

A NUMERICAL STUDY OF THE PARAMETERIZATION
OF DEEP TROPICAL CONVECTION

by

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To,

Carolyn, my wife,

and,

my parents,

for their love, encouragement, and support.

" It is God who takes water from the earth
 and turns it into drops of rain.
 No one knows how the clouds move or how
 thunder roars through the sky, where God dwells.
 Do you know how clouds float in the sky,
 the work of God's amazing skill?
 Can you shout orders to the clouds and
 make them drench you with rain?
 At God's command amazing things happen,
 wonderful things that we can't understand. "
 (Job ch36 vs27,29, ch37 vs5,16, ch38 v34)

"....Jesus got up and gave an order to the wind and
 the stormy water; they died down, and there was a
 great calm. Then he said to the disciple, "Where
 is your faith?"

But they were amazed and afraid, and said to
 one another, "Who is this man? He gives orders to
 the winds and waves, and they obey him!" "
 (Luke ch8 vs24,25)

ABSTRACT

A two dimensional meso-scale cloud model is used to simulate an ensemble of deep tropical convective clouds, using data collected during the GATE experiment. Convection is initiated using heating of random magnitude, distributed across the domain. Large-scale and radiational forcing, plus surface fluxes of sensible and latent heat, derived from GATE data, are communicated to the model. No initial wind shear is included in the initial integrations. The resulting simulations are used to study the parameterization of deep convection, representing a new approach to the problem.

Comparison is made between the simulation and observational data, using horizontal averages and budgets of dynamical and thermodynamical quantities. The analysis is carried out over the observational time-scale (several hours), and also a few tens of minutes, a scale on which observations can say little. The model is found to be a good representation of the real atmosphere. Details unobtainable with any accuracy from observational data are analysed, to complete the description of the cloud ensemble. The internal cloud structure is determined, being divided into updraughts, downdraughts, and inactive cloud.

The equations governing the effect of convection upon its large-scale environment are derived in full, and compared to approximate forms used by previous workers. These form the basis of many parameterization schemes. The diagnosed "cloud model" is used to study simple representations of Q_1 and Q_2 for use in parameterization schemes, within both large-scale and coarse meso-scale models. Comments are made concerning the importance of shallow convection in determining the vertical structure of Q_2 . The dependence of convective magnitude upon atmospheric structure, and large-scale processes is investigated, in relation to the closure problem of cumulus parameterization.

The effects of shear upon convective magnitude, and vertical transports are briefly considered, and the parameterization of cloud momentum transports studied.

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LIST OF SYMBOLS AND ABBREVIATIONS.

h	height of a constant pressure surface
u	horizontal momentum
w	vertical velocity in height co-ordinates
ω	vertical velocity in pressure co-ordinates
Θ	potential temperature
T	temperature
q	water vapour mixing ratio
l_c	cloud water mixing ratio
l_r	rain water mixing ratio
l_τ	total liquid water mixing ratio
g	acceleration due to gravity
g^*	$g(1+l_\tau)$
ρ	density
p	pressure
Θ_v	virtual potential temperature = $\Theta(1+0.61q)$
Θ_{cl}	cloudy or liquid water virtual potential temperature = $\Theta(1+0.61q-l_\tau)$
Θ_e	equivalent potential temperature
Θ_{es}	saturated equivalent potential temperature
Θ_{esv}	virtual saturated equivalent potential temperature
ϕ_i	initial model value of ϕ
$\overline{\phi}$	domain average of ϕ
ϕ'	$\phi - \overline{\phi}$
$\langle \phi \rangle$	volume integral of ϕ over the domain
$\overline{\phi}^{\text{anvil}}$	ϕ averaged over anvil areas of the domain (as defined in chapter 4)
$\overline{\phi}^{\text{clear}}$	ϕ averaged over clear areas of the domain (as defined in chapter 4)
$\overline{\phi}^c$	ϕ averaged over cloud
$\overline{\phi}^e$	ϕ averaged over cloud environment
UD	updraught(s)
DD	downdraught(s)
IC	inactive cloud
$\overline{\phi}^{uo}$ or ϕ_{uo}	ϕ averaged over updraught air
$\overline{\phi}^{dp}$ or ϕ_{dp}	ϕ averaged over downdraught air
$\overline{\phi}^{ic}$ or ϕ_{ic}	ϕ averaged over inactive cloud
σ	fractional cloud area
$-Q$	condensational drying

$\left(\frac{\partial \phi}{\partial t}\right)_{\text{FOR}}$	external forcing applied to the model
$\left(\frac{\partial \phi}{\partial t}\right)_{\text{LS}}$	large-scale forcing applied to the model
$\left(\frac{\partial \phi}{\partial t}\right)_{\text{RAD}}$	radiative forcing applied to the model
$\left(\frac{\partial \phi}{\partial t}\right)_{\text{BL}}$	boundary layer forcing applied to the model
$\left(\frac{\partial \phi}{\partial t}\right)_{\text{CLD}}$	cloud forcing applied to the model
S_ϕ	turbulence
$Q1$	apparent cumulus heat source $= \frac{L\bar{Q}}{c_p \pi} - \frac{\partial \overline{\omega' \theta'}}{\partial p} + \overline{S_\theta} + \left(\frac{\partial \overline{\theta}}{\partial t}\right)_{\text{BL}} + \left(\frac{\partial \overline{\theta}}{\partial t}\right)_{\text{CLD}}$
$(Q1)_{\text{conv}}$	$= \frac{L\bar{Q}}{c_p \pi} - \frac{\partial \overline{\omega' \theta'}}{\partial p}$
$Q2$	apparent cumulus moisture sink $= -Q - \frac{\partial \overline{\omega' q'}}{\partial p} + S_q + \left(\frac{\partial \overline{q}}{\partial t}\right)_{\text{BL}}$
$(Q2)_{\text{conv}}$	$= -Q - \frac{\partial \overline{\omega' q'}}{\partial p}$
$Q3$	apparent cumulus momentum source or cumulus friction $= -\frac{\partial \overline{\omega' u'}}{\partial p} + \overline{S_u}$
$(Q3)_{\text{conv}}$	$= -\frac{\partial \overline{\omega' u'}}{\partial p}$
$(Q1)_{\text{para}}$	parametric form of $(Q1)_{\text{conv}}$
$(Q2)_{\text{para}}$	parametric form of $(Q2)_{\text{conv}}$
$(Q3)_{\text{para}}$	parametric form of $(Q3)_{\text{conv}}$
$(\phi)_{\theta, q}$	(ϕ) calculated from θ or q field
(CS)	$\sigma \overline{\omega'} \frac{\partial \overline{\phi}}{\partial p}$
(ADV)	$\sigma \left(u \frac{\partial \overline{\phi}}{\partial x} \right)^c$
(ED)	$\sigma \left(\frac{\partial u}{\partial x} (\phi - \overline{\phi}) \right)^c$

$$(BAL) \quad \frac{\partial \sigma \bar{\phi}}{\partial t} + (\omega_b \bar{\phi} - (\omega \phi)_b) \frac{\partial \sigma}{\partial p}$$

$$(EE) \quad -(1-\sigma) \bar{Q}^c + (\bar{q} - q_b) \frac{\partial \sigma}{\partial t} \quad \text{or} \quad \frac{L(1-\sigma) \bar{Q}}{c_p \pi} + (\bar{\theta} - \theta_b) \frac{\partial \sigma}{\partial t}$$

$$(FORC) \quad -\sigma \left(\frac{\partial \phi}{\partial t} \right)_{FOR}^c$$

$$(TURBC) \quad -\sigma \bar{S}_\phi^c$$

$$(CPG) \quad g \sigma \left(\frac{\partial (h - h_x)}{\partial x} \right)^c$$

E entrainment rate

D detrainment rate

CAPE convective available potential energy

rhs right hand side

lhs left hand side

wrt with respect to

π $(p/1000\text{mb})^{-R/c_p}$

CHAPTER 1

INTRODUCTION.

Many scales of motion exist within the troposphere, from synoptic scale waves (with a horizontal length scale, L , of several thousand kilometers), through meso-scale features such as frontal regions ($L \approx 400\text{km}$), to cumulus convection ($L \approx \text{few kilometers}$) and turbulence. In the past 25 years, with the advent of powerful computers, many of these features have been studied using time dependent numerical models.

Due to computational limitations, models representing large-scale flow have grid lengths of 100–200km. Features which exist below this length scale are un-resolved. However much of the kinetic energy of the atmosphere exists on these un-resolved scales. The kinetic energy spectrum of the atmosphere peaks at the scales listed above. It is seen that fronts exist on the grid scale of large-scale models, while cumulus convection are sub-grid scale features, and not represented at all by the grid point parameters. Interaction between various scales are important, and motion on small scales may play an important role in the development of larger scale flows. Hence the need to represent in some manner (to parameterize) these un-resolved features in large-scale models arises.

The physical processes which require parameterization in any model, depends upon the scale of the model. In a numerical cloud model, turbulent effects need to be represented, while for large-scale and coarse meso-scale models cumulus clouds themselves need to be parameterized.

The heat release associated with the change of phase within cumulus convection has important effects upon both synoptic and meso-scale flows. The aim of this thesis is to study the parameterization of convection within models of these scales using a numerical model of a tropical deep cloud ensemble.

1.1 THE IMPORTANCE OF CUMULUS CONVECTION TO LARGE-SCALE FLOW.

Cumulus convection plays an important role in the energy budget of the atmosphere. This was first noted by Riehl and Malkus(1958), and is especially true in the tropics. Convection accounts for much of the ascent in the tropical branch of the Hadley cell, the majority of tropical rainfall being attributed to cumulus convection. Several large-scale and meso-scale features have been identified as being greatly effected by convective heat release, in both the tropics and mid-latitude.

The first such feature to be identified as such was the tropical cyclone. Palmen and Riehl(1957) first noted the importance of latent heat release in the energy budget of tropical cyclones. Cumulus convection is importance in the transformation of a cold core tropical depression (easterly wave), to the warm core structure (Simpson(1947)) of an intense tropical cyclone (Yanai(1961a,b)). Cumulus convection embedded within the motion system of the tropical depression heats the atmosphere, feeding off the large conditional instability of the atmosphere and moisture convergence within the friction layer of the depression. The energy release is much greater than the dissipation within the friction layer. This heat release causes the disturbance to amplify. Early models of tropical cyclones failed to produce an intensifying disturbance unless cumulus heating was parameterized (Charney and Eliassen(1964), Ooyama(1964), Kuo(1965)). Charney and Eliassen termed the mechanism of tropical cyclone growth "Conditional

Instability of the Second Kind" (CISK).

Latent heat release by cumulus convection is also important in the development of mid-latitude baroclinic systems. Smagorinsky(1956) indicated that if a baroclinic flow was to develop correctly, then heating by convection needed to be considered. Danard(1964) found a similar result. Cumulus heating was found to increase the thickness pattern of mid-latitude systems, and also increase secondary flows. Tracton(1973) found that cyclones intensified more rapidly within numerical models when convective heating was included, bringing better agreement with observation. Recent studies by Bosart(1981) suggested that cumulus parameterization is still far from adequate in numerical models. The effects of meso-scale convective systems embedded within mid-latitude flows has been studied by Fritsch and Chappell(1980b), and Maddox et al(1981). These indicated that the effects of such systems were large.

Other large-scale and meso-scale systems effected by cumulus convection are tropical easterly waves (Wallace(1971), Nitta(1972)) and cold fronts (Ross and Orlanski(1978), Williams et al(1981)).

1.2 OBSERVATIONAL STUDIES RELATED TO CUMULUS PARAMETERIZATION.

Cumulus parameterization is thought to be possible because of a scale separation (in both space and time) between convective and large-scale motion. Hence the development of large-scale flow due to convection can be obtained by considering the net effects of an ensemble of cumulus clouds, instead of having to consider each convective element (Arakawa and Schubert(1974)). The ensemble of clouds is specified in terms of the variables of the large scale flow, and is usually considered to be in equilibrium with large-scale processes.

Any cumulus parameterization scheme is required to:-

1. Predict the vertical distribution of the latent heat release by convection, and also the vertical transports of dynamical and thermodynamical quantities by convection.

2. Predict the magnitude of the convective activity (the closure problem).

To facilitate the development of parameterization schemes, and to bring better understanding of cloud - large-scale interaction, several observational data sets have been collected (Marshall Islands, GATE, BOMEX in the tropics, AMTEX, SESAME in mid-latitudes). From such data sets the vertical distribution of heating and moistening due to convection can be diagnosed (Yanai et al(1973), Cho and Ogura(1974), Nitta(1975), Thompson et al(1979), Kuo and Anthes(1984)c). Rather fewer attempts have been made to diagnose the dynamical effects of convection (Cho et al(1979), Stevens(1979)).

However from observations, little information can be obtained concerning the mechanism by which cumulus cloud effects the large-scale, and also how large-scale processes control convection. The observational analysis can be extended through the use of one dimensional cloud models (eg. Simpson and Wiggert(1969), Johnson(1976)). From these cloud scale parameters can be obtained using large-scale data. Ooyama(1971) developed a mathematical theory for use with such models, in which clouds effected the large-scale atmosphere by subsidence within cloud environment, and detrainment. The equations can be used in both a prognostic sense (Arakawa and Schubert(1974)) within parameterization schemes, or in a diagnostic manner on observational data (Yanai et al(1973,1976), Nitta(1975), Shapiro and Stevens(1980)). A brief review of observational results, related to cumulus parameterization is presented in chapter 2.

A further use of observational data is the semi-prognostic test of parameterization schemes. Here large-scale variables are prescribed and the cumulus effects calculated. However these are not allowed to feed back upon the large-scale. The sequence of cumulus heating and moistening can then be compared to values obtained directly from observations. Examples of such techniques are found in Lord(1982), Krishnamurti et al(1980,1983).

Several doubts must be expressed concerning the use of simple cloud models in such studies. They are crude, containing little dynamical information. Many do not include any downdraught

structure, and when represented it is usually modelled by an inverted plume (eg. Johnson(1976)), the downward mass flux being related to that of the updraught. The details of such a relationship are not well known, and probably vary between cloud types. Also the nature of the inflow into downdraughts is unclear (cloudy or environmental air?). Cloud microphysics are also poorly represented, water loading playing an important role in updraught and downdraught buoyancy. Also in parameterization schemes, the mean effect of convection is required, and it is unclear whether the models represent the mean life cycle of a cloud.

Shear is known to organize convection (Newton and Newton(1959), Browning and Ludlam(1962)), and several distinct types of convection are known to occur within the atmosphere. Moncrieff has developed two dimensional analytical models to represent various flow configurations (Moncrieff and Green(1972)). Each type has different transport properties of dynamical and thermodynamical variables (Moncrieff(1981)). Use of such models in observational studies may be better than one dimensional models, although as yet the thermodynamics of such models are crude.

Further questions must be raised concerning the availability of observational data. Few complete data sets exist. Hence in semi-prognostic tests, parameterization schemes can only be compared to a limited set of possible atmospheric conditions. Hence a comprehensive picture of how well a parameterization scheme performs cannot be obtained.

The closure problem is far from being well understood. In the tropics large-scale dynamical processes appear dominant in controlling convective activity, rather than thermodynamic factors (Malkus and Williams(1963), Reed and Recker(1971), Soong and Frank(1983)). In mid-latitudes there is more evidence for the thermodynamic sounding controlling convection, together with dynamical processes (Zawadzki et al(1981)).

1.3 THE DEVELOPMENT OF PARAMETERIZATION SCHEMES.

The first parameterization schemes were developed in the early 1960's and were crude, representing stop-gap measures until the processes involved were better understood. Manabe et al(1965) adjusted the atmospheric thermodynamic profile back to a Θ_e curve when convection occurred. This scheme was used in GCM's and synoptic scale forecast models. It was not very successful, producing far too much rain at the surface (Krishnamurti et al(1980) compares various forms of these schemes with GATE rainfall data). Other early schemes were based upon penetrative convection, and used in tropical cyclone models (Ooyama(1964)). These had an integral constraint upon the total heat release (usually related to the moisture convergence in the friction layer). However the vertical distribution was calculated more by "intuition" of the individual researcher than by hard facts.

Kuo(1965,1974) developed a scheme along these lines, in which convection rained out a fraction, $(1-b)$, of the moisture convergence into a column of the atmosphere. The vertical distribution of heating was specified by the cloud environment temperature difference (cloud temperature being specified by a moist adiabat from cloud base). Heating of the large-scale was assumed to occur by mixing of the cloud and environment air. Problems with this scheme were the determination of b , and also that b cannot be negative, as sometimes observed in the tropics (Thompson et al(1979)).

The development of better schemes occurred with the increase in observational data in the early 1970's. Mass flux schemes were developed using Ooyama's theory and simple cloud models. These predicted the mass flux within convective clouds from large-scale variables. Arakawa and Schubert(1974) developed a complex scheme using an ensemble of analytical cloud models, which are assumed to remain in quasi-equilibrium with large-scale processes. This scheme has had some success in large-scale models, although is expensive to implement.

Anthes(1977) developed Kuo's original scheme, using a simple cloud model to specify the vertical distribution of heating. Further work has been carried out in recent years to provide a better definition of the parameter b (Krishnamurti et al(1983), Kuo and

Anthes(1984)c), and to include the effects of downdraughts (Johnson(1984), Molinari and Corsetti(1984)).

Miller and Moncrieff(1983) have developed a scheme using dynamical analytical cloud models (Moncrieff and Green(1972)) with a formulation similar to that of Ooyama(1971). Convective activity is predicted by the available potential energy within the atmospheric thermodynamic profile.

Recently Betts(1983) has suggested a return to adjustment schemes, based upon the observations of convective atmospheres. This scheme is still in the early stage of development (Betts and Miller(1984)), but has the advantage over schemes using cloud models of simplicity.

1.4 MESO-SCALE FEATURES.

With the increase of observational data has come the recognition that meso-scale convective features exist within the atmosphere. In the tropics, Squall lines (Houze(1977), Zipser(1977)) and organized cloud clusters (Leary(1979), Ogura et al(1979)) have been identified, mainly through GATE observations (reviewed by Houze and Betts(1981)). In mid-latitudes Squall lines have also been observed (Newton(1967)), together with supercells (Marwitz(1972a)) over the mid-western United States, and multi-celled clusters (Marwitz(1972b)), Heymsfield et al(1980)).

As well as convective towers embedded within these systems, convective processes are also seen on the meso-scale. Deep layers of stratiform cloud are seen in the tropics, extending from just above the freezing level to the top of the troposphere. These are associated with stratiform rain and contain meso-scale updraughts in the cloud layer, and meso-scale downdraughts driven by rain evaporation below this (Zipser(1977), Gamache and Houze(1982)). Although vertical motions are weak ($w < 10 \text{ cm s}^{-1}$), the area of such systems is large (having horizontal dimensions of several hundreds of kilometers), and they appear to play an important role in the mass balance of the atmosphere, and also its heat and moisture budget

(Johnson(1980,1984), Leary and Houze(1980), Molinari and Corsetti(1985)). The distribution of heating is different from that associated with convective towers, having a maxima in the upper troposphere, instead of the mid-troposphere.

A few modelling studies of mid-latitude meso-scale features have been attempted using coarse meso-scale models (Kreitzberg and Perkey(1976), Fritsch and Chappell(1980b)). Their effect upon the large-scale flow was found to be large.

The parameterization of such features is not well understood. They exist on the grid scale of present large-scale models, and it is unclear which processes need to be parameterized, and which are resolved by the model. The magnitude of cumulus towers embedded within the meso-scale circulation may be controlled by both meso-scale and large-scale processes, making the closure problem more difficult.

1.5 COARSE MESO-SCALE MODELS.

Since the mid-1970's coarse meso-scale models have been developed for use both as research and forecasting tools (Kaplan and Paine(1973), Kreitzberg et al(1974), Tapp and White(1976), Carpenter(1979)). These have grid lengths of between 10 and 50km, with domains of up to several thousand kilometers in length.

This brings new problems when the parameterization of convection is considered. Each grid point now does not contain an ensemble of clouds but perhaps only one or two convective towers. Also the convective effect will probably extend beyond the grid box where the convection occurs. The scale of the convective effect is still not well known. Cloud motion between grid boxes also has to be considered. Meso-scale model time steps are usually less than cloud lifetimes (on the order of 30 minutes). Hence the time evolution of a convective cloud may be resolved, and this may be of some importance.

The closure problem of cumulus convection on these scales is also unclear. With such short time scales being considered, convective activity does not necessarily balance the thermodynamic effects of the large-scale processes, as on larger space and time scales.

Several parameterization schemes for use with meso-scale models have been developed employing one dimensional cloud models, and estimating convective activity from the available potential energy contained within the atmospheric thermodynamic profile (Kreitzberg and Perkey(1976), Fritsch and Chappell(1980a)). These have been used in the study of mid-latitude meso-scale systems.

1.6 NUMERICAL CLOUD MODELS AND PARAMETERIZATION.

One of the aims in developing numerical cloud models was to aid the development of parameterization schemes. However only a few limited studies have been carried out with this aim specifically in mind. Mainly cloud models have been used to study the problem of cloud dynamics, modelling individual clouds only.

The first cloud model was published by Malkus and Witt(1959), being two dimensional, and was used to study dry convection. Many similar models followed (Lilly(1962), Ogura(1962), Orville(1968)), either being slab-symmetric, or axi-symmetric, and dry or containing only simple representation of liquid water processes. Later as computational power increased, more complex models were developed (Schlesinger(1973), Takeda(1971)), with better representations of cloud microphysics. In the early 1970's three dimensional models were developed (Steiner(1973), Miller and Pearce(1974), Schlesinger(1975)). Much has been achieved in understanding the dynamics of cumulonimbus storms with such models.

Using numerical models in studying the parameterization problem has several advantages over observational data. Many of the problems raised in section(1.2) are overcome. Clouds are explicitly represented, both updraught and downdraughts being present. The interaction of these features can be studied. Also the numerical model allows a better representation of cloud microphysics, and hence

important processes such as water loading. Hopefully cloud activity will respond to large-scale processes prescribed to the model, and so the question of closure can be considered with more freedom than with observational data. Cloud models have also been found to be able to represent the organization of convection under the action of wind shear, and so provide a method by which the consequences of such effects, for parameterization, can be studied.

Numerical models also provide much better spatial and temporal resolution of the convective atmosphere than observations ever could. Hence the cloud - large-scale interaction can be studied in much greater detail (as well as cloud-cloud interactions). Time dependent effects can also be considered, these perhaps being important in the parameterization of convection within coarse meso-scale models.

Cloud models can also provide a more comprehensive view of the convective response to large-scale atmospheric structure and processes (if the model is a reasonable representation of the real atmosphere). Hence through comparison of numerical cloud models to parameterization schemes in semi-prognostic tests, a much fuller view of the ability of a particular scheme to represent atmospheric processes could be obtained.

If, as well as resolving convective towers, meso-scale features can also be simulated, then a greater understanding of how to parameterize such features could be obtained.

To study the parameterization of convection within larger scale models, the development of cumulus ensemble models is required, in which several clouds exist simultaneously, and also several different cloud sizes. This implies that large-domains are required, perhaps several hundreds of kilometers long, with a horizontal resolution of around 1km. This restricts modelling studies of this kind to be of a two dimensional nature. However despite the possible advantages of the use of such models, few ensemble models of deep convection have been developed.

Hill(1974) developed a two dimensional ensemble model of shallow convection over Florida, with a domain size of 30km to investigate the spacing of cloud. The same model was used to investigate a deep

convective "ensemble" (Hill(1977)). However because of the limited domain size, only one deep convective cloud existed at any one time.

Soong and Tao(1980) used a two dimensional model with a domain of 62km (horizontal resolution=1km, vertical resolution=0.3km) to study a rainband observed during GATE, and to investigate the response of convection to changes in large-scale forcing. They briefly compared their results to those of a simple Kuo scheme. Using a similar model, momentum transports of a GATE line convection event were studied (Soong and Tao(1984)). Brief comparison was made to the cloud momentum transport scheme developed by Schneider and Lindzen(1976). Here a domain of 120km was employed.

Krueger(1985) has developed a two dimensional ensemble model with a domain of 30km, with a third order turbulence closure scheme to study the interaction of convection with the sub-cloud layer.

1.7 AIM OF THIS STUDY.

In this thesis a two dimensional model of an ensemble of tropical deep convective clouds is developed, using as a starting point the numerical cloud model first described by Miller and Pearce(1974). A domain of 260km is used, with a horizontal resolution of 1km in the horizontal, and 50mb in the vertical, sufficient to resolve individual clouds and their transport properties.

The aims of the study are:-

1. To demonstrate that such a model, with simple representations of large-scale, boundary layer, and radiative processes (prescribed from observational data), is able to represent the convective response of the atmosphere, and so show it is a useful tool for studying cumulus parameterization.
2. Having achieved (1), to use the model to study various aspects and problems of cumulus parameterization discussed above.

This thesis has begun by describing and discussing the general nature of the cumulus parameterization problem. In chapter 2 techniques used for diagnosing the effects of convection upon the atmosphere from observational data are described, and their application to data from a meso-scale cloud model discussed. A brief review of observational data, in relation to cumulus parameterization is presented, and a more detailed description is given of the relevant aspects of the Thompson et al(1979) GATE data set.

Chapter 3 briefly describes the model, and the development of the ensemble model (unsheared). In chapter 4 comparison is made between domain averaged thermodynamic quantities and observation data, to check that the model is a reasonable representation of the atmosphere. Studies of cloud transport properties on the temporal scale of cloud lifetime are also made, with relevance to parameterization within meso-scale models.

Chapter 5 develops a mathematical theory of cumulus convection following Ooyama(1971). However all terms are retained, approximations for use with analytical cloud models not being required. In the second half of the chapter internal cloud structure is described, and a brief comparison is made to one dimensional cloud models.

Chapter 6 presents a diagnosis of the equations developed in chapter 5, and studies the interaction of convection with its large-scale environment. Simple representation of the vertical distributions of convective heating and drying, for use in parameterization schemes are considered. In chapter 7, the second half of the cumulus parameterization problem is considered, that of closure. The relationship between convective and large-scale processes is studied.

Chapter 8 gives a brief (and limited) study of convective momentum transports, and their parameterization. The thesis concludes with a summary, and suggestions for further work.

CHAPTER 2

DATA ANALYSIS, MESO-SCALE MODELS, AND OBSERVATIONS.

In this chapter the methods used for calculating convective effects upon the atmosphere are described. These techniques can be used on both observational data and results of numerical models. The limitations of observational data are discussed, and the use of analytical cloud models to overcome these is briefly described. Results from observational studies are presented, concentrating upon the GATE data set of Thompson, Payne, Recker, and Reed (here after referred to as TPRR(1979)).

2.1 TECHNIQUES OF DATA ANALYSIS.

In this section the techniques used to diagnose the effects of convection upon the large-scale atmosphere are described. Such information is required if the parameterization of cumulus convection within large-scale models is to be carried out correctly. In the past these methods have been applied to observational data eg. Reed and Recker(1971), Yanai et al(1973), Ogura and Cho (1973), Nitta(1975), TPRR(1979), and also to the simulated dynamical and thermodynamical fields of numerical models (Soong and Tao(1980), Dudhia(1984)) as in this study.

2.1.1 Equations Applicable To Area Averaged Quantities.

The equations of fluid dynamics are usually written in a form applicable to a parcel of fluid. Those for potential temperature (θ), mixing ratio (q), and horizontal momentum are (in two dimensions),

$$\frac{D\theta}{Dt} = \frac{LQ}{c_p \pi} + S_\theta - \left(\frac{\partial \theta}{\partial t} \right)_{RAD} \quad \text{eqn(2.1)}$$

$$\frac{Dq}{Dt} = -Q + S_q \quad \text{eqn(2.2)}$$

$$\frac{Du}{Dt} = f_v - g \frac{\partial h}{\partial x} + S_u \quad \text{eqn(2.3)}$$

where $\frac{D\phi}{Dt}$ is the total derivative, $\frac{\partial \phi}{\partial t} + u \frac{\partial \phi}{\partial x} + \omega \frac{\partial \phi}{\partial p}$,

h is the height of a constant pressure surface,

$-Q$ is the moisture source/sink due to

condensation,

$\left(\frac{\partial \theta}{\partial t} \right)_{RAD}$ is the radiational source/sink of heat,

S_ϕ are turbulence effects,

and π is $(p/p_0)^{-R/c_p}$, with $p_0=1000\text{mb}$.

These equations are not necessarily applicable to observational data. This is usually collected with a data point spacing of greater than 50km, when studying cloud ensembles and the atmosphere on the large-scale. Hence the measurements must be viewed as being representative of the atmosphere over a large area. A similar situation exists in large-scale numerical models, which usually have a grid spacing of greater than 100km. Grid point values are taken to be average values taken over the entire grid box. Thus before eqns(2.1) to (2.3) can be applied to observations, or used in a large-scale model, they must be averaged over some horizontal area A .

Each variable ϕ is split into an averaged component, and an eddy component, such that,

$$\phi = \bar{\phi} + \phi' \quad \text{eqn(2.4a)}$$

$$\text{where } \bar{\phi} = \frac{1}{A} \iint_A \phi \, dA \quad \text{eqn(2.4b)}$$

$$\text{and } \bar{\phi'} = 0 \quad \text{eqn(2.4c)}$$

$\bar{\phi}$ is associated with the observed values or grid point value in a numerical model, while ϕ' is associated with convection occurring within the area A. Such motions are unresolved by observations, and grid point values. Such a philosophy is only applicable if the scale of the eddies is much smaller than that of the area A. If the eddy scale and A are of similar size, then it is not correct to assume that the area average defines the environment in which the eddy exists.

Applying this to the potential temperature equation (written in flux form), the result is,

$$\begin{aligned} \frac{\partial \bar{\theta}}{\partial t} + \frac{\partial \theta'}{\partial t} + \frac{\partial \bar{u}\bar{\theta}}{\partial x} + \frac{\partial \bar{u}\theta'}{\partial x} + \frac{\partial u'\bar{\theta}}{\partial x} + \frac{\partial u'\theta'}{\partial x} + \frac{\partial \bar{w}\bar{\theta}}{\partial p} + \frac{\partial \bar{w}\theta'}{\partial p} + \frac{\partial w'\bar{\theta}}{\partial p} + \frac{\partial w'\theta'}{\partial p} \\ = \frac{LQ}{C_p\pi} + S_\theta - \left(\frac{\partial \theta}{\partial t}\right)_{RAD} \end{aligned} \quad \text{eqn(2.5)}$$

Applying the averaging operator, eqn(2.4(b)) to eqn(2.5), assuming A is constant, and collecting eddy terms on the rhs,

$$\begin{aligned} \frac{\partial \bar{\theta}}{\partial t} + \frac{\partial \bar{u}\bar{\theta}}{\partial x} + \frac{\partial \bar{w}\bar{\theta}}{\partial p} = \frac{LQ}{C_p\pi} - \frac{\partial \bar{u}\theta'}{\partial x} - \frac{\partial \bar{u}\theta'}{\partial x} - \frac{\partial \bar{u}\theta'}{\partial x} \\ - \frac{\partial \bar{w}\theta'}{\partial p} + \bar{S}_\theta - \left(\frac{\partial \theta}{\partial t}\right)_{RAD} \end{aligned} \quad \text{eqn(2.6)}$$

It is usually assumed that there is little correlation between velocities, and thermodynamic variables in the horizontal, while they are well correlated in the vertical. Thus the horizontal derivatives on the rhs are usually neglected when compared to $-\partial \bar{w}\theta'/\partial p$. Such terms may become important when the scale of area A approaches the eddy scale. However in such a case, the technique used above for averaging the equation may be inappropriate. Eqn(2.6) simplifies to,

$$\frac{\partial \bar{\theta}}{\partial t} + \frac{\partial \bar{u}\bar{\theta}}{\partial x} + \frac{\partial \bar{w}\bar{\theta}}{\partial p} = \frac{LQ}{C_p\pi} - \frac{\partial \bar{w}\theta'}{\partial p} + \bar{S}_\theta - \left(\frac{\partial \theta}{\partial t}\right)_{RAD} \quad \text{eqn(2.7)}$$

A similar analysis for q, and u yields,

$$\frac{\partial \bar{q}}{\partial t} + \frac{\partial \bar{u}\bar{q}}{\partial x} + \frac{\partial \bar{w}\bar{q}}{\partial p} = -Q - \frac{\partial \bar{w}q'}{\partial p} + \bar{S}_q \quad \text{eqn(2.8)}$$

and,

$$\begin{aligned} \frac{\partial \bar{u}}{\partial t} + \frac{\partial \bar{u}^2}{\partial x} + \frac{\partial \bar{u}\bar{w}}{\partial p} + g \frac{\partial (\bar{h} - h_x)}{\partial x} - f\bar{v} = -\frac{\partial \bar{w}u'}{\partial p} - g \frac{\partial \bar{h}}{\partial x} \\ + \bar{S}_u \end{aligned} \quad \text{eqn(2.9)}$$

The rhs of eqn(2.7) and eqn(2.8) are usually denoted by,

$$\pi Q1 - \pi \left(\frac{\partial \theta}{\partial t}\right)_{RAD} = \pi \left\{ \frac{LQ}{C_p\pi} - \frac{\partial \bar{w}\theta'}{\partial p} + \bar{S}_\theta \right\} - \pi \left(\frac{\partial \theta}{\partial t}\right)_{RAD} \quad \text{eqn(2.10)}$$

and,

$$Q2 = -Q - \frac{\partial \bar{w}q'}{\partial p} + \bar{S}_q \quad \text{eqn(2.11)}$$

where Q_1 is the apparent heat source,
 and Q_2 is the apparent moisture source,
 due to convection (Yanai et al(1973)).

Little work has been carried out on the effects of convection upon dynamical fields, and then much of it has been carried out using the vertical component of vorticity (eg Cho et al(1979)). In this study the effect of convection upon horizontal momentum is described in chapter 8. The rhs of eqn(2.9) is denoted by Q_3 , the apparent source of horizontal momentum due to convection, or cumulus friction (Schneider and Lindzen(1976)).

Convection is seen to modify the large-scale thermodynamic fields through moist processes (condensation/evaporation), and vertical transports due to eddy correlation (vertical eddy flux divergence), while it effects the horizontal momentum fields through vertical transports, and eddy pressure gradients. Turbulent effects are included in the definition of Q_1 , Q_2 , and Q_3 , as these are also unresolved by both the large-scale structure of the atmosphere, and by the mesh of large-scale models.

There are two ways in which the area averaged equations set can be utilized:-

1. As the basis for a prognostic model. In this case the terms on the rhs have to be parameterized in terms of the grid point values. This will be studied in detail in chapters 5 and 6.
2. Diagnostically. The lhs is calculated using observed values to give the values of the rhs. The equations are used in this way in the analysis of observational data to obtain the results reviewed later in this chapter, and the analysis of the meso-scale model utilized in this study, in chapter 3 and 4.

It is important to note that an alternative to potential temperature, used by many workers as a quantity conserved in dry motion, is the static energy, s , defined by,

$$s = c_p T + gz \quad \text{eqn(2.12)}$$

It can be shown that,

$$ds \approx c_p T d(\ln \Theta) \quad \text{eqn(2.13)}$$

(s is only approximately conserved in dry motion, whereas Θ is totally conserved.) The equivalent form of eqn(2.7) is,

$$\frac{\partial \bar{s}}{\partial t} + \frac{\partial \overline{u\bar{s}}}{\partial x} + \frac{\partial \overline{\omega\bar{s}}}{\partial p} = LQ - \frac{\partial \overline{\omega's'}}{\partial p} + \bar{S}_s - \left(\frac{\partial \bar{s}}{\partial t} \right)_{\text{RAD}} \quad \text{eqn(2.14)}$$

This equation will be used briefly in the next sub-section. In the remainder of this study, Θ is used, consistent with the thermodynamic equation within the numerical model.

2.1.2 Limitations Of Observational Data.

Diagnostically, the area averaged equation can be used to calculate Q1, and Q2. However the components of these, ie. the contribution due to moist processes, and that due to vertical eddy transports, cannot be calculated independently. This is a limitation of observational data; it can give little information on the mechanisms by which convection affects the larger scale.

Although individual eddy fluxes cannot be obtained, their sum can. This is the eddy flux divergence of total energy. Multiplying eqn(2.8) by L, and adding to eqn(2.14), the result is,

$$c_p \pi Q1 + LQ2 = F_h = - \frac{\partial \overline{\omega'h'}}{\partial p} = - \frac{\partial \overline{\omega's'}}{\partial p} - L \frac{\partial \overline{\omega'q'}}{\partial p} \quad \text{eqn(2.15)}$$

where h is the moist static energy of

the atmosphere defined by,

$$h = c_p T + g z + L q = s + L q \quad \text{eqn(2.16)}$$

h is a measure of the total energy content of the atmosphere, and is approximately conserved in both moist and dry motions.

Static energy has been used in this derivation because it is easier to see the relationship between s, q, and h. However an equivalent analysis may be performed for Θ . In this case,

$$F_{\Theta_e} = - \pi \frac{\partial \overline{\omega'\Theta'}}{\partial p} - \frac{L}{c_p} \frac{\partial \overline{\omega'q'}}{\partial p} \quad \text{eqn(2.17)}$$

Here the F_{Θ_e} is related to the vertical transport of equivalent potential temperature, Θ_e defined as,

$$\Theta_e = \Theta \exp \left[\frac{Lq}{c_p T_s} \right] \quad \text{eqn(2.18)}$$

where q is the mixing ratio of the parcel,
and T_s is the temperature of the parcel at saturation.

θ_e is the potential temperature of a parcel which is lifted through saturation until $q = 0$, all condensed water falling from the parcel. θ_e is also a measure of the total energy of the atmosphere, and is conserved in both dry and moist ascent. It can be shown that,

$$dh \simeq c_p T d(\ln \theta_e) \quad \text{eqn (2.19)}$$

Hence F , and F are directly related to each other, although units are different.

Using the area averaged equations, little more can be achieved in analysing observational data. Further information can be gained through the use of analytical cloud models. Such methods have been used by Yanai et al(1973), Ogura and Cho(1973), Nitta(1975), and Johnson(1976). However results obtained will be dependent on the type of analytical cloud model used, and the cloud height spectrum assumed for the ensemble.

