PARAMETERIZATION OF THE INTERACTIONS BETWEEN BOUNDARY-LAYER CLOUDS AND BOUNDARY-LAYER TURBULENCE

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1. INTRODUCTION

Boundary-layer clouds are those that are in direct turbulent communication with the earth's surface; they include stratus, stratocumulus, and shallow cumulus clouds. They are particularly characteristic of the eastern subtropical oceans (Neiburger, 1960; Bugaev, 1973; Agee and Dowell, 1973; Arakawa, 1974), the Arctic ocean in summer (Herman and Goody, 1976; Curry and Herman, 1985), and the cold outbreaks associated with extratropical cyclones (Lenschow and Agee, 1976; Chou and Atlas, 1982; Stage, 1983). From the perspective of weather forecasting, boundary-layer clouds are important because of their high frequency of occurence (Hahn *et al.*, 1982, 1984), and their impacts on aviation, agriculture, military activities, and recreation. From the perspective of climate modeling, they are important because they tend to cool the planet by reflecting the incoming solar radiation without diminishing the outgoing terrestrial radiation; and also because they act to deepen the moist layer that feeds through the trades to fuel the tropical precipitation centers. The last ten years have seen an increasing recognition of the need for a boundary-layer cloud parameterization suitable for use in general circulation models (e.g., Arakawa, 1974). The most important element of such a parameterization is the coupling between the boundary-layer clouds and the boundary-layer turbulence.

A variety of approaches to parameterization have emerged, including empirical methods (Slingo, 1980; Ramanathan and Dickinson, 1980) and simple physical models (Randall, 1976; Suarez et al., 1982); Randall et al., 1985). All of these efforts have been strongly influenced by the paper of Lilly (1968), who proposed a simple but elegant theoretical model for the cloud-topped marine layers that are found over the cool waters of the eastern subtropical oceans, where largescale sinking motion occurs in association with the subtropical highs. Lilly used the empirically supported assumption that these layers are well mixed in their conservative thermodyanmic properties. The mixing is due to small-scale turbulence, which is maintained by buoyant convection in a uniformly saturated cloud layer. The convection is driven primarily by radiative cooling, which is concentrated at cloud top. The cloud-top level is also marked by a strong upward increase in temperature and a sharp upward decrease in absolute humidity. Turbulent entrainment progressively incorporates the dry warm inversion-layer air into the clouds. The large-scale sinking motion carries free atmospheric air downward to cloud top, where its properties are abruptly transformed during the entrainment process: it is cooled, moistened, and horizontally accelerated as it descends through a vertical distance of just a few meters. In a steady state or time average, the mass budget of the marine layer balances entrainment against large-scale divergence; the moisture budget balances surface evaporation against entrainment drying; the temperature budget balances radiative cooling primarily against entrainment warming; and the momentum budget balances the ageostrophic wind against surface drag and entrainment of free-atmospheric momentum. The entrainment rate itself is that required to allow a balance between production of turbulence kinetic energy (TKE) by convection and destruction of TKE by dissipation and buoyant comsumption. Turbulence, and particularly turbulent entrainment, thus plays a key role in all aspects of the boundary-layer cloud's evolution. Lilly's model has been evaluated in light of a number of field studies of boundary-layer clouds that have been conducted during the past ten years, including AMTEX (Lenschow and Agee, 1974), the California Stratus Experiment (Brost et al., 1976; Albrecht et al., 1985), JASIN (Royal Society of London, 1983), EPOCS (Hanson, 1983), the Cardington Experiment (Roach et al., 1982; Caughey et al., 1982; Slingo et al., 1982), and the North Sea measurements reported by Nicholls (1984). In addition, satellite observations have documented the very widespread occurance of boundary layer clouds, and have revealed important spatial and temporal variations that escaped the notice of groundbased observers (Bugaev, 1973; Minnis and Harrison, 1984a, b, c). On the whole, the observations show that Lilly's model sucessfully captures many of the most basic characteristics of cloudy boundary layers. It applies not only to the subtropical marine stratocumulus regimes that are so prominent off the west coasts in summer, but also to the stratocumulus layers that accompany cold-air outbreaks off the east coasts of North America and Asia in winter. The observations confirm that entrainment plays the important role envisioned by Lilly; that buoyant convection is concentrated in the cloud layer, just as the model predicts; and that radiative cooling is indeed strongly peaked near cloud top. At the same time, however, the observations make it clear that several additional physical ingredients are needed to improve the model's realism. These ingredients include partial cloudiness, which practically every satellite image reveals on both the micro and mesocales (Coakley and Bretherton, 1982); drizzle, which can produce a downward moisture flux comparable to the upward turbulent moisture flux (Brost et al., 1982a, b; Nicholls, 1984; Albrecht et al., 1985); shear generation of turbulence kinetic energy, which can be comparable in importance to convective generation (Brost et al., 1982a, b); solar radiation, which produces significant daytime warming in the interior of the cloud layer, and can lead to strong diurnal effects (Nicholls, 1984); and incomplete mixing, which is often particularly noticeable in the moisture sounding (Roach et al., 1982). Further observations are planned for the near future as part of FIRE (the First ISCCP Regional Experiment; ISCCP is the International Satellite Cloud Climatology Project); a discussion of these plans is given by Randall et al. (1984).

The body of this paper begins with a discussion of the mechanisms that produce turbulence in boundary-layer clouds. Special emphasis is placed on the cloud-top top level, where many fast processes operate simultaneously. Then we examine the ways in which the turbulence feeds back to modify the boundary-layer cloud's structure. After this preparation, parameterization for GCMs is discussed, beginning with an examination of how boundary-layer clouds are represented (or not) in the current ECMWF model, and continuing with a description of the boundary-layer cloud parameterization in the UCLA model. The paper closes with some suggestions concerning future parameterization efforts.

Boundary-layer clouds that occur in a conditionally unstable layer are discussed by M. Tiedtke elsewhere in this Seminar Proceedings; such clouds will not be considered here.

