The simulation of a fair weather marine
boundary layer in GATE
using a three-dimensional model

By S. NICHOLLS\(^1\) and M. A. LEMONE
National Center for Atmospheric Research\(^2\) Boulder, Colorado, 80307, USA
and G. SOMMERIA
Laboratoire de Meteorologie Dynamique, CNRS, Paris, France

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SUMMARY

A detailed three-dimensional model has been used to simulate a cumulus-topped tropical oceanic
boundary layer. A particularly dense set of observational data from surface-based and airborne instru-
mentation collected during one day of the GARP Atlantic Tropical Experiment (GATE) has been used
enabling comprehensive initial and boundary conditions to be specified. This also permits detailed compari-
sions to be drawn between the model and observations.

The model is found to reproduce the structure of the observed undisturbed boundary layer quite
realistically. The clouds developed in the model are similar to those observed in both amount and physical
size and the modelled turbulent fluxes in the subcloud layer agree well with observations. Also, the evolution
of the modelled boundary layer is similar to that in which the observations were made. The model confirms
that the major effects of the turbulent mixing are to warm the subcloud layer but to moisten and cool the
cloud layer. The limited scale of the model (2 km × 2 km × 2 km) does, however, lead to systematic dif-
ferences between measured and modelled variances.

The model’s turbulent kinetic energy balance for the coupled cloud-subcloud layers is presented.
Although reliable measurements are only available in the subcloud layer, they confirm that the model’s
values are qualitatively correct except when very close to the bottom boundary.

The model’s temperature and humidity profiles are strongly affected by mixing occurring across cloud-
base throughout much of the subcloud layer. This is reflected in the shape of the \(\theta_e\) profile which displays a
characteristic minimum midway through the subcloud layer. When scaled using cloudbase parameters,
the resulting profiles are similar to those obtained in more unstable conditions over land.

1. INTRODUCTION

In recent years considerable progress has been made in numerically modelling the
turbulent atmospheric boundary layer (e.g. Deardorff (1972, 1974), Wyngaard et al. (1974),
Sommeria (1976)). Models can produce a wealth of detail about the structure and behaviour
of boundary layers and are often used to test different ideas concerning the physical processes
occurring in boundary layers. However, there have been very few comparisons between
model predictions and observations although satisfactory agreement is clearly a vital
requirement if a model is to be used as a testbed for, say, new parametrization schemes.
There are a number of reasons why such comparisons have not been forthcoming which
stem basically from the difficulty in obtaining suitable turbulence measurements throughout
the region of interest and the computer resources required for each simulation. In addition,
‘ideal’ conditions (e.g. cloudfree, neutral or very strong buoyant mixing) are often chosen
for numerical simulation and they rarely correspond to those in which measurements can
readily be obtained.

The GARP Atlantic Tropical Experiment (GATE) provided a unique opportunity
to study the structure and evolution of the undisturbed tropical boundary layer over the
ocean by assembling a large, diverse set of instrumentation in a fairly small area. In addition
to the usual fixed point measurements, missions were flown with several aircraft whose

gust-probe systems made measurements of turbulent fluctuations throughout the depth
of the boundary layer. Because of the density of the measurements in both space and time,
especially those made by the aircraft, the data provide unparalleled opportunities for

\(^1\) Now at Meteorological Research Flight, Farnborough, UK
\(^2\) The National Center for Atmospheric Research is sponsored by the National Science Foundation.
the testing of small-scale turbulence simulation models. Results from some of these flights have already been discussed by Nicholls and LeMone (1980) with particular reference to the interaction between shallow cumulus convection and the subcloud layer.

The purpose of this paper is first to describe, in some detail, the simulation of one particular undisturbed period during GATE (on 10 September 1974) using the model developed by Sommeria (1976).

The observations are used to specify the initial and boundary conditions for the model and then to test the characteristics of subcloud layer turbulence generated by the model. The physical characteristics of clouds produced by the model and the evolution of the mean profiles can also be compared with observations. A previous comparison with a limited set of data (Sommeria and LeMone, 1978), obtained in rather different conditions in the trade wind zone, (Puerto Rico area) has already shown some encouraging agreement. The major differences in the dynamics between the two cases are largely due to the considerably decreased windspeed in the GATE simulation (2–3 m s⁻¹ compared with 15 m s⁻¹ in the trade wind case).

This first step may then be considered as a validation of the model (for a given set of conditions). If this is considered successful, the second step is to make use of the model results as an aid to the further interpretation of the experimental data. Thus the effects of cloud-subcloud layer coupling mechanisms become clearer and quantities which involve measurements beyond the capabilities of present instruments (e.g. in-cloud temperature and humidity or pressure-velocity correlations) can at least be included in the discussion and their importance assessed.

2. DESCRIPTION OF THE MODEL

The model used in this experiment is a slightly modified version of the one developed by Sommeria (1976) in collaboration with J. W. Deardorff. A comprehensive specification is contained in Sommeria (1976) and only a brief outline appears below.

The main purpose of the model is to describe the detailed evolution of the turbulent meteorological field within a three-dimensional grid network covering a small domain of the atmosphere (a cube 2 x 2 x 2 km with a grid interval of 50 m). The basic variables include the u, v, w components of velocity, the virtual potential temperature θ_v, the specific humidity q, the specific content q_1 of cloud liquid water, and the deviation p' of the pressure from a prescribed large-scale field. The usual set of meteorological equations for moist air with the anelastic approximation for density is used. This comprises the equations of state and continuity, a Navier-Stokes equation with a Coriolis term and a buoyancy term including the effects of water vapour and cloud droplet loading. A thermodynamic equation including latent heat release together with conservation equations for water vapour and liquid water including condensation and evaporation processes complete the set.