The mathematics of such techniques are directly related to the formulation used in many parameterization of deep convection, and will be described in detail in chapter 5. The area averaged vertical mass flux, $\bar{\omega}$, is written as,

$$\bar{\omega} = \sigma \bar{\omega}^c + (1-\sigma) \bar{\omega}^e \quad \text{eqn (2.20)}$$

where $\sigma = a/A$, the fractional area of cloud within the domain (usually assumed to be much less than 1),

$\bar{\omega}^c$ is the average omega velocity within clouds,
and $\bar{\omega}^e$ is the average omega velocity within cloud environment.

For thermodynamic variables, and horizontal momentum,

$$\bar{\phi} = \sigma \bar{\phi}^c + (1-\sigma) \bar{\phi}^e = \bar{\phi}^e + \sigma (\bar{\phi}^c - \bar{\phi}^e) \quad \text{eqn (2.21)}$$

and so $\bar{\phi} \simeq \bar{\phi}^e$ as $\sigma \ll 1$ and $\bar{\phi}^c \gg (\bar{\phi}^c - \bar{\phi}^e)$

Using this formulation, and shown later, the vertical eddy flux divergence can be written as,

$$-\frac{\partial \bar{\omega} \bar{\phi}}{\partial p} = -\frac{\partial \sigma (\bar{\omega} \bar{\phi}^c - \bar{\omega}^c \bar{\phi}^e)}{\partial p} \quad \text{eqn (2.22)}$$

Relationships are assumed between cloud and large-scale variables, and so cloud properties are calculated. Hence the eddy flux divergences may be estimated separately from the effects of moist processes.

2.1.3 Vertical Profiles And Budgets.

Q1 and Q2 can be calculated as profiles in the vertical if Θ , q , u , and ω are known as functions of height or pressure. The distribution of the cloud effects in the vertical is very important and affects the future evolution of the atmosphere. The form of the distribution is one of the main problems with which the parameterization of convection is concerned.

Vertical integrals of Q1 and Q2 provide information on the precipitation at the surface, and also on surface fluxes of temperature and moisture. Such budgets allow the mean properties of convection to be studied.

Integrating eqn(2.10) from cloud top (at pressure p_t) to the surface (at pressure p_s), the resulting equation is,

$$c_p \pi \int_{p_t}^{p_s} (Q1 - \overline{\left(\frac{\partial \Theta}{\partial t}\right)_{rad}}) dp = L \int_{p_t}^{p_s} Q dp - c_p \pi (\omega' \Theta')_{p_s} \quad \text{eqn(2.23)}$$

$$= P + H$$

where $(\omega' \Theta')_{p_s}$ is the eddy flux at the surface,
and assuming that $(\omega' \Theta')_{p_t}$ is zero.

The vertical integral of condensation and evaporation is equal to the surface precipitation, P , while the eddy flux at the surface is equal to the heat through the surface, H . Both P and H can be estimated independently of Q1, and so the budget analysis provides an independent check on the calculation of Q1. A similar analysis may be applied to Q2.

2.1.4 Analysis Of A Meso-scale Numerical Model.

Horizontal averages over the domain of the model are seen

equivalent to observed quantities, or grid point values in a large-scale numerical model. Both budgets and vertical profiles are used to study convection. With several clouds within the domain at the same time, study of each individual cloud and its effects is impractical.

Using a numerical model much more can be achieved than with an observational data set. Assumptions do not have to be made concerning analytical cloud models, and the dependence of cloud variables upon environmental fields. Cloud values can be obtained directly from the numerical model. This gives greater confidence in the study of the interaction of cloud with their environment.

Budgets of the meso-scale model used in this study, have been studied by Longbottom(1984), using a mid-latitude storm simulation. The budgets used centred space and time differencing schemes. Although the time scheme is different from that utilized in the model(see chapter 3), and the turbulence and filter terms have to be analysed using fields already corrected for such effects, budget errors are found to be small. Budgets computed in this study employ similar techniques, and errors are again small (the difference between $L\bar{Q}/c_p$ calculated from the Θ and q equation being $\approx 1\%$).

Data is available with much greater frequency from a numerical model than in the case of the observations. Hence as well as horizontal averages, time averages are also used. This will smooth out any high frequency component in the horizontally averaged data. This does not cause any disadvantage when parameterization is considered. Parameterization is concerned with the mean effect of an ensemble of cloud over time-scales of the order of cloud lifetimes. Thus knowledge of fine time detail is not required. Time averages of model data will be more representative of the state of the atmosphere over a time period between 0 and T, than a single observation at T/2. Hence a time average model would seem better than unaveraged observational data.

There are differences between Q1 and Q2 as calculated from observations and the meso-scale model. The aim of this study is to look at the mechanism by which an ensemble of deep convective clouds affect their environment. To represent such an ensemble, the domain

must be large enough to contain several deep convective clouds simultaneously. To achieve this (due to computational limitations) the resolution of the grid (1km) is such that small boundary layer clouds are poorly represented, if at all.

Such clouds are important, "ventilating" the sub-cloud layer (the atmosphere below cloud base of deep convection), distributing surface fluxes throughout the layer and increasing it's depth, feeding deep convection. However because they are not well represented within the model, their effects are not present. A crude parameterization of their effect is incorporated in the model, represented by $(\partial\theta/\partial t)_{BL}$, and $(\partial q/\partial t)_{BL}$, and is discussed in chapter 3.

Hence quantities analysed for the meso-scale model, equivalent to those from observational data are,

$$Q1 = \frac{L\bar{Q}}{C_p\pi} - \left(\frac{\partial \overline{\omega'\theta'}}{\partial p} - \left(\frac{\partial \bar{\theta}}{\partial t} \right)_{BL} \right) + \bar{S}_\theta \quad \text{eqn(2.24)}$$

$$Q2 = -\bar{Q} - \left(\frac{\partial \overline{\omega'q'}}{\partial p} - \left(\frac{\partial \bar{q}}{\partial t} \right)_{BL} \right) + \bar{S}_q \quad \text{eqn(2.25)}$$

The effects of resolved cloud processes within the model, upon the environment are designated by $(Q1)_{conv}$, and $(Q2)_{conv}$, defined as,

$$(Q1)_{conv} = \frac{L\bar{Q}}{C_p\pi} - \frac{\partial \overline{\omega'\theta'}}{\partial p} \quad \text{eqn(2.26)}$$

$$(Q2)_{conv} = -\bar{Q} - \frac{\partial \overline{\omega'q'}}{\partial p} \quad \text{eqn(2.27)}$$

Further comment as to whether the model resolves the effects of shallow convection are made in chapters 4 and 6.

2.1.5 Units.

There are several different units in which horizontal average and budgets rates of change of thermodynamic quantities may be expressed. The most obvious are K/s for temperature and g/kg/s for moisture. However these numbers are small; of the order of 10^{-3} to 10^{-4} for atmospheric processes. Daily rates give figures which are much easier to visualise.

An alternative to the above units is that of energy, obtained by multiplying temperature and moisture changes by c_p and L respectively. The quantities are then referred to as sensible and latent heat. A common unit for these is Wm^{-2} . Surface rain rates are often re-expressed from mm/day (equivalent to $mm^3/mm^2/day$) to Wm^{-2} , making comparison with energy budgets easier. Re-expression into terms of energy is useful as moisture plays an important role in atmospheric energetics, through condensation.

Moisture changes may also be expressed in terms of temperature by dividing latent heat by c_p .

In this study, extensive use is made of all these units, and it is useful to note conversion factors between them (table 2.1).

2.2 BRIEF REVIEW OF OBSERVATIONAL RESULTS.

Much work has been previously published on the effects of tropical convection upon the atmosphere as diagnosed from observational data. Three main data sets have been used, collected primarily for this purpose:-

1. The Marshall Island data set. This was collected in the tropical Pacific between April and June 1956. Important studies are Nitta(1972), Yanai et al(1973 and 1982), Ogura and Cho(1973).
2. The "Barbados Oceanographic and Meteorological Experiment" (BOMEX) (1969). This lies within the Atlantic trade wind region.
3. The "GARP Atlantic Tropical Experiment" (GATE), carried out between June and September 1974. Data from phase (30th August to 18th September) has been used most in these studies (TPRR(1979), Johnson(1980), Ogura et al(1979)). A review of GATE results is given in Houze and Betts(1981).

$\frac{\partial T}{\partial t}$	1K/day $\equiv 5.9W/m^2$ (over a 50mb thick layer)
$\frac{\partial q}{\partial t}$	1g/kg/day $\equiv 2.5K/day$ (over a 50mb thick layer) $\equiv 14.8W/m^2$
Surface rainfall	1mm/day $\equiv 29W/m^2$

Table 2.1 Conversion between units
used in budget studies
of the atmosphere.

Other data sets have been used in studying convection, eg. a composite easterly wave data for the Pacific (Reed and Recker(1971)), and VIMHEX data (Betts(1973)).

It would be impossible to give a full description of all results obtained. In section(2.3.1) an over view is given of the main observational results relating to the parameterization problem. Section(2.3.2) describes the TPRR composite easterly wave data set collected during phase of GATE. Data from this is used in the meso-scale model employed in this study (see chapter 3). Comparisons are made between the data and model results in chapter 4, to check that the simulation is reasonable. Many features of tropical convection seen in GATE are typical of those seen in other observational studies.

Thermodynamic results are presented here. A brief review of observational studies of the dynamical effects of convection is presented in chapter 8.

2.2.1 Overview.

The study of parameterization can be divided into main sections:-

- a. mechanisms by which cloud effect their environment,
- b. the relationship between the intensity of convective activity and large-scale processes.

In this section the main features of observational results are presented under these two headings.

(a) Mechanisms by which clouds effect their environment.

Earlier parameterization schemes (eq. Manabe et al(1965)), suggested that under the action of deep convection, the atmospheric profile approached a moist adiabatic equilibrium structure (usually one with the mean of the atmosphere). Studies of Marshall Island

data, and Reed and Reckers 1971 work on a composite tropical wave over the Pacific, showed that such a structure was not characteristic of the convective atmosphere.

Yanai et al(1973) applied a bulk one dimensional cumulus model to Marshall Island data to determine cloud properties, and the mechanisms by which cloud affect their environment. Nitta(1975) extended the method by employing a spectrum of clouds, each type having a different top height. Both methods found that the upward mass flux of air in cloud exceeded that measured on the large-scale. The difference is accounted for by subsidence in cloud environment. The shape and magnitude of Q1 and Q2 was found to be explainable in terms of this subsidence ie. the term,

$$\bar{m}^c \frac{\partial \bar{\phi}}{\partial p} \quad \text{eqn (2.28)}$$

where $\bar{m}^c$ is the mass flux within cloud,
and ϕ is θ or q .

Better agreement was obtained if the evaporation of detrained cloud water was taken into account. Generally Q1 has a maximum in the mid-troposphere, while Q2 peaks at around 700mb.

Johnson(1980) applied a similar bulk cloud model to GATE data (TPRR data set) with similar results. However he also showed that the contribution to m by downdraughts, on both the cloud and meso-scale cannot be neglected.

(b) Relationship between convection and large-scale processes

For parameterization to be possible it is necessary to assume that cloud and large-scale processes can be separated, and that some relation between the two exists.

Reed and Recker(1971) showed that the magnitude of convective rainfall is linked to synoptic scale moisture convergence. This idea had already been used in Kuo's 1965 parameterization scheme, and was the main idea behind CISK. However most work had been carried out assuming that only the boundary layer convergence was related to the convective rainfall. Reed and Recker showed that much of the

convergence occurred above the boundary layer.

Cho and Ogura(1974) showed a correlation between the mass flux inside deep convection and low level convergence. Yanai et al(1976) showed a similar correlation between the mean vertical motion over a large area and deep convective mass flux. No correlation is found for smaller cloud.

Simple parcel theory suggests that the convective available potential energy (CAPE) of the atmosphere, defined as,

$$CAPE = \int_{p_t}^{p_b} \frac{1}{\rho} \frac{\theta'_v}{\bar{\theta}_v} dp \quad \text{eqn(2.29)}$$

where $\theta'_v = \bar{\theta}_v^c - \bar{\theta}_v$,

$\bar{\theta}_v$ is the temperature along a moist adiabat

$\bar{\theta}_v^c$ is the cloud potential temperature,

and p_t , and p_b are the pressures at cloud top and bottom respectively,

controls the magnitude of convective updraughts. However TPRR(1979) showed that when convective rainfall is maximum, CAPE is a minimum. Hence over the tropical oceans the expected relationship between CAPE and convection is called into question. Mid-latitudes show more evidence for thermodynamic control by the atmospheric profile.

2.2.2 The TPRR Data Set.

In this study, data used to initialize the model and to provide values for large-scale cooling and moistening is taken from the TPRR data set. This is a study of easterly waves over the GATE area. Observational data collected during phase III of GATE (30th August to 18th September 1974) was composited to form an average description of one cycle of the wave (the period being approximately 80 hours). A detailed description of the results is to be found in TPRR(1979) together with descriptions of the types of observations used, and techniques of their analysis. Comparison between convective effects simulated by the meso-scale model and those observed is made in

chapter 4, and so a brief description of the data set is provided here. It also serves to highlight in greater detail some of the observational results described in section(2.3.1).

(a) Dynamical structure of an easterly wave.

Burpee(1972) suggested that easterly waves observed over the tropics were caused by joint baroclinic-barotropic instability of the mid-tropospheric easterly jet, which develops over Africa during the summer months. Norquist et al(1977) showed this to be true using GATE data. Wallace(1971), and Nitta(1972) studying waves over the Pacific, suggested that diabatic effects were important in the growth of these features. In the eastern Atlantic, diabatic effects are only important to wave growth over the land. Over the oceans they tend to oppose wave growth.

The 80 hour period of the GATE waves is divided into eight categories, each 10 hours in duration. Category 4 contains the trough of the wave, category 8 the ridge.

Average zonal and meridional wind fields, for the centre of the GATE B-scale array (fig 2.1) are shown in fig 2.2. The zonal wind shows a low level westerly monsoon flow, with a maximum just above the surface. Strong easterlies with a maximum around 600-700mb are seen (a southward extension of the easterly jet centred at 15°N). An upper level jet is also seen. Flow relative to the wave is from the west, except at upper levels.

The mean meridional wind is weak, and oscillates with height. Southerlies are seen below 800mb, around 400mb, and in the upper levels, with northerlies at other levels. A correlation is seen with the mean divergence profile (fig 2.3(a)), southerlies being correlated with convergence, northerlies with divergence. This suggests a relationship between the meridional and Hadley circulations.

The majority of the vertical mass flux within the ITCZ occurs within cumulus convection, and divergence layers are associated with cloud top outflow. Hence the divergence profile suggests 3 cloud populations. The evolution of this profile is shown in fig 2.3(b). As the trough passes over the GATE area the divergence at 550 and

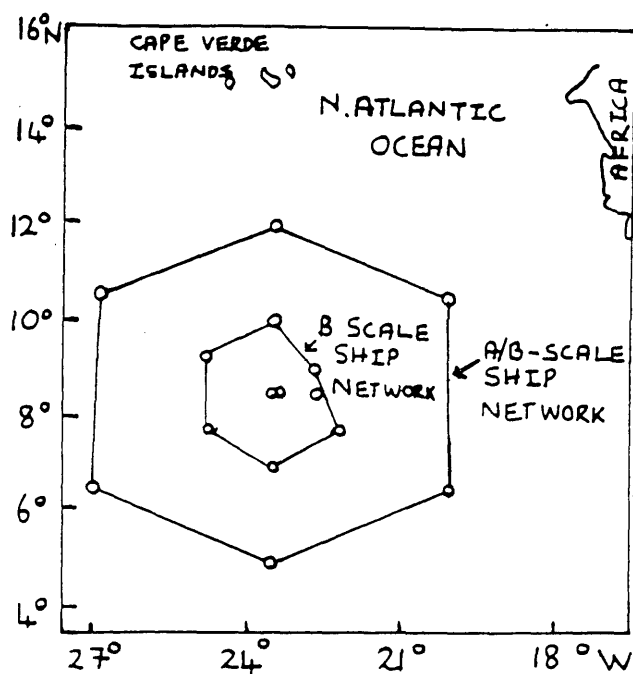


Fig 2.1 GATE ship observation network (fig 1 of TPRR(1979)).

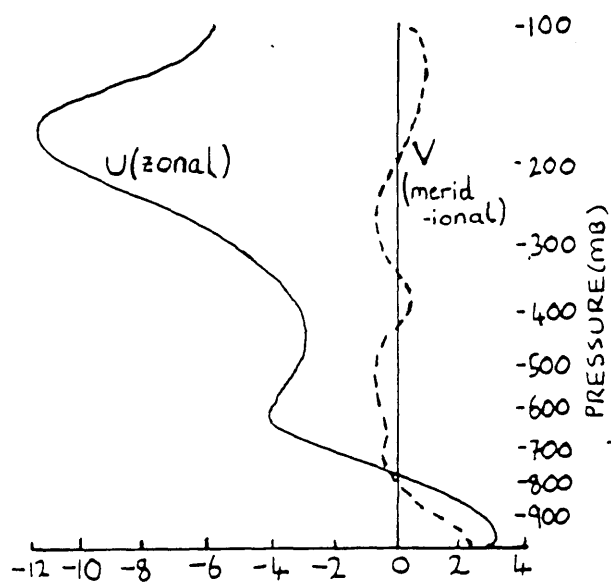


Fig 2.2 Mean zonal and meridional wind components for centre of B-scale network for a composite easterly wave (fig 3 of TPRR(1979)).

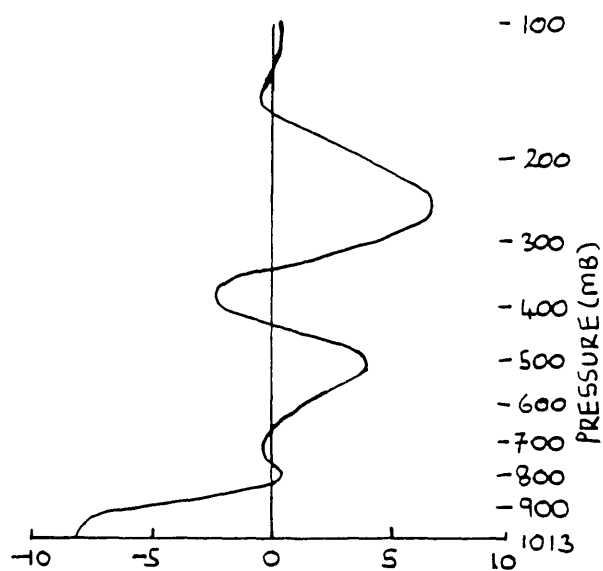


Fig 2.3(a) Profile of mean horizontal divergence for a composite easterly wave (fig 5 of TPRR(1979)).

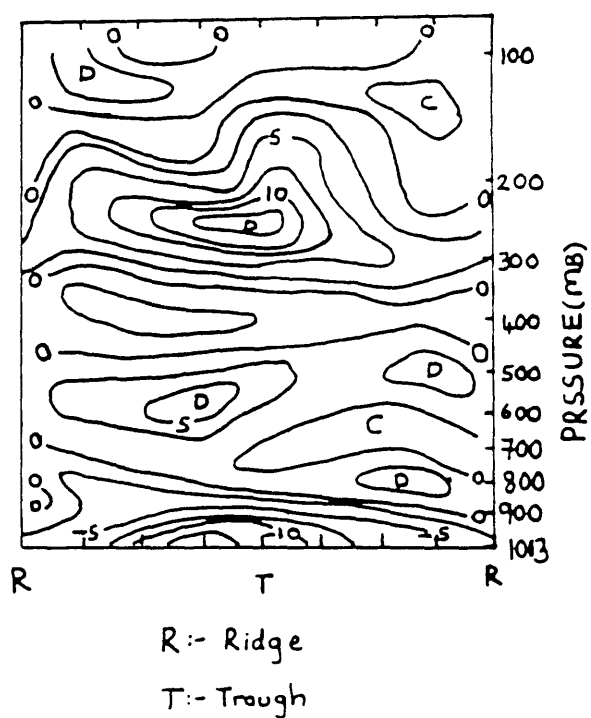


Fig 2.3(b) Evolution of horizontal divergence over one wave cycle (fig 13 of TPRR(1979)).

250mb intensifies, associated with an increase in the mean vertical velocity over the region (fig 2.4). After the trough has passed, activity decreases, and a new divergence layer at 800mb develops.

As the wave passes over the GATE area, large variations are seen in the meridional wind profile, of the same order of magnitude as the mean profile. The upper and lower level flows are correlated, the wave sloping eastward with height. Variations in the zonal velocity are not as great. As convection intensifies, the low level jet decreases, and increases as the ridge moves over the system.

(b) Budget analysis.

Some information concerning the magnitude of convection during various phases of the wave have been indicated when briefly studying the dynamical structure of the wave. In this section observational budgets of moisture and sensible heat are described, showing the relationship between convection, large-scale convergence, surface fluxes and CAPE.

Moisture and sensible heat budgets for all 8 categories of the wave are shown in fig 2.5(a), and (b), together with the mean over all the wave. It is clear that convection is linked to large-scale convergence of moisture and divergence of temperature.

From the moisture budget, convection is a minimum over the ridge of the wave. Precipitation increases as the moisture convergence does. However for the first half of the wave large-scale forcing and surface fluxes exceed the rate of precipitation, causing storage of water vapour in the atmosphere. As the trough passes, large-scale forcing decreases, whereas convection continues. During this half of the wave the atmosphere loses water vapour stored in the first half.

Surface evaporation increases with convective activity, although the correlation is not perfect. Study of daily values showed that the magnitude depended on wind speeds. Southerly winds ahead of the ridge maintained the high evaporation rate as convection decreased.

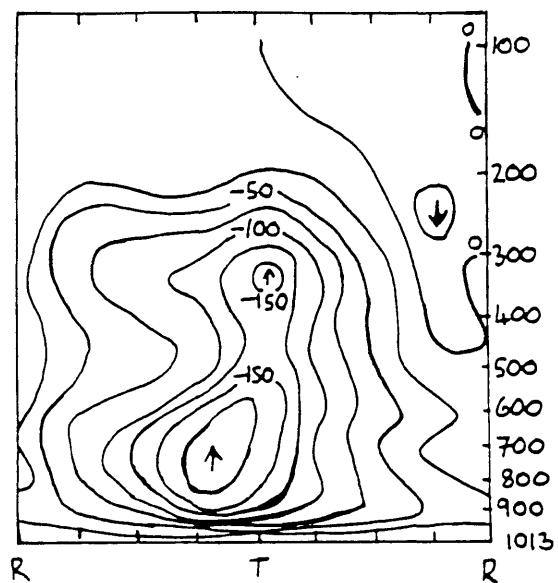
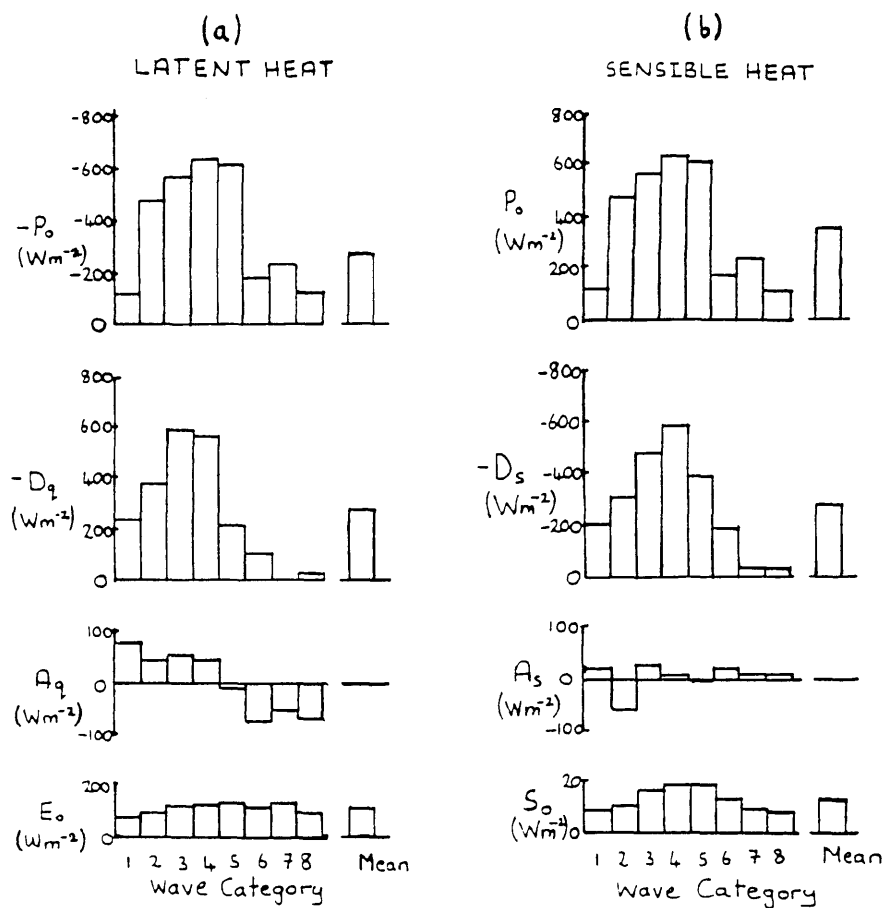


Fig 2.4 Evolution of $\bar{\omega}$ over one wave cycle
(fig 14 of TPRR(1979)).



P_o - Precipitation rate

$-D_q, -D_s$ - Large-scale forcing

A_q, A_s - Storage rate

E_o - Surface evaporation rate

S_o - Surface sensible heat flux

Fig 2.5 (a) Latent heat budget over one wave cycle of the TPRR data set.

(b) Sensible heat budget over one wave cycle of the TPRR data set.

The sensible heat budget shows a similar relationship between large-scale forcing and precipitation. Surface fluxes are relatively un-important in the budget. No systematic storage is observed.

For the moisture budget, $3/4$ of the precipitation is provided by the large-scale moisture convergence, the rest by surface fluxes. For sensible heat $3/4$ of the energy release by convection is balanced by convergence, the rest being balanced by radiation.

Fig 2.6 shows the relationship between surface rainfall and the CAPE for the wave. As precipitation increases, CAPE decreases, due to the convection drying the lower layers of the atmosphere. The nature of this relationship is probably modified when entrainment is taken into account. This would reduce the value of CAPE over the ridge due to the atmosphere being drier than in the trough. However the results indicate that the magnitude of the convection is primarily controlled by the large-scale convergence, and not CAPE.

(c) Thermodynamic vertical structure.

Fig 2.7 shows the profile of temperature and mixing ratio for the GATE area at 2.5 hours. The lowest 100mb is extremely warm and moist, surface temperature being 27°C , with a Θ_e of between 348 and 350K. When lifted such air loses buoyancy between 200 and 100mb. Air in the lowest 150mb is able to undergo buoyant moist ascent if lifted through its saturation level. Thus it is easy to see that the atmospheric profile is able to support active deep convection, such as that observed in GATE.

The TPRR data set is a composite of many observation sites over the GATE domain. Hence no convection actually grew in the soundings which will be used to initialize the numerical model. To confirm that these reasonably represent the convective tropical atmosphere, fig 2.7 also includes a sounding from the ship Researcher, taken before the formation of a line convection event. Comparison is favourable, although more detail is seen in the Researcher profiles, especially for dew point temperature. Both show evidence of a mixed layer, although it is deeper in the Researcher data. TPRR predicts cloud base at 960mb, Researcher at 930mb, giving a difference in the moist adiabats of only a few tenths of a degree.

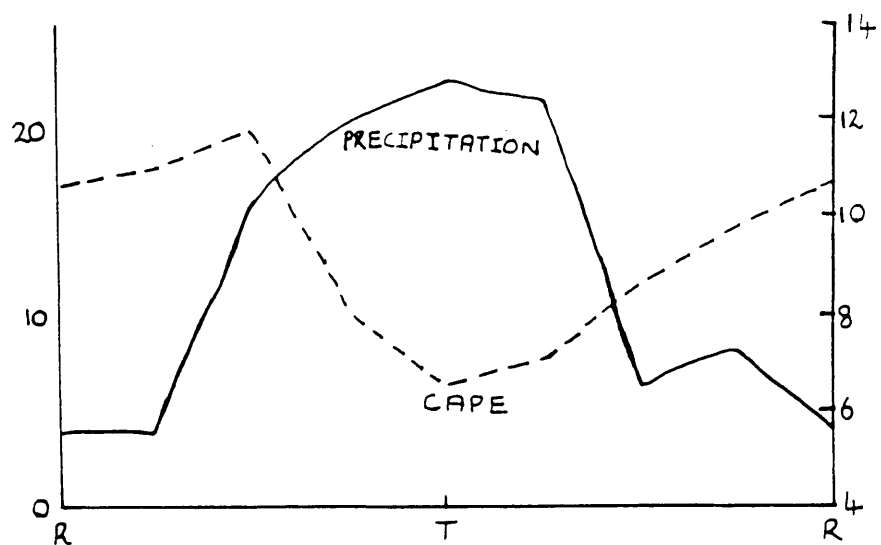


Fig 2.6 Relationship between surface rainfall and CAPE over one wave cycle (fig 19 of TPRR(1979)).

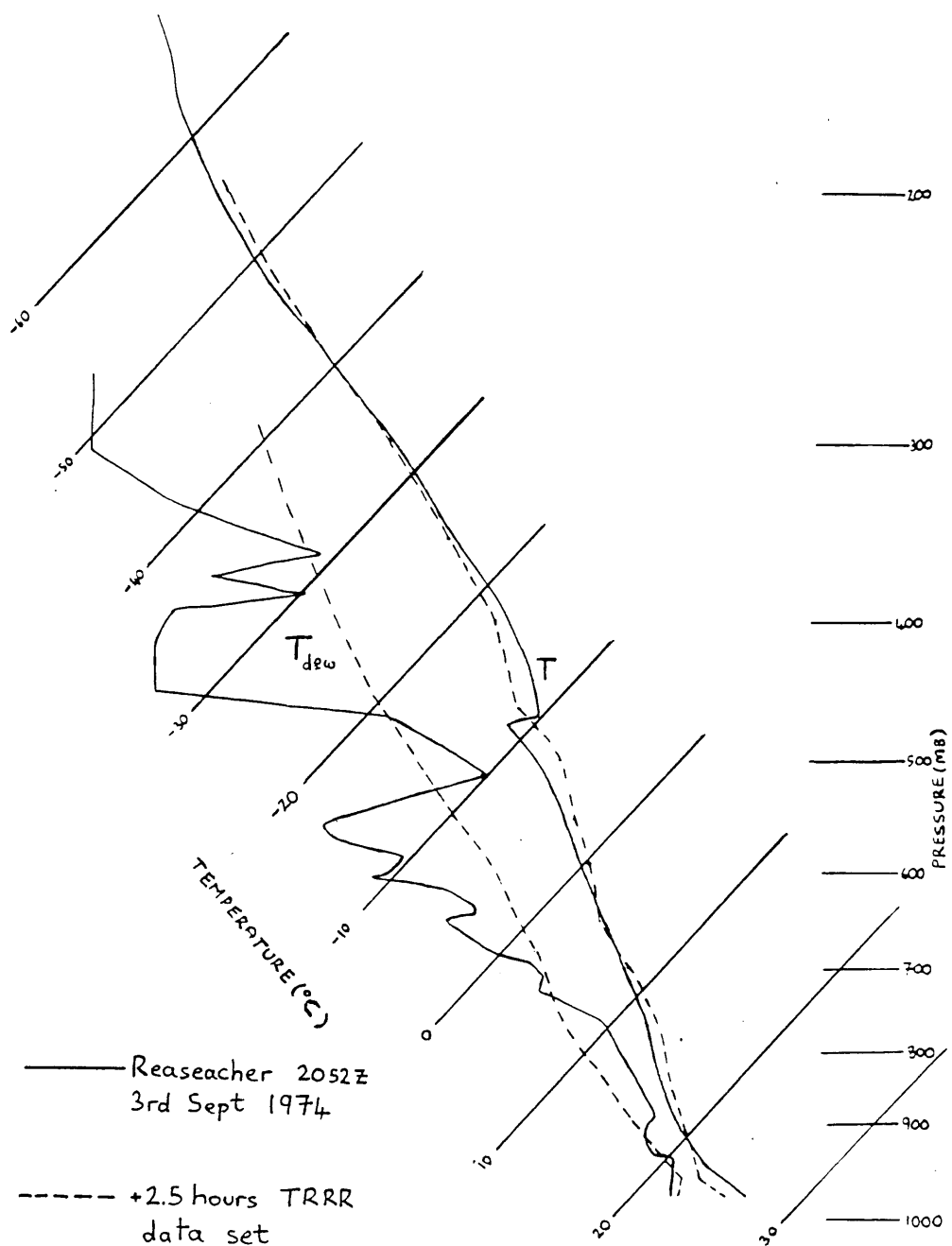


Fig 2.7 Thermodynamic profiles for 2.5 hours of the TPRR data set, and a single profile from the GATE ship Researcher.

Deviations of T and q from the 2.5 hour profile are shown in fig 2.8(a) and (b). T' never exceeds 1K, showing a cooling in the lower levels and a warming above 600mb, the exact shape being controlled by the balance between Q_1 and large-scale forcing. After the passage of the trough, the magnitudes of the deviations decrease. During this period, the budget results showed that convection is suppressed.

q' shows a drying in the lowest 100mb, as the magnitude of the convection increases. Above this moistening occurs. This is too large to be caused by surface drying alone, and so the large-scale convergence field must contribute to it (the storage of water, as indicated by the budgets). As the wave passes from trough to ridge, the lowest 100mb re-moistens, and convection dries the atmosphere above.

These features are understood if the balance of Q_2 and large-scale forcing is studied (fig 2.9(a) and (b)). These follow each other closely at all levels. However in the first four wave categories the large-scale forcing generally exceeds the convective effect, and so leads to the storage observed. For the rest of the wave, as large-scale forcing decays convection dominates, and drying occurs above 900mb. Below 900mb the opposite is true.

Q_1 is generally larger than convergence effects. However the lower atmosphere is observed to cool. This does not seem to be due to downdraught outflow, as Q_1 does not indicate that this exists. Radiation cools the tropical atmosphere, and so a combination of radiative and large-scale forcing must exceed Q_1 in the lower layers of the atmosphere. (TPRR did not calculate a radiative cooling profile from GATE data, but used that of Dopplnick(1970) to check surface energy budgets.)

These results again indicate that the large-scale convergence fields controls the convection. Also convection, in general, tries to balance the large-scale effects throughout the entire depth of the atmosphere. However the exact nature of the balance is difficult to formulate (the "closure" problem of parameterization schemes), due to the storage of water vapour by the atmosphere when convection is growing.

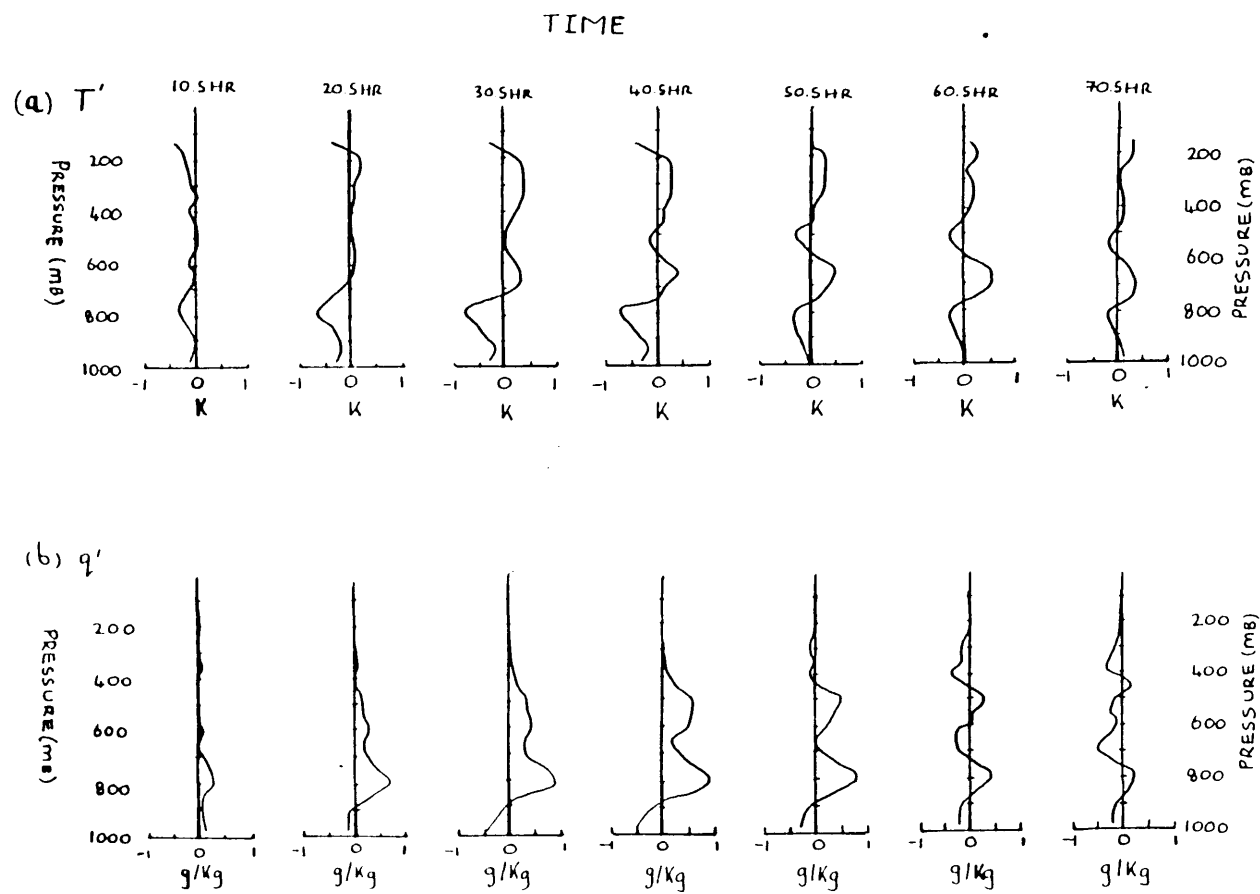


Fig 2.8 Deviations of T and q from the 2.5 profile of the TPRR data set over one wave cycle.