2. WHY CLOUDS ARE TURBULENT

2.1 The phenomenology of cloud top: A parameterizer's nightmare

The typical observed situation near the top of a boundary-layer cloud field can be summarized as follows. A strong upward increase in temperature is accompanied by a rapid upward decrease in mixing ratio, and intense vertical shear of the horizontal wind. Locally, the temperature may increase upward by one degree per meter! The cloud-top itself is bumpy, and in some cases its downward excursions can penetrate through the entire cloud layer. The cloud is then said to be broken. The bumpiness of cloud top reflects the turbulent character of the air within the cloud layer. This implies that the cloud fraction is closely related to the turbulence within the cloud layer. The turbulence is effectively confined below cloud top; at higher levels it is throughly quenched by the strongly stable stratification. For practical purposes, the cloud top and the boundary layer top coincide.

The upper layer is so dry that, in spite of its warmth, it emits longwave radiation only weakly downward towards cloud top. The clouds themselves are nearly opaque to infrared radiation, and so they emit strongly upward. Therefore, a short distance above cloud top the net long-wave flux is strongly upward. A short distance below, inside the clouds, the net flux is practically zero. Since the net upward radiation increases sharply with height across the cloud-top region, there is a strong radiative cooling there. Cooling rates on the order of 10 K hr^{-1} have been observed (Singlo et al., 1982). The horizontally averaged thickness of the radiatively cooled layer depends on both the bumpiness of cloud top and the optical extinction length associated with cloud absorption of infrared radiation. Observations show that this thickness can range from about ten meters to several hundred meters, depending on conditions (e.g., Platt, 1976; Curry and Herman, 1985).

To a large extent, it is the concentrated radiative cooling that accounts for the rapid downward decrease of the temperature near cloud top. In other words, cloud-top radiative cooling promotes the formation of the temperature inversion. For reasons to be discussed later, the presence of the inversion is favorable for the continued existence of the cloud. There is thus a kind of symbiotic relationship between the cloud and the inversion.

The temperature fluctuations associated with the turbulent eddies in the cloud layer are on the order of 0.5 K; the concentrated radiative cooling near cloud top can produce temperature changes of this magnitude in about 10^2 seconds. Since the lifetimes of the largest eddies are on the order of 10^3 seconds, the cooling can very readily influence the eddy dynamics.

In summary, the cloud top is a region of complex and very vigourous interactions among turbulence, radiation, and evaporative cooling, compressed into a layer only about 100 meters thick. It is quite impossible to explicitly resolve these interactions in a general circulation model. A parametric approach is absolutely necessary.

2.2 Turbulence kinetic energy sources and sinks

Not only boundary-layer clouds, but practically all clouds are turbulent. The reasons can be seen by exmaining the equation for conservation of turbulence kinetic energy (TKE), which can be written as

$$\rho \frac{\partial e}{\partial t} + \frac{\partial Tr}{\partial t} = \frac{g \, F_{sv}}{c_p \, T} + \underset{\sim}{F}_V \cdot \frac{\partial \underset{\sim}{V}}{\partial z} - D_s \tag{1}$$

Here ρ is density; e is the TKE per unit mass; Tr is the upward flux of TKE due to pressure-velocity correlations and triple-velocity correlations; F_x represents the upward turbulent flux of a quantity x, i.e.,

$$F_x = \rho \, \overline{w'x'} \quad ; \tag{2}$$

g is the acceleration of gravity; s_v is the virtual dry static energy; and D_s is the rate of dissipation per unit mass. The virtual dry static energy is given by

$$s_v = c_p \, T_v + gz \tag{3}$$

where z is height; and T_v is the virtual temperature, defined by

$$T_v \equiv T(1 + qR_w/R_d)/[1 + (q + \ell)] \ , \tag{4}$$

where T is temperature, q is the mixing ratio of water vapor, R_w and R_d are the respective gas constants for water vapor and dry air, and ℓ is the mixing ratio of liquid water. For simplicity, advection by the mean flow is neglected in (1), and the possibility of ice is ignored in (4). A derivation of (1) is given (in slightly different form) by Monin and Yaglom (1971). According to (1), TKE can be generated by either buoyant convection ($F_{sv} > 0$) or by (down gradient) turbulent momentum transport in the presence of shear. Both strong convection and intense shear commonly accompany cloud systems.

Shear of the mean horizontal wind at cloud top, whose importance has been emphasized by Brost et al., (1982a, b), simply reflects an approximate thermal-wind balance in the presence of a sloping density discontinuity (the inversion). Horizontal gradients in the inversion height are bound to be very common, and the observed strong shear is a straightforward consequence. There are obvious analogies between the dynamical balances at cloud top and those that occur in fronts. Brost et al. (1982a,b) have emphasized the importance of shear for the cases observed in the California Stratus Experiment, although Nicholls (1984) found that horizontally homogeneous cloud sheets over the North Sea were associated with very little shear.

The buoyancy term of (1) represents the effects of convection. Both latent heat effects and radiation promote convection in clouds. Radiation destabilizes by cooling the cloud top through longwave emission, and warming the cloud interior through solar absorption.

The virtual dry static energy flux (hereafter, the buoyancy flux) depends in an interesting way on the cloud amount. In order to explore this dependence, it is necessary (though perhaps a bit painful) to introduce a number of thermodynamic variables and some relations among them. We consider turbulent fuctuations on isobaric surfaces, and neglect the very small contributions to the static energy fluctuations from height fluctations on these surfaces. The virtual dry static energy flux is then just times the virtual temperature flux.

To sufficient accuracy, fluctuations of virtual temperature satisfy

$$T_v' = T' + \overline{T}\,(\delta q' - \ell') \; , \tag{5}$$

where $\delta \equiv (R_w - R_d)/R_d \simeq 0.608$. The buoyancy flux can then be obtained from

$$F_{sv} = c_p\,F_T + \varepsilon L\,(\delta F_q - F_\ell) \; . \tag{6}$$

Here $\varepsilon \equiv c_p\,T/L$, and L is the latent heat of condensation, whose temperature dependence will be neglected for simplicity. According to (6), the buoyancy flux includes contributions not only from the flux of temperature itself, but also from the fluxes of water vapor and liquid water. An upward water vapor flux contributes positively to the buoyancy flux, because water vapor is lighter than dry air; but an upward liquid water flux contributes negatively, because liquid water is much more dense than dry air.