These equations are spatially averaged over the grid volumes. This implies a statistical treatment for the grid-averaged sub grid-scale turbulent processes. This treatment is based on a simplified closure of the second order moment equations together with a representation of sub-grid condensation as described in Sommeria and Deardorff (1977). In contrast to most other boundary layer models which generally use horizontal mean or ensemble mean equations, the parametrized sub grid-scale processes represent here only a small fraction of the directly computed grid-scale turbulent processes (above the lowest level). The number of arbitrary or empirical assumptions made in such a model is, therefore, considerably reduced. The actual specification of these assumptions was left unchanged from previous simulations of marine boundary layers, i.e. no ‘tuning’ of the model was attempted prior to this integration.

Other characteristics of the model which are of interest for this study involve the boundary conditions and the definition of the variables inside a grid volume. At the vertical sides of the domain under consideration, cyclic boundary conditions are prescribed for the local variables. However, some externally prescribed large-scale effects can be taken into account.
in the evolution of the horizontally averaged variables (Sommeria, 1976) thus allowing large scale subsidence or advection to be represented. At the upper boundary the vertical gradients of the variables are prescribed according to the observed values. The surface fluxes are calculated from the specified surface variables and the computed values at the first level of the model by a surface layer parametization based upon the results of Businger et al. (1971) and Dyer and Hicks (1970). The validity of this approach is dependent upon the assumption that a constant flux or 'surface' layer extends from the sea surface to the first level at which model variables are computed. A simple space-staggered grid is used, with scalar variables computed at the centre of grid volumes and the components of velocity at the centres of the corresponding faces. The levels at which all variables (except \( w \)) are defined are consequently odd multiples of the half grid height. For example the lower boundary conditions for all variables (except \( w \)) are applied at the level 25 m above the surface.

3. General Considerations

The general philosophy behind comparisons of this type has been outlined by Sommeria and LeMone (1978), but a brief restatement of the problems as they apply to this particular work is in order.

In making any comparison between model and experimental results, it is important to bear in mind the limitations of both. In many instances, quantities evaluated from model results are not directly comparable to corresponding quantities calculated from aircraft measurements for a number of reasons outlined below.

The model can only consider a limited array of grid points due to limitations imposed by present computer size. A grid interval of 50 m ensures that the scales responsible for the major part of the transfer process in the boundary layer are adequately represented, but is also sufficiently small to allow an effective parametrization of the sub grid-scale processes (Sommeria, 1976; Sommeria and Deardorff, 1977). However, the model domain of \( 2 \times 2 \times 2 \text{ km} \) and the periodic boundary conditions employed in the model distort the larger eddies and limit the scales over which comparisons with data can be made.

Data derived from aircraft measurements are necessarily averaged along a flight path typically at least 30 km in length in order to obtain a sufficient sample. Fluctuation components are derived in the model by the removal of the area mean. For the aircraft data, they are deviations from a best fit line since small trends are usually observed over a distance of 35 km. Therefore an overbar represents a \( 2 \times 2 \text{ km} \) horizontal area average for results from the model and a (nominal) 35 km linear average for the aircraft data. The aircraft data therefore contain information on scales up to an order of magnitude greater than those considered by the model. In attempting to overcome this problem, Sommeria and LeMone (1978) used filtered aircraft data to ensure that only those scales represented in the model remained. No such filtering was attempted in this simulation so the spectral distribution of quantities being compared is of considerable importance. The grid interval of the model was selected to ensure that the scales responsible for vertical turbulent transfer were adequately resolved.

Finally, no large-scale or diurnal effects, which are often observed as slow variations in experimental measurements, are considered by this model.

4. Specification of Initial and Boundary Conditions

The period chosen for the simulation was part of 10 September 1974, during which two aircraft (the UK C-130\(^1\) and the DC-6\(^2\)) took a detailed series of measurements in a horizontally uniform area within the so-called 'GATE C-scale' triangle (see Fig. 1). Descriptions of the aircraft instrument systems are contained in Nicholls and LeMone

\(^{1}\) UK C-130 is operated by the Meteorological Office's Met. Research Flight, RAE, Farnborough, UK.

\(^{2}\) DC-6 was operated by the National Oceanographic and Atmospheric Administration, USA.
(1980), Nicholls (1978) and Bean et al. (1976). The actual measurements are described in
detail below. Briefly, a well-mixed subcloud layer was observed, capped by a weakly stable
layer. A 5–10% area cover of shallow non-precipitating cumulus whose tops extended to a
maximum height of 1·1 km remained virtually unchanged throughout the flight. The mean
cloud base height was observed to be 580 m. Windspeeds were light, subcloud layer mixing
being driven primarily by the action of buoyancy (see section 6d). No apparent mesoscale
organization of the clouds was observed.

Most of the data used to specify the initial and boundary conditions were derived from
the aircraft measurements. The aircraft repeated an L-type pattern (see Fig. 1) at heights
from 15–970 m between 1230 and 1610 GMT when the DC-6 departed, leaving the UK C-130
to perform a closely spaced sea-surface temperature mapping pattern at 150 m in the same
location, until 1800 GMT. The two aircraft were thus able to make a dense series of measure-
ments throughout a relatively small volume (approximately 35 × 35 × 1 km) for a period
of approximately 5 h. These measurements provide good estimates of horizontally averaged
parameters which are necessary to describe the model initial conditions, together with
good vertical resolution.

Additional sources of data were also considered in defining the initial profiles but,
since a comparison between all pertinent measurements has been presented by Barnes
et al (1980), further discussion will be kept to a minimum.

(a) Surface pressure

Specified externally in the model, a constant value of 1014·8 mb was chosen from measure-
ments taken on the HECLA, DALLAS and METEOR (from WMO Surface Data Set)
shown in Fig. 2a. The semi-diurnal pressure wave which dominates Fig. 2a is unimportant
in determining the small scale dynamics of the boundary layer and is not modelled.
(b) Sea surface temperature

Measurements of sea-surface temperature are not sufficiently accurate for defining the lower boundary conditions in the model since the surface fluxes are very sensitive to the value of the air-sea temperature difference. A surface temperature of 27.8°C was chosen to ensure that the prescribed surface fluxes (see section 4) were close to those measured by the aircraft. Fig. 2b shows that this value is very close to the reported measurements (from WMO Surface Data Set). Sea-surface temperature was set constant across the model domain. The sea-surface temperature measurements obtained by the UK C-130 shows a gradual increase of about 0.5°C from the southeast to the northwest corners of the area shown in Fig. 1.