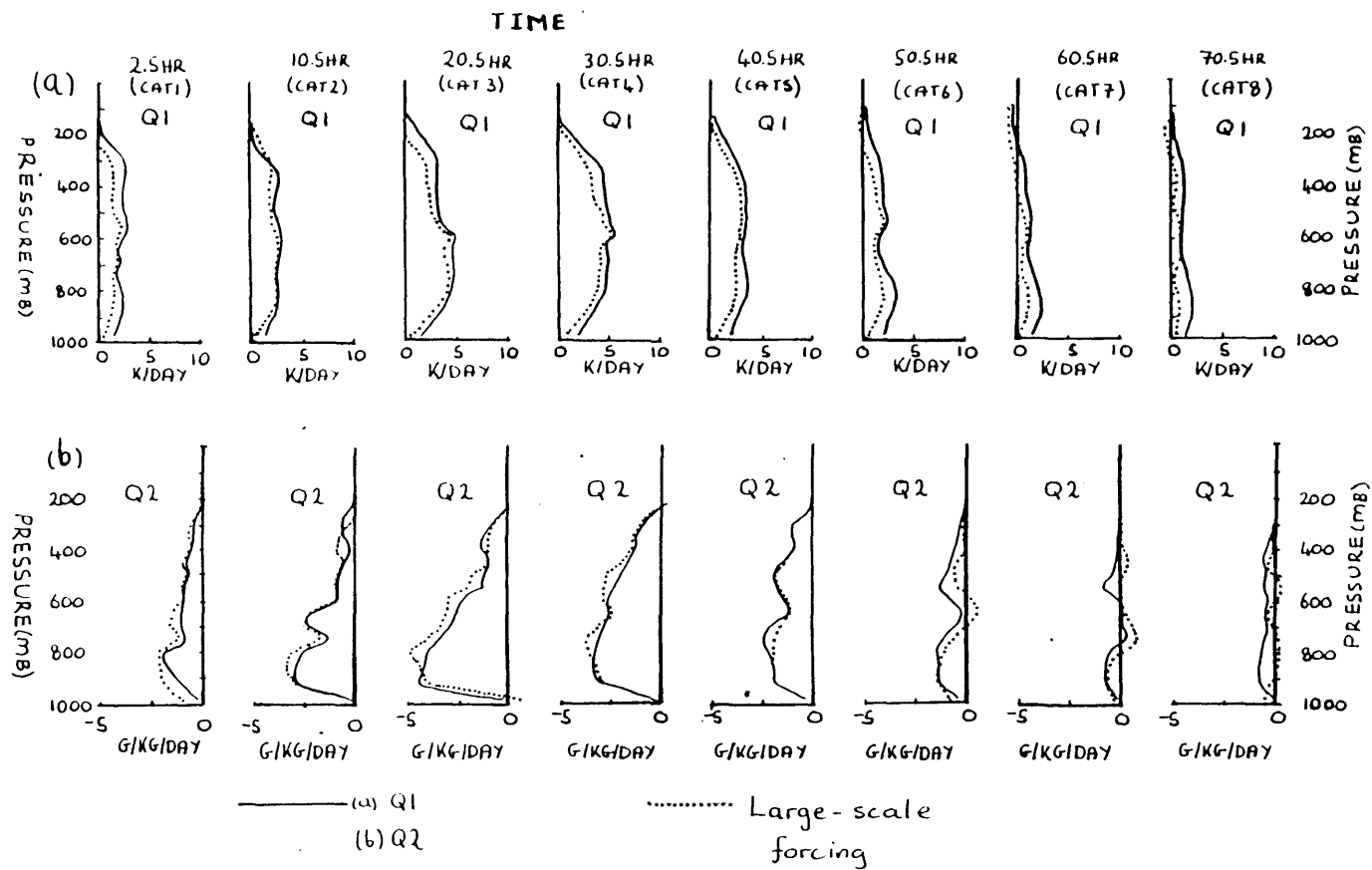


Fig 2.9 Comparison of convective effect and large-scale forcing (a) Temperature (b) Moisture

2.3 SUMMARY.

The mathematical techniques used to analyse the effect of convection upon the large-scale atmosphere, using data collected over a large area have been described. Modification to these techniques when they are applied to a model meso-scale of deep convection have been discussed. Such techniques will be used throughout this study.

A brief review of observational data sets used to study convection in the past was presented, together with a more detailed description of the TPRR GATE data set (TPRR(1979)). This will be used in chapters 3 and 4 to initiate the model, and check if the resulting convection reasonably represents the real atmosphere.

CHAPTER 3

MODELLING OF A DEEP CLOUD ENSEMBLE

In this chapter the model employed in producing a two dimensional simulation of a tropical cloud ensemble is briefly described. The development of such a simulation, together with the problems encountered and the solutions chosen, is presented.

3.1 THE MODEL

3.1.1 Equation Set

A meso-scale numerical model is used to simulate an ensemble of deep cumulus clouds. The model was developed in the early 1970's and is first described in Miller and Pearce (1974). In it's 3D form it has been used to study both tropical convective systems (Moncrieff and Miller (1976), Miller and Betts (1977)), and mid-latitude storms (Miller (1978), Thorpe and Miller (1978)). In a 2D form, such as the one used in this study, mid-latitude storms have been studied (Thorpe, Miller, Moncrieff (1982)).

The primitive equations are used, transformed to pressure co-ordinates. The Boussinesq approximation is made in the vertical momentum equation and non-hydrostatic terms retained. The transformation to pressure co-ordinates and the approximations used are discussed in Miller (1974). The resulting equation set is written in flux form. The momentum and mass continuity equations are:-

Horizontal momentum

$$\frac{Du}{Dt} + g \frac{\partial(h-h_r)}{\partial x} = S_u \quad \text{eqn(3.1)}$$

Vertical momentum

$$\frac{D}{Dt} \left(\frac{\omega}{p_r g} \right) + g \left(\frac{\theta_v - \theta_r}{\theta_r} - l_c - l_r \right) + p_r g^2 \frac{\partial(h-h_r)}{\partial p} = S_\omega \quad \text{eqn(3.2)}$$

Mass continuity

$$\frac{\partial u}{\partial x} + \frac{\partial \omega}{\partial p} = 0 \quad \text{eqn(3.3)}$$

From the two momentum equations and the continuity equation a diagnostic Poisson type equation is formed for $(h-h_r)$. This is solved by using a fourier transform with a recurrence technique in the vertical.

Together with the two velocity components and the height field, potential temperature, moisture, cloud and rain water are also stored within each grid box (the positions of these variables are illustrated in fig 3.1). The equations for these variables are:-

Potential temperature

$$\frac{D\theta}{Dt} = \frac{LQ}{c_p \pi} + S_\theta \quad \text{eqn(3.4)}$$

Moisture

$$\frac{Dq}{Dt} = -Q + S_q \quad \text{eqn(3.5)}$$

Cloud water

$$\frac{Dl_c}{Dt} = Q - \text{PROD} + S_{l_c} \quad \text{eqn(3.6)}$$

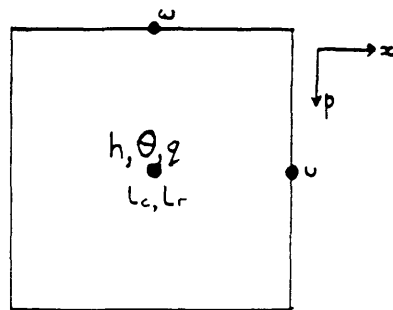


Fig 3.1 Position of variables in a model grid box.

Rain water

$$\frac{Dl_r}{Dt} = \text{PROD} - g \frac{\partial(p_r l_r v_t)}{\partial p} + S_{l_r} \quad \text{eqn(3.7)}$$

where v_t is the terminal fall speed of rain water given by,

$$v_t = 5.32 l_r^{0.2} \quad \text{eqn(3.8)}$$

Equations (3.6) and (3.7) apply to any grid box which becomes saturated. All water above saturation is converted into cloud and rain water. The cloud microphysics is parameterized using the Kessler (1969) scheme, with modifications due to Liu and Orville (1969). No ice phase is included. The scheme is represented by the term PROD in the rhs of eq(3.6) and (3.7). It is composed of three terms:-

(a) Evaporation of raindrops

$$P_1 = \beta (q - q_{\text{sat}}) \quad \text{eqn(3.9)}$$

(b) Accretion of cloud water by raindrops

$$P_2 = \gamma l_c l_r^{0.95} \quad \text{eqn(3.10)}$$

(c) Conversion of cloud water to rain water

$$P_3 = \alpha (l_c - l_{\text{crit}}) \quad \text{eqn(3.11)}$$

where l_{crit} is the value of cloud water below which no conversion takes place.

A typical value for α, β, γ , is 10^{-3} , while l_{crit} is taken to be 0.5g/kg.

Subgrid scale motion, denoted in eqns (3.1) to (3.7) by S_ϕ is parameterized by a Reynold stress turbulence scheme, developed by Brugge and described in some detail in Brugge and Moncrieff (1985). It is based on previous work by Lilly (1962), Smagorinsky (1963),

Deardorff (1970,1972) and Schlesinger (1978), and is typical of schemes used in convection modelling. The scheme acts upon deviations from a reference profile, forcing grid point values back to this reference value.

3.1.2 Numerical Techniques

Space derivatives are represented by a second order centred differencing scheme. The exception to this is the rain water, in which an upstream differencing technique is used to inhibit the formation of negative values in the rain water field.

A mixed time integration scheme is employed, using alternatively a Euler forward step and a Euler backward step (Miller and Pearce (1974)). The resulting scheme is slightly unstable with respect to advection of waves. However the application of a 17 point numerical filter every 16 timesteps keeps short waves from amplifying.

Periodic boundary conditions are applied in the horizontal with a four point overlap. At the upper and lower boundaries omega is set to zero. This is realistic as one represents the tropopause, the other the surface. At these boundaries the nonhydrostatic pressure gradient balances the buoyancy. This provides a gradient boundary condition for $(h-h_b)$ which is used in solving the Poisson equation for $(h-h_r)$.

For the subgrid scale motion vertical derivatives are set to zero at the upper and lower boundaries. This implies that subgrid scale motion conserves momentum, temperature, and moisture when integrated over the domain.

The continuity equation of the model has no prognostic capacity (it contains no $\partial/\partial t$ term). In deriving this equation approximations are made which filter sound waves from the equation set (note that the anelastic approximation is not made - see Miller (1974), Miller and White (1984)). The upper and lower boundary conditions on omega filter the quickly moving surface gravity waves from the equation set. This means that the length of the model timestep, as determined by the CFL condition,

$$\Delta t = \frac{\Delta L}{C_{max}} \quad \text{eqn(3.12)}$$

where ΔL is the length of a grid box,
 and $C_{\max}$ is the maximum speed found
 in the model,

is determined by the slower propagating internal gravity wave,
 ($C \sim 20\text{m/s}$), allowing timesteps of a few tens of seconds, if ΔL is of
 the order of a kilometer.

3.2 SIMULATION OF A DEEP CLOUD ENSEMBLE

The mesoscale model is used to simulate a two dimensional deep cloud ensemble. Similar experiments have been carried out by Hill (1974) and Soong and Tao (1980). However both these studies used domains of limited size, 30km, and 65km respectively, which allows only one cloud to develop at any time. Hence these studies were limited in that the possibility of cloud/cloud interactions were not allowed.

In this study a two dimensional domain of 260km by 900mb (surface pressure is 1000mb) is used. This allows the possibility of several cumulonimbus clouds to exist simultaneously. A horizontal grid length of 1000m, with one of 50mb in the vertical is employed, together with a timestep of 15 seconds. The equations are integrated forward for 9 hours, allowing time for several cloud to grow and decay.

3.2.1 Initial Thermodynamic Profile

An initial profile derived from the TPRR data set (described in chapter two) is used. The profile represents the state of the atmosphere on the mesoscale 20.5 hours after the passage of a ridge of an easterly wave over the GATE area. During the next twenty hours convective activity increases as the trough of the wave moves across the GATE domain. Fig 3.2 shows the temperature and mixing ratio values extrapolated to model pressure levels.

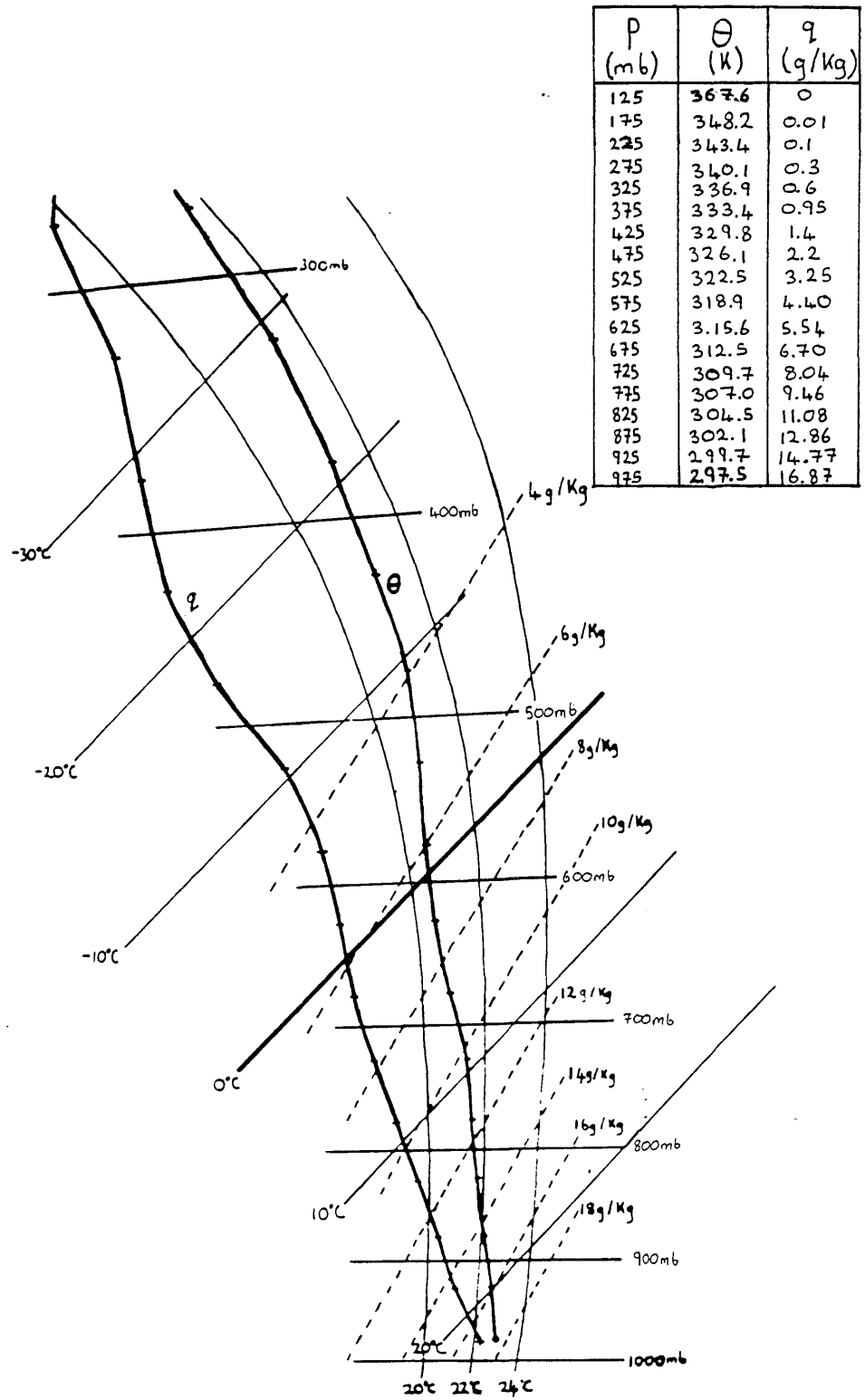


Fig 3.2 T-θ gram of initial model profiles (based upon TPRR 20.5 hours).

The lowest model layers are very moist, air being lifted from 975mb saturating at 965mb, with a Θ_e of 347.5K. Such a parcel would reach 225mb under moist ascent (ignoring mixing and water loading), indicating that deep convection is possible. A plot of Θ_e and Θ_{es} (fig 3.3) indicates that air from the lowest three model levels have positive CAPE values (a typical value of CAPE for such a sounding is 1500-2000J/kg).

Although the profile has been extrapolated to give values of temperature and mixing ratio at model levels, it shows similar features to the observed profile (fig 2.7 of chapter 2). However due to the vertical resolution of the model (50mb), warm moist air between 975mb and the observed surface pressure of 1013mb is not seen by the model. The observed temperatures at 982mb and 1013mb were 23°C and 27°C, with a mixing ratio of 17g/kg throughout this layer. The temperature and mixing ratio of the lowest model layer, centered on 975mb are 22.2°C and 16.9g/kg. Thus the model sounding underestimates the amount of warm, moist air near the surface. (Note that the lower boundary of the model is 1000mb, while the mean observed surface pressure was 1013mb. Thus much of the moist air lies below the lower boundary of the model!)

3.2.2 Wind Profile

For these initial simulations no initial vertical wind shear is introduced. The interaction between such a shear and cloud is poorly understood, and it's inclusion complicates the dynamics of the system.

The simulation is used to study the parameterization of deep convection. However to the present few attempts to parameterize cloud/shear interaction have been made. The inclusion of shear would complicate the comparison between model results and existing parameterization schemes. However the cloud/shear interaction may have important consequences for the parameterization of convection, and a shear is included in later simulations described briefly in chapter 8.

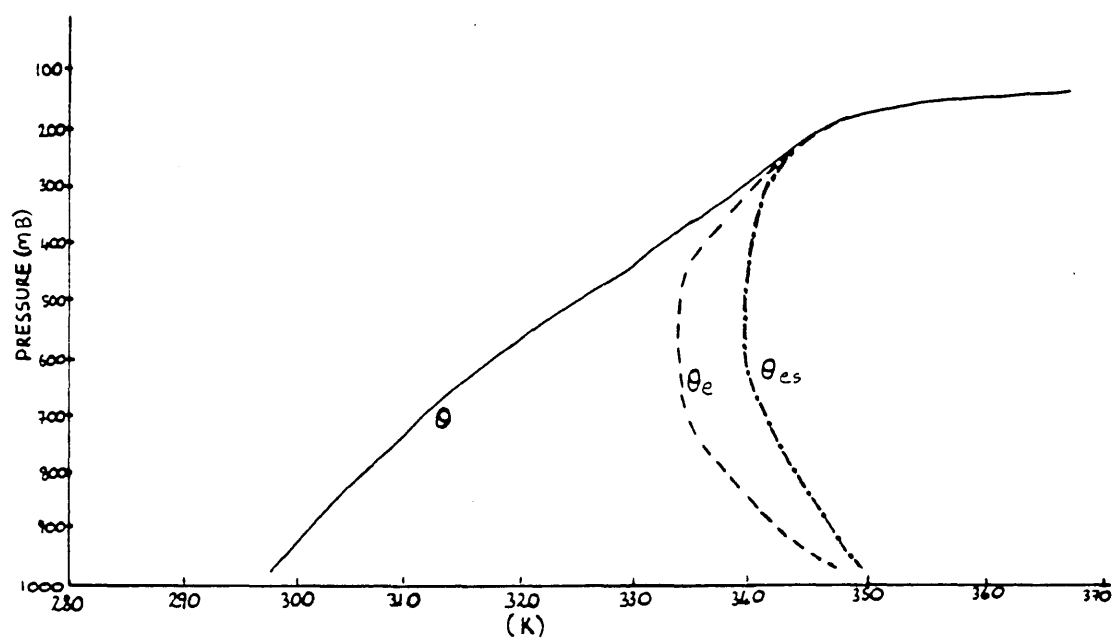


Fig 3.3 Thermodynamic structure of the model initial profile.

3.2.3 Cloud Forcing

It is usual in a single cloud simulation to initialize the convection by supplying a buoyant bubble (which represents an initial convective cell) or a cold pool at the surface (representing downdraught air). Both of these methods specify the position of the resulting convection. In simulating an ensemble of clouds this may not be desirable.

An ensemble consists of several clouds which are interacting with their environment and with each other. The scale on which such clouds are separated influences both these interactions. Such a scale is best determined by the simulation, rather than by an external process.

Previous workers have used random number generators to provide temperature perturbations to force convection, choosing a different random number for each grid box. Hill (1974,1977) applied such perturbations to the lowest layer of his model, their average magnitude being based on observations of surface fluxes. They were applied over a period of 15 minutes. Soong and Tao (1980,1984) used perturbations with an average of zero at 450m and the surface of their model. These ranged from -0.5 to 0.5K and were applied over 5 minutes. Soong and Tao state in their 1980 paper that " the random disturbances do not control the amount or intensity of deep or shallow clouds ". This would seem an ideal situation. However several problems were encountered when trying to initialize convection.

The literature contains an absence of work on the initial development of deep convection. However over tropical oceans they are often seen to develop from fields of boundary layer cumulus. Experiments were carried out using random number perturbations with a horizontal scale of 4km, applied at levels 17 and 16 (925 and 875mb). This roughly corresponds to the top of the boundary layer, to which the small cumulus are confined. The mean value of the random forcing was zero. To obtain deep convection it was found necessary to keep the forcing fixed for a period of 30 minutes (approximately the lifetime of a small cumulus cloud). Perturbations in the range -0.12 to 0.12 K/minute were used. Deep convection developed within the first few hours of the simulation. This cloud forcing is used throughout the rest of the study.

Convection continued during the whole of the simulation (up to 9 hours) if cloud forcing was applied throughout. However if withdrawn the convection decayed on a scale of 2 to 3 hours. The introduction of a horizontal wind field may have aided the continuation of convection by causing convergence at the head of convectively induced downdraughts, producing lifting which may be sufficient to induce new convection.

3.2.4 Large Scale, Radiative And Surface Forcing

Deep convection warms and dries the atmosphere, stabilizing it and so inhibiting further convection. However in the ITCZ (in which the GATE area lies), the net large scale vertical motion is upward. This causes the atmosphere to cool and moisten, decreasing atmospheric stability and so allowing the possibility of further convection. Hence the large scale and convective motions act against each other, maintaining the conditional instability of the atmosphere.

Soong and Tao (1980) studied the response of deep convection to changes in large scale forcing. They found that the convective heating and moistening balanced the large scale thermodynamic effects. Without such large scale ascent no deep clouds formed within their simulations. Dudhia (1984) carried out similar motion using a 3D version of the model used in this study. He studied a rainband observed over the GATE area and found that the strength of the convection was linked to the strength of large scale ascent.

As discussed in chapter 1, one of the major concerns of the parameterization of deep convection is what controls the magnitude of such motions, and how is it related to the magnitude of the large scale motions. It is important that large scale forcing be included in a study of this kind, where control mechanisms are investigated. Also if the simulation is to be extended over several cloud lifetimes, it is important to balance the thermodynamic effect of the clouds by the large scale forcing.

Such large scale forcing can be applied in two ways: (a) communicate the observed mean large scale vertical motion to the model and use the domain mean $\bar{\theta}$ and $\bar{q}$ values to form the terms

$$\frac{\partial \bar{\theta}}{\partial t} \propto -\bar{\omega}_{obs} \frac{\partial \bar{\theta}}{\partial p} \quad \text{eqn (3.13)}$$

$$\frac{\partial \bar{q}}{\partial t} \propto -\bar{\omega}_{obs} \frac{\partial \bar{q}}{\partial p} \quad \text{eqn (3.14)}$$

This is physically realistic. However if $\bar{\theta}$ and $\bar{q}$ depart from the observed values the forcing will depart from that observed, making comparison of model results to observations difficult.

(b) Communicate the observed $(\bar{\omega} \partial \bar{\theta} / \partial p)$ and $(\bar{\omega} \partial \bar{q} / \partial p)$ to the model, applying them as a cooling and moistening rate to the rhs of eq(3.4) and (3.5). This makes comparison between the model and observations easier.

In this study the second method was used. Cooling and moistening rates (figs 3.4 and 3.5 respectively) were derived from the TPRR data set for the period +20.5 hours to +30.5 hours. The magnitude of the forcing was changed after the first hour and then every two hours, simulating the passage of an easterly wave over the GATE area. The cooling has a maximum around 600mb of -10K/day, while the moistening has a much lower level maximum at 850mb of 4g/kg/day. Vertically integrated values are shown in table (3.1). The moistening is constant throughout the period simulated, while the cooling increases in magnitude by 10% of the original value.

Radiative cooling (fig 3.6) also plays a part in the tropospheric energy budget. A profile calculated by Cox (1979) for the GATE area is communicated to the model. This has a net value of -128W/m^2 when integrated vertically, a quarter of the cooling due to large scale forcing.

Surface fluxes are also introduced into the model, the sensible and latent heat fluxes being 10W/m^2 and 125W/m^2 respectively. In the real atmosphere these are distributed through the boundary layer by shallow non precipitating convection (section 2.1.4 of chapter 2). To achieve a crude parameterization of this, the surface flux heating and moistening are distributed equally between 975 and 925mb. The heating

Fig 3.4 Cooling due to large-scale ascent (based upon TPRR 28.5 hours).

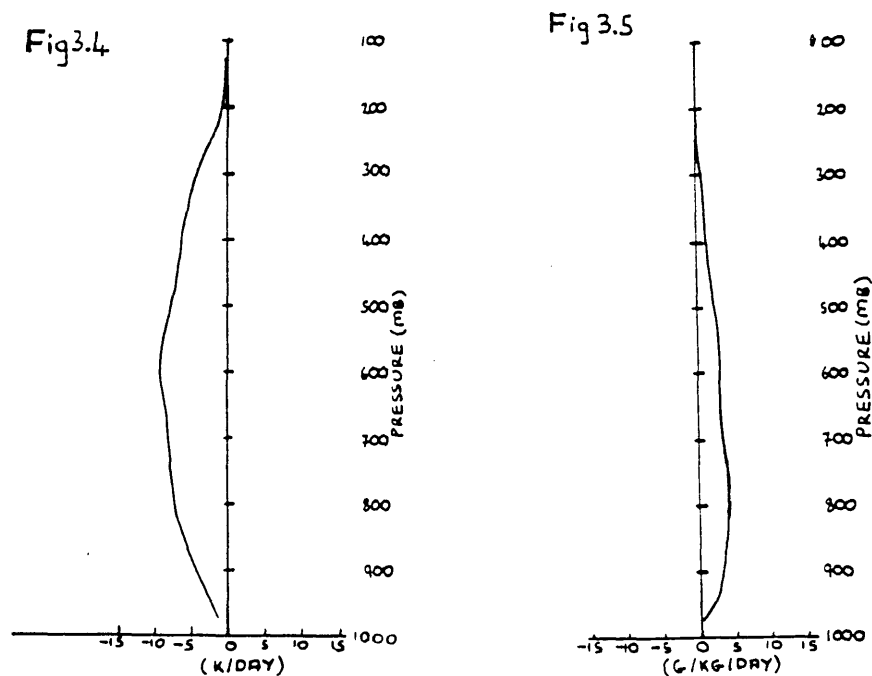


Fig 3.5 Moistening due to large-scale ascent (based upon TPRR 28.5 hours).

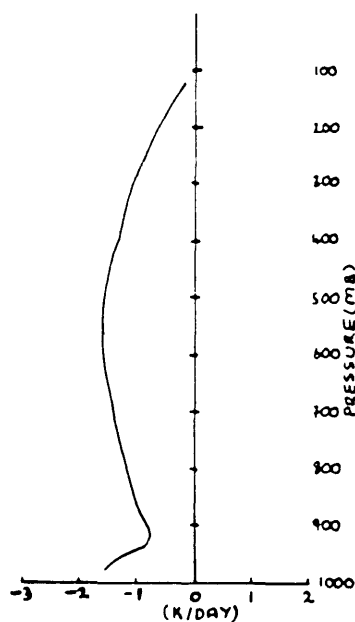


Fig 3.6 Cooling due to radiation (based on Cox et al(1979)).

Time (hour)	$\left(\bar{\omega} \frac{\partial \bar{\theta}}{\partial p} \right)_{obs}$ (W/m ²)	$\left(\bar{\omega} \frac{\partial \bar{q}}{\partial p} \right)_{obs}$ (W/m ²)
+22.5	-504	-548
+24.5	-527	-549
+26.5	-549	-547
+28.5	-572	-548

Table 3.1
Vertically integrated large-scale
cooling and moistening for the
GATE composite easterly wave of
TPRR(1979), for +22.5 to +28.5
hours.

and moistening rates at these levels are 0.86K/day and 4.3g/kg/day.

3.2.5 Initial Simulation

Using the initial profile and the thermodynamic profiles described in the previous sections the model was integrated forward for 9 hours, allowing ten deep convective clouds to form and decay.

(a) Filtering of the initial profile

From analysing the initial simulation it became clear that the turbulence contains a large 4 gridlength wave component in the vertical. Investigating this it was found that under the action of a $\partial^2/\partial p^2$ operator the initial sounding contained a similar wave. This was smoothed out as the simulation continued. However reference was made to the initial sounding in the turbulence parameterization scheme, which acts upon deviations from it using a similar operator. This caused the mean turbulence structure to have a similar wave structure in the vertical. At levels where this was a maximum, the turbulence was dominant in balancing the large scale cooling, leaving the clouds to play a secondary role.

To prevent this from occurring the initial profile was filtered to remove the wave. When reintroduced to the model, the wave component in the turbulence was much reduced. This illustrates that numerical models are very sensitive to "noise" in the initial conditions, and such components should be removed.

(b) Improved simulation (sim A)

The improved simulation (sim A) produced a surface rainfall rate between 1 and 9 hours of 580W/m², in good agreement with the observed rainfall of 560W/m². The apparent cumulus heating over this period is 590W/m², which compares to a net external heat flux of -665w/m². Thus the convection is balancing the external forcing.

Such a balance is indicated by the deviation of θ from the initial sounding (fig 3.7). Deviations are less than 0.5K throughout the simulation. The domain mean mixing ratio (fig 3.8) indicates that the boundary layer dries rapidly under the action of the convection, 50% of this occurring in the first three hours of the simulation. However some form of steady state is reached between 5 and 9 hours, the magnitude of the drying being approximately 1.25g/kg. Observations show that the drying over the GATE area for the same period was 0.35g/kg. However 0.3 g/kg of this occurred between +20.5 and +24.5 hours, the period in which most of the drying occurred within the simulation. Between 800 and 200mb the domain shows a moistening of up to 0.5g/kg, whereas observations only indicate moistening up to 0.2g/kg.

Profiles of Q1 and Q2 (here including the effects of turbulence and boundary layer forcing) agree well with the observations. Q1 (fig 3.9) is relatively flat between 800 and 300mb. Q2 (fig 3.10) peaks around 900mb as observed, but does not reduce much towards the surface. Magnitudes are also similar.

From this brief comparison between model and observed results, it would seem that the simulation is a reasonable representation of the mesoscale atmosphere over the GATE area.

3.2.6 Balance Between Eddy Flux Divergence And Turbulence

The contribution of condensational drying, vertical eddy flux divergence, and turbulence to Q2 are shown in fig 3.11 averaged over the period 7 to 9 hours of sim A. Similar structures are seen throughout the simulation. In the lowest 250mb a balance between turbulence and eddy flux divergence is apparent. However turbulence is expected to play a minor role in balancing large scale processes. Here it is obvious that this is not the case.

The vertical component of the turbulence has a form similar to $\partial^2(\phi - \phi_{ref})/\partial p^2$ where ϕ_{ref} is a reference profile. In this case ϕ_{ref} is given by the initial profile. Integrating such a quantity across a level of the domain, the result is,

$$\frac{\partial^2(\bar{\phi} - \phi_{ref})}{\partial p^2} \quad eqn(3.15)$$

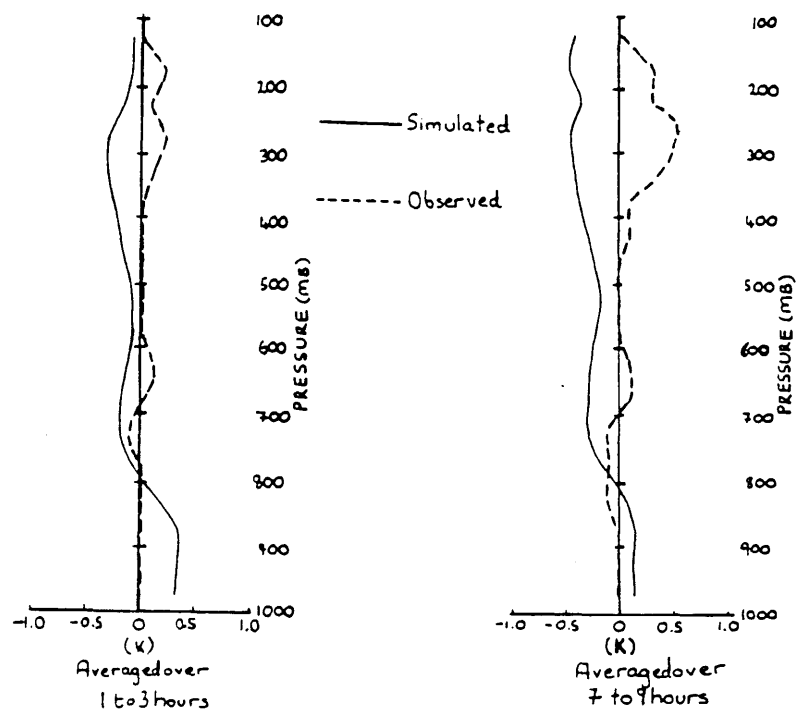


Fig 3.7 ($\bar{T} - T_r$) for sim A compared to observations.

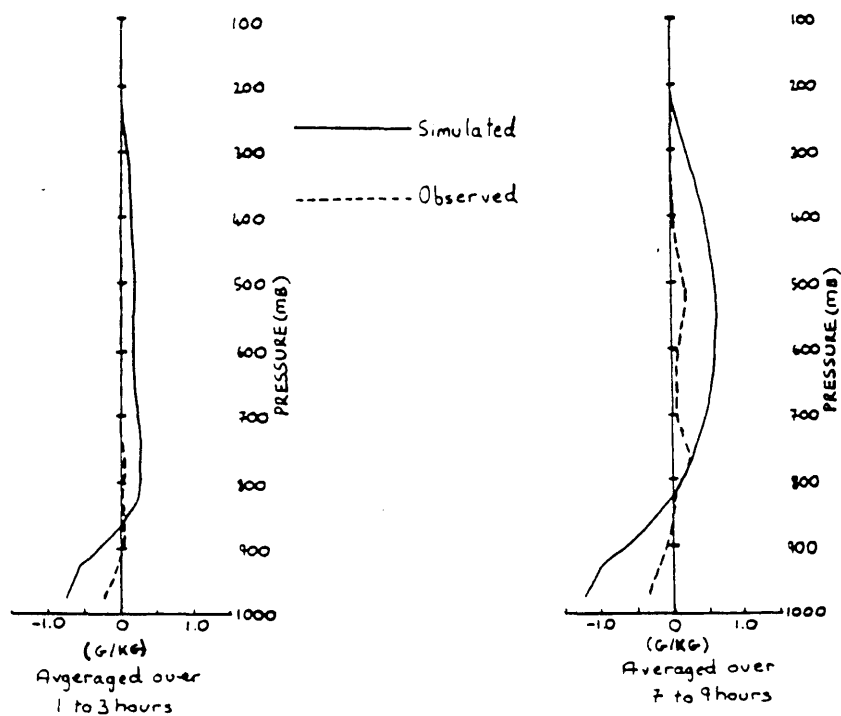


Fig 3.8 ($\bar{q} - q_r$) for sim A compared to observations.

Fig 3.9 Q1 for sim A averaged between 7 and 9 hours, compared to observations.

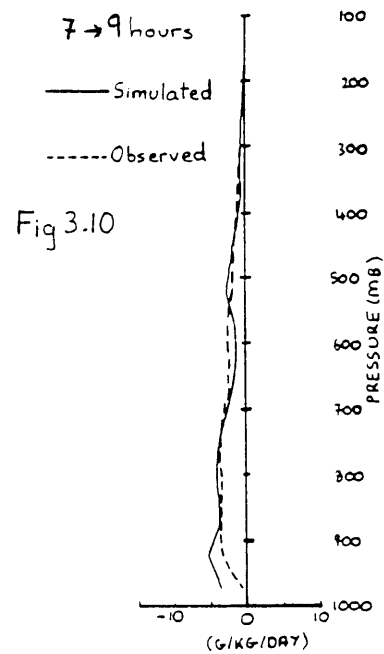
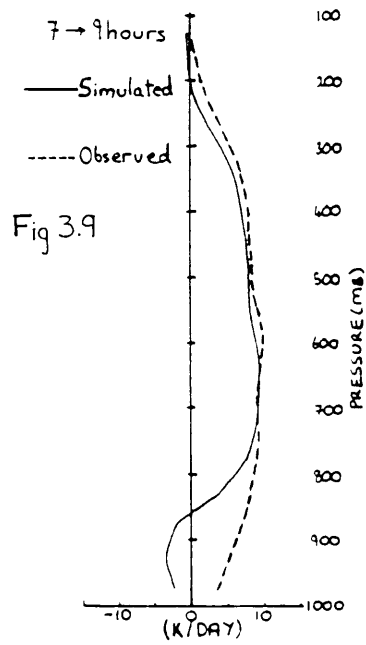


Fig 3.10 Q2 for sim A averaged between 7 and 9 hours, compared to observations.