We now introduce the moist static energy, which is given by

$$h = c_p\,T + qz + Lq \; , \tag{7}$$

and the total mixing ratio, which is denoted by

$$r \equiv q + \ell \; . \tag{8}$$

The moist static energy is approximately conserved under both moist adiabatic and dry adiabatic processes. The total mixing ratio is simularly conserved, provided that any liquid water drops are small enough to be carried along with the motion of the air. This condition is violated in the presence of drizzle; a correction has been suggested by Nicholls (1984), but will not be considered here. By use of (7) and (8) in (6), we can express the buoyancy flux in terms of the fluxes of moist static energy, total mixing ratio, and liquid water:

$$F_{sv} = F_h - (1 - \delta\varepsilon)L\,F_r + [1 - (1 + \delta)\varepsilon]L\,F_\ell \; . \tag{9}$$

This result aplies whatever the fractional cloudiness may be. If there is no cloud at all, the liquid water flux vanishes, so (9) simplifies to a relationship between the buoyancy flux and the moist static energy and total water fluxes, i.e.

$$(F_{sv})_{clear} = F_h - (1 - \delta\varepsilon)\, L\, F_r. \qquad (10)$$

Lilly (1968) showed that a similar simplification can be made for case of a uniformly cloudy layer. To do this, he used the important and empirically verified assumption that within cloudy air the vapor mixing ratio, q, is equal to the saturation mixing ratio, q*. In that case, water vapor fluctuations on isobaric surfces are proporational to temperature fluctuations, and from this it follows that the flux of water vapor is proportional to the moist static energy flux:

$$LF_q = \gamma F_h/(1 + \gamma) \quad . \qquad (11)$$

Here γ is defined by

$$\gamma \equiv \frac{L}{c_p} \left(\frac{\partial q^*}{\partial T}\right)_p \qquad (12)$$

An expression for γ can be derived from the Clausius-Clapeyron equation. If (11) holds everywhere throughout the layer, i.e. if the layer is uniformly cloudly, then it is possible to express the flux of liquid water in terms of the fluxes of moist static energy and total mixing ratio:

$$LF_\ell = LF_r - \left(\frac{\gamma}{1 + \gamma}\right) F_h \qquad (13)$$

By use of (13) in (9), we can show that in a uniformly cloudy layer the buoyancy flux satisfies

$$(F_{sv})_{cloudy} = \beta F_h - \varepsilon\, L\, F_r \quad , \qquad (14)$$

where

$$\beta \equiv [1 + (1 + \delta)\, \gamma\varepsilon]/(1 + \gamma) \quad . \qquad (15)$$

Comparing (10) and (14), we see that F_{sv} can be expressed as a linear combination of F_h and F_r for both clear and fully cloudy layers. What about partly cloudy layers? To analyze this case, it is necessary to adopt a model for the distributions of temperature, water vapor, and liquid water on isobaric surfaces within the cloud layer. This is a subject of current research.

2.3 Entrainment

By definition, entrainment is the active advance of turbulence into a non-turbulent region. It can and often does occur in the absence of clouds, but for reasons that will become clear it tends to be particularly vigorous when clouds are present. The advancing boundary of an entraining turbulent region is typically quite sharp and well-defined. There is a net flux of mass across the boundary because it is moving into the quiet air, relative to the mean flow. This is the entrainment mass flux, E. The air that crosses the boundary becomes turbulent, i.e. it acquires turbulence energy. At the same time its other properties are often transformed as well. Air that is entrained into a boundary layer cloud is typically moistened, cooled, and horizontally accelerated. These transformations are caused by sharp convergence of the turbulent fluxes of moisture, sensible heat, and momentum. For example, the turbulent moisture flux decreases sharply upward at the top of a stratocumulus layer. This moisture flux convergence deposits moisture in the entrained air, increasing its mixing ratio from dry free atmospheric values to a humid boundary-layer values. Mathematically, this can be expressed by

$$E\ \Delta r = -(F_r)_B \quad , \tag{16}$$

where Δr is the difference in mean mixing ratio between the free atmosphere and the cloud layer, and subscript B denotes a level slightly below the boundarylayer top. The moisture balance expressed by (16) is valid for a sufficiently thin layer. A derivation is gien by Lilly (1968). Similar balances for moist static energy and momentum are expressed by

$$E\Delta h = -(F_h)_B + \Delta R \quad , \tag{17}$$

and

$$E\Delta\underset{\sim}{V} = -(F_{\underset{\sim}{V}})_B \quad , \tag{18}$$

respectively. In (17), R is the net upward radiation, and so $\Delta R > 0$ corresponds to radiative cooling. For an active subtropical stratocumulus layer, typical values are as follows:

$E \sim 0.05$ kg m^{-2} s^{-1}, $\Delta r \sim -3$ g kg^{-1}, $\Delta h/c_p \sim 5$ K, $\Delta R \sim 70$ W m^{-2},
$|\Delta\underset{\sim}{V}| \sim 5$ m s^{-1}.

By using (16-17) in (14), we find that at the top of an overcast boundary layer the buoyancy flux satisfies

$$(F_{sv})_B = -E\ (\beta\Delta h - \varepsilon L\Delta r) + \beta\Delta R \quad . \tag{19}$$

The problem of determining the entrainment rate is still controversial and is badly in need of observational input. Most of the entrainment parameterizations that have been suggested are based on an assumed balance between certain terms of the turbulence kinetic energy equation discussed in Section 2.2. The two most widely used parameterizations are "Eulerian partitioning", proposed by Kraus and Schaller (1978); and "Process partitioning," proposed by Manins and Turner (1978) and Stage and Businger (1981a, b). Randall (1984) has recently discussed the relationships between these two theories. Observational work and/or large-eddy simulations are needed to test these ideas, or to suggest new approaches.

3. FEEDBACK LOOPS: HOW TURBULENCE AFFECTS THE CLOUDS

The importance of turbulent entrainment in the mass, temperature, moisture, and momentum budgets of cloudy boundary layers has already been discussed. We now examine the key roles of entrainment in the dynamics of stratocumulus formation, deepening, and dissipation. Some interesting feedbacks are at work.