(c) Potential temperature profile

This was obtained from the aircraft data and is shown in Fig. 3a. The measurements span a period from 1230 to 1700 GMT, during which time the mean mixed layer potential temperature was observed to increase. This was mainly due to the changing pressure values illustrated in Fig. 2a but since this pressure change is not modelled, a value consistent with the prescribed surface pressure was selected. The points shown in Fig. 2a are each an average value from a 35 km measurement run calculated assuming a constant surface pressure of 1014.8 mb. The value of $\theta = 298.6$ K chosen for the mixed layer is the same as the average
Figure 3(a), (b), (c) Horizontally averaged temperature and humidity profiles. The ● represent the mean horizontal average measurements made by the aircraft. The full curves are the initial profiles specified in the model ($t = 0$) and the dashed curves show the evolution after 3 h simulated time. (b) shows the initial profile of relative humidity.

Figure 3(d) Horizontally averaged wind profiles. The ● represent the horizontal average measurements made by the aircraft (error bars represent the standard deviation of the mean), —— are the geostrophic profiles externally applied in the model. The full curves are the initial profiles specified. The modelled profiles after 3 h of simulation are shown by the dashed lines.

(10 m) value measured from the ships over the period 1200–1700 GMT, as reported by Barnes et al (1980) ($T = 26.7^\circ$C, $\bar{\theta} = 298.6$K).

The mixed layer top was fairly well defined by these data alone. The gradients of $\bar{\theta}$ above the mixed layer were determined from soundings obtained by the UK C-130 on ascents and descents and the structure sonde data from the METEOR and FAY for the period 1200–1700 GMT (not shown here).

(d) Specific humidity profile

Figure 3c shows the specific humidity profile. The value of $q$ in the mixed layer is
almost constant, although a decrease is observed above 500 m. A constant mixed layer value of 16.0 g kg\(^{-1}\) was chosen, in close agreement with the UK C-130 measurements. Near surface ship observations show an average value of \(\bar{q} = 16.8\) g kg\(^{-1}\). This probably reflects the large gradient in \(\bar{q}\) close to the surface which is sometimes evident in tethered balloon soundings. Many tethered balloon and structure sonde soundings also show a marked decrease in \(\bar{q}\) above the mixed layer to about 800 m above which the decrease with height is substantially less. This compares favourably with the aircraft data and consequently the profile has been specified as shown. The upper part of the profile above 800 m was defined by aircraft ascents, structure sonde profiles, and radiosonde ascents all of which consistently show a slowly decreasing relative humidity between 75 and 80%. Figure 3b shows the mean relative humidity profile corresponding to \(\bar{\theta}\) and \(\bar{q}\) profiles defined previously.

(e) Wind profiles

The aircraft data show that both \(u\) and \(v\) remain roughly constant throughout the mixed layer with some shear occurring above, mainly in the region 500–700 m (Fig. 3d). In the mixed layer, \(u\) is taken to be constant at 2.5 m s\(^{-1}\) and \(v\) constant at 0 m s\(^{-1}\). This is in reasonable agreement with observations of the surface wind which was generally 2.5 m s\(^{-1}\) in the C-scale array and veered from 240° at 1200 GMT to 300° by 1800 GMT. The profiles above the mixed layer, which fit the aircraft observations, imply a veering to 330° up to 100 m with little change in speed. Above this there is no further change. This is supported by measurements made by a tethered balloon system aboard the DALLAS which show a veering to 325° up to a height of 1000 metres with little change in speed. These profiles are also in qualitative agreement with the surface, 850 mb, and 700 mb streamline charts for 1200 and 1800 GMT.

The model was initialized with the profiles shown in Fig. 3d. The geostrophic wind was chosen equal to the actual wind in the upper part of the boundary layer. These conditions ensure that during the integration wind profiles alter significantly only within the mixed layer.

(f) Mean vertical velocity

Horizontal divergence estimates using radio sonde data from the corner ships of the C-scale triangle have been calculated by B. Brümmer (private communication) for the period 0600–1200 and 1200–1700 GMT but are very small (\(\sim 10^{-5}\) s\(^{-1}\)) throughout the boundary layer and probably lie within the limits of error involved in making such a calculation. The divergence (and by implication the mean vertical velocity) was therefore set to zero throughout the model domain. This is further supported by the mean value of divergence from a number of undisturbed periods also calculated by Brümmer (1978), which also remains very small (\(-10^{-5} < \text{divergence} < 10^{-5}\) s\(^{-1}\)) throughout the cloud and subcloud layers.

(g) Large-scale advective effects

No large-scale advective effects were included in this integration.

(h) Radiative profiles

Unfortunately, no direct radiative flux measurements were available from the DC-6 or the UK C-130. The radiative heating profiles have therefore been calculated theoretically by methods developed at Colorado State University (Cox, et al., 1976). For the purpose of the calculation, a standard tropical atmosphere is assumed above 100 mb. This has been coupled to average rawinsonde soundings for 10 September from the C-scale ships and to the profiles of \(\bar{\theta}\) and \(\bar{q}\) given previously for the lowest 2 km. Cloud-free conditions are assumed, which is expected to be a reasonable approximation given that only a 5–10% area coverage of shallow cumulus was observed in the area with medium level clouds absent. Although it is difficult to be certain of the effects of cirrus on the calculation, extensive upper cloud was not present in the area on this particular day.

Figure 4 shows the calculated radiative heating and cooling rates which are on average of two calculations for 1330 and 1630 GMT (between which times the aircraft measurements were made) on this day at this latitude. The longwave cooling in the lowest kilometer can
be compared with other airborne measurements obtained by Ellingson (GATE Workshop Report, 1977) which are an average for cloud-free conditions throughout GATE. The agreement is very close.