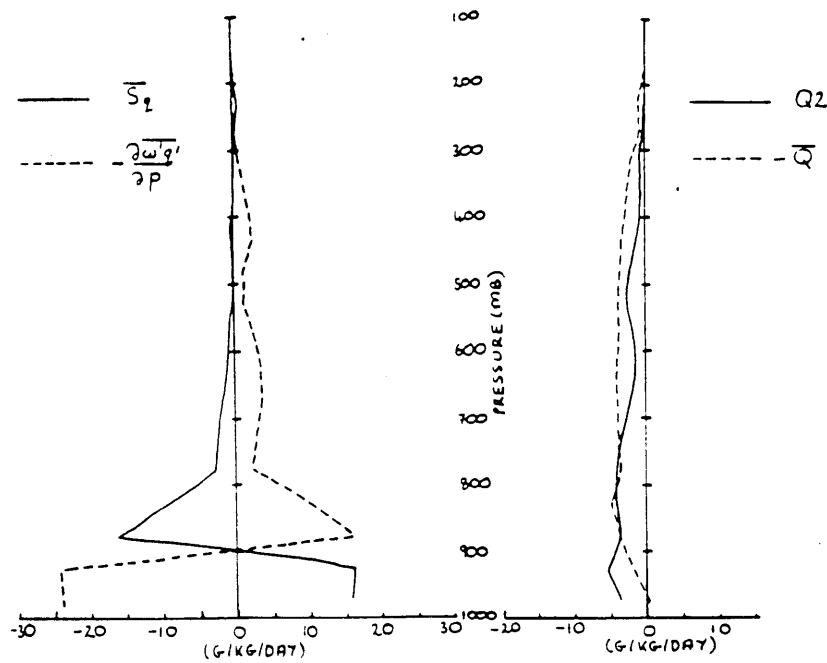


Fig 3.11 Composition of Q2 averaged between 7 and 9 hours for sim A.

Thus the turbulence structure in the vertical may be understood by studying the deviation of $\bar{q}$ from the initial profile (fig 3.8). From the shape of this profile, the turbulence parameterization scheme can be seen to add moisture to the lower domain, and remove it from above (fig 3.12), (because of the upper and lower boundary conditions on q in the turbulence scheme it conserves moisture). This balances the vertical eddy flux divergence of moisture.

Such a balance is physically unrealistic. By drying out the boundary layer, the convection stabilizes the atmosphere. Moisture is replaced by surface fluxes and large scale ascent. However with turbulence replacing a large proportion of the moisture, the convection may remain active over a longer time scale, or be of greater magnitude.

To reduce the effect of the turbulence scheme the reference profile was changed to $\bar{q}$ (sim B). This had the effect of greatly reducing the turbulence contribution to Q_2 (fig 3.13). The shape of Q_2 was little changed (fig 3.14). This resulted from a reduction in eddy flux divergence and condensational drying in the lower domain. This indicates that the simulation is relatively insensitive to the nature of the turbulence parameterization scheme.

Rainfall reduced between sim A and sim B from 430W/m^2 to 340W/m^2 , averaged over the first three hours of the simulation, indicating that the turbulence scheme aided convection as speculated above.

3.2.7 Difficulties In Comparing A 2D Simulation With Observations.

(a) The problem of initial convection(?)

It has already been noted that observations over the GATE area show a drying of the lowest layers between +20.5 and +22.5 hours. The simulation described shows a similar drying (at 975 and 925mb) in the first 5 hours, the magnitude being much larger than that observed. Why is there such a large difference in this drying when other

Fig 3.12 The influence of $(\bar{q}-q_1)$ on the turbulence parameterization scheme.

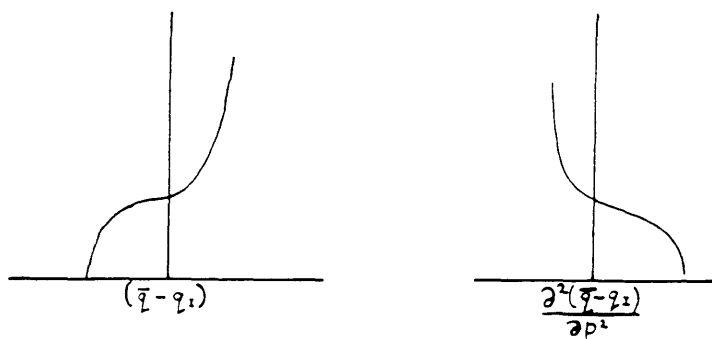


Fig 3.13 $\bar{S}_q$ for sim B averaged between 7 and 9 hours.

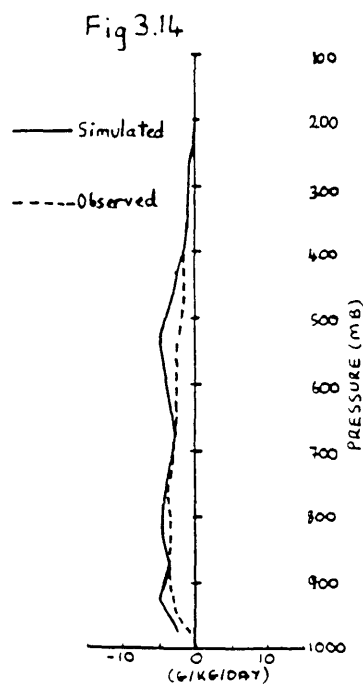


Fig 3.14 Q_2 for sim B averaged between 7 and 9 hours.

diagnosed fields (eg Q1, Q2) indicate that the simulation is a reasonable representation of the GATE region?

A reasonable answer is that the drying is caused by the initialization of convection. The initial profile has a large value of CAPE ($\approx 1500\text{J/kg}$). It is able to support active deep convection. In the real atmosphere such convection will be inhibited by the stabilizing effect of previous convection. In the simulation this does not occur and so the initial convection is too active.

(b) Attempts to "spin-up" the model

Several attempts were made to reduce the magnitude of the boundary layer drying. The basic philosophy is similar to the spin-up procedures used in large scale models. Convection is allowed to grow within the simulation, but extra forcing is applied to force the mean thermodynamic profile back to the initial profile. Hence after the spin-up phase the model will contain convection and the model atmosphere will be little changed from its initial value. This convection will inhibit future convection, preventing a large drying of the boundary layer.

In a first attempt moisture was added to the initial profile at 975, and 925mb, (+0.5 and +1.0g/kg), in an attempt to allow the model to reach a steady state with a moisture profile closer to the observed profile at +20.5hr (sim C). After 5 hours the drying at 975mb was less than sim B (fig 3.15), compared to the observed profile. However comparing the drying from the simulation initial profile, sim C showed a greater drying, together with a much greater rainfall (fig 3.16).

A second attempt was made applying newtonian moistening/drying to all grid points such that,

$$\frac{\partial q}{\partial t} \propto \frac{(q - \bar{q})}{\tau} \quad \text{eqn (3.17)}$$

where τ is a relaxation time.

This was applied for the first three hours of the simulation. Several values of τ were tried, but only those of 30 minutes or less had any appreciable effect. The forcing provided moistening in the lowest 100mb and drying above. The low level moisture source

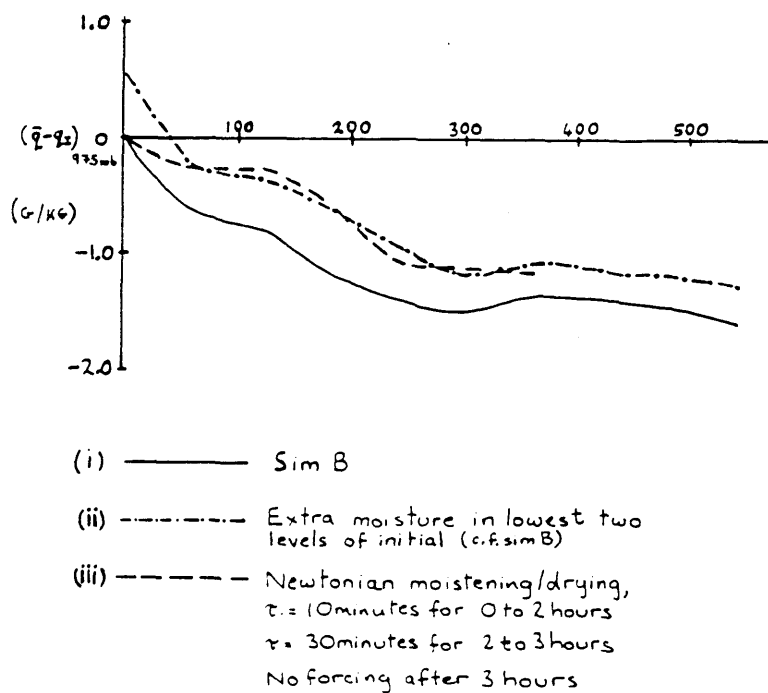


Fig 3.15 $(\bar{q} - q_r)$ at 975mb (from the initial profile of sim B) for various simulations indicated.

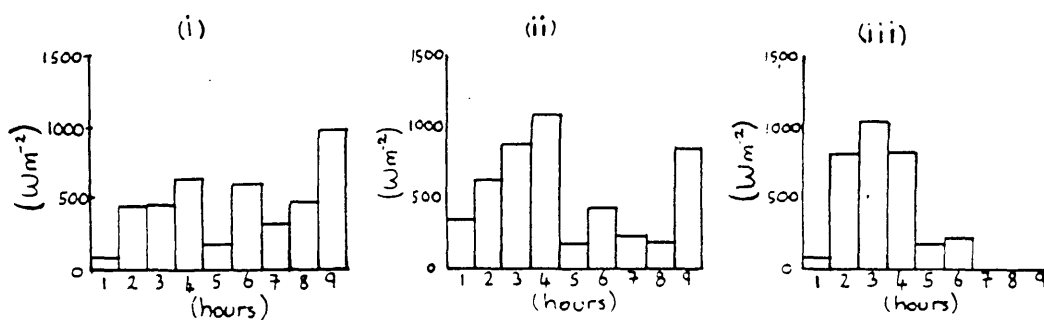


Fig 3.16 Surface rainfall rate averaged every hour for the simulations of fig 3.15

increased the magnitude of the initial convection (fig 3.16). However the drying was reduced ,but only for as long as the forcing continued. When removed the mixing ratio decreased to a value near to that observed in previous simulations (fig 3.15).

Neither of these two initialization procedures worked very effectively in reducing the magnitude of the low level drying. This suggests that the drying is not purely an initialization problem, which would be reduced by the previous convection (if this were so, then the Newtonian forcing technique should have shown a marked reduction in the drying when the forcing was switched off due to the presence of previous convection). Also as the observations show a drying at 975mb over the period covered by the simulation, it can be argued that the simulation is correct in drying the lowest layer of the atmosphere. However the problem of it's magnitude must be returned to.

(c) The use of a two dimensional domain to study the real atmosphere

The observations are taken over a three dimensional area, whereas the simulation is two dimensional. The scale of cloud separation within the simulation is of the order of 50km. Observational studies of convection over the GATE area show similar scales for cloud separation (fig 3.17 - part of fig 3 of Martin, Auvine and Suchman(1985) - anvils around 200mb, previously associated with cumulonimbi, have a scale separation of around 50km).

Hence in reality clouds have a much larger area from which to draw moisture than in a two dimensional simulation. If the observed and simulated clouds are of the same order of magnitude, then a greater drying would be expected to occur within the simulation.

From this it can be argued that in order for a two dimensional simulation to have a similar domain averaged mixing ratio to that observed, the domain could not support as active convection as observed. However if the simulation is as active as observed, then the low level drying could be compensated by a greater large scale moistening. This would make comparison with observed energy budgets difficult. However such methods (e.g Newtonian moistening) do not completely compensate for drying and cause the convection to become

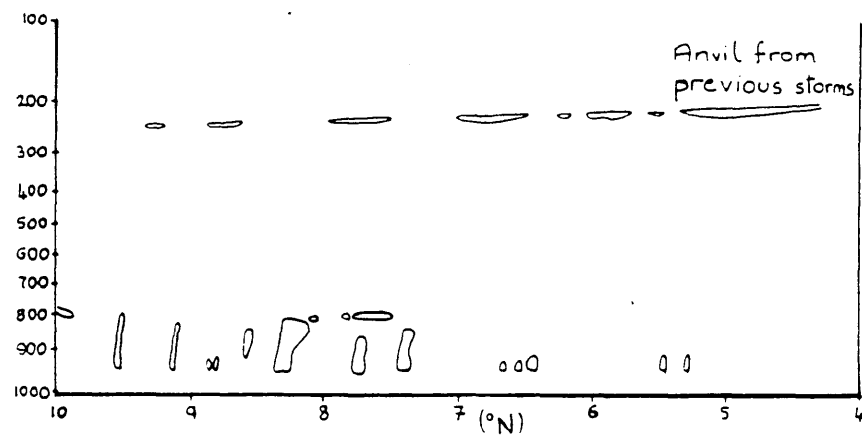


Fig 3.17 Section along $23^{\circ} 30'W$ for 1200 18/9/74 showing position of cloud ($1^{\circ} \approx 100km$). (Fig 3 of Martin, Auvine, Suchman(1985))

more active (the forcing feeds the convection).

(d) The nature of the low level drying in a two dimensional atmosphere

Simulations B and C dry the lowest layer of the model (975mb) to a value 1.25 to 1.5g/kg below that of the initial profile of sim B (+20.5). This occurs even though sim C had 0.5g/kg more moisture at this level. This indicates that the two dimensional atmosphere prefers to be in such a state.

Two further simulations were carried out, each with different moisture profiles at 975 and 925mb. These are summarised below;

sim D	975 and 925mb dried by 1g/kg below sim B profile
sim E	975 and 925mb dried by 3g/kg below sim B profile

The deviation of from the initial value of sim B for all four integrations is shown in fig 3.18 for 975mb. Sim D dries out to -1.5g/kg before reaching a quasi-steady state. Sim E moistens to around -2.0g/kg by 9hours. Sim D contains active convection, as sim B and C during their first five hours. The convection in sim E builds gradually to a similar value at the end of nine hours (fig 3.19).

This drying occurs during the first five hours of the simulation. The rate at which convection removes moisture from the boundary layer exceeds the rate at which it is replaced by surface fluxes and large scale forcing. This situation is ultimately disastrous for the convection. CAPE is linked to the moisture content of the lowest layers of the atmosphere. Hence after the continuous convection a decrease is seen in the surface rainfall during the fifth hour. During the period when the moisture profile shows a quasi-steady state behaviour the convection is oscillatory in nature.

The steady state value can be seen as representing a boundary between periods of active convection and those when little convection occurs. In the fifth hour drying reaches a maximum, while rainfall is reduced (convection is suppressed). In the sixth hour the drying is a minimum and active convection occurs.

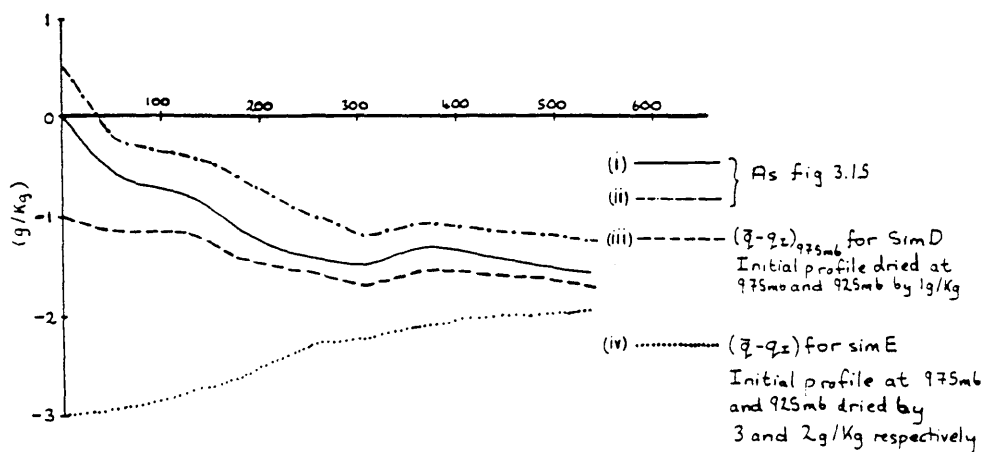


Fig 3.18 $(\bar{q}-q_s)$ at 975mb (from the initial profile of sim B) for various simulations indicated.

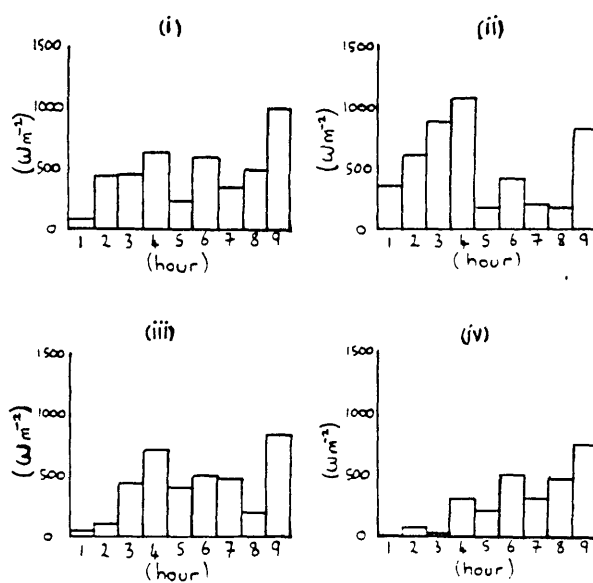


Fig 3.19 Surface rainfall rate averaged every hour for the simulations of fig 3.18.

(e) Do similar processes occur in the real atmosphere ?

Does an equivalent process occur in the real atmosphere? The drying over the GATE area approaches a much moister steady state than the simulations. Hence CAPE is larger for the real atmosphere than in the simulation. This allows more active convection or more continuous convection. In either case a greater rainfall might be expected to be observed than those simulated. However these agree to within 100 to 50 W/m² over the nine hours of the simulation.

This may indicate that convection only occurs over a limited area of the GATE region at any one time, causing a much larger drying than the average mixing ratio over the entire GATE area suggests. Some indication for this is seen in GATE radar data, and also satellite images of the tropics. Thus the two dimensional simulations may not be as bad a representation of the real atmosphere as the low level drying suggests.

3.2.8 Model Sensitivity To Changes In The Rain Water Scheme.

This study is concerned with moist processes occurring within deep convective clouds, and their effect upon the atmosphere. In the simulation cloud microphysics are parameterized as described in section(3.1.1). This scheme contains four constants, the values of which determine the production and evaporation of rain water. Thus the surface rainfall and so the net energy release due to convection, will have some dependence upon these parameters.

Several 3 hour integrations were carried out to determine the sensitivity of the model to doubling of these parameters, from the values used in sim B. These were;

$$\alpha = 0.001 \text{ s}^{-1}$$

$$\beta = 0.001 \text{ s}^{-1}$$

$$\gamma = 0.001 \text{ s}^{-1}$$

$$l_{\text{crit}} = 0.5 \text{ g/kg}$$

Table 3.2 compares budgets of condensational heating and surface rainfall, averaged between one and three hours for these simulations.

Changes from sim B are small (less than 20%), indicating that the model is relatively insensitive to these changes. With increased evaporation, and decreased autoconversion (due to increasing l_{crit}), approximately 10% less rain is seen at the surface. With more accretion and autoconversion, rainfall increases by 20%. When all parameters are doubled the change is 10%.

Changes from the vertical distribution of condensational heating of sim B are shown in fig 3.20 again averaged between one and three hours. Differences of less than 3K/day are seen, and little change is seen in downdraught cooling. The vertical structure is more sensitive to changes in α and γ , than l_{crit} and β .

These results show that the net effect of convection within the model is relatively insensitive to the values used for the parameters of the cloud microphysics scheme. During the remainder of this study, all simulations use the values of sim B.

3.3 SUMMARY

The development of a two dimensional tropical cloud ensemble has been described. Problems relating to the balance of physical processes within such a model have been discussed, together with those concerning the comparison of a two dimensional simulation with observations collected in a three dimensional atmosphere. Results have been found to be insensitive to doubling of the constants within the cloud microphysics parameterization scheme. From these studies it would appear that a two dimensional model provides a reasonable representation of the real atmosphere, and so is a useful tool in the study of cloud ensembles.

In the next chapter a detailed comparison is made between the simulation and data provided by the TPRR data set. The temporal and spatial structure of the ensemble are also discussed.

Parameter changed and it's new value	$\frac{LQ}{C_p}$ (W/m ²)	Surface rainfall rate (W/m ²)
Auto-conversion $\alpha = 0.002s^{-1}$	568	547
Auto-conversion $l_{crit} = 1.0g/kg$	458	421
Evaporation $\beta = 0.002s^{-1}$	479	423
Accretion $\gamma = 0.002s^{-1}$	556	565
All values doubled	526	514
Sim B	478	474

Table 3.2 Comparison of 3 hour integrations with various rainfall parameters.

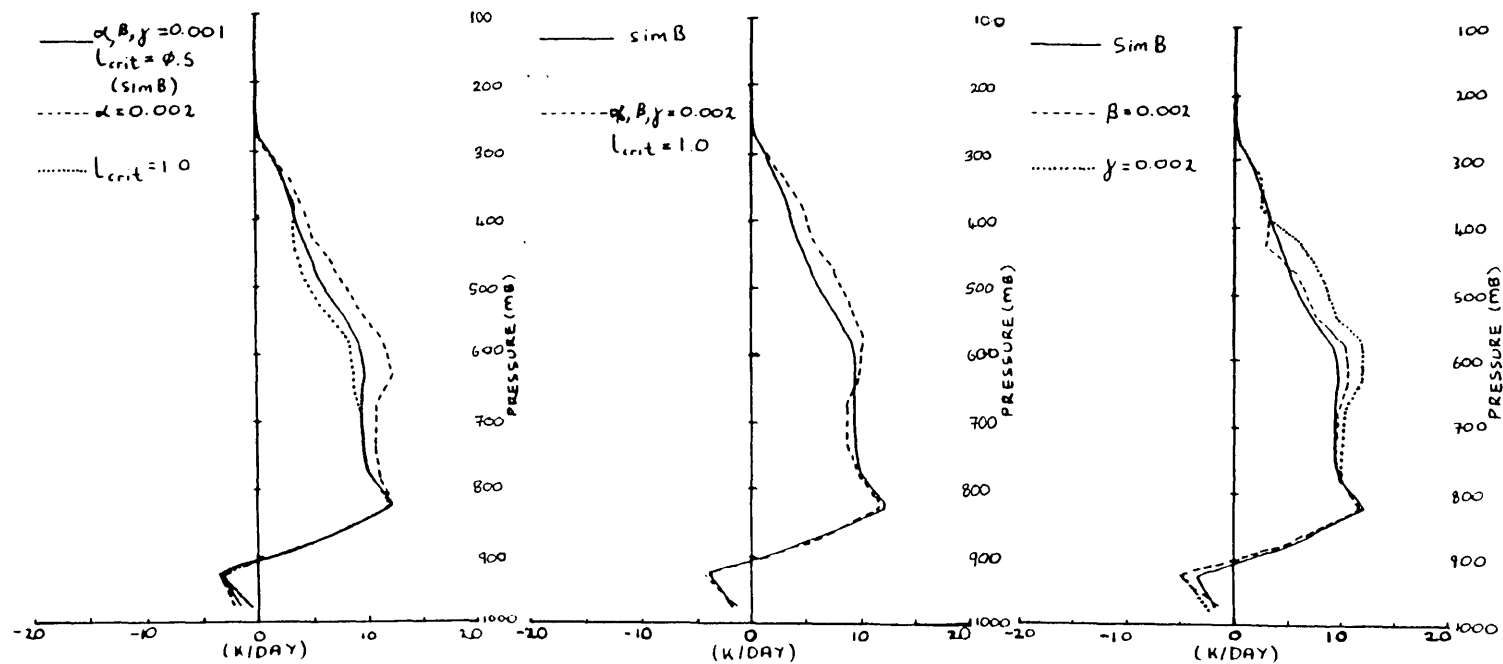


Fig 3.20 Sensitivity of the vertical structure of $L\bar{Q}/c_p$ to changes in parameters of the cloud microphysics scheme (changes from the values of sim B indicated).

CHAPTER 4

THE STRUCTURE OF A DEEP CLOUD ENSEMBLE (UNSHEARED)

A comparison is made between a simulation of a tropical cloud ensemble (simB of chapter 3) and the TPRR data set (described in chapter 2). Quantities which cannot be directly calculated from observations of the real atmosphere (eg condensational heating, eddy transports) are diagnosed to complete the description of the cloud ensemble. Time averages are taken over a time scale comparable to that of the observations and, also a few tens of minutes, for which at present observations can make little comment. It is found that within the simulation distinct "cloudy " and "clear" regions develop. These are described and discussed.

4.1 COMPARISON OF THE SIMULATION WITH OBSERVATIONS

In this section a comparison is made between the TPRR data set, and quantities diagnosed from a simulated cloud ensemble. The development and nature of the simulation has been described and discussed in chapter 3. Here simB of chapter 3 is used. The simulated clouds will be used to study the parameterization of convection in chapters 5, 6, 7, and 8. It is therefore necessary to check if they behave reasonably. Also such a study gives further

understanding of why the observations appear as they do. Condensational heating and eddy transports cannot be calculated directly from observation. However using the simulation such quantities are easily calculated, and provide a more complete view of the ensemble.

The simulation discussed here does not include any initial vertical wind shear. Cloud-shear interaction is poorly understood at present, and it was felt that at this stage shear would complicate matters. However as will be seen good agreement between the simulated thermodynamic fields and observations indicate that the effects of shear on such quantities are small. The effects of wind shear will be discussed in chapter 8.

Clouds develop within the first hour of the simulation, some reaching upper levels. Deep clouds continue to develop throughout the simulation (which was continued for nine hours). The initial development of shallow clouds across the domain (fig 4.1) appears to be uniform. Thus the random number forcing does not seem to introduce any preferred initial space structure (see also section(4.2.2)). The initial separation of convective elements was 15 to 20km.

4.1.1 Budget Studies

In this section budgets over the entire model domain are presented and compared to those of the GATE area, calculated in TPRR. Observations in GATE were taken over a variety of time scales. Upper air data was collected every 3 to 6 hours, while radar, boundary layer, and surface data were collected more frequently. However the data set available had been extrapolated to give a data point every 2 hours. Budgets are presented on this two hour time scale, by averaging over a two hour period centered on the observational data point, and also on an eight time scale. Table 4.1 contains budgeted quantities for simB, together with observed GATE data.

It should be noted that a different radiation profile from that used by TPRR in their analysis is used in the simulation. TPRR used a profile calculated by Dopplick(1970), based on climatological data for

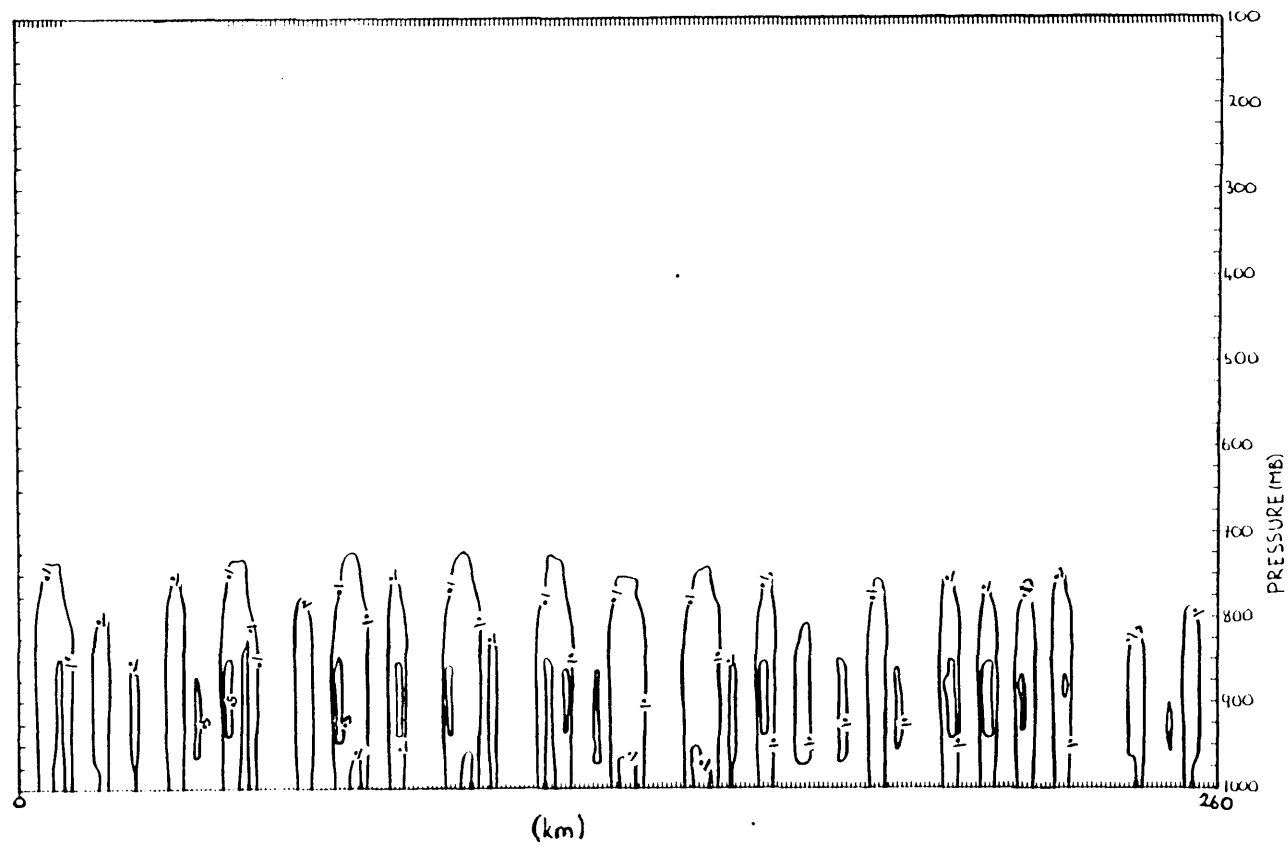


Fig 4.1 Initial development of clouds across the domain.

Time (hr)	$\langle c_p T \rangle$ (Jm^{-2})		$\langle c_p \frac{\partial T}{\partial t} \rangle$ (Wm^{-2})	LS forcing, radiational cooling (Wm^{-2})	Cloud and boundary layer forcing(Wm^{-2})	$\langle c_p \pi (Q1)_{\text{conv}} \rangle$ (Wm^{-2})	
	Observed	Simulated				Observed	Simulated
1-3	28184×10^5	24183×10^5	-202	-631	- 44	651	472
3-5	24185×10^5	24170×10^5	-275	-655	- 2	668	381
5-7	24186×10^5	24159×10^5	-128	-677	-132	684	417
7-9	24187×10^5	24154×10^5	57	-700	- 78	700	835
1-9	24187×10^5	24168×10^5	-115	-666	- 1	676	527

Sensible heat budget

Time (hr)	$\langle L_q \rangle$ (Jm^{-2})		$\langle L_q \frac{\partial q}{\partial t} \rangle$ (Wm^{-2})	LS forcing, radiational cooling (Wm^{-2})	Cloud and boundary layer forcing(Wm^{-2})	$\langle L (Q2)_{\text{conv}} \rangle$ (Wm^{-2})		Surface rainfall (Wm^{-2})	
	Observed	Simulated				Observed	Simulated	Observed	Simulated
1-3	12356×10^4	12731×10^4	197	548	127	-632	-478		473
3-5	12411×10^4	12858×10^4	290	549	127	-634	-386		409
5-7	12421×10^4	13030×10^4	265	547	127	-635	-409		466
7-9	12458×10^4	13106×10^4	-160	548	127	-637	-835		745
1-9	12411×10^4	12931×10^4	149	548	127	-635	-527	557	523

Latent heat budget

Table 4.1 Energy budgets for sim B averaged over 2 hour periods and between 1 and 9 hours.

the tropics. That used in the simulation was one calculated by Cox(1979) using data collected during phase III of GATE (the phase in which the observational data lies). The differences in the vertical structure of these two profiles can be seen in fig 4.2. However these are never greater than 1K and have little effect on the mean convective effect within the simulation. The budgeted values for both profiles is 128W/m^2 ; so little account has to be taken of the change when comparing simulated and observed budgets.

The simulation is more variable than the observed atmosphere. The TPRR data set represents the mean state of the GATE B-scale area. This has a horizontal length of 330km, similar to the length of the model domain. However the observational data is three dimensional, and so more clouds may exist within the area at a given time with greater overlap in cloud life spans than in the two dimensional model. Hence the increased variability of the model might be expected. (Further discussion of the comparison of 2D and 3D data is carried out in chapter 3, section(3.2.7)).

A further point to be considered when looking at observational data is, what is the magnitude of any error bars? A feel for errors in the observed budgets can be obtained by comparing diagnosed budgets of Q1 and Q2 from upper air data, surface observations of precipitation, and surface flux data. Using this technique errors in Q1 and Q2 calculated by TPRR(1979) are believed to be between 10 and 20%.

The simulation is less active than observation suggests. Over 1 to 9 hours $\langle(Q1)_{\text{conv}}\rangle$, and $\langle(Q2)_{\text{conv}}\rangle$ are approximately 100W/m^2 less than observed values. However the simulated rainfall is only 25W/m^2 less than the mean value observed between +20.5 and +30.5 hours. These differences are similar to the estimated errors in the observations. Although there is some dependence of the convective magnitude upon the peak value of the cloud forcing, the simulated magnitude of the convection might be expected to be of lower magnitude due to the drying of the lower domain as described in chapter 3. This leads to less potential energy being available for the moist processes of convection.

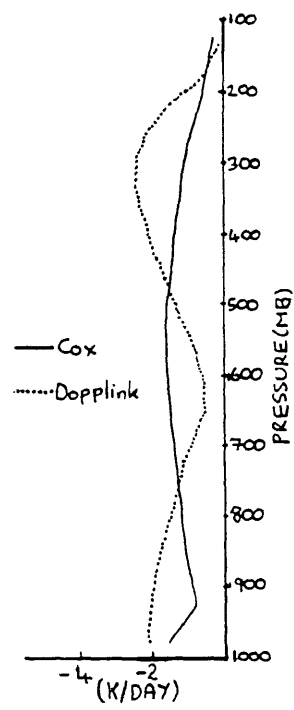


Fig 4.2 Comparison of Dopplink and Cox radiation profiles.

Changes in the simulated $\langle c_T \rangle$ and $\langle L_q \rangle$ with time are negative and positive between 1 and 7 hours indicating that the convection does not quite balance the forcing applied to the simulation. The opposite is true between 7 and 9 hours. Observations show that $\langle c_T \rangle$ increases with time, indicating that convective effects exceed the large scale and radiational cooling. $\langle L_q \rangle$ also increases with time showing that convection does not quite balance the large scale forcing and surface fluxes of moisture.

Both the simulated $\langle c_T \rangle$ and $\langle L_q \rangle$ show the effect of their being little convection in the first hour of the simulation (rainfall over the first hour is only 80 W/m^2). The change from their initial values (0.24195×10^{10} , and $0.12255 \times 10^8 \text{ J/m}^2$ respectively), those of the 1 to 3 hour average, show cooling and drying approximately twice as large as those observed over a similar period throughout the remainder of the simulation. The magnitude of such changes decreases once active convection exists, showing that a balance between convection and large scale forcing exists in the simulation, as in the real atmosphere.

The variability of convection within the simulation has already been commented upon. It may be thought that this is linked to the magnitude of the imposed cloud forcing. Although this is small averaged over the entire simulation, over 2 hour periods it can become large. However on this time scale, no direct correlation between cloud forcing and convective magnitude is seen.

Over the entire simulation $(Q1)_{\text{conv}}$, $(Q2)_{\text{conv}}$, and surface rainfall agree well with observations. Is this independent of the short time scale structure? To investigate an integration of the model (simF) was carried out using a different random number forcing to that used in simB. Comparison is made in table (4.2). SimF also shows unsteady behaviour with no ideal correlation between cloud forcing and convection. Again the cloud forcing integrates to near zero over the entire simulation. 1 to 9 hour averages agree well with each other, and with observation. Hence budgets over the entire simulation suggest that such averages are useful when considering the mean effect of a cloud ensemble on the atmosphere in which it is embedded.

Time (hr)	Sim B		Sim F	
	$\langle c_p \pi(Q1)_{conv} \rangle$	$\langle L(Q2)_{conv} \rangle$	$\langle c_p \pi(Q1)_{conv} \rangle$	$\langle L(Q2)_{conv} \rangle$
1-3	473	-478	299	-303
3-5	381	-386	968	-983
5-7	417	-409	368	-355
7-9	835	-834	423	-415
1-9	527	-527	515	-514

Table 4.2 Comparison of simulations with different random cloud forcings.

4.1.2 Cloud Height Spectrum

No observational cloud spectrum is available for this period of the GATE experiment. The nature of the spectrum may be inferred from a vertical profile of horizontal divergence (fig 2.3 of chapter 2). This shows divergence between 800 and 400mb, and around 250mb.

Most vertical mass flux within the ITCZ occurs within cumulus convection (deep convection forms the ascending branch of the Hadley cell). Thus it is usual to associate layers of horizontal divergence with cloud top outflow. In this case, there is evidence for two distinct cloud populations, with cloud tops between 800 and 500mb, and at 250mb. Visual observations made during GATE suggest cloud layers existed, but give no indication of whether these occurred in distinct layers.

The model used in this study is cyclic, and so the domain averaged omega is zero at all levels, together with the horizontal divergence. Thus it is not possible to link divergence and cloud population. However cloud top pressures can be observed and a cloud spectrum calculated (fig 4.3).

There are many cloud tops below 800mb. These will have little effect on the net condensational heating/drying of the atmosphere, producing negligible amounts of rain. In reality they contribute to Q1 and Q2 through the vertical eddy flux divergence term, ventilating the boundary layer, supplying heat and moisture to the bottom of the cloud layer, where it is transported throughout the depth of the troposphere by deep convection. It is difficult to know whether low cloud achieves this within the model. Such clouds are poorly resolved, both their vertical and horizontal scales being similar to the grid lengths of the domain. Hence their transport properties derived from the simulation should be viewed with some caution. (Section(4.2.2) contains further discussion of the transport properties of such clouds).