3.1 Thin-cloud instability

As discussed in Section 2, the opacity of the clouds with respect to infrared radiation leads to strong radiative cooling within the first hundred meters below cloud top. In a geometrically thin cloud, liquid water mixing ratios tend to be small, so the cloud is optically thin as well; it is also likely to be broken. For these reasons, the radiative cooling at the top of a thin cloud layer is weaker than that at the top of a thicker cloud layer, all other things being equal. An increase in the radiative cooling of the boundary layer tends to produce a decrease in the temperature and so an increase in the relative humidity. This means that once a thin cloud has formed, the cooling it produces will tend to deepen it with time. This is a kind of "bootstrapping" mechanism; Randall and Suarez (1985) have termed it "thincloud instability," or TCI. It applies only to thin clouds because once the cloud depth exceeds some threshold value, on the order of 100 meters, the cloud has become as optically thick as it can be; further deepening has little effect on the radiative cooling rate, and so the feedback mechanism no longer operates.

3.2 Cloud-top entrainment instability

Squires (1958) pointed out that air entrained at the tops of cumulus clouds is chilled and moistened by the evaporation of liquid water drops, as it mixes with the cloudy air around it, and that this cooling might cause the entrained air to become negatively buoyant, so that it sinks through the cloud under the action of the buoyancy force. Lilly (1968) recognized that the same process could be important in stratocumulus clouds, and he suggested that it would lead to rapid entrainment and a break-up of the cloud sheet. This process has come to be called cloud-top entrainment instability, or CTEI. Lilly suggested that CTEI occurs if the wetbulb potential temperature decreases upward at cloud top. This is equivalent to an upward decrease in the moist static energy. A sufficiently strong inversion could thus prevent CTEI. Randall (1976, 1980a) showed that a more accurate criterion for the onset of CTEI is that the upward increase in the virtual dry static energy at cloud top be less than a critical value, given by

$$(\Delta s_v)_{crit} = \left[\frac{1 - (1 + \delta)\varepsilon}{1 + \gamma}\right] L(q^*_{B+} - q_{B+}) \quad . \tag{20}$$

Here the subscript B+ denotes the air above cloud top. According to (20), $(\Delta s_v)_{crit}$ increases as the relative humidity of the air above cloud top decreases.

During the past few years, several observational and modeling studies have shed some light on the existence of CTEI, the conditions required for it to operate, and its effects on the cloud layer. Deardorff (1980a, 1980b) used a large-eddy model (a high-resolution three-dimensional numerical model capable of explicitly resolving the most important turbulent eddies) to simulate CTEI and its consequences for a stratocumulus deck. According to his model results CTEI does occur, it begins at the time predicted by the stability criterion discussed above, and it leads to a rapid decrease in the cloud fraction and a rapid transformation of the cloud-layer stratification from a saturated moist adiabat to an unsaturated conditionally unstable sounding. Moeng and Arakawa (1980) used a two-dimensional moist second-order closure model with 5 km horizontal resolution to simulate the cloud regimes that occur between the west coast of North America and the ITCZ. The orientation of the domain was along the observed tradewind flow, from northeast to southwest. They found a stratocumulus regime in the northeastern part of the domain, where the prescribed sea surface temperatures were low and the prescribed large-scale subsidence was strong. Toward the southwest, the gradually warming sea-surface temperatures and the gradually weakening large-scale subsidence led eventually to an abrupt transition in the cloud form, from solid overcast to broken clouds, accompanied by energetic mesoscale eddies. The transition occured where the CTEI stability criterion was first violated. Further numerical simulations of CTEI have been reported by Chen and Cotton (1983) and Mason (1985).

3.3 Cloud deepening through entrainment

Because the entrainment of warm dry air across cloud top tends to decrease the relative humidity at a given level in the marine layer, it tends to raise the cloud base level. At the same time, however, the entrainment tends to raise the cloud top level, and the possibility exists that the cloud top may rise faster than the cloud base. When this occurs, the entrainment of warm dry air actually tends to cause a thickening of the cloud with time! Randall (1984b) has discussed this counterintuitive process, and termed it "cloud deepening through entrainment", or CDE for short. As discussed in Section 2, the presence of deep cloud tends to favor rapid entrainment. This suggests that when CDE occurs, a positive feedback loop operates to deepen the cloud. Unlike TCI, this mechanism does not depend on an increase of cloud opacity with cloud depth, and so it can operate even for clouds that are already optically thick. Randall (1984b) presents some evidence that CDE is associated with particularly deep clouds in the UCLA GCM.

A further implication of CDE is that cloud-top entrainment instability need not necessarily lead to a destruction of the cloud layer or a decease in the fractional cloudiness. When CDE and CTEI occur together, the latter leads to rapid entrainment, just as originally envisioned, but the entrainment builds the cloud deck rather than destroying it. This suggests that solid cloud sheets can sometimes persist in the presence of inversions that are unstable with respect to CTEI.

3.4 Discussion

It seems likely that the three feedback mechanisms discussed in this Section help to determine where and when boundary-layer clouds form, deepen, and dissipate. The formation process is aided, if not initiated, by thin-cloud instability. Once the cloud has become optically thick, it will tend to become particularly deep if entrainment itself is favorable for deepening, i.e. if CDE occurs. This is especially true if CDE is accompanied by cloud-top entrainment instability. In the absence of CDE, however, it seems likely that CTEI drastically reduces the fractional cloudiness, and triggers the onset of shallow cumulus convection. When some unfavorable external agency such as unusually strong large-scale subsidence succeeds in thinning a cloud layer to a shadow of its former self, thin-cloud instability can finish the job by triggering a final catastrophic retreat of the boundary layer top to below the condensation level.

4. PARAMETERIZATION FOR GENERAL CIRCULATION MODELS

4.1. Specifications

On the basis of current observational knowledge and theoretical understanding, as summarized in the preceding Sections, it is possible to draw up specifications to be met by a satisfactory boundary-layer cloud parameterization. From an operational point of view, the parameterization should be capable of predicting cloud occurence, cloud fraction, the heights of cloud top and cloud base, and the fluctuations of these variables on diurnal, synoptic, and seasonal time scales. The physical mechanisms that must be parameterized to achieve these operational goals are as follows:

FIRST-ORDER SPECIFICATIONS:

- represent the effects of entrainment in the conservation equations for boundary-layer mass, temperature, mixing ratio, and momentum;
- include the effects of cloud-top radiative cooling on the bound-

ary-layer temperature and on the entrainment rate;

- take into account the role of CTEI in limiting the extent of stratiform clouds.