Radiative effects were decoupled from other processes in the model and the mean cooling rate (dashed line in Fig. 4) was applied as a prescribed constant at each level in the model throughout the integration.

(i) Sea-surface roughness

The surface fluxes are prescribed in the model by the Businger-Dyer formulation (see section 2 and Sommeria, 1976). This specifies the fluxes of heat, water vapour, and momentum in the surface layer through transfer coefficients which are dependent on stability, height above the surface, and surface roughness ($z_0$). As roughness measurements were not available, $z_0$ was adjusted (with the stability parameter defined by extrapolating the airborne flux measurements to the surface) until 10 m transfer coefficients were obtained in closest agreement with those measured by other workers during GATE, namely, $C_D, C_F = 1.4 \times 10^{-3}, C_H = 1.6 \times 10^{-3}$ (GATE Workshop Report, 1977). Using these values, the chosen mean profiles and sea-surface temperature give surface fluxes which are in good agreement with those obtained by extrapolating the aircraft measurements to the surface. The value of $z_0$ obtained by this adjustment was $2 \times 10^{-4}$ m which is not inappropriate for these conditions (e.g. Monin and Yaglom, 1971, p. 295).

5. The evolution of the mean profiles

The integration was initialized with the horizontally uniform conditions specified previously. No clouds were included initially; these were allowed to develop with time. Convection was initiated by applying small random temperature perturbations at the lowest level of the model.

Since the simulated turbulence will not initially be in balance with the prescribed mean profiles, a period of mutual adjustment must occur. This is best illustrated by Fig. 5, which indicates the development of clouds in the model and shows that the adjustment process takes place relatively quickly, on a time scale comparable with that derived from a consideration of the turbulence ($\tau_T = (h/w^*) = 10^3$ s) rather than on a time scale associated with gross changes in mean subcloud layer properties (e.g. due to the subcloud layer drying out). The evolution of the mean profiles shows that such a time scale, $\tau_T$, is of the order of a day, for if a time scale is defined by considering the heating of the subcloud layer by convection alone, then
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Figure 5. Modelled fractional area cloud coverage throughout the integration. The tick marks on the upper axis show the intervals at which turbulence statistics were selected for comparisons with data.

\[ \frac{\partial \theta_v}{\partial t} \approx \frac{\partial}{\partial z} w' \theta_v' \approx 1.2 \overline{w' \theta_v'/h} \]

assuming a cloudbase flux of \(-0.2 \overline{w' \theta_v'/h}\) (see section 6a).

If

\[ \overline{w' \theta_v'/h} = C_g \Delta u_0 \Delta \theta_v \]

where \( \Delta \) represents the difference between the sea surface and 10 m (it is assumed that \( C_H = C_E = C_g \)) then \( \tau_L \), which is now the length of time required to heat the mixed layer by an amount \( \Delta \theta_v \), is defined by

\[ \Delta \theta_v / \tau_L \approx 1.2 C_g \Delta u_0 \Delta \theta_v / h \]

i.e.

\[ \tau_L \approx \frac{h}{1.2 C_g \Delta u_0} = \frac{600}{1.2 \times 1.5 \times 10^{-3} \times 2.5} = 1.3 \times 10^5 \text{s} \]

\[ = 1.5 \text{d} \]

Since \( \tau_L \gg \tau_T \) a quasi-equilibrium state is achieved in which the mean turbulence structure is only changing very slowly and is in balance with the mean profiles. This corresponds to the undisturbed tropical boundary layer in which the observations were made.

Figure 6 shows the evolution of the subcloud layer \( \theta_v \) profiles during the adjustment period (the initial profile is not realistic due to the slightly different depths of the initial imposed constant \( \bar{\theta} \) and \( \bar{\theta} \) profiles although this anomaly is quickly mixed away) and helps illustrate the mechanism by which the turbulent adjustment occurs in the model. In the first stages of the simulation, rising convective elements remain positively buoyant throughout the depth of what will become the mixed layer, leading to large vertical velocities at the stable layer base, and subsequent vigorous cumulus formation (as seen in Fig. 5). With such a large mass flux at cloud base, compensating subsidence is also strong between clouds modifying the \( \theta_v \) profile in the subcloud layer. This reduces the excess buoyancy of rising convective elements near cloudbase which thereby reduces cloud activity and subsidence between clouds such that eventually a steady state is obtained. In fact, there does appear to be some overcompensation leading to relatively small cloud activity...
Figure 6. Modelled mean virtual potential temperature profiles in the subcloud layer.

Figure 7. Evolution of temperature and humidity in the subcloud layer. The full curves represent the modelled values while the • show the estimates available from aircraft data (together with the estimated errors in each). The dashed curve in (a) shows the effect of correcting for the observed horizontal temperature advection (not included in the simulation).
for the period $t = 0.7 - 1.0 \, \text{h}$. The adjustment process is complete after approximately one hour of simulated time. This is further supported by the time development of other indicators of turbulent activity, e.g. total turbulent energy, variances, and fluxes in the subcloud layer which have also become more or less steady after approximately one hour.

Figure 6 also suggests that the subcloud layer $\theta_s$ profile retains a characteristic shape once the quasi-equilibrium state is established with a minimum midway through the subcloud layer. This is discussed further in section 7.

The changes in the mean profiles of $\bar{\theta}, \bar{q}, \bar{u}$ and $\bar{v}$ at the end of the integration ($t = 3.0 \, \text{h}$) are shown in Figs. 3a, 3b, 3c and 3d. The main effects are the moistening and cooling of the cloud layer leading to some destabilization of the temperature profile there. Changes in the modelled subcloud layer are small with $\bar{\theta}$ increasing slowly but $\bar{q}$ remaining almost unchanged. The rate of change of $\bar{T}$ and $\bar{q}$ in the modelled subcloud layer calculated over the final 2 hours of the simulation agree well with the observed variation as shown in Fig. 7 (although the errors in estimating $\partial \bar{T}/\partial t$ and $\partial \bar{q}/\partial t$ from measurements are considerable). The warming due to turbulent transport is not completely offset by the radiative cooling in the subcloud layer and a slow increase in temperature occurs.