The cloud top "height" spectrum is continuous up to 600mb, with a gap between 600 and 500mb, followed by a peak between 500 and 400mb. A further peak occurs between 300 and 200mb. These layers agree well with the observed divergence profile. The upper level peak in the

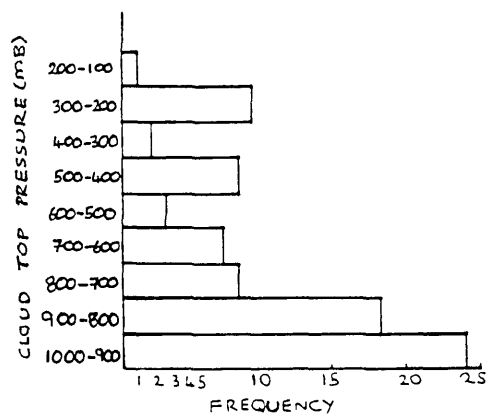


Fig 4.3 Cloud height spectrum.

cloud spectrum corresponds to the peak in divergence at 250mb. However the mid level peak is approximately 100mb higher than observed in the divergence profile. This may be caused by the lack of a vertical wind shear within the simulation. The observed velocity field contains a low level shear, which would inhibit weaker convection and bring the mid-tropospheric maximum in the cloud spectrum down to low levels. The continuous cloud spectrum between 800 and 400mb agrees with the deep divergence layer between the same levels.

Medium level clouds (cloud top between 700 and 400mb), have an active life of 30 to 60 minutes, and produce up to 10mm of rain at the surface (at one grid point). Their remnants last for no more than an hour before completely dissipating. Deep convection (clouds top between 400 and 100mb), have an active life of 60 to 90minutes, giving between 30 and 100mm of rain. Although remnant towers dissipate rapidly, cloud in the upper model layers persist throughout the simulation once they form, covering approximately 25% of the domain at the end of the integration. A dissipation time scale is difficult to state, due to new cumulonimbus strengthening and extending existing cloud as the simulation continues. A further reason for the long lifetime of such cloud may be the radiative cooling in the upper layers of the domain (up to 1K/Day).

These high level cloud features are similar, in some respects, to anvils observed in the tropics. These persist for much longer than an individual cloud lifetime. However significant differences are apparent. Organized large tropical anvils are associated with organized cloud systems (squall lines, meso-scale cloud clusters). They are deep (up to 400mb), and composed of ice. Ice falls from such cloud, melts, evaporates, cools the environment and causes subsidence (Zipser(1977), Houze(1977)). This is observed to occur over a large region, approximately 100 to 200km in length.

Clouds in the simulation do not appear to be organized in nature, the "anvil" being 150 to 200mb thick, and composed of liquid water (the cloud microphysics within the model does not include a representation of the ice phase). No precipitation is associated with this cloud; the cloud water content is usually less than 0.5g/kg, not allowing any auto-conversion of cloud water to rain water. Hence the

model anvil cloud is not associated with subsidence as observed in the tropics. The lack of precipitation from anvil cloud will also contribute to their long lifetime.

The calculated cloud spectrum gives confirmation that the horizontal divergence over the GATE area is associated with the outflow layers of convective clouds. The observed divergence is calculated from the net vertical motion. This is upward, and so the cloud spectrum indicates that ascent observed to occur over the ITCZ occurs within cumulus convection.

Layers of convergence are also seen in the observed divergence profile, between the surface and 850mb, and around 400mb. The lower convergence layer is associated with inflow into the convective clouds. This is probably not true for the upper layer. Horizontal divergence is calculated from the area averaged omega field. As stated earlier, there is good reason to suppose that this reflects the average mass flux within cloud. Around the 400mb level, the medium level cloud will have less influence on this average, leaving the stronger signal of deep convection. Hence an increase in $\overline{\omega}$ might be expected, leading to a convergent layer when horizontal divergence is calculated.

Why should the simulation contain a similar cloud height spectrum to that observed? It cannot really be said that the random cloud forcing employed accurately simulates the true forcing mechanism. The spectrum must be determined by the thermodynamic structure of the atmosphere. This will be discussed further after this structure has been studied. The spectrum may also be influenced by shear. While shear would appear to be essential for organized convective cloud (Ludlam(1963)), smaller cloud may be inhibited (Kuo(1963)). Thus the presence of shear may reduce the maximum cloud top height of the medium level cloud, and so bring closer agreement with the observed mid-level divergence.

4.1.3 Vertical Structure Of The Model Atmosphere.

Budgets contain information as to how much energy is released by

convective processes within the atmosphere. However they make no statement concerning how this is distributed in the vertical. This is very important and has a large influence on the future development of motions within the atmosphere. This is illustrated by work done on hurricane development.

Riehl and Malkus(1961) and Yanai(1964(a)), observed that latent heat release within deep convection is an important mechanism in driving the hurricane circulation. Ooyama(1964), Charney and Eliassen(1964), and Kuo(1965), studied hurricane development utilizing models in which the effects of convection were parameterized. Such studies were found to reasonably represent the hurricane circulation. Syono and Yamasaki(1966), and Yamasaki(1968), considered the vertical distribution of the latent heating. They found that the nature of the final circulation was dependent upon this parameter.

In this section, vertical profiles of horizontally averaged quantities are presented, building up a detailed picture of the mean state of the simulated atmosphere. Comparison is made to observations recorded in the TPRR data set.

(a)Potential temperature.

The domain average($\bar{T}-T_c$) profile, between 1 and 3, and 7 and 9 hours is shown in fig 4.4. Initially cooling is uniform with height above 800mb, but later warming develops in mid-levels. Cooling in the lowest is usually associated with outflow from convective scale downdraughts. A cooling of 1K is seen after 9 hours, a rate of $\approx 3K/day$. Considering the forcings applied to the model in the lowest layer (2K/day), 1K/day is attributed to downdraughts. The reason for the warming around 850mb is unclear. This will be returned to in section(4.2.1).

Profiles of $\partial \bar{T} / \partial t$ (fig 4.5) agree with the budgets presented earlier. Between 1 and 7 hours, cooling is dominant, while between 7 and 9 warming. This is due to the inbalance of convective and external forcing on a 2 hour time scale. However $\partial \bar{T} / \partial t$ never exceeds 5K/day, and $(\bar{T}-T_c)$ is never greater than 1K, even though the forcing applied to the model reaches 12K/day, indicating that some balance is reached between convection and the forcing.

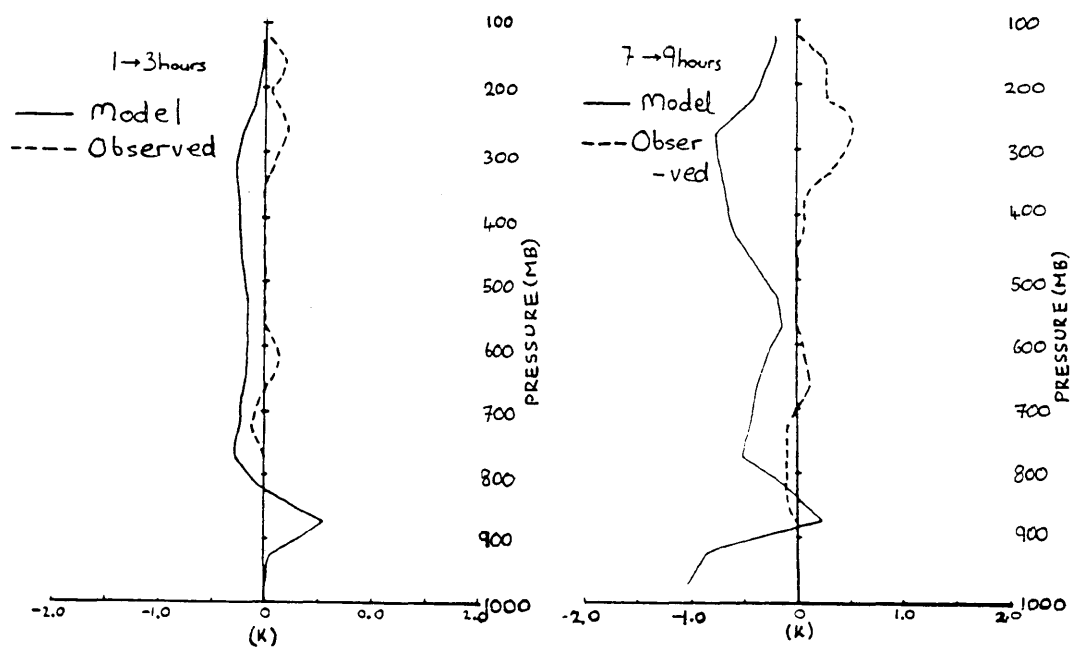


Fig 4.4 ($\bar{T} - T_1$) averaged between 1 and 3, 7 and 9 hours.

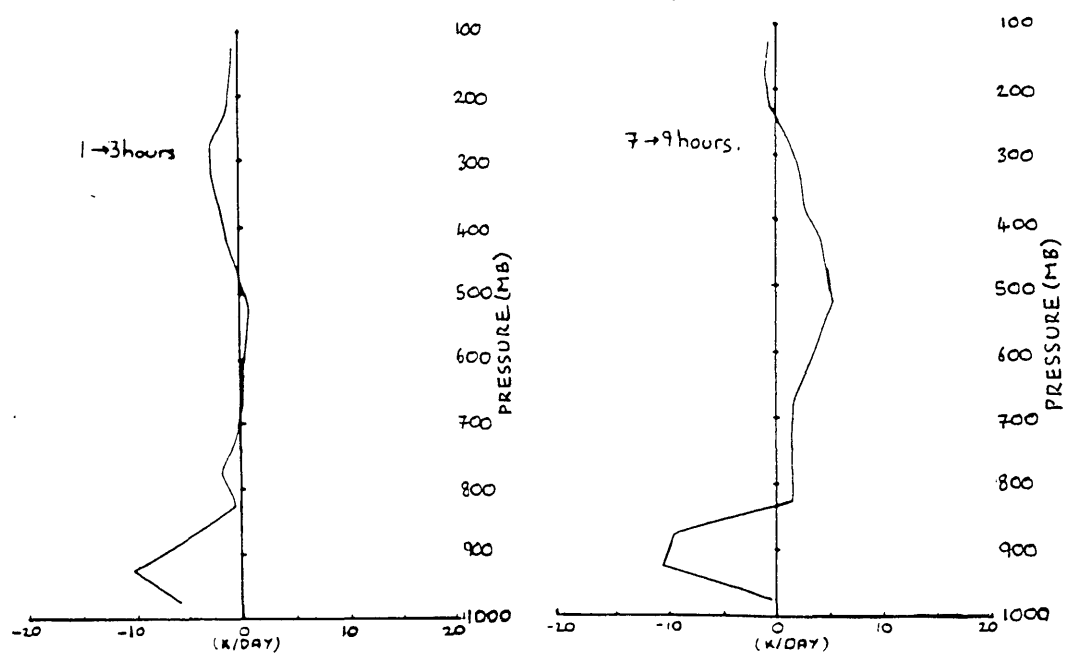


Fig 4.5 $\partial \bar{T} / \partial t$ averaged between 1 and 3, 7 and 9 hours.

After 5 hours, most of the change in $(\bar{q}-q_r)$ occurs above 500mb. Budgets show that the simulated convection is not as active as observed. The cooling in the upper layers indicates that the convection is not deep enough. This causes discrepancy between the model and the observations (fig 4.4). An upper level cooling is seen in the observation between +2.5 and +10.5 hours. During this period little deep convection exists, and so the large-scale and radiational forcings dominate the upper layers.

The simulated "warming" between 500 and 600mb may correspond to the observed warming between 600 and 700mb. If $\partial\bar{T}/\partial t$ is considered over 1 to 9 hours, then it is seen to be positive at these mid-levels.

No low level cooling is seen in the observations. Again this is a consequence of comparing a 2D domain with a 3D atmosphere (see chapter 3). The cold pool produced by the simulated downdraughts, has less area to spread into than in the real atmosphere. Thus when averaging over the domain a greater signal is seen. Another contributing factor linked to this scale problem, may be that the observed data is a composite of radio-sonde ascents collected from 20 points over the GATE domain. Many of these ascents may not pass through downdraught cold pools, and so show no cooling near the surface.

(b) Mixing ratio

Drying occurs in the lowest 100mb, with moistening above (fig 4.6). This agrees well with the observed structure, although magnitudes are larger in the simulated results, for reasons already discussed in chapter 3.

$\partial\bar{q}/\partial t$ (fig 4.7) during 1 to 5 hours indicates drying and moistening as indicated by $(\bar{q}-q_r)$. However between 5 and 9 hours, this structure is replaced by one in which either the entire domain moistens (5 to 7 hours), or dries (7 to 9 hours). This shows that the drying of the surface layer in the first 5 hours of the simulation produces a quasi-steady state, atmospheric structure being in a state which the 2D model prefers to the initial sounding, (as suggested in chapter 3). After the structure has developed, oscillations occur around it.

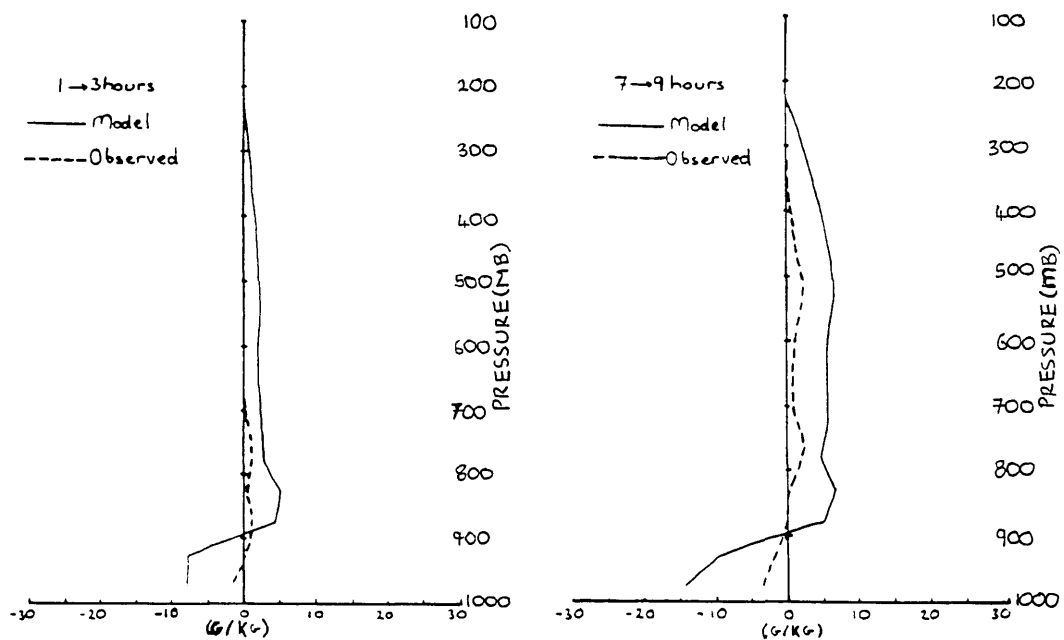


Fig 4.6 $(\bar{q} - q_T)$ averaged between 1 and 3, 7 and 9 hours.

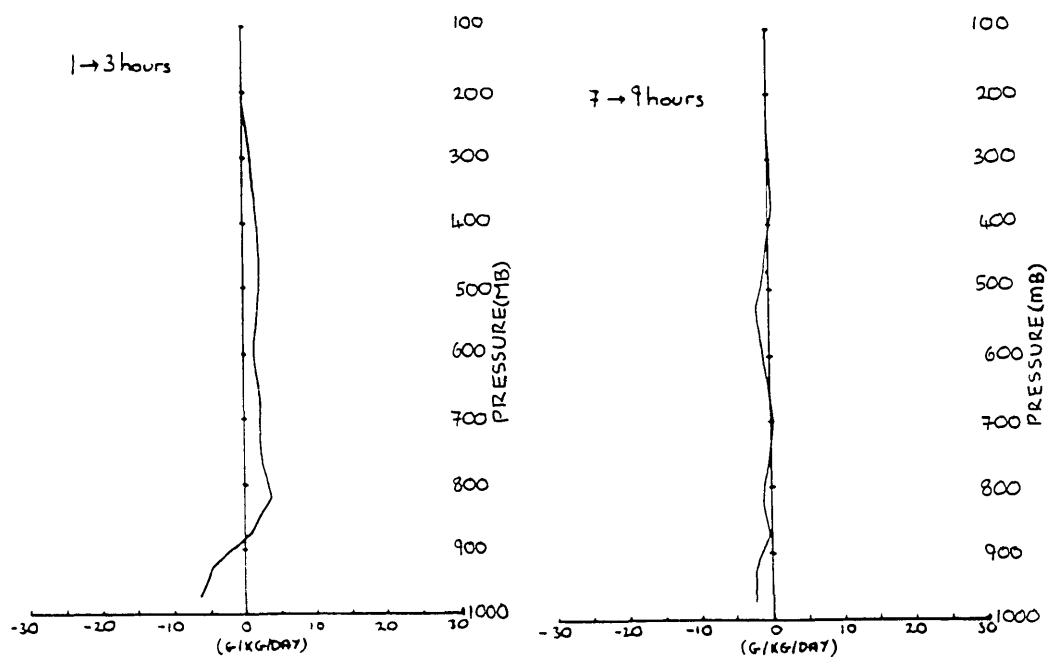


Fig 4.7 $\partial \bar{q} / \partial t$ averaged between 1 and 3, 7 and 9 hours.

(c) Apparent sources/sinks of heat and moisture (Q1, Q2).

The general shape of Q1 is similar to the observed profile, although magnitudes vary between the 2 hour averaging periods. Over the entire simulation magnitudes agree well with the observations, although above 500mb the simulated profile falls off to zero more rapidly than observations suggest (fig 4.8). This agrees with the observations made when the potential temperature field was studied. This may be caused by a lack of meso-scale convective processes within the simulation, which heat the atmosphere at these levels (Johnson(1984)).

The peak of the profile occurs between 500 and 600mb, slightly higher than the observed maxima, around 600mb. The observations show no cooling in the lowest 150mb as the simulation suggests. This is usually attributed to the evaporation of falling rain in convective scale downdraughts. This discrepancy is not surprising, as the observed Q1 is diagnosed from the θ field. Over the GATE area, this shows no "cold pool" due to downdraught air at the surface.

Q2, (fig 4.9), has a similar shape to the observed profile. Magnitudes are similar, but again the simulated profile decays more rapidly with height.

Other contributions to Q1, and Q2 as well as the convective contributions are included in figs 4.8, and 4.9. Observationally it is easier to calculate Q1 and Q2 due to difficulties in obtaining accurate measurements of radiative cooling and surface fluxes.

(d) Condensational heating, eddy transports and turbulence.

Q1 and Q2 are composed of condensational heating/drying, vertical eddy flux divergence, a contribution from the boundary layer forcing applied to the model, and the effects of the turbulence parameterization scheme. As previously explained, (chapter 2), these terms cannot be calculated as separate quantities from observations of the large-scale. However they are easily separated when analysing a numerical model which explicitly represents clouds. These terms are shown in fig 4.10 averaged between 1 and 9 hours. Each two hour average also shows the same structure.

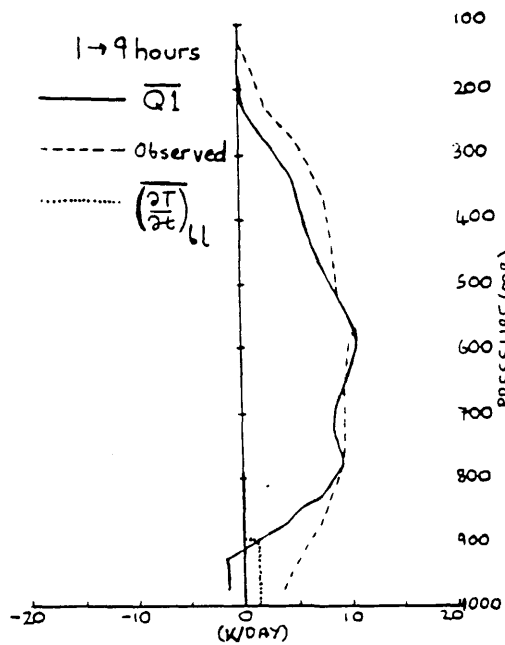


Fig 4.8 Q_1 averaged between
1 and 9 hours.

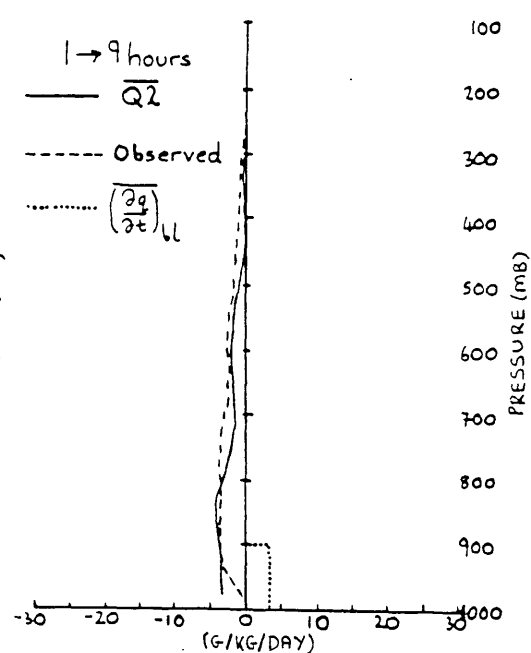


Fig 4.9 Q_2 averaged between
1 and 9 hours.

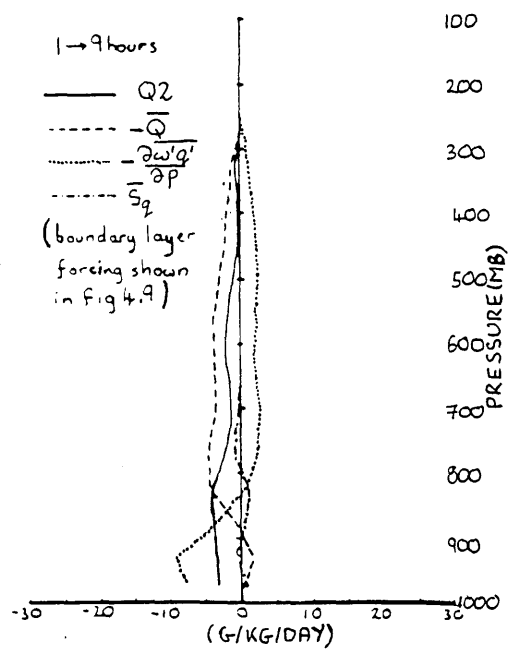
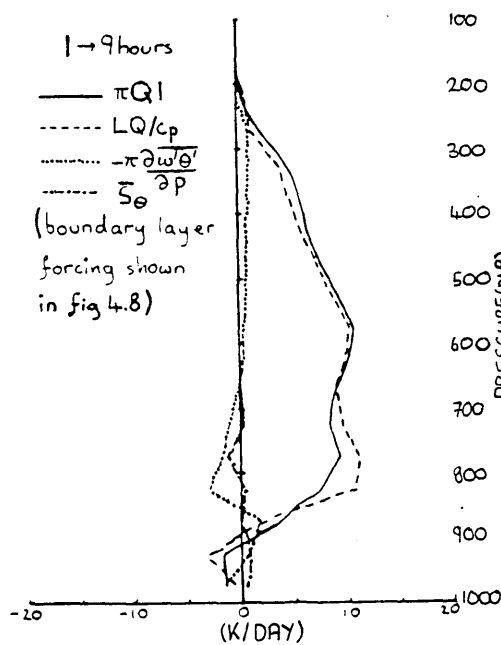


Fig 4.10 Components of Q_1 , Q_2 averaged between 1 and 9 hours.

Condensational heating and evaporative cooling (referred to as moist processes) are the main contributors to Q1. Maximum heating occurs between 800 and 500mb. The effects of downdraughts can be seen below 800mb by the rapid decrease in the heating rate, and also cooling at the surface. The effects of turbulence are negligible, and the vertical eddy flux divergence plays a secondary role, at most being only 20% of the magnitude of the moist processes. The eddies cool the atmosphere between 800 and 600mb, and warm it above. Below 850mb a warming occurs down to 950mb, with a cooling in the lowest model layer. In general convective motions remove heat from the lowest part of the atmosphere, where the energy release due to moist processes is maximum, and distributes it in the upper part of the domain. The structure of eddy transports will be discussed in more detail in chapter 6.

In Q2, moist processes and eddy transports play more of an equal role (again the effect of turbulence is negligible). Moist processes dry the domain above 900mb, and moisten it below this. The eddy transports of moisture, dry the atmosphere below 850mb and moisten it above. $-\partial\overline{\omega'q'}/\partial p$ is the dominant term in Q2 below 800mb. Thus convective motions redistribute moisture to a much greater extent than temperature.

The vertical eddy flux of "total energy",

$$-\pi \frac{\partial \overline{\omega' \Theta'}}{\partial p} - \frac{L}{c_p} \frac{\partial \overline{\omega' q'}}{\partial p} \quad \text{eqn (4.1)}$$

is shown in fig 4.11, together with a profile diagnosed from Q1 and Q2 of the TPRR data set. The simulated eddy flux is much larger than that observed within the lower layers of the atmosphere. However inclusion of the heating and moistening due to surface fluxes into the simulated profile bring closer agreement. The convection removes energy below 800mb, and redistributes it above this level. Good agreement with the simulated and observed fields suggests that the structure of individual eddy transports of temperature, and moisture are also correct.

As reported in chapter 2, previous workers have tried to calculate eddy flux divergences from observational data using one dimensional analytical cloud models. An example of such work is that

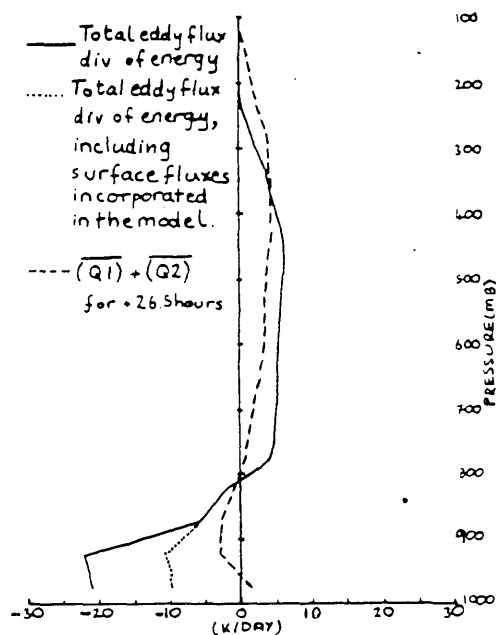


Fig 4.11 Total eddy flux of energy averaged between 1 and 9 hours (compared to profile calculated from observations).

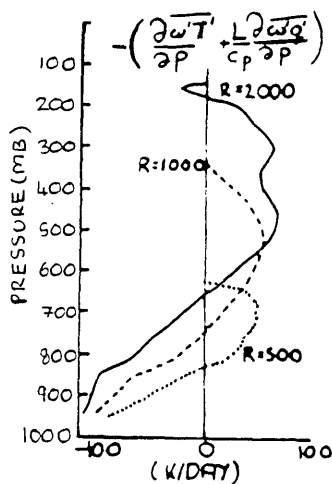


Fig 4.12 Total eddy flux of energy of a one dimensional cloud model for a tropical hurricane sounding (Anthes(1977)).

of Anthes(1977). He used a mean hurricane for the Gulf of Mexico (Hebert and Jordon,(1959)). This has a similar structure to the atmosphere over the GATE area, and so similar convective transports might be expected to occur. However Anthes results seem to over estimate the convective transports (fig 4.12). They remove energy equivalent to 100K/day from the surface layers, compared to less than -20K/day for the numerical model.

Although Anthes model did not contain convective downdraughts, it would seem that his cloud model is much too active. As the model is similar to those used by other workers, perhaps convective effects diagnosed using such models are generally too large, and so might be viewed with some caution.

(e) Liquid water budget.

This budget is difficult to obtain from observations, as detailed information on the liquid water content of the atmosphere is not readily available. Johnson(1980) has attempted to estimate the storage of liquid water for the composite easterly wave of TPRR, using ground based and satellite pictures of cloud cover. He found that the storage of liquid water was important during wave categories where cloud amount changed rapidly (categories 1, 5, and 6). During other periods of the wave it was not important. However Johnson was cautious about his results, due to inaccuracies in cloud cover and cloud water content measurements.

The water budget of the meso-scale model can easily be studied; both cloud and rain water being carried as variables within the model. The area averaged model equations for liquid water substance are

$$\begin{array}{l} \text{CLOUD} \\ \text{WATER} \end{array} \frac{D\bar{l}_c}{Dt} = \bar{Q} - \overline{\text{PROD}} - \frac{\partial \overline{\omega' l'_c}}{\partial p} + \bar{S}_{l_c} \quad \text{eqn(4.2)}$$

$$\begin{array}{l} \text{RAIN} \\ \text{WATER} \end{array} \frac{D\bar{l}_r}{Dt} = \overline{\text{PROD}} - \overline{\text{FALL}} - \frac{\partial \overline{\omega' l'_r}}{\partial p} + \bar{S}_{l_r} \quad \text{eqn(4.3)}$$

with terms as defined in chapter 3.

Adding these two equations together gives an equation for the total liquid water within the domain,

$$\frac{D(\bar{l}_c + \bar{l}_r)}{Dt} = \bar{Q} - \text{FALL} - \frac{\partial \bar{\omega}' \bar{l}_c'}{\partial p} - \frac{\partial \bar{\omega}' \bar{l}_r'}{\partial p} + \bar{S}_c + \bar{S}_r \quad \text{eqn(4.4)}$$

Integrating over the entire domain, the eddy terms average to zero. Also for a cyclic domain both horizontal and vertical derivatives on the lhs of eqn(4.4) are zero. Hence the liquid water budget for the domain (in terms of latent energy) is

$$L \left\langle \frac{\partial (l_c + l_r)}{\partial t} \right\rangle = L \langle Q \rangle - L \langle \text{FALL} \rangle + L \langle S_c \rangle + L \langle S_r \rangle \quad \text{eqn(4.5)}$$

$\langle \text{FALL} \rangle$ is equal to the surface rainfall rate.

Between 1 and 9 hours $\langle LQ \rangle$ and $-\langle \text{FALL} \rangle$ are the major terms in the budget, having values of 528Wm^{-2} and 523Wm^{-2} respectively. Turbulent effects can be neglected. Hence little storage of liquid water occurs in the domain ($\approx 3\%$ of $\langle L \partial q / \partial t \rangle$), confirming Johnson's results.

$\langle L(l_c + l_r) \rangle$ is only $\approx 1\%$ of $\langle Lq \rangle$, rain water contributing $\approx 10\%$ of the total liquid water content. The contribution of liquid water to the total moisture content of the domain increases above 400mb, to $\approx 30\%$ at 225mb. In this region remnant cloud is seen, which covers $\approx 25\%$ of the domain by 9 hours. Little water vapour exists at these levels of the atmosphere, and so the ratio is increased. **Fig 4.13** shows the vertical profiles of the main terms in eqn(4.4). Time dependent terms can be neglected at all levels of the atmosphere, as can turbulent effects. $-\text{FALL}$ removes rain water above 700mb, increasing the amount below. It is seen that eddy transports of liquid water play a large role in determining the net distribution of liquid water (and so net condensational energy release) within the vertical.

Eddy transports remove cloud water from the lowest 300mb of the domain, increasing cloud water content above this. Rain water transports have a different structure. The depth of the drying layer is greater, and rain water content is increased in the lowest 100mb of the domain. These difference are due to downraughts, which contain more rain water than cloud water below 600mb.

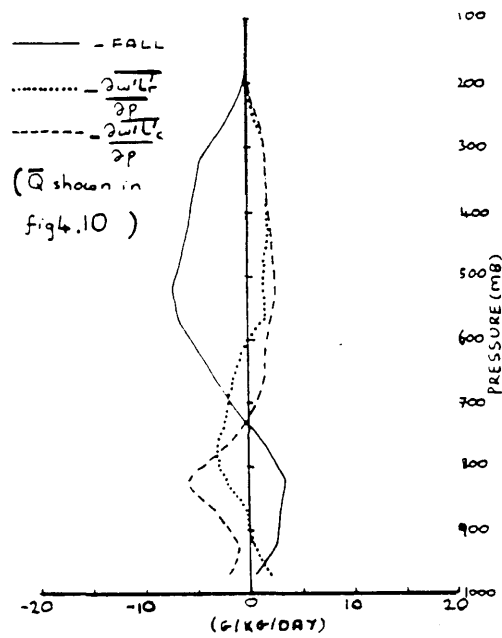


Fig 4.13 Liquid water budget of model, averaged between 1 and 9 hours.

4.1.4 Cloud Height Spectrum (continued).

The question of why two distinct cloud populations exist is now reconsidered. It was seen that a gap exists in the spectrum between 400 and 300mb. This would seem to be caused by the formation of a stable layer between 500 and 600mb. The domain averaged temperature profile shows a reduction in the cooling occurring in this region. Between 7 and 9 hours, $\partial \bar{T} / \partial t$ indicates warming occurring in this layer. When unaveraged fields are studied a warm, dry layer surrounding deep convection is seen. Deviations from the initial profiles are at a maximum 0.5K, and -0.3g/kg. This extends up to 50km away from the clouds. There is some indication of such a layer in the observations between 600 and 700mb.

Such a layer will provide a stabilizing influence for weaker convection. When their cloud tops enter the warm layer, their buoyancy is reduced, and their ascent rate decreased (fig 4.14). Hence a gap exists within the cloud spectrum, those clouds which would have reached between 400 and 300mb being confined to the lower levels by the stable layer.

Thus some evidence exists towards the idea that large clouds have a controlling influence on smaller clouds, through their effect upon the atmospheric thermodynamic structure. Further evidence for this is seen in the Q1 profile, which peaks between 500 and 600mb. Deep convection is seen to produce 3 to 10 times more rain than medium height convection, and so Q1 mainly represents the heating due to deep convection. Hence the stable layer can be attributed to deep convection. Unfortunately detailed study of the effects of each cloud type is difficult to carry out due to the time dependent nature of the simulation, although further comments on the contribution of various cloud types is made in the next chapter.

4.1.5 Summary.

The results presented show that the two dimensional simulation agrees quite well with the data obtained from the GATE experiment. Hence cloud structure, and the parametric techniques by which

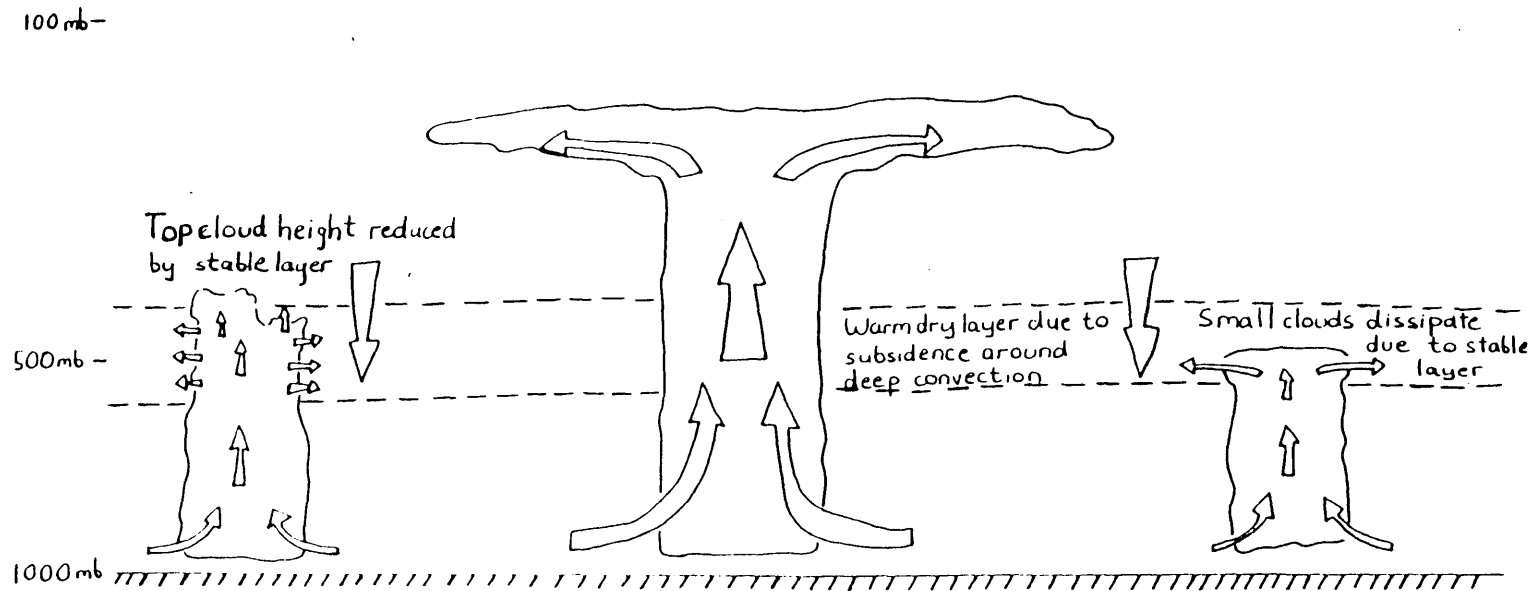


Fig 4.14 Formation of warm layer by deep convection and control of weaker convection.

convection is represented in large-scale models may be studied with some confidence (see chapters 5 and 6).

There is more downdraught effect within the simulation than observations suggest. This is due to the comparison of a 2D model with 3D observational data (see chapter 3). Perhaps a further contributing factor is a sampling error in the GATE data. If many of the radio-sonde ascents did not pass through downdraught cold pools, this would cause a further reduction in the downdraught signal.

It has been commented upon that the transport properties of convection calculated using analytical cloud models and observational data, appear to be much too active compared to the simulated cloud. Hence the results obtained from such techniques should perhaps be viewed with some caution.

4.2 SPATIAL AND TEMPORAL STRUCTURE OF THE ENSEMBLE.

The structure of the ensemble averaged over the entire domain, and over several hours has been described and discussed. However a detailed spatial and temporal structure exists within the simulation. Averages over various parts of the domain show distinct regions exist. Time averages over 20 minute periods show a much more detailed and variable structure than the 2 hour averages (although these show variations in magnitude between averaging periods).

Fig 4.15 is a plot of cloud and rain area within the domain against time for several layers of the model. Twelve deep convective clouds exist during the simulation. There is a greater overlap in cloud life spans in the first 4 hours of the integration. In the latter part of the simulation, one large cloud dominates, although overlap in life spans still occur. The other distinguishing feature of the simulation is the formation of "anvil", and "clear" regions. Deep convection seems to be confined to certain well defined areas of the domain, which contained previous storms. Anvils from deep convection tend to merge over these regions as the simulation progresses.

