurnal variations in the state of the mixed layer and in the mixed layer convective fluxes.

It is possible to expand the present study in several directions. The radiation calculation could be coupled to the mixed layer computations. In addition, the atmospheric boundary layer model could be coupled to an oceanic boundary layer model. Another important possibility is to relax the assumption of horizontal homogeneity and to include a $\mathbf{v} \cdot \nabla p_a$ term in (25), a $\mathbf{v} \cdot \nabla b_{hm}$ term in (1) and a $\mathbf{v} \cdot \nabla (q + b)_{hm}$ term in (2). Computation could then be made on a two-dimensional grid of points covering the eastern North Pacific Ocean. For the rapid horizontal variations in sea surface temperature and large-scale divergence typical of coastal California, a rough calculation shows that these horizontal advection terms are probably not negligible.

Lilly’s stratocumulus model applies to completely overcast conditions while cumulus parameterization theories generally assume the fractional area of cloudiness is small. It remains a challenging problem to develop a general parameterization theory which can handle stratocumulus, trade cumulus, and the continuous transition from one to the other.

Finally, the time integration of the stratocumulus model has produced diurnal variations for which there is apparently little present observational data. It would seem reasonable to design an observational program with the specific purpose of testing the predictions of the theory. This could perhaps be done with a single well-instrumented aircraft (e.g. Lenschow, 1973).

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