If the horizontal temperature gradients measured by the aircraft are used to account for horizontal advection (which was not modelled), the agreement becomes marginally better (Fig. 7a). No corresponding significant horizontal humidity gradient was detected. Although one should not perhaps overemphasize the comparison of the evolution of mean profiles in such a case study in the absence of a more precise account of mesoscale and large scale effects, the model does succeed in reproducing rates of heating and moistening in the subcloud layer which are in good agreement with the observations.

6. COMPARISON OF MODELLED TURBULENT QUANTITIES WITH OBSERVATIONS

Throughout this section all of the model results have been taken from the final two hours of the simulation, i.e. during the quasi-equilibrium period.

(a) Vertical turbulent fluxes

Figures 8 and 9 show the behaviour of the vertical turbulent fluxes at each level in the model. These represent an average over the whole domain and include both computed scale and sub grid-scale contributions.

Results have been selected from the model by taking averages over 40 time steps (200 s) at intervals of 200 time steps (1000 s) from $t = 1.17 \, \text{h}$ to $2.94 \, \text{h}$ (these times are

![Figure 8](image-url)

Figure 8. Comparison of measured and modelled heat and moisture fluxes from the model (squares and full curves) and aircraft observations (triangles and dashed curves). The range of observed values is indicated by bars; the shaded area denotes the range of modelled values. Curves are fitted by eye.
marked in Fig. 5. The range of modelled values is represented by shading. The observations are represented by a mean value and bars which enclose all the individual points at any particular level. The variability of the modelled profiles during the integration is comparable, in some sense, to the scatter observed in the aircraft measurements. The degree of correspondence to be expected between the model average and a 30 km linear average in the real atmosphere is not obvious and clearly depends on the spectral distribution of energy of the parameters in question. However, for the second order moments examined in this study, the dispersion of the model results and experimental points is similar. This is partly due to the fact that the major contributions to these parameters occur at scales resolvable by both the aircraft and the model as shown in section 6b.

Figure 8 shows the variation of $w^T_w$, $w^q_w$, and $w^T^T$ with height. Note that, although there is some variation in the model-derived $w^q_w$ and $w^T^T$ values in the subcloud layer (which is linked to the degree of cloud activity), there is very little change in the $w^T_w$ profile. However, in order that the subcloud $\vartheta$ profile remain similar throughout the integration, it is necessary for the $w^T_w$ profile to be non-linear to account for radiative cooling, which changes with height in the mixed layer. This is observed in the model where $\partial w^T_w/\partial z$ is largest near the surface and decreases upwards throughout the mixed layer, in accordance with the prescribed radiative cooling profile.

A linear fit to the subcloud $w^T_w$ profile (a common construction in many simple one-dimensional mixed layer models) may be extrapolated to give a cloudbase virtual heat flux equal to $-0.2$ times the surface value. This behaviour has also been noticed in previous simulations (loc cit) and appears to be independent of modelled cloud activity. Above cloudbase, each of the fluxes is sensitive to the degree of cloud activity as has previously been noticed in simulations by Sommeria (1976), and Sommeria and LeMone (1978).

In general, the agreement between the aircraft observations and model predictions in the subcloud layer is excellent. Both show little convergence of water vapour in the subcloud layer, the main effects being the warming of the subcloud layer and the cooling and moistening of the cloud layer. The modelled temperature flux appears to be slightly underestimated perhaps a result of too low an input surface value.

Figures 9a and b show the momentum flux profiles. Again, the agreement between aircraft results and model values is good despite the very small numerical values of both $u'w'$ and $v'w'$. The maximum in the $v'w'$ profile near cloudbase appears in both sets of results.

(b) Variances

Figure 10 shows the variation of vertical velocity variance with height. Changes in
cloud activity which can be seen to alter the profile in the cloud layer again seem to have little effect on the values in the subcloud layer (as noted by Nicholls and LeMone, 1980) where a maximum is consistently observed near $z = 250$ m.

Profiles of model-derived variances for the velocity components $u$ and $v$, temperature $T$ and specific humidity $q$ are compared with observed values in Fig. 11. The model variances were obtained as a series of 20 min averages from 1.7 to 3.0 h. The figure shows that the model underestimated the variances for all variables but temperature. The underestimate is probably due to the limited range of wavelengths that can be resolved by the model when compared with those present in the aircraft data. To illustrate this, a composite spectrum of $q$ in the east-west direction from aircraft measurements made during the sea surface temperature mapping pattern at 150 m appears in Fig. 12a. This spectrum was averaged from nine individual runs, each 5 min (30 km) long. Note that the major peak occurs at a wavelength of about 10 km, which is considerably larger than the model domain (2 km). The $u$ and $v$ spectra show similar behaviour (e.g. Nicholls and LeMone, 1980) which would also explain the underestimate of $u'^2$ and $v'^2$ by the model. The cause of the overestimation of temperature variance is presently unknown. In contrast, the wavenumber band containing significant energy in the $w$-spectrum based on the same nine runs is largely resolved by the model as shown in Fig. 12b. Here the peak of the spectrum occurs at a wavelength of about 700 m. Inspection of the cospectra of $u, v, T$ and $q$ with $w$ also reveals that the same wavenumber band contributes most to the covariances, as seen in Fig. 12c. Despite there being considerable energy on 10 km scales in the $u, v, T$ and $q$ spectra, the contributions to the covariances at these wavenumbers remain quite small. This is a common feature observed both during GATE (e.g. Nicholls and LeMone, 1980) and elsewhere (e.g. Nicholls and Readings, 1981), suggesting that although there is considerable energy on longer scales, these are not directly involved in the energy transfer processes within the subcloud layer. There is therefore some justification for believing that the turbulent fluxes are modelled reasonably well despite the failure of the model to reproduce the observed variances.
Figure 11. Profiles of variance for east-west velocity $u$, the north-south velocity $v$, temperature $T$, and specific humidity $q$, for the model (squares) and from aircraft observations (triangles). The range of observed values is indicated by bars, the standard deviation by arrows. The shaded area shows the range of the model averages. Range not available for $u^{2T}$. 
Figure 12. (a) Frequency weighted composite $q$-spectrum. The data were obtained from nine runs by the UK C130 at an altitude of 150 m. (b) As (a) but for $w$-spectrum. (c) As (a) but for the $wq$-cospectrum.
However, although the range of wavenumbers contributing to the covariances are largely resolved throughout the subcloud layer, the range does extend slightly past 2 km so that a fraction of the measured covariance does occur on non-resolvable scales. This fraction varies between 10–25%, the greater values being at higher levels. Clearly then, at least part of the good agreement between the model and the observations is due to the tendency of the model to accumulate energy at resolvable scales which should, in the real atmosphere, be transferred to scales larger than the simulated domain.