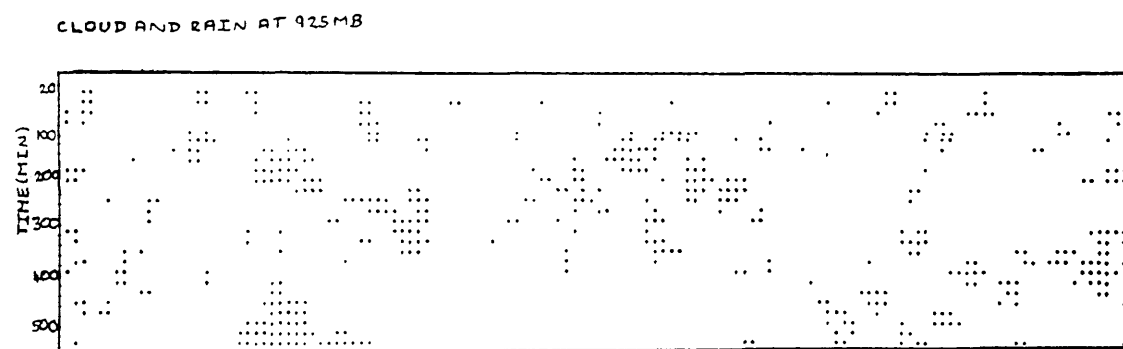
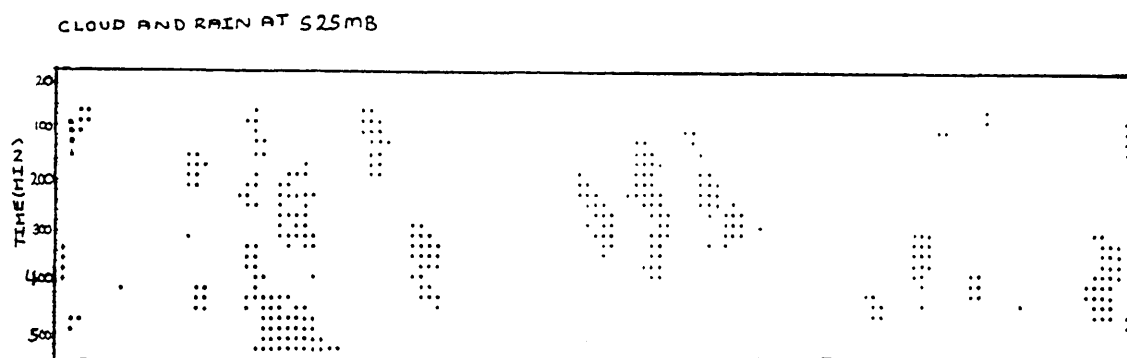
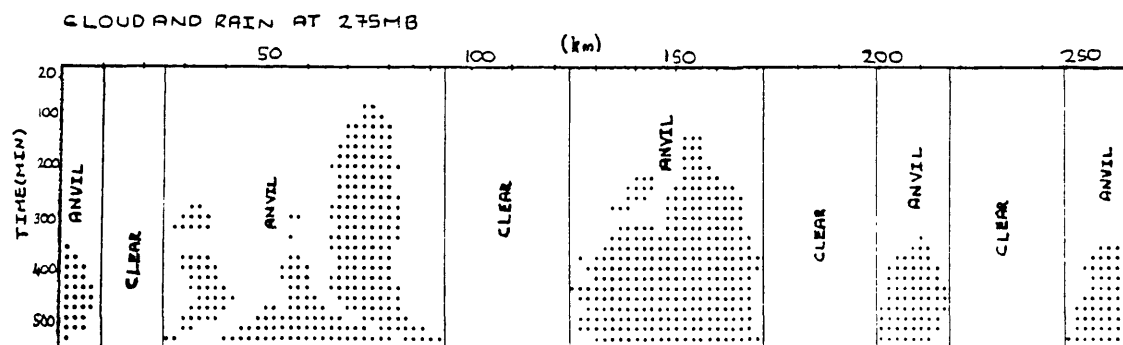


Fig 4.15 Cloud area against time plots (showing formation of "anvil" and "clear" areas).

4.2.1 Temporal Structure.

With the meso-scale model, the effect of convection upon it's environment on time scales shorter than a few hours can be considered. On a temporal scale of a few tens of minutes, time dependent effects become important. Such analysis gives a more detailed understanding of the processes involved in convection. Does such detailed knowledge have any importance when considering the parameterization of convection?

The time step of GCM's is of the order of 30 minutes, less than one cloud lifetime. However the grid length is typically 100km. The area represented by each grid point is probably large enough for many cloud life spans to overlap, reducing the effect of time dependent terms. Thus in such a model, when considering the parameterization of convection, the assumption of a steady state response of convection to the large-scale fields is perhaps reasonable.

If coarse meso-scale models are considered, then a different situation exists. Both the time step, and horizontal resolution are much smaller than in a GCM (typically a few tens of minutes, and a few tens of kilometers). Hence the statistical properties of a cloud ensemble used in GCM's cannot be used. Each grid box may only contain one cloud, and time dependent effects may become important.

Using the same analysis techniques used previously, model fields are diagnosed, quantities being averaged over 20 minute periods. These are plotted on pressure-time graphs. It is important to remember that each time point does not represent one specific time, but an average over the period indicated.

(a) Vertical structure.

Fig 4.16 shows the domain averaged temperature fields, together with $\partial \bar{T} / \partial t$. Although the upper level cooling is a continuous process, the picture is not so steady for the middle and lower layers of the atmosphere. Periods of overall heating and cooling exist, extending over the period of an hour. Temperature deviations from initial values vary by several tenths of a degree, a significant fraction of it's average over a period of hours. A warming is seen to occur

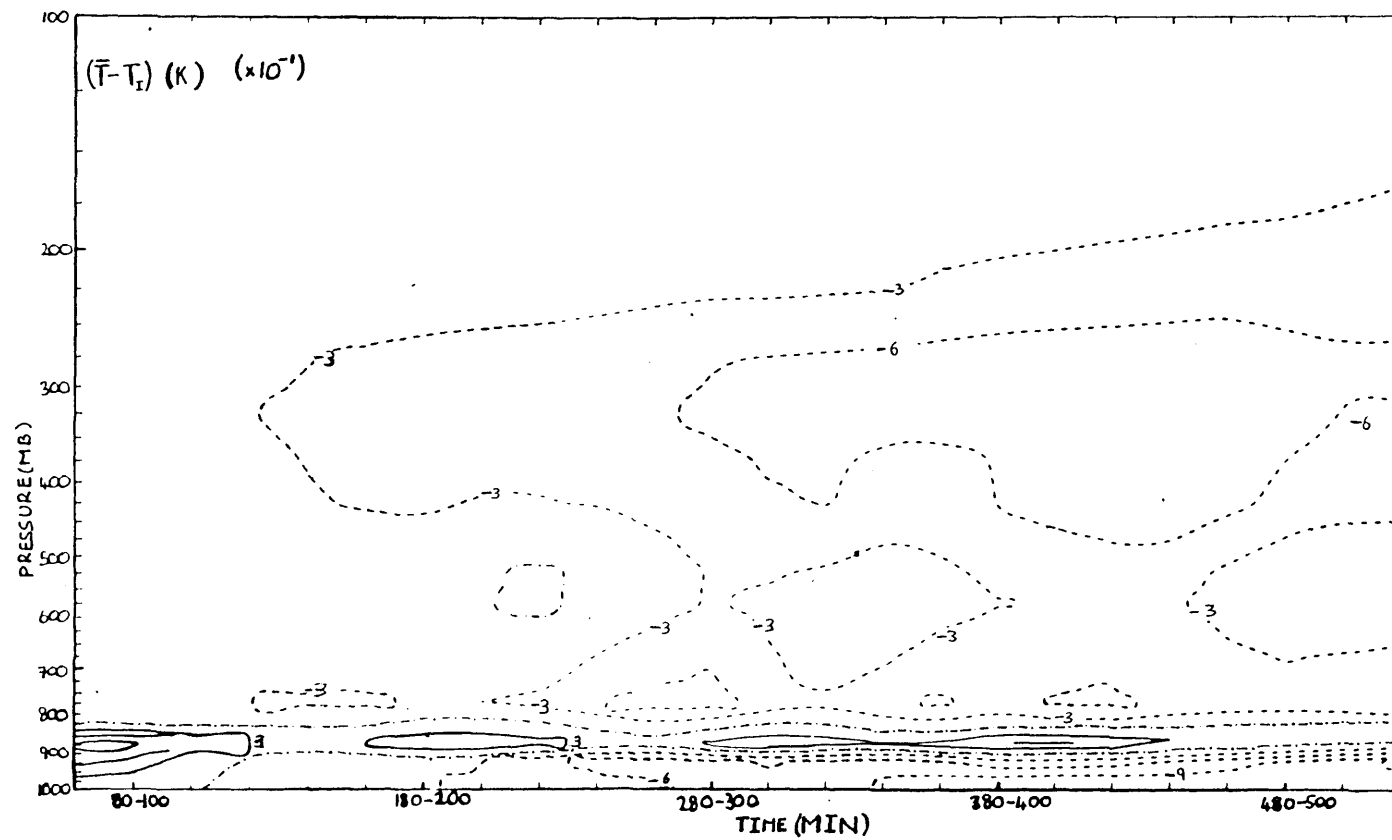


Fig 4.16(a) $(\bar{T} - T_1)$ 20 minute averages.

(contour every 0.3 K)

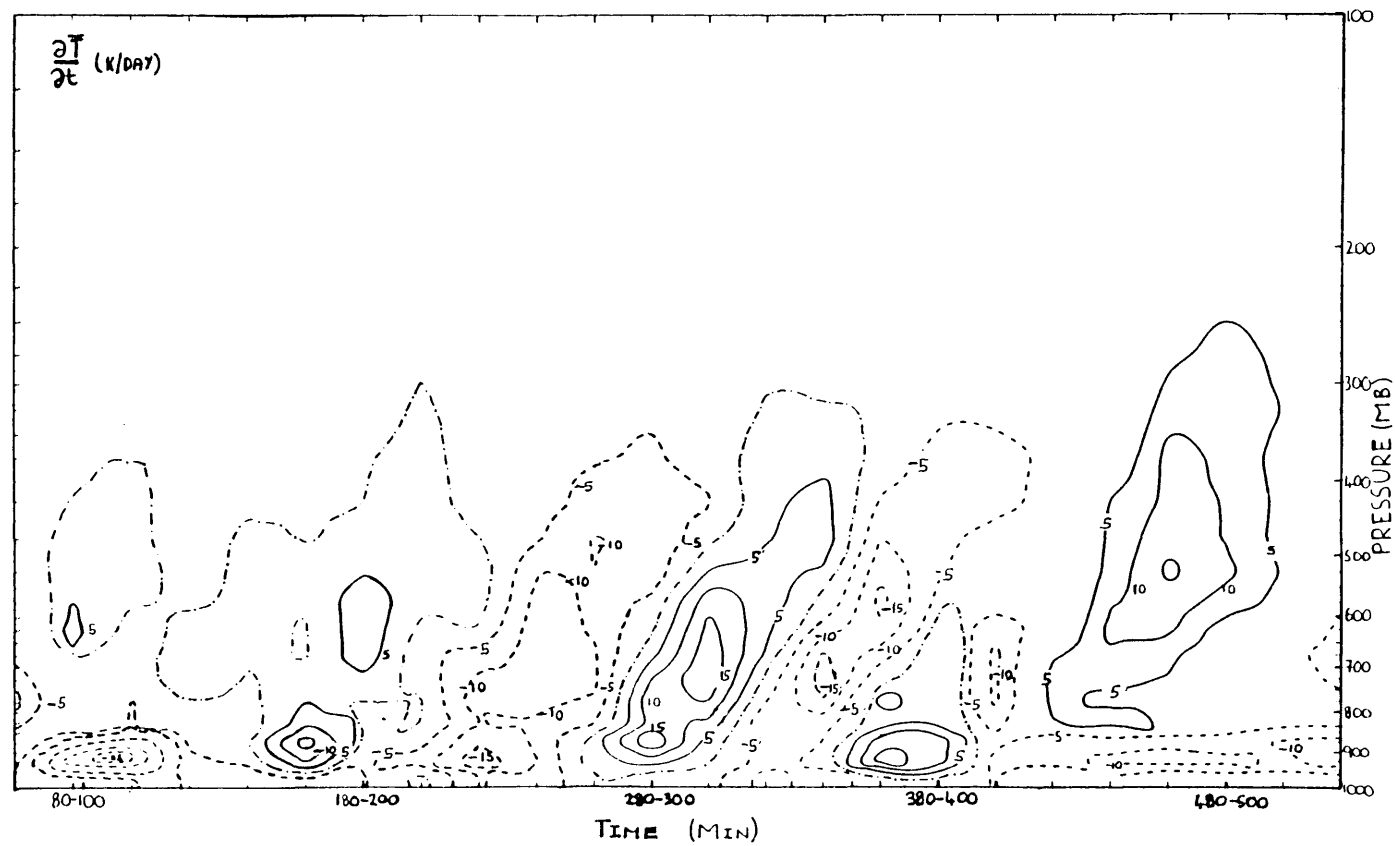


Fig 4.16(b) $\partial T/\partial t$ 20 minute averages.

(contour every 5K/day)

between 500 and 600mb towards the end of the simulation. The warm layer around 875mb, seen in the two hour averages, is shown on the pressure plots. This seems to be linked to the cloud forcing applied in this region.

Mixing ratio also shows unsteady behaviour (fig 4.17). The deviation from the initial model profile indicates that the lowest model layers dry continuously throughout the simulation. However $\partial \bar{q} / \partial t$ indicates that periods of moistening do occur within this layer (eg 320 to 340 minutes). $\partial \bar{T} / \partial t$, and $\partial \bar{q} / \partial t$ are negatively correlated above 900mb, the correlation not being clear below this. This would suggest that moist processes are involved to a significant extent in contributing to these two quantities.

This is clearly seen when the condensational heating is considered (fig 4.18). Heating and drying is well correlated to convective events as expected. Convection appears continuous between 1 and 4 hours of the simulation, with heating between 900 and 225mb, and associated downdraught cooling below. However after this time individual convective events seen to dominate the domain. Three major events occur, one starting between 260 and 280 minutes, and the other between 360 and 380, and 420 and 440 minutes. Some overlap occurs, especially during the last two events. Downdraught coolings are seen to form 20 to 40 minutes after the start of convection. These are clearly associated with cooling and moistening of the lowest layers of the atmosphere.

In addition to heating a large depth of the domain, cooling due to moist processes also occurs, up to pressures between 400 and 300mb. Examples of this feature are at 260 to 280 minutes, and 360 to 380 minutes. They occur immediately after strong condensational energy release. Peak coolings of -5K/day are seen, compared to 20-30K/day warmings during active convection. The depths of the features are too deep to be associated with convective scale downdraughts. They are due to evaporation of the remnant cloud and rain water left after active convection has finished. Hence convective motions play a part in cooling and moistening large depths of the atmosphere, together with the large-scale effects. The convective contribution can be up to 50% of the large-scale. Cooling and moistening is also seen at the top of the model domain, between 200 and 100mb. From the cloud area

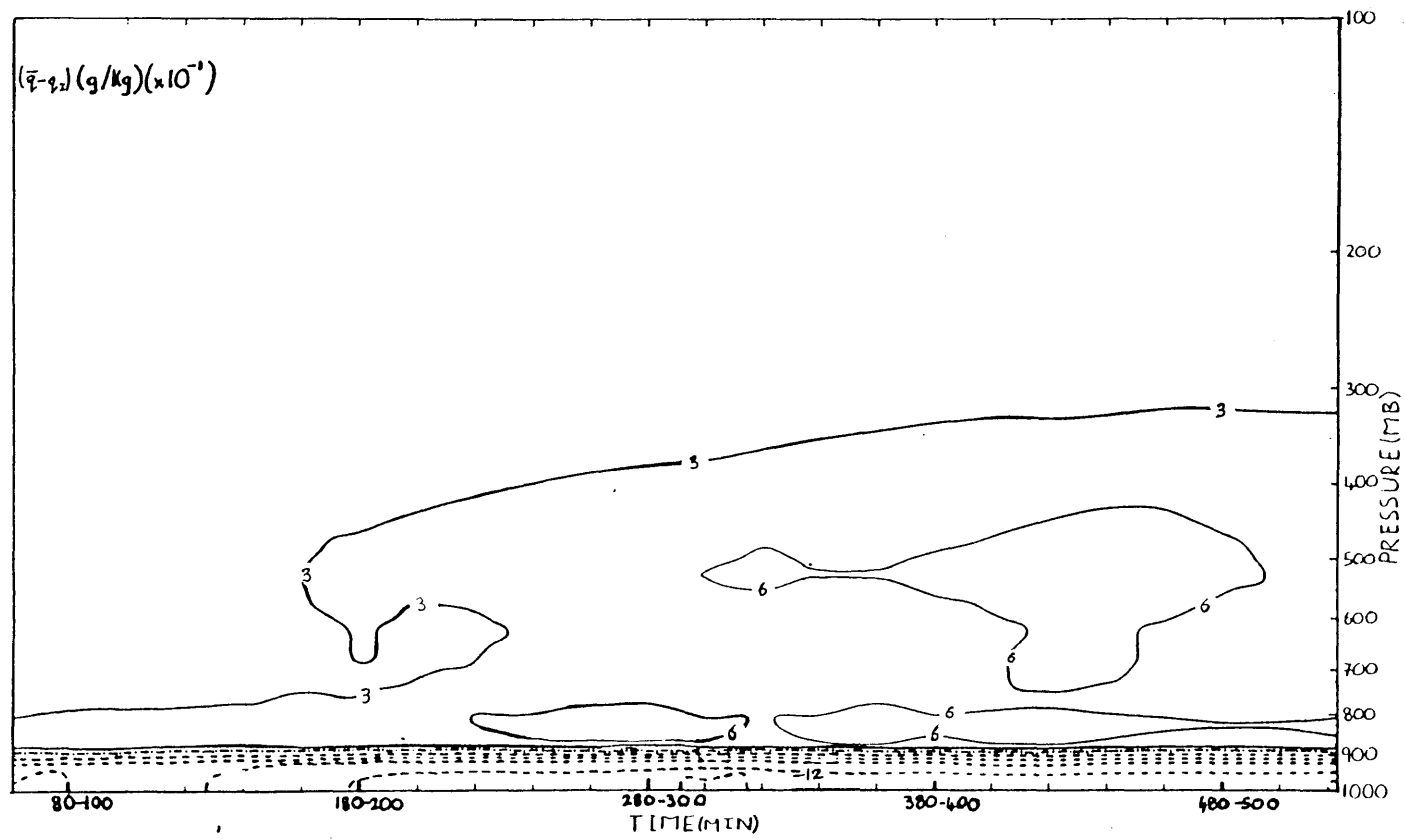


Fig 4.17(a) $(\bar{q} - q_I)$ 20 minute averages.

(contour every 0.3 g/kg)

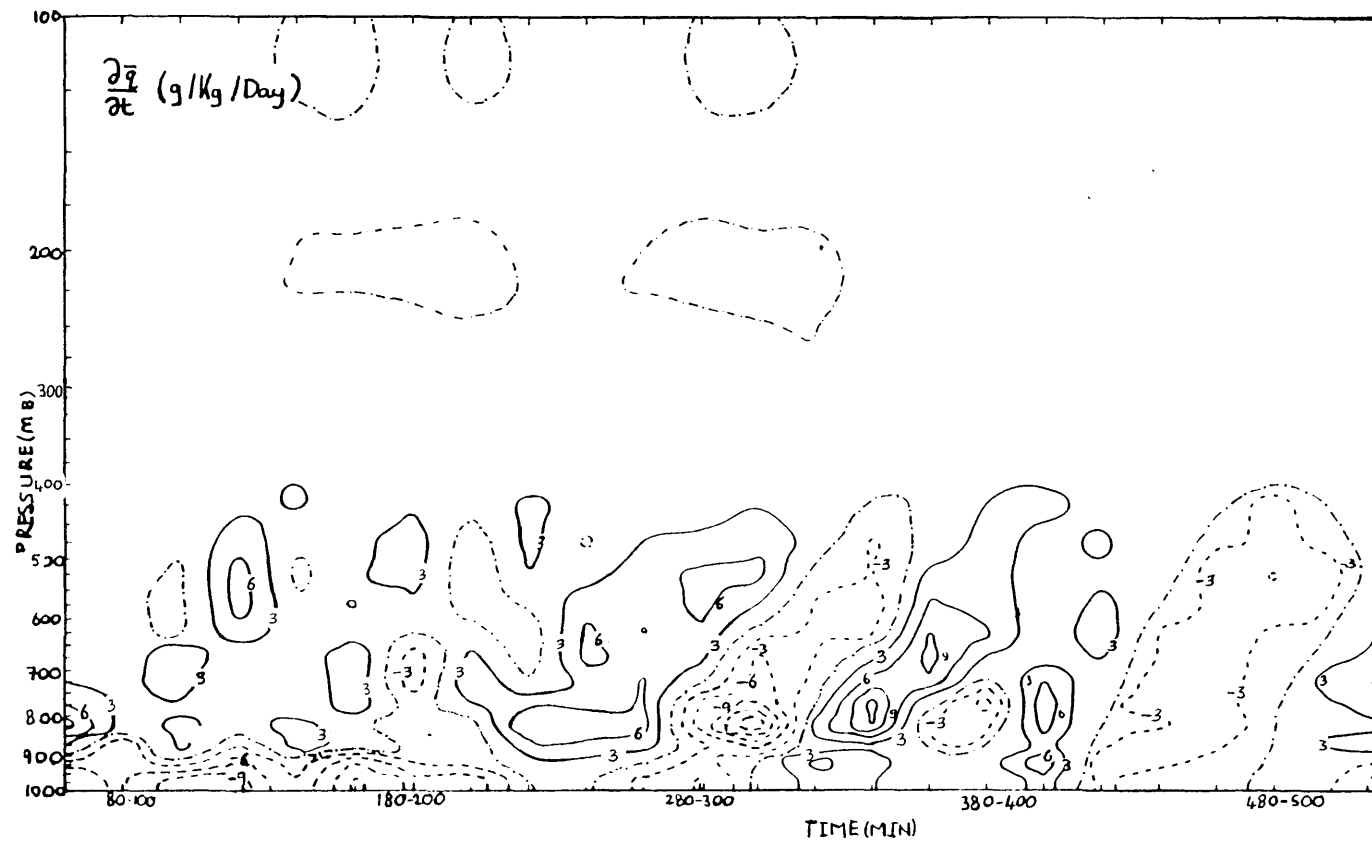


Fig 4.17(b) $\partial \bar{q} / \partial t$ 20 minute averages.
 (contour every 3 g/Kg/day)

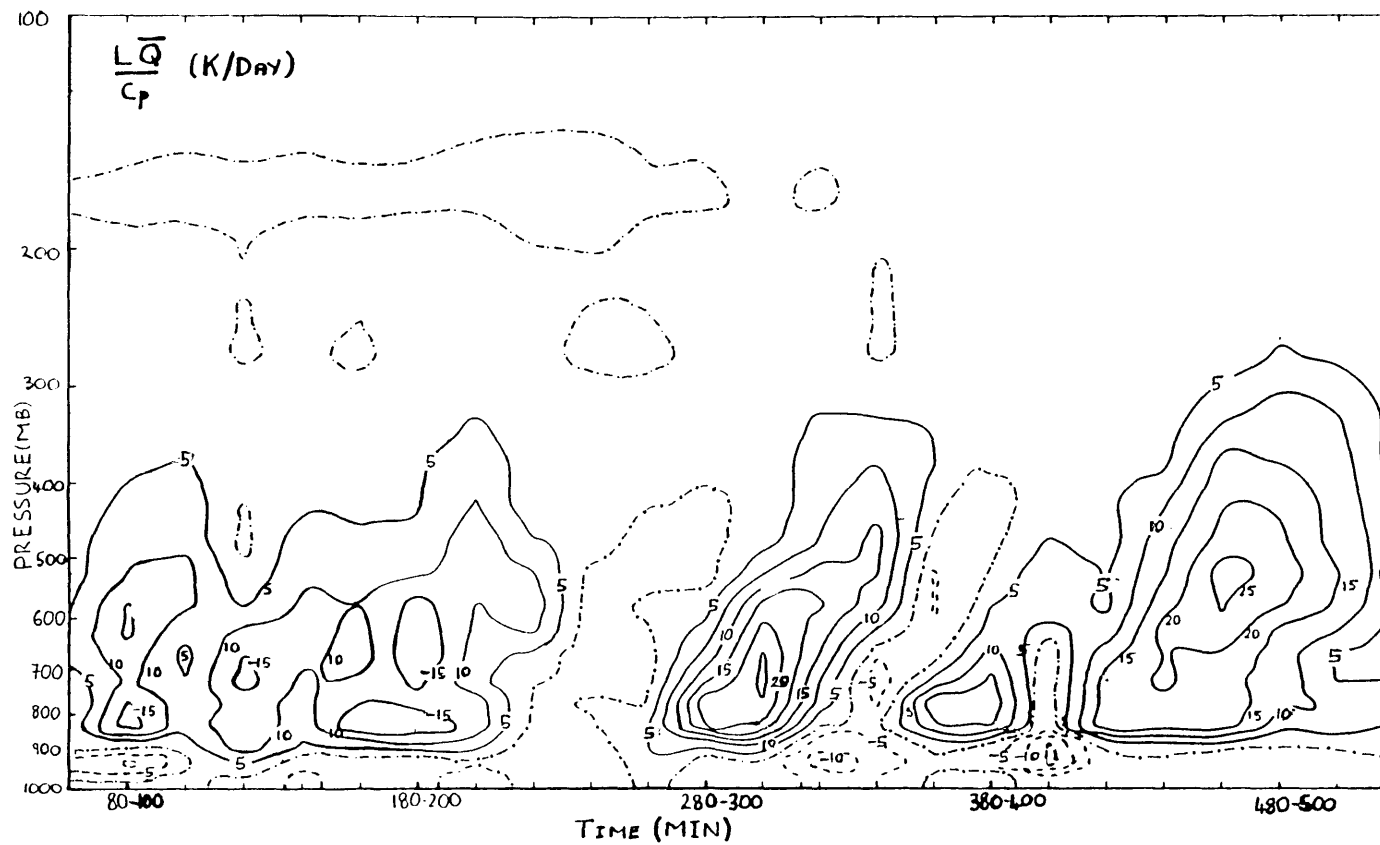


Fig 4.18 $L\bar{Q}/c_p$, 20 minute averages.

(contour every 5K/day)

against time plots, anvil cloud exist in this region, and so the cooling can be attributed to evaporation of such cloud.

The vertical eddy flux of temperature and moisture are shown in fig 4.19 and 4.20. The structure observed in the 8 hour average is present, and this seems related to the periods of latent heat release. The vertical transport of temperature has peak heating rates of 3 to 6K/day, much smaller than those due to condensation. However they are larger than observed in the longer time averages. The equivalent term for moisture has maximum moistening rates of 3 to 6g/kg/day, approximately 50% of the condensational drying rate.

Good correlation between eddy transports and condensational events proves that their association with convection is correct. During active and growing phases of convection, they remove energy from the lower troposphere and distribute it in the upper half of the domain. However the pressure-time plots show a different behaviour associated with the decay of convection. Examples of it are seen around 340 to 360 minutes, and 240 to 260 minutes. Here eddy transports cool and dry the atmosphere. Without knowledge of the internal cloud structure it is difficult to interpret this result. These will be presented in chapter 5, and this feature returned to in chapter 6.

(b) Budgets.

Budgets of energy over twenty minute periods are presented in fig 4.21(a). Again continuous convection is seen over the first 5 hours of the simulation. The net energy release is similar in magnitude to that of the large-scale moisture forcing, indicating that balance between the two exists. Oscillations do occur during this time, indicating that several convective cells grow and decay. The convective events which occur in the latter half of the integration have peak energy releases in excess of the applied forcing. However for much of this period $\langle \partial T / \partial t \rangle$ and $\langle \partial q / \partial t \rangle$ are negative and positive respectively, indicating that large-scale effects still exceed the convective energy release.

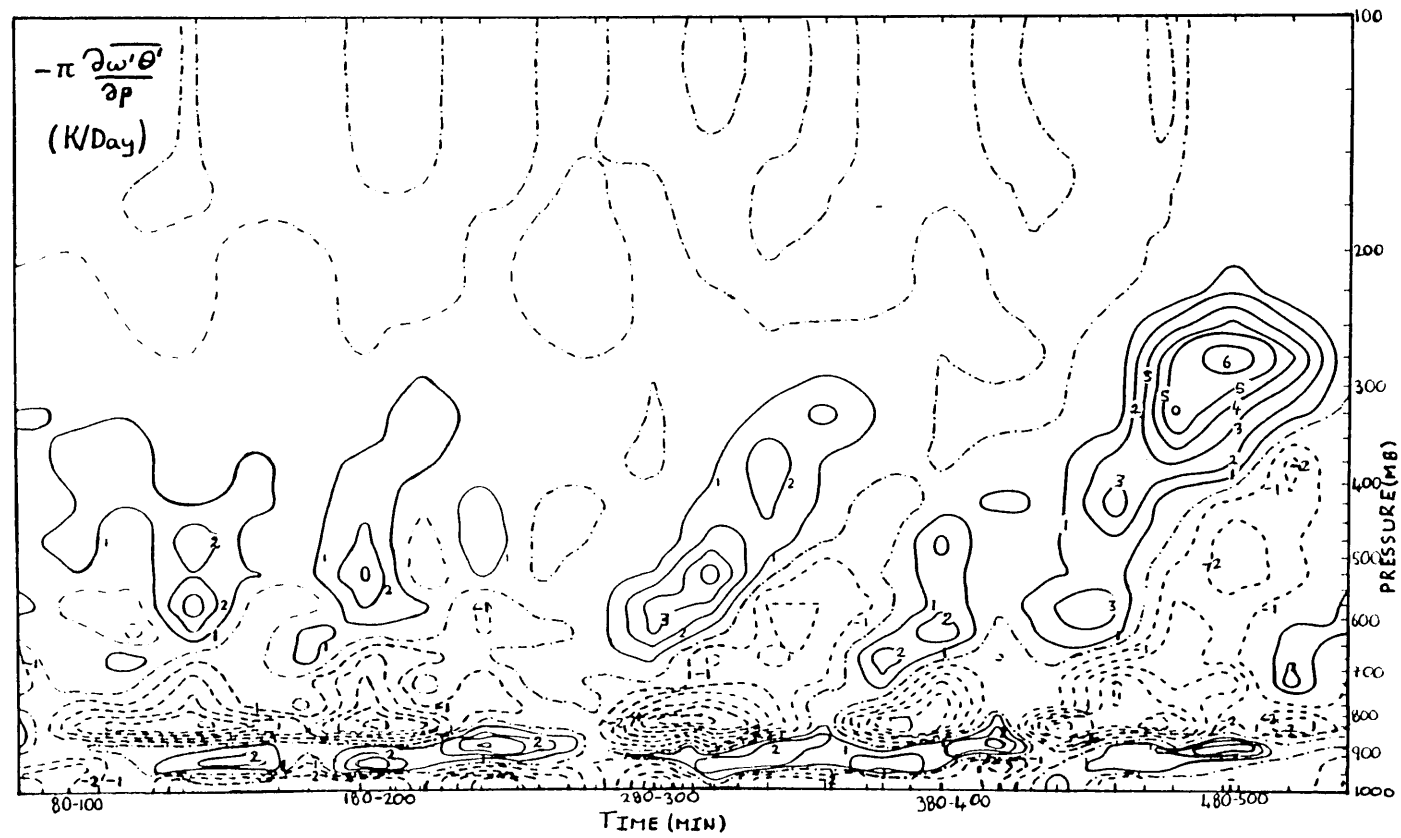


Fig 4.19 $-\pi \frac{\partial \bar{\omega}' \theta'}{\partial p}$ 20 minute averages.
(contour every 1 K/day)

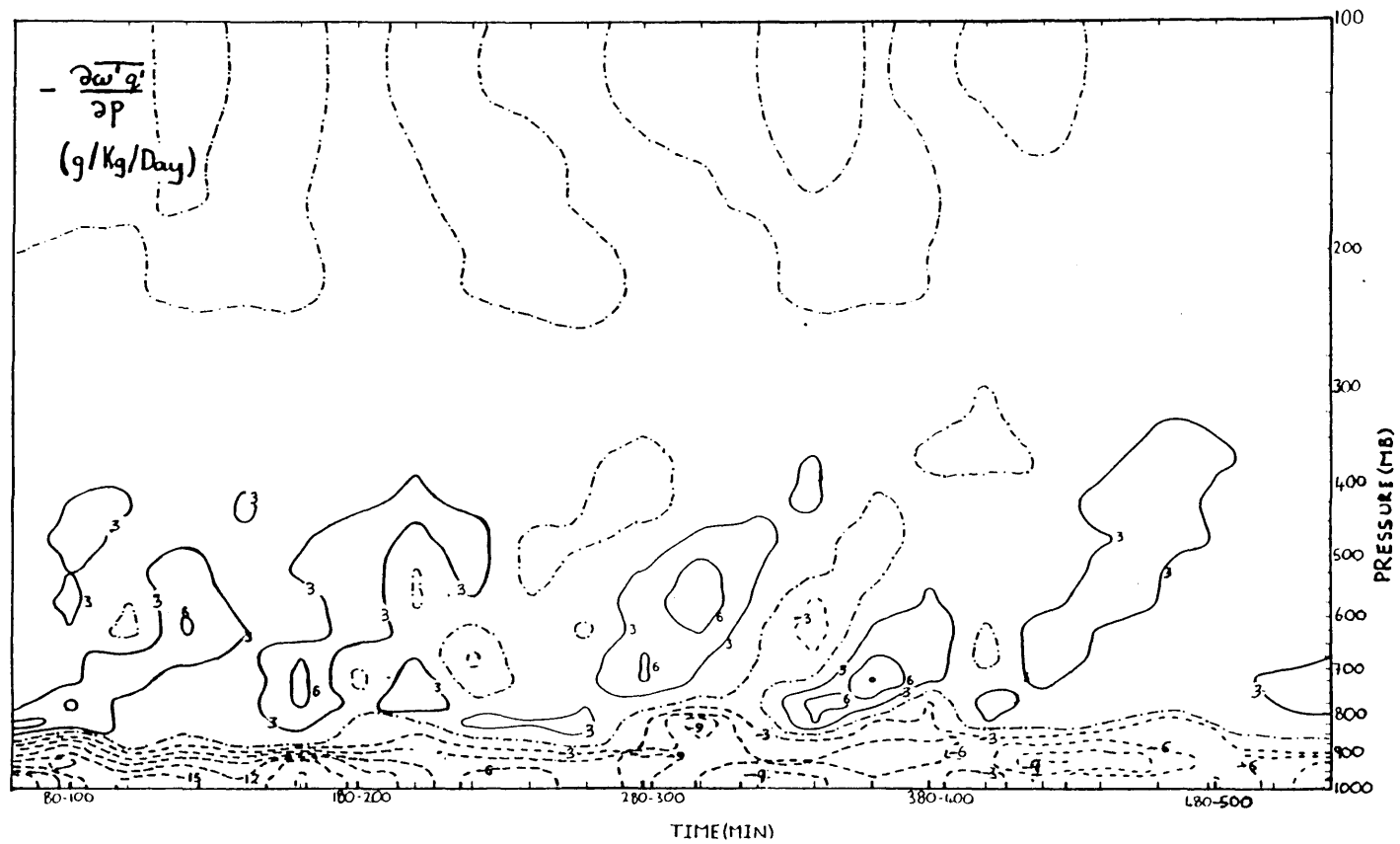


Fig 4.20 $-\frac{\partial \overline{\omega'q'}}{\partial p}$ 20 minute averages.
 (contour every 3g/Kg/day)

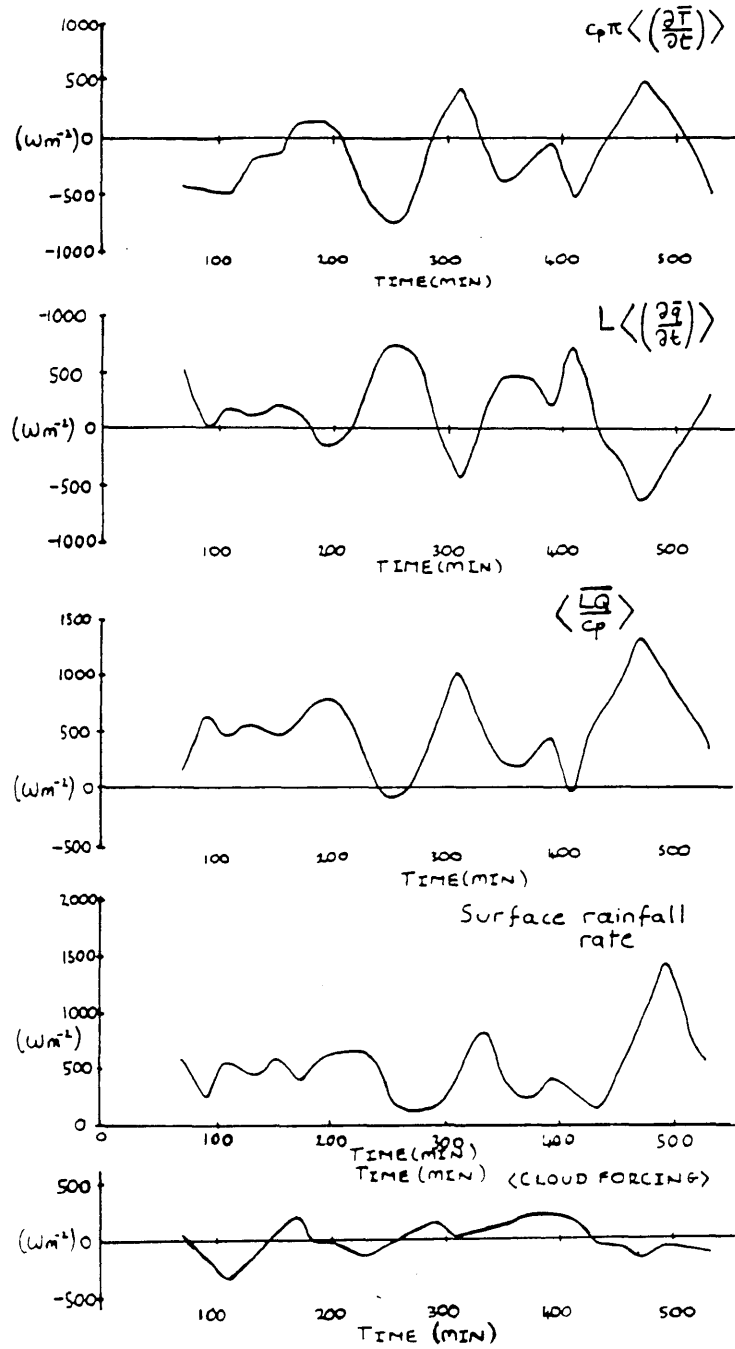


Fig 4.21(a) Budgets of $\frac{\partial \bar{T}}{\partial t}$, $\frac{\partial \bar{q}}{\partial t}$, $\bar{Q}$, rainfall and cloud forcing.