By incorporating entrainment and cloud-top radiative cooling, a parameterization makes it possible for the GCM to simulate the basic balances (discussed in Section 2) that allow a stratocumulus layer to maintain itself in the face of persistent large-scale sinking motion. It also becomes possible to simulate thin-cloud instability and cloud deepening through entrainment. Later in this Section, we present evidence that at least a simple parameterization of CTEI is needed to prevent excessive boundary-layer cloud amounts; this is the motivation behind the third specification listed above. Together, the first-order specifications allow at least a partially succesful simulation of the global distribution of boundary-layer clouds.

However, in order to represent the effects of fractional cloudiness and diurnal forcing, the following additional specifications must be met:

SECOND-ORDER SPECIFICATIONS:

- take into account the effects of small-scale inhomogeneities in cloud water on the radiation field;
- include the effects of partial cloudiness on the liquid water and buoyancy fluxes;
- represent the effects of solar absorption on the entrainment rate.

No existing parameterization fulfills any of these second-order specifications.

4.2 Boundary-layer clouds in the ECMWF model

In the ECMWF model, vertical diffusion can operate at all levels, but a thin surface layer and two additional layers in the lowest kilometer are provided in order to crudely resolve the boundary layer, and the effects of vertical diffusion are particularly important for these lowest model layers. Surface transfer coefficients and eddy diffusion coefficients for sensible heat, moisture, and momentum are predicted, using the Richardson-number theory developed by Louis (1979). The depth of the PBL is not explicitly determined, but can be inferred from the vertical profiles of the diffusive fluxes. Cloudiness in the PBL is detected using an empirical scheme based or ideas developed by J. Slingo (1980). Partial cloudiness is permitted, and is taken into account in the radiation parameterization. Because the boundary layer clouds influence the solar and terrestrial radiation parameterazation, they influence the time evolution of the temperature sounding. In this way, they can indirectly affect the vertical diffusion coefficients through the Richardson number, but only on space and time scales that are explicity resolved by the model. There is no direct, fast, parameterized coupling between the radiative effects of the clouds and the boundary-layer turbulence parameterization.

Cloudiness and latent heat effects are not directly linked in the ECMWF model; it is possible to have cloudiness without any latent heat effects, and vice versa. The turbulence parameterization does not recognize latent heat effects, except indirectly through the effects of latent heat on the temperature sounding and Richardson number. Dry conservative variables such as potential temperature are conserved during mixing, even when saturation occurs in the turbulent air. There is no direct, fast, parametric coupling between latent heat effects and the boundary-layer turbulence parameterization.

The model is quite capable of simulating a deepening of the PBL, e.g. as forced by surface heating. However, it cannot be said to simulate entrainment. To reconcile these seemingly contradictory statements, consider how the turbulent layer will deepen in response to a warming of the ground below an initially stable sounding, and for simplicity suppose that the vertical resoluton is infinite. At t=0, a well mixed (isentropic) layer rests below an inversion, which is surmounted by the stable free atmosphere. A warming of the ground tends to destabilize the PBL, and vertical mixing redistributes the sensible heat so as to maintain an isentropic structure below inversion. However, Richardson number effects prevent turbulent exchange from modifying the air above the inversion until the potential temperature of the ground (and the mixed layer) increases to equal that of the free atmosphere above the inversion. After this stage has been reached, the effects of the turbulence advance continuously to higher levels, in such a way that at all times the potential temperature of the ground is equal to that of the PBL top. This process has been termed "encroachment" (Carson and Smith, 1984). It is important to note that as long as a finite inversion exists, the turbulent layer does not deepen.

In a typical marine stratocumulus situation, the sea surface temperature is essentially fixed, and the inverison is very strong. Radiation tends to cool the layer with time, and large-scale sinking motion tries to push the inversion down to the sea surface. The turbulence parameterization of the ECMWF model will try to conteract the radiative cooling by removing sensible heat from the ocean, but the effects of the turbulence will not extend upward through the inversion because of the Richardson number effect, and so the model will be unable to resist the inexorable squashing of the PBL cloud layer by subsidence. The model cannot simulate the entrainment process in which the effects of the subsidence are balanced by very sharp turbulent flux convergences at the PBL top. We now examine the boundary-layer cloudiness simulated by the ECMWF model in 50-day "winter" and "summer" runs at T63 (runs BVH and BV9). The "winter" run begins January 17th, and means will be discussed for the month of February. The "summer" run begins June 15, and means will be discussed for the month of July. The global mean cloudiness for both runs is slightly less than 50%, which is perfectly satisfactory in view of current observations; and the zonal mean cloudiness distributions also appear to be quite reasonable, with a tropical maximum corresponding to the ITCZ, subtropical minima reflecting the desert regions, and further maxima at high latitudes.

Fig. 1 shows the distribution of total cloudiness simulated by the model in the last thirty days of the summer integeration. Unfortunately, the history tapes do not contain information about the vertical distribution of the simulated cloudiness, but it is clear that the model has not simulated the vast sheets of stratocumulus cloud that are observed to occur off the west coasts of North America, South America, South Africa, Europe, and Australia. There is a cloudiness maximum off the coast of California, but it is not as extensive as observed. Off the west coasts of South America and South Africa, there are very distinct cloudiness minima (less than 10%) just where maxima are observed to occur. In a sense, this is not surprising, because these are regions of large-scale sinking motion and cold sea surface temperatures. Some of us were taught in introductory meteorology courses of the distant past that such conditions are unfavorable for cloud formation. The ECMWF model seems to have taken this lesson to heart.

The winter total cloudiness results are shown in Fig. 2. For the marine subtropical stratocumulus regimes, anomalous minima are again simulated. The cloudiness maxima that are observed to be associated with the winter storm tracks near the east coasts of North America and Asia are displaced towards the northwest. The simulated storm-track precipitation maxima (not shown) are similarly displaced. These errors in the cloudiness and precipitation fields are consistent with the conclusions of Arpe and Klinker (1985), who focused on dynamical fields.