(c) Properties of the cloud field

Due to limitations imposed by the airborne instrumentation, comparisons between modelled and observed clouds are restricted to simple macrophysical quantities.

The modelled cloud area coverage is approximately constant at 4–5% from $t = 1$ h to 3 h (see Fig. 5). An analysis of the upward and downward looking radiation thermometers on the UK C-130 reveals an average fractional area coverage of between 5 and 10%, depending on the threshold value used to discriminate between cloud and no cloud. Some of this variation may be due to the instruments seeing the upper parts of clouds which are indistinguishable from lower cloud because of the response times of the sensors (1–2 s). This would lead to a larger measured fractional area coverage. Neither the model nor the data show any significant change in the average cloud amount with time (once a steady state is reached).

Figure 13. Modelled horizontally averaged liquid water content (as Fig. 8). Also shown are the cloudbase and cloud top heights observed from the aircraft (arrowed).

Figure 13 shows the modelled horizontally averaged liquid water content, thereby defining the extent of the cloud layer. Although there were no measurements of liquid water made on the aircraft, the heights of cloud base ($h$) and cloud top predicted by the model agree closely with observations from the aircraft. The modelled cloudbase did not change significantly throughout the integration, although cloudbase was observed to rise slowly throughout the flight (by an average $20\,\text{m}\,\text{h}^{-1}$). Such a change would only be marginally resolvable in the model over a 3 h period, but is consistent with the slow warming trend in the subcloud layer (e.g. Fig. 7a).

Although the modelled total cloud amount remained fairly steady with time, the size distribution changed significantly. During the periods 1:17 to 1:80 and 2:10 to 3:00 h,
there were a number of small clouds scattered throughout the domain whereas from 1.80 to 2.10 h one substantial cloud (with a lifetime of 46 min simulated time) dominated the model. It is interesting to note that the modelled subcloud layer fluxes are quite insensitive to this change in cloud size distribution and remain virtually unchanged.

Figure 14 shows the relationship produced by the model between the maximum area covered by an individual cloud at the level of maximum cloudiness (note that this is not necessarily the same as the level at which the average liquid water content is a maximum) and the lifetime of the cloud. Such a relationship is useful for parametrization of cloud convection in large scale models. Results from a previous simulation of an undisturbed trade wind boundary layer (Sommeria and LeMone, 1978; Beniston and Sommeria, 1981) are also shown for comparison. Clearly, the model produces a reasonable relation between cloud size and lifetime, although there is a slight difference between the two simulations. It is not known whether this is due to actual differences in the dynamics in the two cases or to the inclusion of a more sophisticated treatment of the sub-grid-scale condensation process (Sommeria and Deardorff, 1977) in the GATE simulation. Nor is it possible to verify the accuracy of the modelled size-lifetime relations due to the lack of any suitable experimental data. However since clouds in the Puerto Rico simulation were subject to larger shear, it is reasonable that they were also the shorter lived.

(d) Turbulent kinetic energy budget

Comparisons can also be made between model derived and measured terms of the turbulent kinetic energy (TKE) balance. This may be expressed (e.g. Lenschow, 1974) using standard notation as

$$\frac{\partial \overline{\mathcal{E}}}{\partial t} = (g/\bar{u}_e)\overline{w^2}\bar{e} - \frac{\partial}{\partial z} \overline{w'\mathcal{E}} - \frac{\partial}{\partial z} (\overline{w'p'/\bar{\rho}}) + (\tau/\bar{\rho}) \frac{\partial \bar{e}}{\partial z} - \varepsilon$$

where the terms on the right hand side of this equation will be referred to as the buoyancy
production (BP), turbulent transport (TT), pressure correlation (PC), shear production (SP) and viscous dissipation (D) respectively.

Estimates of each of the terms in the equation can be derived from the aircraft measurements with the exception of the pressure correlation (e.g. Pennell and LeMone, 1974). The viscous dissipation term, $\varepsilon$, was derived from the high frequency velocity spectra by application of the Kolmogorov inertial subrange formulae (e.g. Nicholls and Readings, 1981). An average of the values obtained from all three velocity components was used as the best $\varepsilon$ estimate. Vertical derivatives in the SP and TT terms were calculated by first fitting smooth curves to the averaged data points and then differentiating the curves. Each of the modelled terms are averages from $t = 1.94$ h to $t = 2.78$ h. The BP, PC, SP and D terms were obtained directly from the model, the others being obtained as a single residual.

Figure 15. Comparison of measured (symbols) and modelled (curves) terms in the turbulent kinetic energy budget. See text for an explanation of the abbreviations. Symbols represent the following measurements: ○ - BP, ▲ - SP, X - D, ○ - TT.