Surface rainfall lags the convective energy release by up to twenty minutes. This represents the fall time of the rain once created in the clouds. The contribution of the cloud forcing to the energy budget of the domain is also shown in fig 4.21(a). The maximum contribution is less than 100Wm^{-2} . No firm correlation exists between periods when latent heat release is a maximum and the maxima of the cloud forcing. Indeed, during the last large energy release due to convection, the cloud forcing is negative.

Fig 4.21(b) shows the relationship between liquid water storage and condensational heat release on the twenty minute time scale. Unlike the eight hour average, liquid water storage can be large. When convective events occur, such storage dominates that of water vapour. The majority of the liquid water falls out of the atmosphere, as $\partial \bar{q} / \partial t$ hardly exceeds the moistening provided by the external forcing.

4.2.2 Spatial Structure.

It has already been commented upon that a study of cloud area against time, indicates that distinct "anvil" and "clear" regions exist. Anvil areas contain both deep and shallow convection, whereas no deep convection is seen over the clear areas. Why does such a structure exist?

The obvious reason is that the cloud forcing has some preferred horizontal scale, confining large warmings to certain areas of the domain. However when the cloud forcing is studied, it appears to be evenly distributed throughout the domain. Although there is some correlation between large heating rates and deep convection, large heatings occur with the "clear" areas. Hence it would seem that convection within the anvil regions is suppressing convection within the clear areas.

Condensational heating and vertical eddy transports were diagnosed for the anvil and clear areas, showing that each has a distinct physical structure.

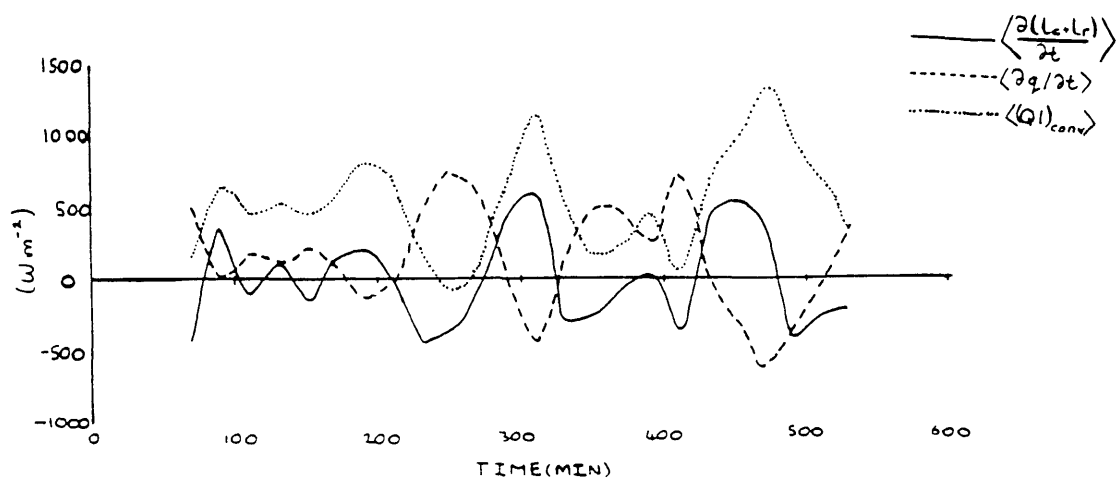


Fig 4.21(b) Budgets of water vapour and liquid storage.

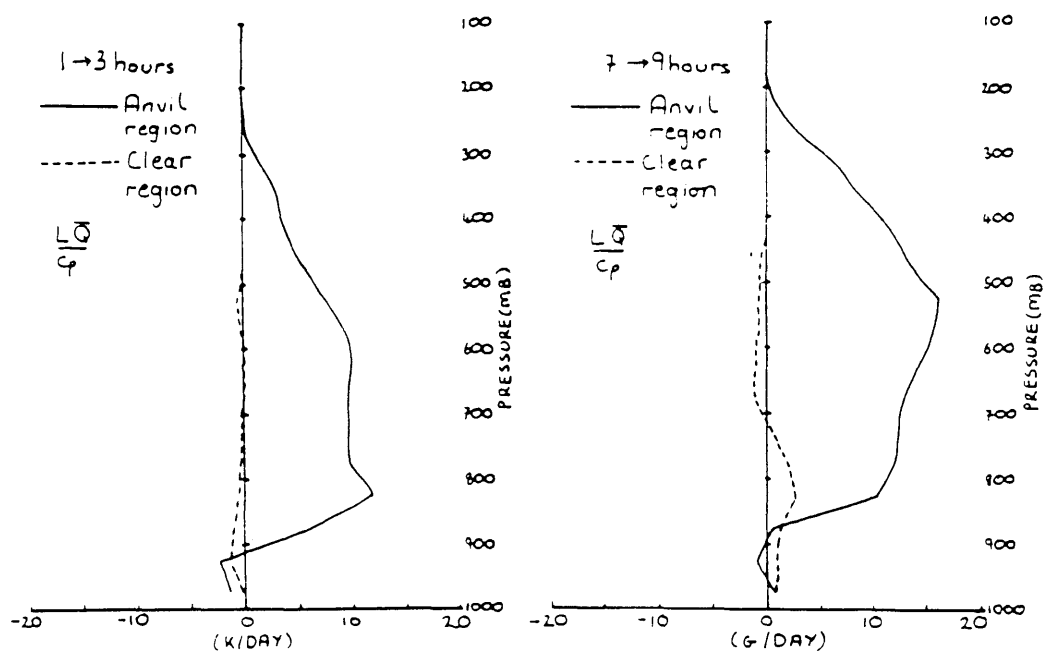


Fig 4.22 $L\bar{Q}/c_p$ for anvil and clear regions of the domain averaged between 1 and 3, 7 and 9 hours.

(a) Vertical structure.

All deep convection occurs under the anvils, together with the majority of the condensational energy release. Fig 4.22 compares this quantity for anvil and clear areas for 1 to 3 hours and 7 to 9 hours. The anvil average shows all the features in domain averaged quantity. In the clear area, little convection occurs within the first 6 hours, in fact the net moist processes energy release is negative. This must be due to evaporation of cloud transports into the clear areas of the domain by the circulation linking the two areas. After 6 hours, deeper convection occurs with a positive energy release up to 700mb, with cooling above.

The eddy flux divergence of temperature and moisture for anvil and clear areas are compared in figs 4.23, and 4.24 respectively. Again most of the eddy flux divergence occurs in the anvil areas, associated with the deep convection present there. In the clear region in the first 6 hours of the simulation, the eddies remove heat and moisture from the lowest 100mb, and redistribute it between 900 and 800mb. This action is usually associated with shallow, non-precipitating convection. A large number of shallow clouds exist within the cloud height spectrum of section(4.1.2). Hence it would seem, at first sight, that the model has sufficient vertical resolution to model their transport properties. However this is quite surprising as the majority of these features exist on the grid scale. Further comments on this will be made in section(6.4) of chapter 6. As convection in the clear region increases in depth, the eddy transports begin to look similar to those for deep convection, but of smaller vertical extent.

The reason for this formation of the two distinct areas is seen when the deviation of mixing ratio from the initial profile is considered (fig 4.25). The initial convection dries out the lowest 100mb of the clear air three times more than the anvil region. This reduces the value of CAPE drastically, and so deep convection cannot exist over these areas. Towards the end of the simulation, the boundary layer of the clear area recovers to a similar state to that of the anvil area (7 to 9 hours). The atmosphere above 900mb is moister in the anvil than in the clear regions. An increase in q' above 900mb is seen as convection within the clear regions deepens.

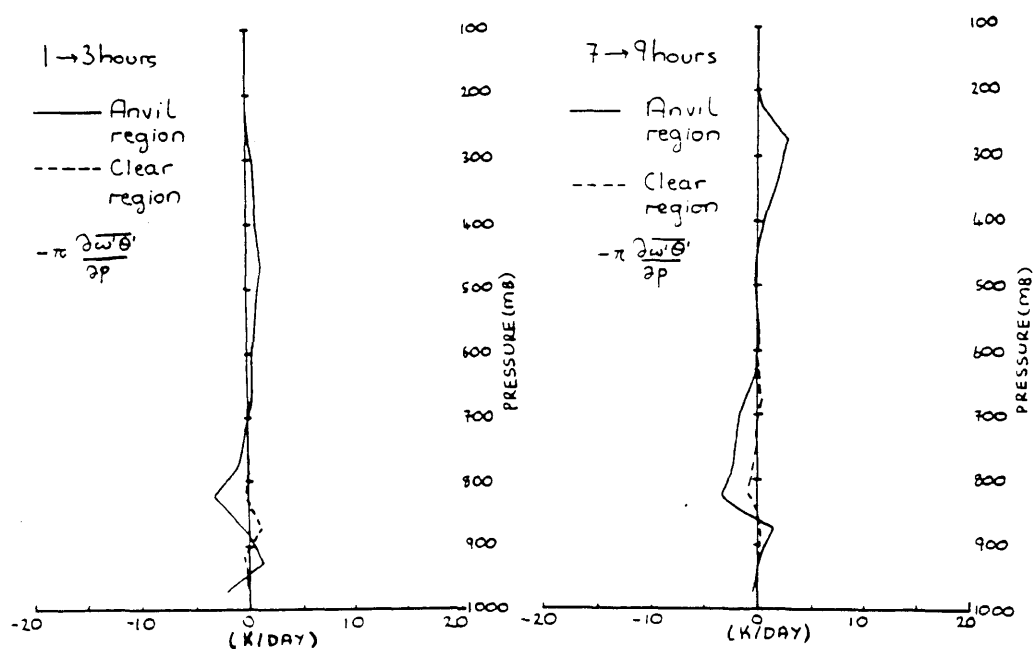


Fig 4.23 $-\pi \overline{\partial \omega' \theta'} / \partial p$ for anvil and clear regions of the domain averaged between 1 and 3, 7 and 9 hours.

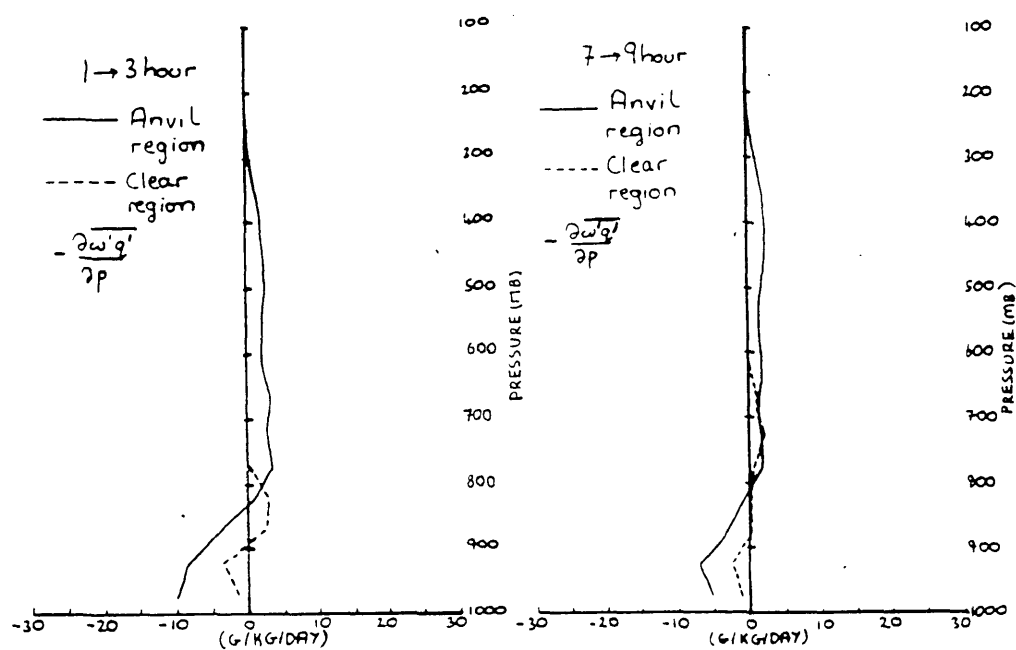


Fig 4.24 $-\partial \overline{\omega' q'} / \partial p$ for anvil and clear regions of the domain averaged between 1 and 3, 7 and 9 hours.

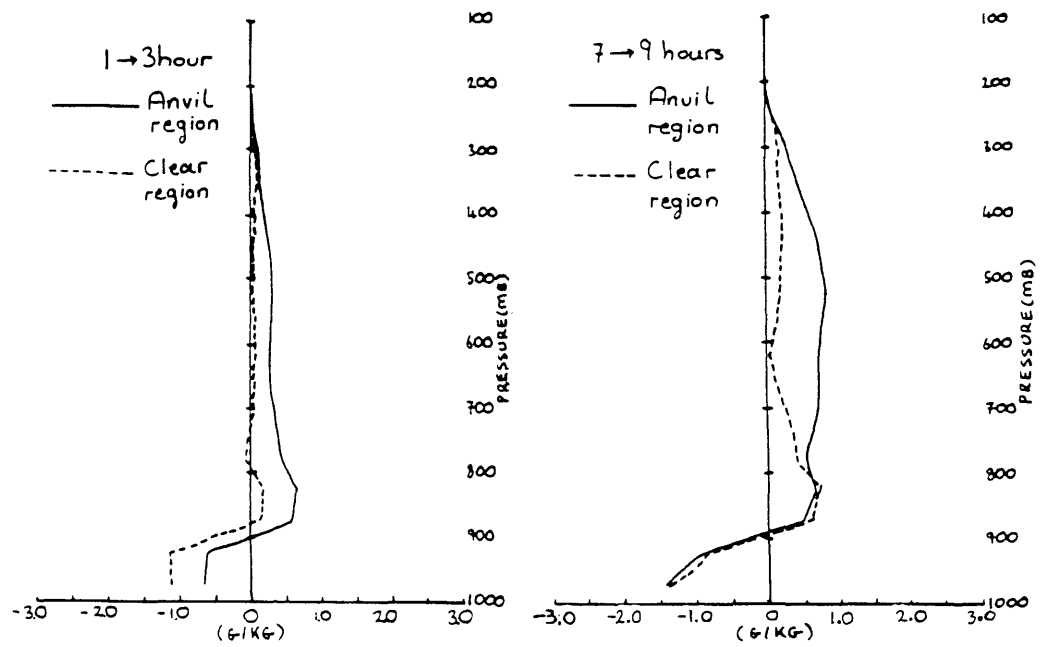


Fig 4.25 $(\bar{q} - q_r)$ for anvil and clear regions of the domain averaged between 1 and 3, 7 and 9 hours.

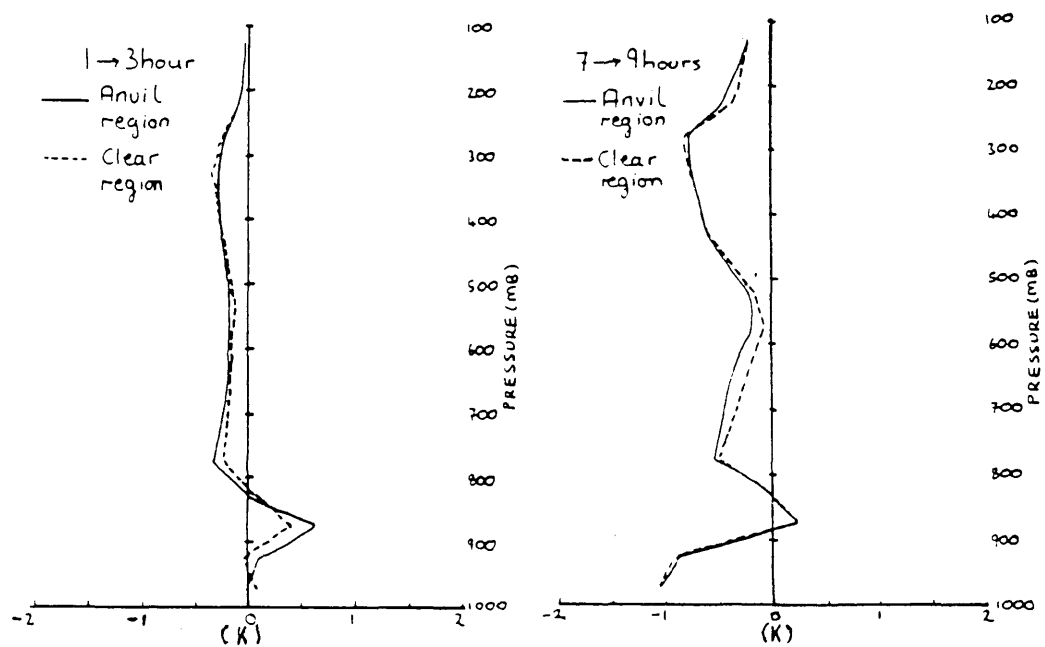


Fig 4.26 $(\bar{T} - T_r)$ for anvil and clear regions of the domain averaged between 1 and 3, 7 and 9 hours.

The deviation of temperature from the initial profile is similar for both areas (fig 4.26). The cooling of the lower layers of the clear area develops more gradually than that of the anvil regions. This would imply that the cold air from convective scale downdraughts gradually spreads over the entire domain. The cooling between 500 and 600mb is slightly less in the clear area, even though there is no large condensational heating. The evaporation of cloud air within anvil regions would account for this cooling.

(b) Relationship between anvil and clear regions.

The fact that the thermodynamic structure of both anvil and clear areas are similar, suggests that the same physical processes are responsible for determining their structure. Over the domain it has been shown that a balance between convection and large-scale forcing determines the atmospheric structure. However only anvil regions contain significant amounts of convection. Hence convective effects must be transferred between anvil and clear regions by the flow set-up by the convection.

This is clearly seen when pressure-time plots of $\partial \bar{T} / \partial t$ are considered for both regions (fig 4.27). For the entire domain $\partial \bar{T} / \partial t$ and condensational heating are well correlated. The same is true for the anvil regions. However $\partial \bar{T} / \partial t$ is similar for both anvil and clear regions. There is some evidence for a lag between the two regions. For example, the cooling around 900mb is seen to take place at 400 to 420 minutes in the anvil average, but at 440 to 460 minutes in the clear regions. Similarly the heating of 440 to 460 minutes in the anvil average occurs at twenty minutes later in the clear average. Hence the effect of convection propagates from the anvil region across the domain. Similar features are seen in $\partial \bar{q} / \partial t$.

Significant flow occurs across the boundaries between the two regions. Fig 4.28 shows budgets of $(Q_2)_{\text{conv}}$, $\partial \bar{q} / \partial t$, $-(\partial u q / \partial x)$, $-\pi(\partial v \theta / \partial x)$, and $-\pi(\partial \omega \theta / \partial p)$ for the anvil region. Large horizontal convergences of heat and moisture occur when convection is active. Vertical transport of temperature cools the anvils areas during this time, the upward mass transport in the cloud exceeding the downward motion in the cloud environment within the anvil region. The horizontal convergence occurs in the lowest 100mb of the domain.

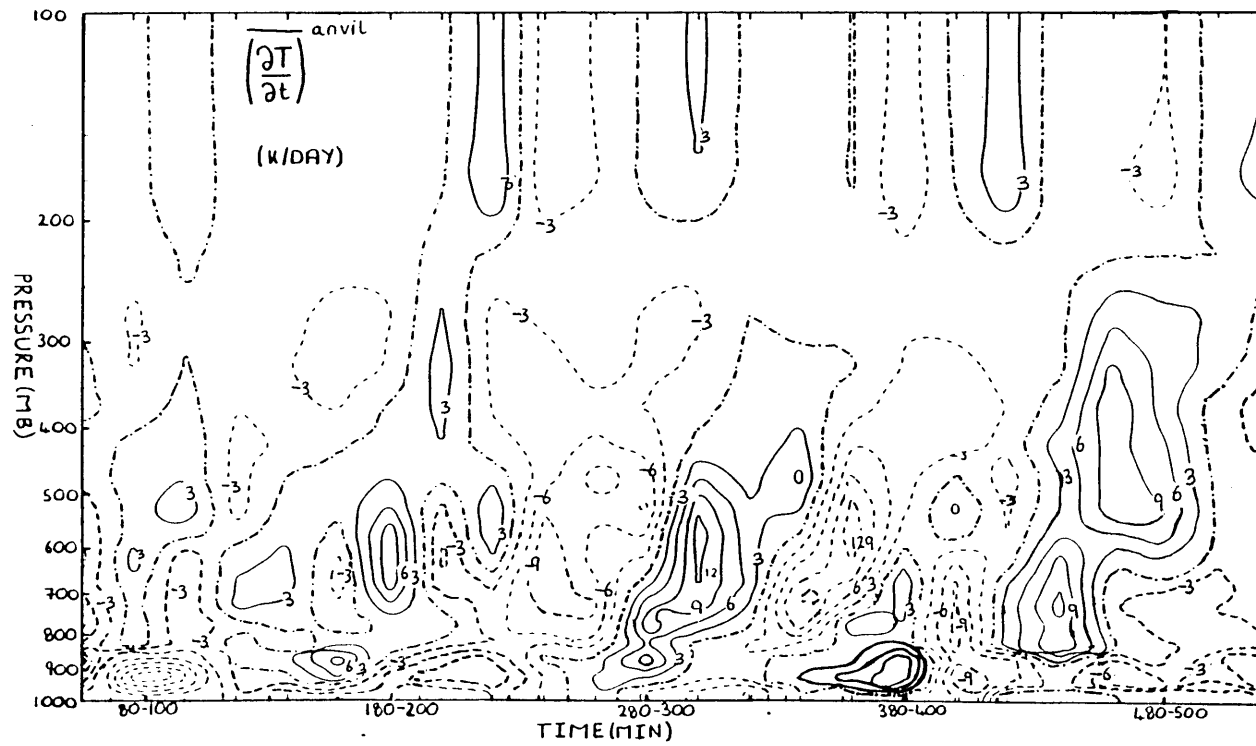


Fig 4.27(a) $\overline{\partial T/\partial t}$ for anvil regions of the domain
(twenty minute averages).
(contour every 3 K/day)

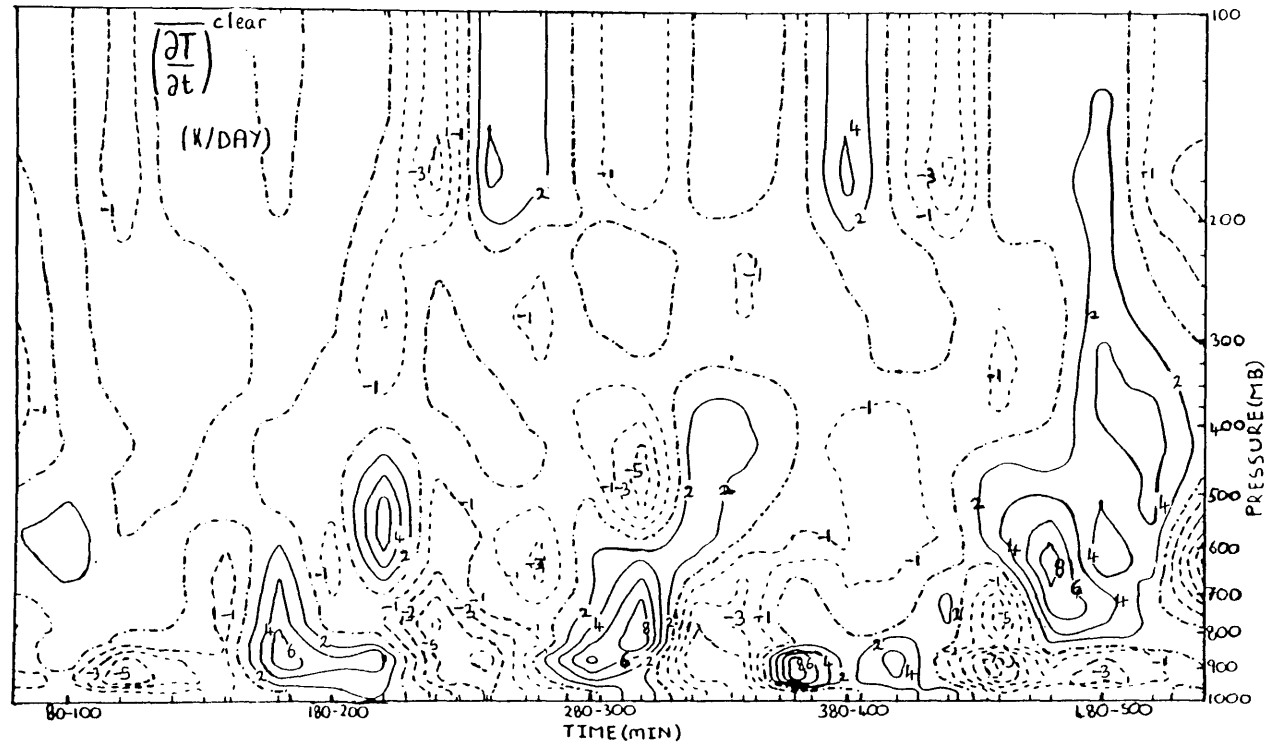


Fig 4.27(b) $\overline{\partial T / \partial t}$ for clear regions of the domain
 (twenty minute averages)
 (contour every 2 K/day)

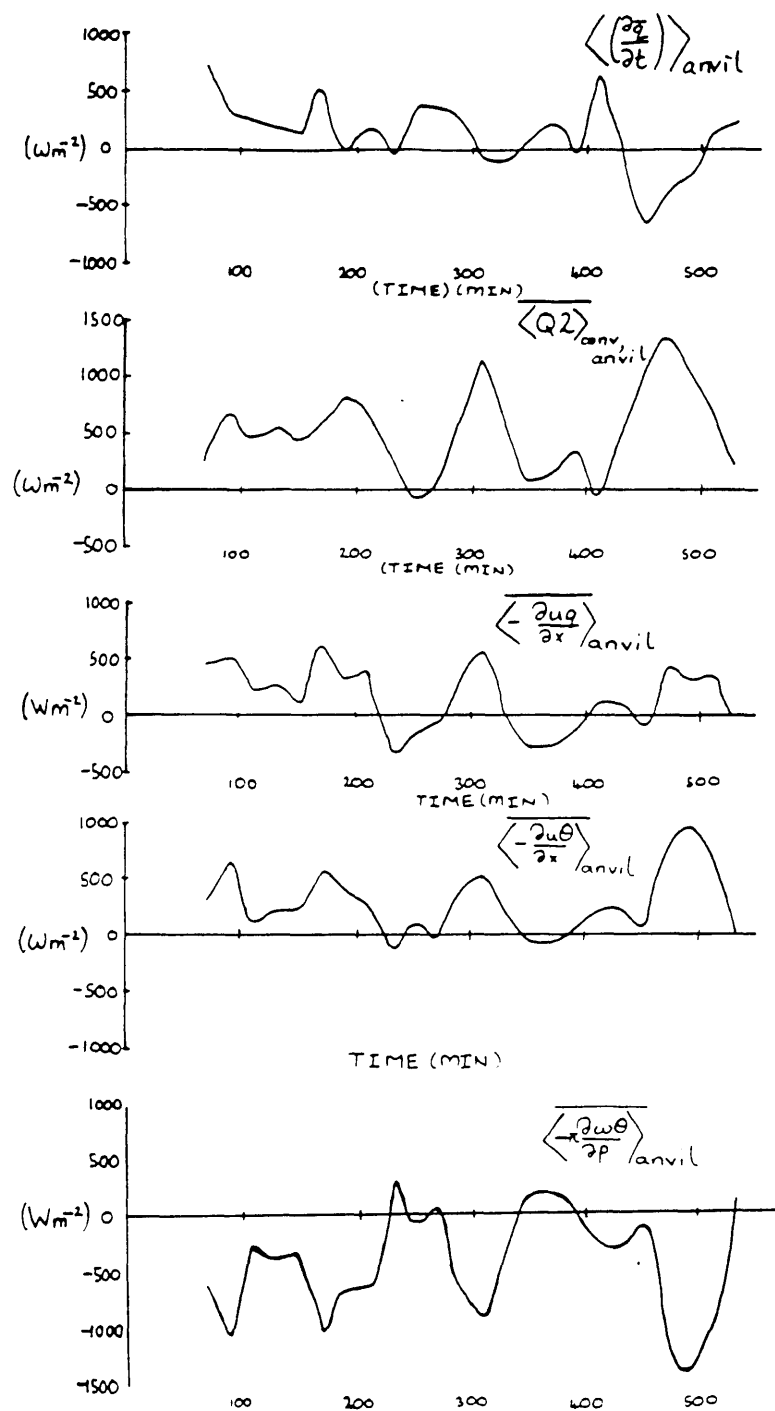


Fig 4.28 Budgets of thermodynamic quantities over anvil regions.

At times when convection decays, the anvil area dries and warms, linked to a net downward motion. Horizontal divergence is seen in the lowest 100mb.

The net mass flux over the domain must be zero due to the cyclic nature of the domain. Thus the effect within the clear is opposite to that in the anvil regions. Table 4.3 gives a summary of the effects induced by convection. These are not dependent upon the cyclic nature of the model. Large-scale values for vertical velocity are a few centimeters per second, compared to tens of meters per second for updraughts. Thus much of the upward mass flux in cloud is balanced by motion within their environment. If the scale on which the observations were made was near the cloud scale, then similar effects as those observed within the model would be seen.

The results show that different but linked physical processes occur in different regions of the model. The moist convective processes produce a circulation which propagates throughout the domain. The balances occurring within the domain, between different physical processes, differs between the two areas defined.

4.2.3 Further Discussion And Summary.

The study of the temporal and spatial structure of the convection within the simulation illustrates some interesting points with regard to the parameterization of convection.

(a) The time dependent nature of deep convection is not seen in the observations. This may imply that not enough convection is included in the domain average. In the first 5 hours, the effect of convection is more continuous. Here more clouds are seen, with a greater overlap in cloud life spans than later in the simulation. With a three dimensional simulation, many more clouds might be expected within the domain. Therefore the time dependence is not seen.

Region	Convective activity	Net vertical motion	Effect on thermodynamic structure	Large-scale forcing
Anvil	Active	Upward	Cooling and moistening	Cooling and moistening
	Decaying	Downward	Warming and drying	Cooling and moistening
Clear	Active	Downward	Warming and drying	Cooling and moistening
	Decaying	Upward	Cooling and moistening	Cooling and moistening

Table 4.3 Summary of the effects of the circulation induced between anvil and clear areas due to deep convection.

However the evolution of large-scale and coarse meso-scale models is towards finer spatial and temporal resolutions. Hence the number of clouds to be parameterized within a gridbox become fewer. Thus time dependent effects become important and have to be incorporated within such schemes (although the effect of their impact is unclear).

(b) The circulation which exists between anvil and clear areas illustrates an important point which parameterization techniques miss out. This is that the horizontal flow across gridbox boundaries due to convection might be important.

In the present formulation of many deep convective schemes, the mass flux balance within a gridbox is written,

$$\bar{m} = \bar{m}^c + \bar{m}^e \quad \text{eqn(4.6)}$$

where $\bar{m}$ is the grid point vertical mass flux,
defined by the model,

$\bar{m}^c$ is the cloud mass flux, predicted by
the parameterization scheme,

and $\bar{m}^e$ is the environmental mass flux.

The cloud circulation is assumed to be confined within the gridbox, ie. all vertical mass transports due to clouds are compensated by vertical mass transports within the clear air, except for a residual which is assumed to form the grid point mass flux. No flow due to convection across gridbox boundaries is considered.

However the definition of anvil and clear areas illustrates that, when the area averaged over is on the cloud scale, such horizontal flows are very important. Thus representation of horizontal eddy fluxes due to convection need to be incorporated into parameterization schemes.

4.3 SUMMARY

In this chapter, comparison has been made between the simulated cloud ensemble and observations of the GATE area. Good agreement is seen. This gives confidence that the transport properties analysed

from the model accurately represent those occurring in the real atmosphere.

The temporal structure of the ensemble has been presented. Although due to having too few clouds within the domain, they show important and new aspects of the effect of convection upon the atmosphere. Although deep convection warms and dries a large depth of the domain during it's growing and mature phase, while decaying cooling and moistening occurs, not only in the lower layers of the domain as expected from downdraughts, but up to 400mb. This will be studied further in later chapters.

The spatial structure illustrates that the horizontal flow produced by convection has important effects upon the temperature and moisture budgets of the atmosphere.

A more detailed study of the effects of convection upon the atmosphere requires information on the internal cloud structure. This is presented within the next chapter.

CHAPTER 5

PARAMETRIC THEORY AND DEFINITION OF THE MEAN ENSEMBLE CLOUD

In this chapter the mathematical theory of the parameterization of precipitating convection is developed. Expressions for $(Q1)_{conv}$, and $(Q2)_{conv}$ are obtained in terms of cloud and large-scale parameters. Although this has been carried out by several previous workers, several approximations were made concerning the use of one dimensional cloud models. The use of a numerical model to specify cloud structure makes the use of approximations unnecessary.

Using the simulation described in chapters 3 and 4, the internal cloud structure of updraughts, downdraughts, and inactive cloud is derived, averaged over the last eight hours of the simulation. The contribution of deep convection to the "mean ensemble cloud" is discussed, and a brief comparison is made to the one dimensional cloud model of Anthes(1977). Brief comments are also made concerning the twenty minute average structure of the ensemble.

5.1 THE THEORY OF CONVECTIVE PARAMETERIZATION.

In this section a mathematical theory of the parameterization of precipitating convection is developed. This has been carried out by

several previous workers (Ooyama(1971), Yanai(1971b), and Arakawa(1971)), for use with one dimensional cloud models, (which specify the cloud variables), and involving several approximations in the course of the derivation. Using the meso-scale model to supply cloud quantities such approximations are not required, and so a fuller analysis may be carried out.

5.1.1 Basic Concepts.

The aim of cumulus parameterization is to deduce the magnitude and structure of $(Q1)_{conv}$ and $(Q2)_{conv}$ (and also $(Q3)_{conv}$) from the large-scale structure of the atmosphere. The starting point for parametric theory are the approximate area averaged equations described in chapter 2;

$$\frac{\partial \bar{\theta}}{\partial t} + \frac{\partial \bar{u}\bar{\theta}}{\partial x} + \frac{\partial \bar{\omega}\bar{\theta}}{\partial p} = \frac{L\bar{Q}}{c_p \pi} - \frac{\partial \bar{\omega}'\bar{\theta}'}{\partial p} + \bar{S}_\theta - \left(\frac{\partial \bar{\theta}}{\partial t} \right)_{RFD} \quad eqn(5.1)$$

$$\frac{\partial \bar{q}}{\partial t} + \frac{\partial \bar{u}\bar{q}}{\partial x} + \frac{\partial \bar{\omega}\bar{q}}{\partial p} = -\bar{Q} - \frac{\partial \bar{\omega}'\bar{q}'}{\partial p} + \bar{S}_q \quad eqn(5.2)$$

(In the model equations, additional terms appear on the rhs representing large-scale forcing, cloud scale forcing, and surface fluxes. These are discussed in chapter 6. They do not effect the derivation of expressions for $(Q1)_{conv}$ and $(Q2)_{conv}$ presented here.)

Consider the area A over which the equations are averaged. The value of Φ averaged over this area can be written as,

$$\bar{\Phi} = \sigma \bar{\Phi}^c + (1-\sigma) \bar{\Phi}^e \quad eqn(5.3)$$

where $\bar{\Phi}^c$ is the value of Φ averaged over
cloud air within A,

$\bar{\Phi}^e$ is the value of Φ averaged over
clear air within A,

and σ is the fractional area of cloud
within the domain,

=a/A, where a is the area of cloud.

It is usual to assume that $\sigma \ll 1$.

Equation(5.3) may be rearranged to give,

$$\bar{\phi} = \bar{\phi}^e + \sigma(\bar{\phi}^c - \bar{\phi}^e)$$

If $(\bar{\phi}^c - \bar{\phi}^e) \ll \bar{\phi}^e$, then as $\sigma \ll 1$,

$$\bar{\phi} \doteq \bar{\phi}^e \quad \text{eqn(5.4)}$$

ie. the structure of the environment is described by the large-scale structure of the atmosphere. This is usually assumed to be true for θ and q , and is very important for parameterization. If it were not so then some estimate of cloud environment structure would need to be made independently of cloud structure. However eqn(5.3) shows that $\bar{\phi}^c$ and $\bar{\phi}^e$ are not independent, and so the problem cannot be solved.

Using the meso-scale model the validity of the approximation may be estimated. Using typical values for average convective towers ($\sigma \approx 5\%$, $\bar{\theta}^e \approx 300\text{K}$, $\bar{q}^e \approx 10\text{g/kg}$, $(\bar{\theta}^c - \bar{\theta}^e) \approx 2\text{K}$, and $(\bar{q}^c - \bar{q}^e) \approx 2\text{g/kg}$) yields errors of less than 1% for θ , and $\approx 1\%$ for q . Because the model domain is 260km in the length, the approximation would seem good for large-scale models.