TOTAL CLOUD COVER, %
JULY MEAN

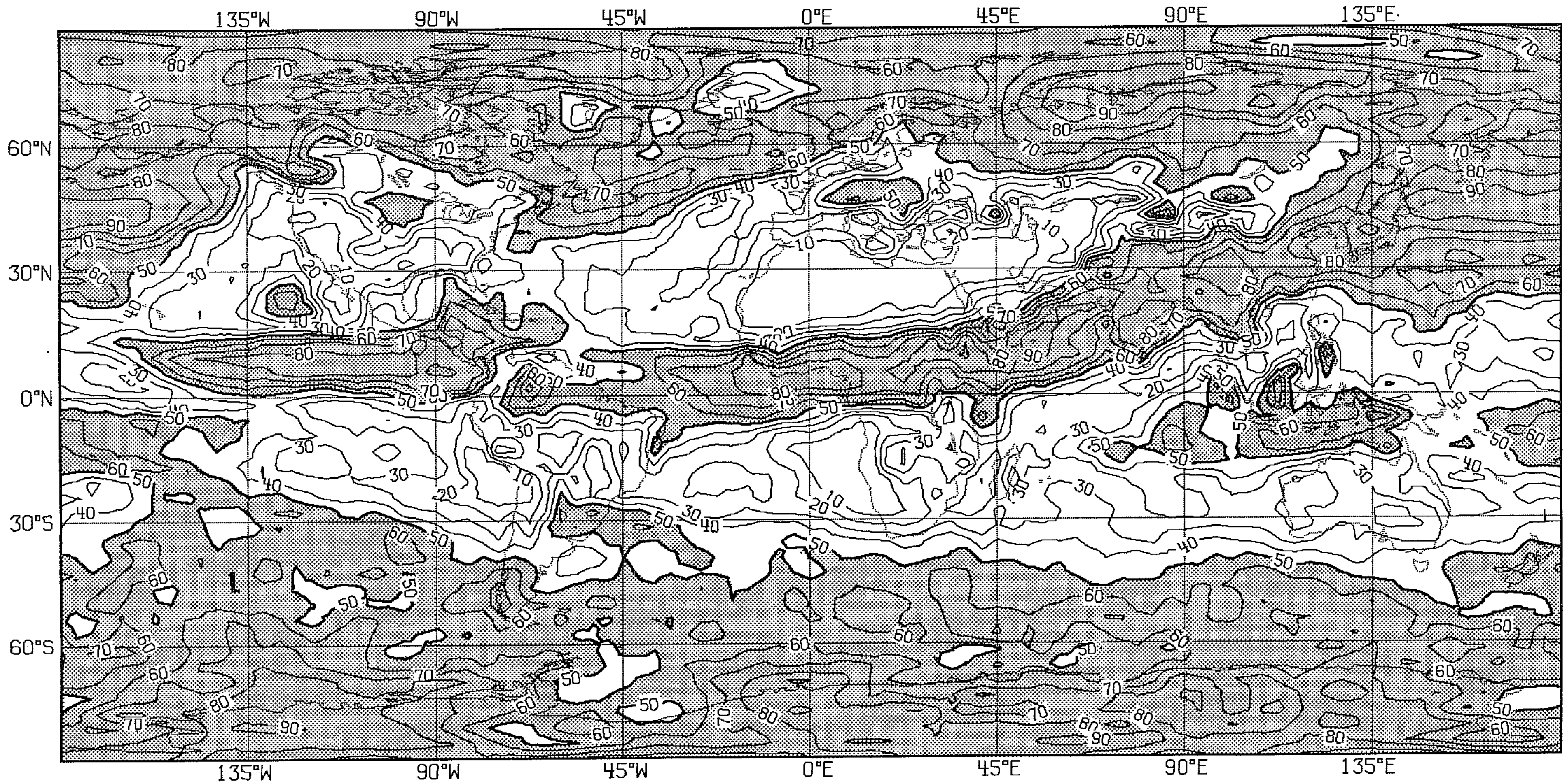

Figure 1. Global distribution of total cloudiness simulated by the ECMWF model for the month of July (the last 30 days of a 50-day run). Values higher than 50% are shaded. The contour interval is 10%.