Figure 15 shows the variation of the modelled quantities with height and compares the corresponding aircraft derived values in the subcloud layer. As was seen earlier (for the $w' T_v$ profiles), the BP term agrees well with the measured values in the subcloud layer and is the dominant source of TKE production. The removal of energy from the model through the viscous dissipation parametrization also conforms closely to the aircraft values and positive shear production at cloud base is also observed. Since the rate of change of TKE is small ($\sim 10^{-3} \text{ m}^2 \text{s}^{-3}$) beneath cloudbase, the residual may be ascribed to the turbulent transport term which does agree reasonably well with measurements throughout much of the subcloud layer. However, this term is expected to be most negative close to the surface (Wyngaard and Coté, 1971) which although supported by the observations does not occur in the model. Similarly, the behaviour of the pressure correlation term near the surface also looks unrealistic (this behaviour is very like that reported by Deardorff (1974) in a simulation of strong convection over land). It appears that some of the sub grid-scale assumptions made in the model become inadequate in the lowest levels where they become dominant terms. However, despite the relatively simple treatment of the sub grid-scale process employed by the model, the main features of the measurements are reproduced in the rest of the subcloud layer. In this region the variation of the terms in the TKE balance with height are qualitatively similar to that found in cloud-free, inversion-capped boundary layers (e.g. Lenschow, 1974).
Instrumental limitations prevented reliable measurements being made in the cloud layer. The modelled terms are relatively small in the first 100 m above cloudbase suggesting this region is dynamically fairly quiet. However, the PC term displays a local maximum just above cloudbase implying that non-hydrostatic pressure fluctuations could be an important component in the cloud-subcloud layer interaction process, even with the small sized clouds generated in this study. Buoyant production by latent heat release peaks midway through the modelled cloud layer and is dissipated mainly through the viscous term although there is a significant loss in the upper part to the pressure correlation term which is probably associated with the production of gravity waves. The residual appears to be rather large in the cloud layer but a major part of this must be due to the turbulent transport term which might be expected to be a significant source term in the upper cloud layer. However, this was one of the least successfully modelled terms in the subcloud layer and a degree of caution must be exercised in the absence of a more thorough investigation of some of the modelled terms.

### 7. The Structure of the Subcloud Layer

During the initial stages of the simulation, the subcloud layer \( \bar{\theta}_v \) profile responded quickly to the exchanges between cloud and subcloud layers eventually reaching the quasi-equilibrium form shown in Fig. 6. This suggests that the shape of the \( \bar{\theta}_v \) profile in the upper

![Graph](image)

**Figure 16** Simulated and measured subcloud \( \bar{\theta}_v \) profiles scaled by cloud-base differences, \( \Delta \bar{\theta}_{vB} \).
subcloud layer is determined primarily by mixing occurring at or near cloudbase. This part of the profile might, therefore, be similar in other corresponding situations if scaled with a suitable cloudbase value. This is tested in Fig. 16 which shows scaled \( \tilde{\theta}_v \) profiles in the subcloud layer. The subcloud layer average, \( \tilde{\theta}_v \), (the tilde denotes a mean throughout the depth of the subcloud layer), has been removed and these values have been scaled by \( \Delta \theta_{v,0} \), which is defined at cloudbase by \( \Delta \theta_{v,0} = (\tilde{\theta}_v - \tilde{\theta}_e) \). The profiles are drawn from the GATE simulation, the Puerto Rico simulation (Sommeria and LeMone, 1978), and VIMHEX data (Betts, 1976 – experimental measurements in strongly heated boundary layers over land). Values of cloudbase and mixed layer scaling parameters are given in Table 1. Notice the order of magnitude increase in \( \Delta \theta_{v,0} \) in the VIMHEX data and the very small values in the GATE and Puerto Rico simulations over the sea.

\[
\begin{array}{|c|c|c|c|c|c|}
\hline
\text{Source} & w^+ & \tilde{\theta}_v & \Delta \theta_{v,0} & \Delta q_{v,0} & \Delta \theta_{v,0} \\
\hline
\text{GATE (} t = 1.84 \text{ h}) & 0.60 & 0.018 & 0.102 & -0.39 & 0.030 \\
\text{GATE (} t = 3.00 \text{ h}) & 0.60 & 0.018 & 0.102 & -0.41 & 0.027 \\
\text{Puerto Rico (} t = 1.47 \text{ h}) & 0.76 & 0.026 & 0.108 & -0.33 & 0.049 \\
\text{VIMHEX (exptl data)} & 2.1 & 0.01 & 0.51 & -1.26 & 0.28 \\
\hline
\end{array}
\]

The scaled \( \tilde{\theta}_v \) profiles shown in Fig. 16 are quite similar, especially in the upper mixed layer where \( \tilde{\theta}_v \) begins to increase strongly above about 0.7 h. A minimum in \( \tilde{\theta}_v \) is also well marked between 0.3 h and 0.5 h a feature which has been observed over land by both Betts (1976) and Johnson (1977), although the temperature differences involved over the ocean in the GATE and Puerto Rico data (~ 0.05°C) preclude its observation using aircraft data.

Since it appears that the \( \tilde{\theta}_v \) profile throughout a large fraction of the subcloud layer is controlled primarily by the mixing occurring near cloudbase, the \( \tilde{\theta}_v \) and \( \tilde{\theta}_q \) profiles should also behave similarly when scaled in an analogous manner. This was first attempted by Betts (1976) using VIMHEX data. Fig. 17 shows the scaled \( \tilde{\theta}_v \) and \( \tilde{\theta}_q \) profiles from a number of simulations together with those measured by Betts. Values of the scaling parameters \( \Delta \theta_{v,0}, \Delta q_{v,0} \) defined in the same manner as \( \Delta \theta_{v,0} \) are given in Table 1.

The scaled profiles are again very similar down to approximately 0.4 h below which the scaled profiles diverge as the effect of the surface becomes dominant (the air-sea surface differences \( \Delta \theta_{v,0}, \Delta q_{v,0} \) are the same sign, whereas the corresponding cloudbase values \( \Delta \theta_{v,0}, \Delta q_{v,0} \) are of opposite signs). There is a small spread in the scaled profiles, which is presumably due to temporary deviations away from the equilibrium value (the profiles shown derived from the models are ‘instantaneous’ i.e. single time step) values. This could explain the differences observed between the VIMHEX data and the simulations since Betts noted that changes in the subcloud (and presumably the cloud) layer structure during the period covered by the observations (1000–1400 GMT) were significant. The model simulations of convection over the ocean were probably a much better approximation to a quasi-equilibrium state. The similarity between the profiles scaled in this manner emphasizes the importance of cloud-subcloud layer exchanges on the profiles throughout the upper two-thirds of the subcloud layer. This type of presentation also illustrates how the usual definition of a ‘transition’ layer between an idealized mixed layer and the cloud layer is rather ambiguous since the thermodynamic variables vary smoothly throughout the subcloud layer when horizontally averaged data is used.