TOTAL CLOUD COVER, %
FEBRUARY MEAN

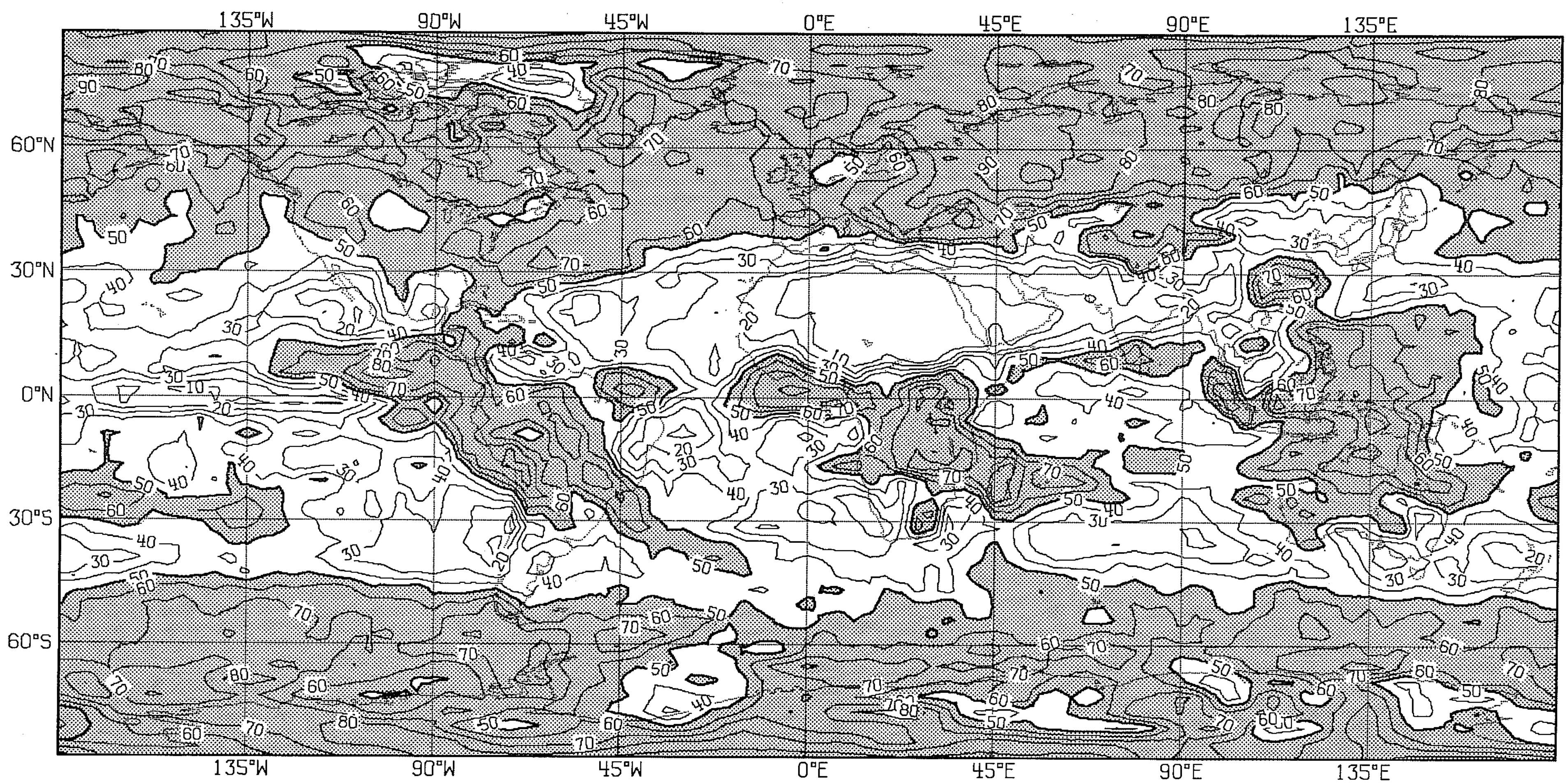

Figure 2. Global distribution of total cloudiness simulated by the ECMWF model for the month of February (the last 30 days of a 50-day run). Values higher than 50% are shaded. The contour interval is 10%.

Both the winter and summer runs produce large cloud amounts in the Artic; the observed strong summer maximum is not simulated.

An analysis of the ensemble-mean of 30 24-hour operational ECMWF forecasts for June 1985 (at T106) shows essentially the same cloudiness results as those discussed above for the summer run. Fig. 3 shows the changes in the temperature and moisture soundings that occur for a "California stratus" gridpoint (25° N, 120° W) in the first 24 hours of the summer run. The low-level temperature inversion is pushed down and the lowest layers are dryed out, under the influence of large-scale sinking motion. The model is unable to oppose the subsidence with entrainment. A realistically deep, cloudy maine layer cannot be sustained.

4.3 Boundary-layer clouds in the UCLA model

The formation of the stratocumulus parameterization in the UCLA GCM has recently been described in detail by Suarez et al. (1982), and an analysis of the results has been presented by Randall et al. (1985). The parameterization is based in part on bulk methods of the type discussed by A. Driedonks elsewhere in these Seminar Proceedings. Only a brief discussion of the scheme will be given here; and emphasis will be on the differences between the UCLA and ECMWF models, in terms of both formulation and results.

No attempt is made to explicitly resolve the vertical structure of the boundary layer. The vertically compact cloud-top entrainment layer cannot be resolved by any conceivable GCM; even large-eddy models that explicity resolve the PBL turbulence in a very limited region cannot currently resolve the entrainment process because of the sharpness of the inversion (Deardorff, 1980b). However, the dynamics of entrainment and the turbulent exchange are very strongly influenced by the inversion, so its effects must be in-

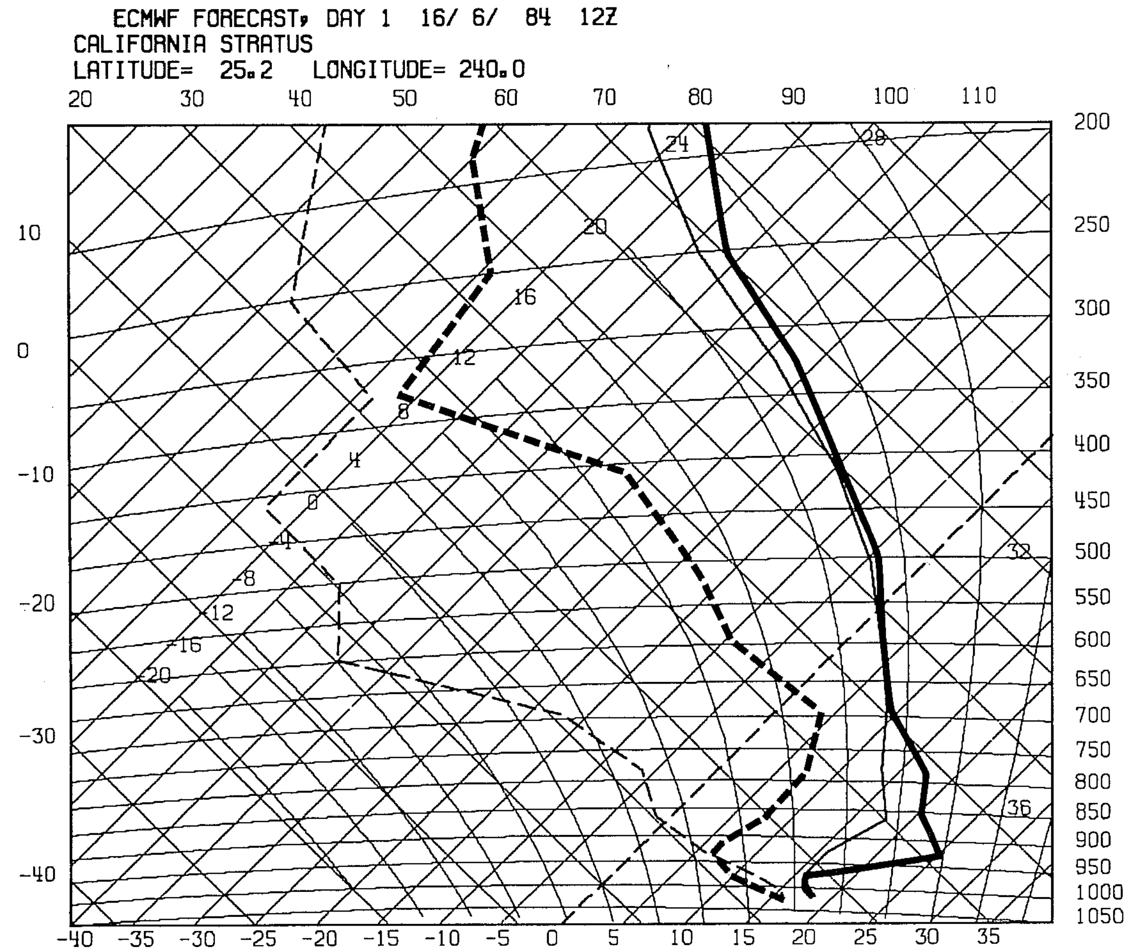

Figure 3. The initial and 24-hour forecast temperature and moisture soundings produced by the ECMWF model for a "California stratus" gridpoint (25° N, 120° W). The solid lines are temperature, and the dashed lines are dew point. The thin lines are the initial condition, and the thick lines are the 24-hour forecast.

cluded somehow. The only solution is to represent this part of the atmosphere's vertical structure parameterically. The parameterization used in the UCLA model is based on idealizing the rapid changes within the inversion as discontinuities. This approach was suggested for use in GCMs by Deardorff (1972). It is assumed that the moist static energy, total mixing ratio, and wind components are vertically uniform inside the PBL, as a result of turbulent mixing. This assumption makes it possible for the GCM to use only one model layer to represent the entire PBL. Without this assumption it would be necessary to parameterize the turbulent exchanges between layers within the PBL; this is at best a difficult task with low resolution, and an expensive calculation with high resolution. Ironically, the success of such efforts is often judged by a multilevel PBL model's ability to generate a well-mixed structure!

The entrainment rate is determined using the is determined using the "Eulerian partitioning" theory discussed by Randall (1984), and is used together with the large-scale convergence and the cumulus mass flux to predict the timeevolution of the PBL depth.

Partial cloudiness is currently not allowed; the PBL is assumed to be either overcast or cloud-free. The presence of a cloud layer is detected by checking for super saturation at the PBL top, using the well-mixed assumption. When clouds are present, the cloud-base level is determined as the level where the total mixing ratio is equal to the saturation vapor mixing ratio, again using the well-mixed assumption.

The PBL clouds interact with both the solar and terrestrial radiation parameterizations, thus affecting the radiative heating and cooling of the boundary layer and the earth's surface. Infrared cooling is taken into account in the entrainment parameterization, but solar warming is currently neglected there.

The effects of latent heating on the entrainment rate are included by computing the contribution of the liquid water flux to the buoyancy flux, using the methods discussed in Section 2.

The effects of CTEI are parameterized very crudely. The criterion discussed in Section 3 is used to detect the instability. When CTEI is predicted to occur, the PBL is assumed to mix with the free atmosphere until the instability is eliminated or the cloud is evaporated, whichever requires less mixing. The PBL depth is assumed to remain fixed during this mixing process, which therefore cannot be interpreted as entrainment. The possible effects of CDE are not considered.

Extensive boundary-layer cloudiness results obtained with the UCLA GCM have recently been published by Randall et al., (1985). Only a brief verbal summary will be given here. The model does produce marine subtripical stratocumulus clouds in approximately the correct locations, although they are much less extensive than observed. CTEI operates in the model to destroy the subtropical cloud sheets as the low-level flow carries them westward and equatorward. In the northern winter, low cloudiness accompanies cold outbreaks on the east coasts of North America and Asia. In the southern winter, a band of low cloudiness rings Antartica. The Arctic summer stratus regime is not simulated. In an experiment in which the CTEI parameterization was turned off, both the low-level cloudiness and the total cloudiness increased dramatically over the globe, to very unrealistic values.

5. CONCLUDING REMARKS

Many of the ideas discussed here were already present in the remarkable 1968 paper by Lilly. It is surprising the Lilly's paper received so little attention during the first several years after it appeared. Perhaps this was because those years saw particularly exciting developments in the glamorous field of cumulus parameterization; cumulonimbus towers are spectacular in a way that stratocumulus turrets can only envy. Ironically, recent studies at ECMWF (Tiedtke, 1983) have shown that the simulated distribution of cumulus precipitation is highly sensitive to the parameterization of boundary-layer clouds.

Although existing boundary-layer cloud parameterizations have demonstrated some skill, they have many very serious deficiencies. There are two kinds of "gaps" that separate our current parameterizations from those that we would like to have. First, there are gaps between the reality that exists in the atmosphere, and our imperfect understanding. Second, there are gaps between our understanding and the more limited ideas that are actually implemented in current parameterizations. Although the first kind of gap is undoubtedly the greatest obstacle to future improvement of some parameterizations, e.g. cumulus parameterization, the second kind of gap appears to be more important for boundary-layer cloud parameterization. Significant progress in the latter can be achieved simply by more complete application of knowledge already in hand. Specific suggestions for future research have been recently given by Randall (1985).

Some attempts have already been made to simulate the observed large-scale structure of the subtropical marine stratocumulus systems (Moeng and Arakawa, 1980; Wakefield and Schubert, 1981; Randall <u>et al</u>., 1985), but with only limited success. It appears, however, that the stage has been set for rapid further progress. The next ten years may well see satisfactory simulations of the marine subtropical stratocumulus regimes and their interactions with the large-scale circulation, as well as a greatly improved capability to numerically forecast the occurence and amount of boundary-layer clouds.

Although this paper has focused on the importance of turbulence for clouds that occur in the planetary boundary layer, it has also pointed out that practically all clouds are turbulent. Our current struggles to parameterize boundary-layer clouds are preparing us for future efforts to parameterize the interactions between turbulence and clouds within the free atmosphere.

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