Finally, the model can be used to investigate the average properties of convective elements, assumptions about which are often made in simple one-dimensional models. Fig. 18 shows the horizontally averaged deviations of \( \theta, q \) and \( \tilde{\theta}_v \) away from the horizontal mean inside rising convective elements (defined by co-located positive \( w^+ \) and \( q^+ \) fluctuations) plotted as a function of height throughout the subcloud layer.
These values are quantitatively very similar to those measured by the aircraft (Nicholls and LeMone (1980), not reproduced here) with the temperature excess becoming negative at a very low level, positive buoyancy being maintained by the moisture excess which increases with altitude. Buoyancy is slightly negative at cloudbase. Such close agreement between the model and the observations suggests that the model could usefully test certain parametrization schemes.

8. CONCLUDING REMARKS

This study shows how a detailed three-dimensional numerical simulation and a comprehensive small scale field observation can complement one another. The case which has been chosen, 10 September 1974 in the C scale area of GATE is representative of a widely occurring set of conditions: an oceanic tropical boundary layer with light winds and a relatively small surface buoyancy flux. The initial and boundary conditions required for the simulation were defined by matching as closely as possible to the available observations and the turbulence (and subsequent cloud field) allowed to develop from initial
random perturbations. After approximately one hour, a quasi-steady state was reached where the turbulence was in balance with the slowly changing mean profiles. This quasi-steady state was found to be similar to the observed undisturbed boundary layer in many respects:

(i) a comparison of statistical quantities in the subcloud layer showed that the modelled fluxes agreed very well with the measurements and confirmed that the main effects of the turbulent mixing were to warm the subcloud layer whilst moistening and cooling the cloud layer;

(ii) the evolution of the subcloud layer thermodynamic profiles was similar to that observed;

(iii) the modelled cloud field covered approximately the same steady, horizontal area as that observed. Cloud base and cloud top heights and the overall distribution of cloud were similar to those seen from the aircraft.

Only the vertical velocity variance was well reproduced by the model. The measured $u, v, T$ and $q$ variances all involved significant contributions from wavelengths of the order of 10 km which were not resolved by the model.

Despite the relatively simple handling of the turbulent kinetic energy balance by the model, most of the terms agreed reasonably well throughout the subcloud although the pressure correlation and turbulent transport terms were not well modelled close to the surface. The balance in the subcloud layer resembles typical cloudfree conditions: the main source terms being buoyancy in the lower part and turbulent transport in the upper part with viscous dissipation as the major sink term. In the cloud layer, the model suggests that the balance is between buoyant production and turbulent transport against losses to the viscous and pressure correlation terms.

The modelled $\theta, \theta_e$ and $\bar{q}$ are strongly affected down to quite low levels in the subcloud layer by mixing occurring at cloudbase. When scaled using appropriate cloudbase para-
meters, the forms of these profiles appear to agree well with previous measurements made in more unstable conditions over land. One characteristic feature is a minimum in $\theta_v$ midway through the subcloud layer. This is too small to be detected over the sea using the present data but has been observed over land.

Although it is tempting to investigate many more detailed interactions using comprehensive data sets produced by small scale models similar to that used here (particularly in view of the difficulty in making and interpreting real measurements), the main limitation of such models, namely the impossibility (at present) of representing turbulent motion outside a scale range of approximately 20 to 1, must be borne in mind.

A valuable future objective would be to test the sensitivity of the model to small changes in either the initial or boundary conditions although the degree of agreement achieved in this and previous studies without the necessity for ‘tuning’ the model to each situation suggests that the model does have considerable predictive capability. However, such a program would be a major undertaking and is beyond the scope of this present study. Progress is currently being made in improving the treatment of the subgrid-scale process within the model and in preparing the model for the latest computers. With these advances, tests such as those mentioned above should become easier to accomplish.

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APPENDIX

$C_D$, $C_E$, $C_H$  Transfer coefficients for momentum, latent and sensible heat

$g$  Acceleration due to gravity

$h$  Subcloud layer depth, i.e. cloudbase height

$k$  Wavenumber $= 1/\lambda$

$p$  Pressure

$q$  Specific humidity

$q_l$  Specific liquid water content

$T$  Temperature

$T_s$  Sea surface temperature

$u$  Wind component east

$v$  Wind component north

$w$  Vertical wind component

$w^*$  A mixed layer scaling velocity $\equiv (gh \bar{w'T'_{so}/T_s})^{1/3}$

$z$  Height

$\Delta$  A difference, defined in the text

$\overline{\varepsilon}$  Turbulent kinetic energy $\equiv \frac{1}{2}(u'^2 + v'^2 + w'^2)$

$\varepsilon$  Rate of dissipation of turbulent kinetic energy

$\lambda$  Wavelength

$\rho$  Density

$\theta$  Potential temperature

$\theta_v$  A mixed layer scaling temperature $\equiv (\bar{w'T'_{so}/w^*})$

$\overline{A}$  Horizontal average of the variable $A$

$A'$  Deviation of the variable $A$ from its horizontal average

$\overline{\overline{A}}$  Horizontal and vertical average in the subcloud layer of the variable, $A$

Subscripts

$0$  Quantity defined at the surface

$b$  Quantity defined at cloudbase ($z = h$)

$p$  An average difference between a convective element and the horizontal average (see text)

$v$  A virtual quantity
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Ellingson, R. 1977 Longwave cooling in the GATE subcloud layer. GATE Workshop Report (see below), 483–487.