What about coarse meso-scale models? Considering a model with a grid length of 15km, and a cloud radius of 5km ($\sigma \approx 35\%$), the above thermodynamic values give errors of $\approx 1\%$, and $\approx 5\%$ for θ and q respectively. Thus the approximation also seems reasonable for such models.

The approximation is not valid for omega. In the ITCZ the mean vertical mass flux ($-\bar{\omega}$) is upward. However it is much smaller than the cloud vertical mass flux, due to much of the motion within the cloud air being compensated by motions within the clear air. Thus both cloud and environmental mass fluxes are important in determining $\bar{\omega}$.

Studies of simulations with initial vertical wind shears, presented in chapter 8, show that the approximation is true for horizontal momentum. In chapter 8 cloud momentum transports are briefly discussed. Here and in the next two chapter, only the thermodynamic effects of clouds are considered.

In parameterizing $(Q1)_{\text{conv}}$ and $(Q2)_{\text{conv}}$ two processes must be considered, (1) vertical eddy flux divergence, and (2) moist processes. The analysis of (1) is common to this work and the methods of previous workers. However for (2), different equations are used, those of previous workers being approximate forms.

5.1.2 Representation Of Vertical Eddy Flux Divergence

The vertical eddy flux of a quantity Φ was defined in chapter 2 as,

$$\overline{\omega'\Phi'} = \overline{\omega\Phi} - \bar{\omega}\bar{\Phi} \quad \text{eqn(5.5)}$$

Previous workers (eg Yanai et al(1973), Nitta(1975)) have expanded the rhs of eqn(5.5) in terms of cloud and large-scale variables. The analysis presented here uses the same techniques as these previous studies, but extends the analysis.

Expanding the rhs of eqn(5.5), using eqn(5.3),

$$\overline{\omega'\Phi'} = (\sigma\overline{\omega\Phi^c} + (1-\sigma)\overline{\omega\Phi^e}) - (\sigma\bar{\omega}^c + (1-\sigma)\bar{\omega}^e)(\sigma\bar{\Phi}^c + (1-\sigma)\bar{\Phi}^e) \quad \text{eqn(5.6)}$$

Several different cloud types may exist within the ensemble. Because ω , Θ , and q are well correlated within cloud air, it is incorrect to write,

$$\overline{\omega\Phi^c} = \bar{\omega}^c\bar{\Phi}^c$$

Using analytical cloud models (as previous workers), the summation over all cloud types is written,

$$\left. \begin{aligned} \sigma\overline{\omega\Phi^c} &= \sum_{i=1}^N \sigma_i \bar{\omega}_i^c \bar{\Phi}_i^c \\ \sigma\bar{\omega}^c &= \sum_{i=1}^N \sigma_i \bar{\omega}_i^c \\ \sigma\bar{\Phi}^c &= \sum_{i=1}^N \sigma_i \bar{\Phi}_i^c \end{aligned} \right\} \quad \text{eqn(5.7)}$$

where the subscript i refers to the i^{th} cloud type.

With a time dependent numerical model, it is difficult to analyse each cloud type, and so an average over all cloud air is used;

$$\left. \begin{aligned} \sigma \overline{\omega \theta^c} &= \sigma \frac{1}{a} \sum_{\text{cloud}} \omega \phi \\ \sigma \overline{\omega^c} &= \sigma \frac{1}{a} \sum_{\text{cloud}} \omega \\ \sigma \overline{\phi^c} &= \sigma \frac{1}{a} \sum_{\text{cloud}} \phi \end{aligned} \right\} \text{eqn(5.8)}$$

where a is the total area of cloud within the domain.

These two sets of summations can be shown to be equivalent. Expanding the summation for $\sigma \overline{\omega \phi^c}$ in eqn(5.7),

$$\sigma \overline{\omega \phi^c} = \frac{a_1}{A} \overline{\omega_1 \phi_1^c} + \frac{a_2}{A} \overline{\omega_2 \phi_2^c} + \dots + \frac{a_N}{A} \overline{\omega_N \phi_N^c} \quad \text{eqn(5.9)}$$

where a_i is the area of cloud type i

In analytical descriptions of cloud ensembles, each cloud type is solved separately. Hence ω and ϕ are correlated and eqn(5.9) can be expressed as,

$$\begin{aligned} \sigma \overline{\omega \phi^c} &= \frac{a_1}{A} \frac{1}{a_1} \sum_{\text{cloud } 1} \omega \phi + \frac{a_2}{A} \frac{1}{a_2} \sum_{\text{cloud } 2} \omega \phi + \dots + \frac{a_N}{A} \frac{1}{a_N} \sum_{\text{cloud } N} \omega \phi \\ &= \frac{1}{A} \sum_{\text{cloud}} \omega \phi = \sigma \frac{1}{a} \sum_{\text{cloud}} \omega \phi \end{aligned}$$

which is equivalent to eqn(5.8).

Rearranging the second term on the rhs of eqn(5.6), and expanding,

$$\begin{aligned} \overline{\omega' \phi'} &= (\sigma \overline{\omega \phi^c} + (1-\sigma) \overline{\omega \phi^e}) - (\overline{\omega^e \phi^e} + \sigma \overline{\phi^e} (\overline{\omega^c} - \overline{\omega^e}) + \sigma \overline{\omega^e} (\overline{\phi^c} - \overline{\phi^e}) \\ &\quad + \sigma^2 (\overline{\omega^c} - \overline{\omega^e}) (\overline{\phi^c} - \overline{\phi^e})) \end{aligned}$$

Now $\sigma \ll 1$, and $|\overline{\phi^c} - \overline{\phi^e}| \gg |\overline{\phi^c} - \overline{\phi^e}|$, so $\sigma^2 (\overline{\omega^c} - \overline{\omega^e}) (\overline{\phi^c} - \overline{\phi^e})$ can be neglected compared to $\sigma \overline{\phi^e} (\overline{\omega^c} - \overline{\omega^e})$. In a similar manner $\sigma \overline{\omega^e} (\overline{\phi^c} - \overline{\phi^e})$ is neglected compared to $\overline{\omega^e \phi^e}$. After further rearrangement, $\overline{\omega' \phi'}$ is given approximately by,

$$\overline{\omega' \phi'} = (\sigma \overline{\omega \phi^c} - \sigma \overline{\omega^e \phi^e}) + ((1-\sigma) \overline{\omega \phi^e} - (1-\sigma) \overline{\omega^e \phi^e}) \quad \text{eqn(5.10)}$$

Previous workers have written,

$$\overline{\omega \phi^e} = \overline{\omega^e \phi^e}$$

and so have neglected the second set of terms in eqn(5.10). The

neglect of any correlation between the vertical motion and the thermodynamic fields within the environment is a necessary one. Neither large-scale fields or cloud models give any information about the nature of this correlation. The neglect of this term would seem reasonable. $\overline{\omega^e}$ is much less than $\overline{\omega^c}$, and so any correlation within cloud environment would be expected to be much smaller than that within cloud air.

Hence neglecting the correlation within the environment, and using eqn(5.4),

$$\overline{\omega'\theta} = (\sigma\overline{\omega\theta^c} - \sigma\overline{\omega^c\theta}) \quad \text{eqn(5.11)}$$

This final form expresses $\overline{\omega'\theta}$ in terms of large-scale and cloud variables as required. Because of the neglect of correlated motions within the environment, eqn(5.11) implies that all vertical transports of θ and q occur within clouds.

A simpler method for obtaining eqn(5.11) is set out below.

$$\begin{aligned} \overline{\omega'\theta} &= \overline{\omega\theta} - \overline{\omega\theta} \\ &= (\sigma\overline{\omega\theta^c} + (1-\sigma)\overline{\omega\theta^e}) - (\sigma\overline{\omega^c\theta} + (1-\sigma)\overline{\omega^e\theta}) \\ &= (\sigma\overline{\omega\theta^c} - \sigma\overline{\omega^c\theta}) + ((1-\sigma)\overline{\omega\theta^e} - (1-\sigma)\overline{\omega^e\theta}) \end{aligned}$$

and again neglecting the correlation in the second set of terms yields,

$$\overline{\omega'\theta} = (\sigma\overline{\omega\theta^c} - \sigma\overline{\omega^c\theta})$$

This shorter method illustrates clearly that the neglect of the correlations within the environment is crucial to obtaining eqn(5.11)

Using this approximate expression for $\overline{\omega'\theta}$, $(Q1)_{\text{para}}$ and $(Q2)_{\text{para}}$ are written as,

$$\left. \begin{aligned} (Q1)_{\text{para}} &= \frac{L\overline{Q}}{c_p\pi} - \frac{\partial(\sigma\overline{\omega\theta^c} - \sigma\overline{\omega^c\theta})}{\partial p} \\ (Q2)_{\text{para}} &= -\overline{Q} - \frac{\partial(\sigma\overline{\omega q^e} - \sigma\overline{\omega^e q})}{\partial p} \end{aligned} \right\} \quad \text{eqn(5.12)}$$

5.1.3 Representation Of Moist Processes.

Having obtained expressions for eddy flux divergence in terms of large-scale and cloud variables, a similar process must be carried out for the terms representing moist processes on the rhs of eqn(5.12). This is carried out by considering budget equations of θ and q over cloudy air. Approximate equations, suitable for use with one dimensional numerical models have been used in the past (eg Arakawa and Schubert(1974)). These are described. However fuller equations are used to derive expressions for $(Q1)_{para}$ and $(Q2)_{para}$ for use with the meso-scale model.

(a) Budget equations over cloud air.

$(L\bar{Q}/c_p\pi)$ in eqn(4.12) can be written as the sum of contributions from cloud and environment:-

$$\frac{L\bar{Q}}{c_p\pi} = \sigma \frac{L\bar{Q}^c}{c_p\pi} + (1-\sigma) \frac{L\bar{Q}^e}{c_p\pi}$$

where $L\bar{Q}^c/c_p\pi$ is the heating(cooling) due to moist processes in cloud,
and $L\bar{Q}^e/c_p\pi$ is the equivalent term for the environment.

The second of these is usually associated with cooling due to rain and cloud water evaporation.

By considering the horizontal average of the theta equation over cloud, an expression for $L\bar{Q}^c/c_p\pi$ is obtained. Cloud area is generally a function of pressure and time ($a=a(p,t)$). Hence for terms involving $\partial/\partial t$, and $\partial/\partial p$, the rule for integrating over variable boundaries must be used:-

$$\frac{\partial}{\partial \alpha} \int_{a(\alpha)}^{b(\alpha)} f(x, \alpha) dx = \int_{a(\alpha)}^{b(\alpha)} \frac{\partial f}{\partial \alpha} dx + f(b, \alpha) \frac{db}{d\alpha} - f(a, \alpha) \frac{da}{d\alpha}$$

Considering the time derivative of θ , integrating over cloud and averaging over the entire domain,

$$\frac{1}{A} \int_{a(p,t)} \frac{\partial \theta}{\partial t} da = \frac{1}{A} \left[\frac{\partial}{\partial t} \int_{a(p,t)} \theta da - \theta_b \frac{\partial a}{\partial t} \right] = \frac{\partial \bar{\theta}}{\partial t} - \theta_b \frac{\partial \sigma}{\partial t}$$

where $\bar{\theta} = \frac{1}{a} \int_a \theta da$

and θ_b is the value of θ at cloud boundary.

Hence the cloud averaged theta equation is,

$$\frac{\partial \sigma \bar{\theta}^c}{\partial t} - \theta_b \frac{\partial \sigma}{\partial t} + \sigma \left(\frac{\partial u \theta}{\partial x} \right)^c + \frac{\partial \sigma \bar{\omega} \theta^c}{\partial p} - (\omega \theta)_b \frac{\partial \sigma}{\partial p} = \sigma \frac{L \bar{Q}^c}{c_p \pi} \quad \text{eqn(5.13)}$$

Similarly integrating over the mixing ratio equation yields,

$$\frac{\partial \sigma \bar{q}^c}{\partial t} - q_b \frac{\partial \sigma}{\partial t} + \sigma \left(\frac{\partial u q}{\partial x} \right)^c + \frac{\partial \sigma \bar{\omega} q^c}{\partial p} - (\omega q)_b \frac{\partial \sigma}{\partial p} = -\sigma \bar{Q}^c \quad \text{eqn(5.14)}$$

These expressions may be substituted into eqn(5.12) for Q. To simplify the resulting expressions for $(Q1)_{para}$, and $(Q2)_{para}$, the cloud averaged continuity equation is used. For an area $a(p,t)$, this is

$$\frac{1}{A} \int_a \frac{\partial u}{\partial x} da + \frac{1}{A} \int_a \frac{\partial \omega}{\partial p} da = \frac{1}{A} \frac{\partial a}{\partial t}$$

which gives,

$$\sigma \left(\frac{\partial u}{\partial x} \right)^c + \frac{\partial \sigma \bar{\omega}^c}{\partial p} - \omega_b \frac{\partial \sigma}{\partial p} = \frac{\partial \sigma}{\partial t} \quad \text{eqn(5.15)}$$

(b) Budget equations for use with analytical models.

Many parameterization schemes use analytical cloud models to specify cloud variables (eg Arakawa and Schubert(1974)). A spectrum of cloud types is assumed to exist. Such schemes use approximate forms of the area averaged equations derived in (a) above. The cloud models used are steady state in nature, and cloud area is constant with height. Thus eqns(5.13), (5.14), and (5.15) reduce to,

$$\sigma \left(\frac{\partial u \theta}{\partial x} \right)^c + \frac{\partial \sigma \bar{\omega} \theta^c}{\partial p} = \sigma \frac{L \bar{Q}^c}{c_p \pi}$$

$$\sigma \left(\frac{\partial u q}{\partial x} \right)^c + \frac{\partial \sigma \bar{\omega} q^c}{\partial p} = -\sigma \bar{Q}^c$$

$$\sigma \left(\frac{\partial u}{\partial x} \right)^c + \frac{\partial \sigma \bar{\omega}^c}{\partial p} = 0$$

One dimensional cloud models (the most common type used) do not contain information concerning horizontal cloud structure. Therefore the horizontal derivatives above cannot be determined. Inflow and outflow are assumed to occur through lateral entrainment and detrainment. Rates for these, E_i and D_i respectively for cloud type i are given by,

$$E_i = + \frac{\partial \sigma_i \bar{\omega}_i^c}{\partial p} \quad \text{if } \left(-\frac{\partial \sigma_i \bar{\omega}_i^c}{\partial p} \right) < 0$$

and

$$D_i = - \frac{\partial \sigma_i \bar{\omega}_i^c}{\partial p} \quad \text{if } \left(-\frac{\partial \sigma_i \bar{\omega}_i^c}{\partial p} \right) > 0$$

Entraining air is assumed to have the potential temperature and mixing ratio of the large-scale atmosphere, and replacing the horizontal derivatives by en/detrainment terms:-

$$\left. \begin{aligned} -E\bar{\theta} + \sum_{\text{detraining cloud}} D_i \bar{\theta}_i^c + \frac{\partial \sigma \bar{\omega} \bar{\theta}^c}{\partial p} &= \sigma \frac{L \bar{Q}^c}{c_p \pi} \\ -E\bar{q} + \sum_{\text{detraining cloud}} D_i \bar{q}_i^c + \frac{\partial \sigma \bar{\omega} \bar{q}^c}{\partial p} &= -\sigma \bar{Q}^c \\ E - D + \frac{\partial \sigma \bar{\omega}^c}{\partial p} &= 0 \end{aligned} \right\} \text{eqn(5.16)}$$

where $E = \sum_{\text{entraining cloud}} E_i$ and $D = \sum_{\text{detraining cloud}} D_i$.

(c) Forms of Q1 and Q2 for use with analytical cloud models.

Using the continuity equation in equation set (5.16), substituting for E in the complementary θ equation, and rearranging,

$$+ \sum_{\text{detraining cloud}} D_i (\bar{\theta}_i^c - \bar{\theta}) - \bar{\theta} \frac{\partial \sigma \bar{\omega}^c}{\partial p} + \frac{\partial \sigma \bar{\omega} \bar{\theta}^c}{\partial p} = \sigma \frac{L \bar{Q}^c}{c_p \pi}$$

Using this substitute for $\sigma L \bar{Q}^c / c_p \pi$ in the expression for $(Q1)_{\text{para}}$ in eqn(5.12), and after some rearrangement,

$$(Q1)_{\text{para}} = (1-\sigma) \frac{L \bar{Q}^c}{c_p \pi} + \sum_{\text{detraining cloud}} D_i (\bar{\theta}_i^c - \bar{\theta}) + \sigma \bar{\omega}^c \frac{\partial \bar{\theta}}{\partial p} \quad \text{eqn(5.17)}$$

Similarly for (Q2) in eqn(5.12),

$$(Q2)_{\text{para}} = -(1-\sigma) \bar{Q}^c + \sum_{\text{detraining cloud}} D_i (\bar{q}_i^c - \bar{q}) + \sigma \bar{\omega}^c \frac{\partial \bar{q}}{\partial p} \quad \text{eqn(5.18)}$$

Equations similar to these have been used to represent $(Q1)_{\text{conv}}$ and $(Q2)_{\text{conv}}$ in several parameterization schemes (Arakawa and Schubert(1974)), and also in several diagnostic studies (eg Yanai et al(1973)). Convection is seen to modify the atmosphere by:-

1. Compensating subsidence, $\sigma \bar{\omega}^c \partial \bar{\phi} / \partial p$, interpreted as the effect of motion within the environment, compensating cloud motions.
2. Detrainment, $\sum_{\text{detraining cloud}} D_i (\bar{\theta}_i^c - \bar{\theta})$.
3. Moist processes within the cloud environment, usually associated with the evaporation of liquid water within cloud environment. Some schemes write this as being proportional

to the detrainment of liquid water from cloud, ie

$$(1-\sigma) \frac{L\bar{Q}^e}{c_p\pi} = - \frac{L}{c_p\pi} \sum_{\text{detraining cloud}} D_i (\bar{T}_{ci}^e + \bar{T}_{ri}^e)$$

Diagnosis of observational data shows that subsidence and evaporation of liquid water are the dominant terms. Detrainment is important in the outflow layer of updraughts, and also at the surface, outflow occurring from downdraughts.

(d) Forms for Q1 and Q2 for use with numerical cloud models.

Using data from a numerical model, both time dependent effects, and horizontal cloud structure can be determined. Hence the approximations made in (c) above are not required. The derivation of this section retains such terms, and the resulting equations used to study parameterization in the proceeding chapter. Summations over all cloud air are carried out due to the difficulty of differentiating between different cloud types in a numerical model.

Following the method of section (c), substituting for $\sigma L\bar{Q}/c_p\pi$ in eqn(5.12) using eqn(5.13), and rearranging,

$$\begin{aligned} (Q1)_{para} = (1-\sigma) \frac{L\bar{Q}^e}{c_p\pi} &+ \frac{\partial \sigma \bar{\theta}^e}{\partial t} - \theta_b \frac{\partial \sigma}{\partial t} + \sigma \left(\bar{u} \frac{\partial \bar{\theta}}{\partial x} \right)^e + \sigma \left(\bar{\theta} \frac{\partial \bar{u}}{\partial x} \right)^e \\ &- (\omega \theta)_b \frac{\partial \sigma}{\partial p} + \sigma \bar{\omega}^e \frac{\partial \bar{\theta}}{\partial p} + \bar{\theta} \frac{\partial \sigma \bar{\omega}^e}{\partial p} \end{aligned}$$

Using eqn(5.15) to substitute for $\partial \sigma \bar{\omega}^e / \partial p$, after further rearrangement, the final form of $(Q1)_{para}$ is obtained:-

$$\begin{aligned} (Q1)_{para} = (1-\sigma) \frac{L\bar{Q}^e}{c_p\pi} &+ (\bar{\theta} - \theta_b) \frac{\partial \sigma}{\partial t} + \frac{\partial \sigma \bar{\theta}^e}{\partial t} + (\omega_b \bar{\theta} - (\omega \theta)_b) \frac{\partial \sigma}{\partial p} \\ &+ \sigma \left(\bar{u} \frac{\partial \bar{\theta}}{\partial x} \right)^e + \sigma \left(\frac{\partial \bar{u}}{\partial x} (\bar{\theta} - \bar{\theta}) \right)^e + \sigma \bar{\omega}^e \frac{\partial \bar{\theta}}{\partial p} \end{aligned} \quad \text{eqn(5.19)}$$

By considering the un-approximated budget equations over cloud air, several additional terms are introduced. $\partial \sigma \bar{\theta}^e / \partial t$ and $(\bar{\theta} - \theta_b) \partial \sigma / \partial t$ are due to the time dependent nature of the ensemble. $(\omega_b \bar{\theta} - (\omega \theta)_b) \partial \sigma / \partial p$ is due to the sloping of cloud boundary with height. Thus vertical as well as horizontal motions transport energy across cloud boundaries. The term is non-zero if the flux of θ through a sloping boundary is

not represented by the flux of $\bar{\theta}$. The detrainment term of eqn(5.17), and (5.18) are replaced by cloud advection, $\sigma \overline{(v \partial \theta / \partial x)^c}$ and an en/detrainment term, $\sigma \overline{(\partial v / \partial x (\theta - \bar{\theta}))^c}$. The advective term arises due to the retention of $\partial / \partial x$ terms. Terms representing moist processes within cloud environment and compensating subsidence are present as previously.

An equivalent expression for $(Q2)_{para}$ can be obtained by the same method:-

$$\begin{aligned} (Q2)_{para} = & -(1-\sigma)\bar{Q}^e + (\bar{q} - q_b) \frac{\partial \sigma}{\partial t} + \frac{\partial \sigma \bar{q}}{\partial t} + (\omega_b \bar{q} - (\omega q)_b) \frac{\partial \sigma}{\partial p} \\ & + \sigma \overline{\left(v \frac{\partial q}{\partial x}\right)^c} + \sigma \overline{\left(\frac{\partial v}{\partial x} (q - \bar{q})\right)^c} + \sigma \bar{\omega}^c \frac{\partial \bar{q}}{\partial p} \end{aligned} \quad eqn(5.20)$$

If the assumption concerning cloud area used in (c) above ie σ is constant wrt p and t , then eqns(5.19), and (5.20) reduce to,

$$\left. \begin{aligned} (Q1)_{para} &= (1-\sigma) \frac{L \bar{Q}^e}{C_p \pi} + \sigma \overline{\left(v \frac{\partial \theta}{\partial x}\right)^c} + \sigma \overline{\left(\frac{\partial v}{\partial x} (\theta - \bar{\theta})\right)^c} + \sigma \bar{\omega}^c \frac{\partial \bar{\theta}}{\partial p} \\ (Q2)_{para} &= -(1-\sigma) \bar{Q}^e + \sigma \overline{\left(v \frac{\partial q}{\partial x}\right)^c} + \sigma \overline{\left(\frac{\partial v}{\partial x} (q - \bar{q})\right)^c} + \sigma \bar{\omega}^c \frac{\partial \bar{q}}{\partial p} \end{aligned} \right\} \quad eqn(5.21)$$

These are equivalent to eqn(5.17), and (5.18).

5.2 A BRIEF DESCRIPTION OF CONVECTIVE PARAMETERIZATION SCHEMES

Several parameterization schemes for precipitating convection are in use. They broadly fall into three categories (1) Adjustment schemes, (2) Kuo type schemes, (3) Mass flux schemes. Most use the theory developed in section 5.1.

5.2.1 Adjustment Schemes.

This type of scheme does not make use of the theory developed previously. They assume that under the action of deep convection adjustment occurs back to some preferred atmospheric structure.

Two types exist;

(a) Moist convective adjustment schemes.

If the atmosphere is conditionally unstable ie $\partial\theta_e/\partial p > 0$, then the atmosphere is adjusted back to the mean θ_e . Two forms exist:-

1. Hard convective adjustment-the entire column is adjusted (eg Krishnamurti et al(1980)).
2. Soft adjustment:-a small area, a , of the column is adjusted (cloudy air), and then mixed with the environment air (eg Manabe et al(1965), Miyokoda et al(1969), Kurihara(1973)).

However the atmosphere is not observed to approach the mean θ_e under the action of convection. Hard adjustment causes excessive cooling and moistening of the lower atmosphere, the surface rainfall being an order of magnitude greater than that observed. Soft adjustment produces the correct magnitude of surface rainfall (Krishnamurti et al(1980)), but proceeds the observed rainfall by one to two days, due to maximum instability occurring before maximum convective activity. Because of this such schemes may not be useful for forecasting models, but may be adequate for climate models.

(b) Saturation point adjustment scheme.

The saturation point of a parcel is a quantity conserved under moist ascent (Betts(1982a)). From observational studies, under the action of convection, the atmosphere is seen to approach quasi-equilibrium structure well defined by the saturation point parameter (Betts(1983)).

In the scheme the atmospheric structure is adjusted to an equilibrium structure on a time scale of one to two hours(Betts and Miller(1984)). The adjusted atmospheric temperature and moisture profiles are derived from this. Work on the scheme is still in progress, but initial results are very favourable.

5.2.2 Kuo Type Schemes.

Large-scale moisture convergence and convection are observed to be linked. This observation was used by Kuo(1965,1974) to provide a constraint on the rainfall due to convection. Other schemes using this approach are Anthes(1977), Lian(1979).

If M_t is the moisture convergence into a column of the atmosphere (including boundary layer fluxes), then the fraction rained out by convection is,

$$R = (1-b)M_t$$

The remainder moistens the atmosphere.

This energy release is assumed to heat the atmosphere according to some distribution function $N(p)$,

$$N(p) = \frac{Q}{\langle Q \rangle}$$

Several forms have been tested. Kuo used,

$$N(p) = \frac{(T_c - \bar{T})}{\langle T_c - \bar{T} \rangle}$$

where T_c is specified by a moist adiabat.

Anthes(1977) used a one dimensional cloud model to specify Q . The two schemes are compared in Kuo and Anthes(1984c). They show good agreement with each other. Inclusion of eddy transports bring closer agreement with observation.

Several empirical formula have been used to calculate the atmospheric drying profile due to convection (Kuo(1974), Anthes(1977)). However a full moisture conservation equation, together with a one dimensional cloud model may be used (Kuo and Anthes(1984c)). The inclusion of eddy transports improves the estimate when compared to observation. However the method over estimates $(Q2)_{conv}$ above 800mb and under estimates below 800mb. It is thought that these discrepancies may be overcome by the inclusion of transports due to shallow convection.

A major problem with the scheme is the determination of b . Kuo(1965) used $b=0.8$ ie. most of the moisture convergence was used to moisten the atmosphere. Observationally $b=0.23-0.26$ (Kuo(1974), Krishnamurti et al(1980)). Krishnamurti also suggests that it is the vertical flux divergence component of M_t which is related to R , the horizontal component being stored within the atmosphere. This allows the storage of water vapour to be modelled well, allowing b to vary with time. This was not the case in the original scheme. Anthes(1977) suggested relating b to the mean relative humidity of the atmosphere with some success. A further problem is that the scheme is not valid after the mature stage of convection, when b is observed to be greater than 1.

5.2.3 Mass Flux Schemes.

These schemes are so called because they attempt to predict the vertical mass flux within convective clouds. Two types of this scheme exist.

(a) Arakawa-Schubert scheme(1974).

This is a more complex scheme, but is more general than the Kuo type scheme. It was developed for use with GCM's. The form of $(Q1)_{conv}$ and $(Q2)_{conv}$ are as presented in section(5.1.4), clouds effecting their environment through compensating subsidence and detrainment of saturated air. The scheme includes a coupled sub-cloud layer.

To link convection and large-scale processes, the vertical velocity at cloud base is determined as a function of the large-scale atmospheric structure. It is assumed that a quasi-steady state exists between the large-scale and cloud-scale processes ie. cumulus clouds react rapidly to the slower changes in large-scale processes. The generation of kinetic energy due to buoyancy forces (the cloud work function), defined as,

$$A = \int_{p_{cloud\ bottom}}^{p_{cloud\ top}} \left(\frac{\sigma \bar{\omega} \bar{\theta}_{vl}^c - \sigma \bar{\omega}^c \bar{\theta}_v}{\bar{\theta}_v} \right) \frac{dp}{\rho g}$$

is assumed to be slowly varying, when compared to the changes induced by the convection or large-scale processes. From such an assumption

it is possible to obtain a value for the cloud base vertical velocity, as a function of cloud type (see Arakawa and Schubert(1974) for full details). Because of the use of this quasi-equilibrium assumption, some workers have interpreted this scheme as an adjustment type scheme.

Lord and Arakawa(1980) verified the closure assumptions using observational data from the tropics. A semi-prognostic test was carried out by Lord(1982), quite good agreement being found between parameterized and observed $(Q1)_{conv}$ and $(Q2)_{conv}$. However the scheme cannot represent storage effects observed within the tropical atmosphere. The original scheme did not include any considerations of downdraught effects, although in some forecasting models which use the schemes, these have been included in a simple manner. A major disadvantage of the scheme is that it is very complicated and expensive to implement in forecasting models.

(b) CAPE schemes.

Both Kuo type and the Arakawa-Schubert schemes use quasi-equilibrium assumptions between the large-scale and cloud-scale processes. However it is observed that in some cases rainfall exceeds the moisture convergence. To overcome this problem several schemes have been developed in which cloud activity is based upon the local CAPE. Kreitzberg and Perkey(1976), and Fritsch and Chappell(1980a) developed schemes for use in meso-scale models. Both used one dimensional models (the latter scheme including downdraughts). The former used subsidence and en/detrainment to modify the atmosphere, while the latter mixed cloud and environment air on a time-scale of 30 to 60 minutes (based on the time required for the average wind to advect a cloud through a grid box).

A new scheme of this type has been developed by Miller and Moncrieff(1983), and tested at ECMWF. This used two dimensional analytical models of steady convection based on the work of Moncrieff and Green(1972). This approach has the advantage that the cloud models specify momentum as well as thermodynamic transports. Downdraughts are also well represented by these models. Although first results are promising, further work is required.

5.3 A NOTE ON AVERAGING TIME SCALES.

In chapter 4 domain averaged quantities were presented averaged over several hours, or tens of minutes. Section(4.2.1) discussed which of these were valid for the study of cloud parameterization in GCM's and coarse meso-scale models.

The area represented by a grid point of a GCM would include many more clouds than represented in the 2D simulation used in this study. Thus although the time step of such models is typically 20 minutes, the time dependent effects seen on this time-scale within the meso-scale are probably not important for GCM's. Their importance is further reduced by the fact that in many forecast models, the convection scheme is only called every hour (a typical lifetime of a convective cloud), and so only the effect of a cloud integrated over it's life span need be considered.

By averaging between one and nine hours of the simulation the results may be more representative of the effect of a larger ensemble of clouds on a shorter time-scale. Such a philosophy is assumed in the study of parameterization in this and the next chapter. In section(4.2.1) it was shown that the eight hour average was independent of the fine time structure of the simulation, and this gives some confirmation to the philosophy described above.

The twenty minute averages used in chapter 4 will also be studied in this and the following chapters. Such results are viewed as being useful when considering parameterization in coarse meso-scale models.

5.4 A DESCRIPTION OF THE "MEAN ENSEMBLE CLOUD".

5.4.1 The "mean Ensemble Cloud".

In section(5.1), $(Q1)_{conv}$ and $(Q2)_{conv}$ were written in terms of cloud variables. Hence study of cloud structure gives further understanding of ensemble transport properties, and also how convection may be parameterized. The accuracy of this diagnosis is dependent upon the accuracy of the cloud structure used. The meso-scale model provides a

good representation of convection, and so provides a suitable "cloud model" for such a study.

One dimensional cloud models have been previously used to diagnose the effect of convection upon the atmosphere (Simpson and Wiggert(1969), Yanai et al(1973), Johnson(1974)), and also to parameterize convection (Arakawa and Schubert(1974)). Although crude they have proved successful in predicting cloud variables. A major problem with such models is that downdraughts, and the processes by which they are formed are generally not represented. Thus the transport properties derived from such models should perhaps be viewed with some caution (see section(4.1.3)). This problem may be overcome in the two dimensional analytical models of Moncrieff and Green(1972), which give a good description of downdraughts.

Further problems arise with the rate at which lateral entrainment/detrainment occurs, and also with the cloud height spectrum. Entrainment/detrainment contributes to determining the maximum height of a convective cloud. Errors in this will produce errors in determining what cloud types exist for a given atmospheric structure. This is linked to the second problem. Few cloud height spectrums have been obtained observationally, and so the spectrums assumed in diagnosing $(Q1)_{conv}$ and $(Q2)_{conv}$ may be incorrect.

A numerical simulation of a cloud ensemble does not suffer from these problems. Downdraughts are well represented, and lateral transports across cloud boundary are implicit. A broad spectrum of cloud types also exists, which agrees quite well with that deduced from the GATE horizontal divergence profile.

Because of the time dependent nature of the simulation it is difficult to determine the structure of each cloud type. Hence in the following description the "mean ensemble cloud" is defined by averaging over all cloudy air on a pressure level. Updraught and downdraught structure are easily obtained. In explaining the structure of the simulated cloud, the averaging processes must be considered.

In parameterization it is the mean effect of an ensemble of clouds which is required, and so the mean ensemble cloud will prove useful in this context. It may be possible to use numerical models to "tune" analytical descriptions of cloud ensembles, and so improve their accuracy.

5.4.2 The Definition Of Cloud.

How is cloud defined when averaging over the meso-scale model? Parameterization is concerned with clouds which transport thermodynamic (and dynamic) quantities ("active cloud"). Hence thin "anvil" cloud need not be considered, as it appears to have few transport properties. It is also interesting to distinguish between updraught and downdraughts.

The first definition of active cloud used was;

UPDRAUGHT $w \gg 1.0\text{m/s}$ l_c or $l_r \gg 0.1\text{g/kg}$

DOWNDRAUGHT $w < 0\text{m/s}$ l_c or $l_r \gg 0.1\text{g/kg}$

Each model grid point was tested with this condition to specify up- and downdraught structure. The lower limit of 1m/s for updraught velocity was chosen by considering the average mass flux of upward moving air. Much cloud existed with w less than this, causing a rapid decrease in the mean vertical mass flux if included in updraught air. No such effect was observed for downdraught air, and so an upper limit of no vertical motion was chosen.

The above definitions caused much of the anvil cloud to be labelled downdraught. Convective downdraughts are associated with heavy rainfall, and water loading is felt to play an important role in driving such motion. Little rain water is observed in anvil cloud, and so the definition of downdraught was changed to,

DOWNDRAUGHT $w < 0\text{m/s}$ $l_r \gg 0.1\text{g/kg}$

This new definition gave satisfactory results. All cloudy air not labelled updraught or downdraught is termed "inactive cloud".

When integrating cloud structure with respect to time, the summation only includes those times when cloud is present. The exceptions to this are σ and ω , which are used in calculating the parametric form of $(Q1)_{conv}$ and $(Q2)_{conv}$. Hence values averaged over all eight hours are needed. As fig 5.1 (σ_t is the fractional time clouds are present) shows, clouds are present throughout much of the eight hour averaging period, and so the two types of average are little different.

5.4.3 Internal Cloud Structure-eight Hour Averages.

In this section the internal cloud structure is described, averaged between one and nine hours.

(a) Cloud area.

The variation of σ with height for updraughts (UD), downdraughts (DD), and inactive cloud (IC) is shown in fig 5.2. For UD σ peaks around 800mb (2.5%). The decrease above this is due to the fact that several cloud types exist within the domain. Thus UD detrainment occurs at different levels.

DD exists throughout the entire depth of the domain, σ_{bb} being approximately constant between 700 and 1000mb. However convective DD would seem to be restricted to the lowest 400mb of the atmosphere. Air designated DD in the upper domain is associated with cloud outflow regions (where gravity waves occur); this will become clearer as the cloud thermodynamic fields are considered.

The majority of cloudy air within the domain is inactive cloud. Anvil cloud is easily seen. The peak between 500 and 600mb corresponds to the outflow layer of medium level convection. The reduction above 800mb corresponds to the rapid reduction in the number of low level clouds above this.

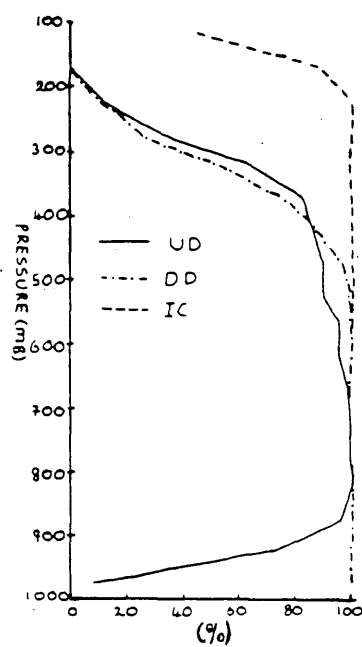


Fig 5.1 Fraction of time UD, DD, IC are present in the 1 to 9 hour averaging period (σ_T).

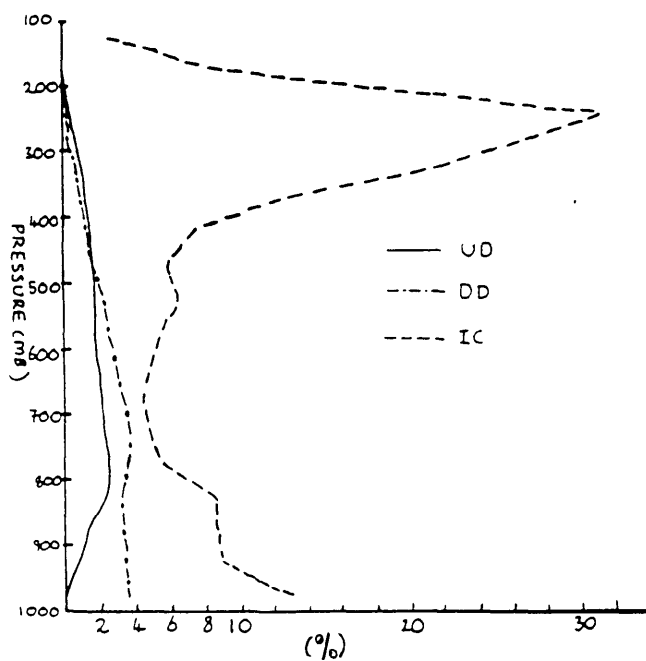


Fig 5.2 Fractional cloud area (σ) of UD, DD, IC (1-9 hours).

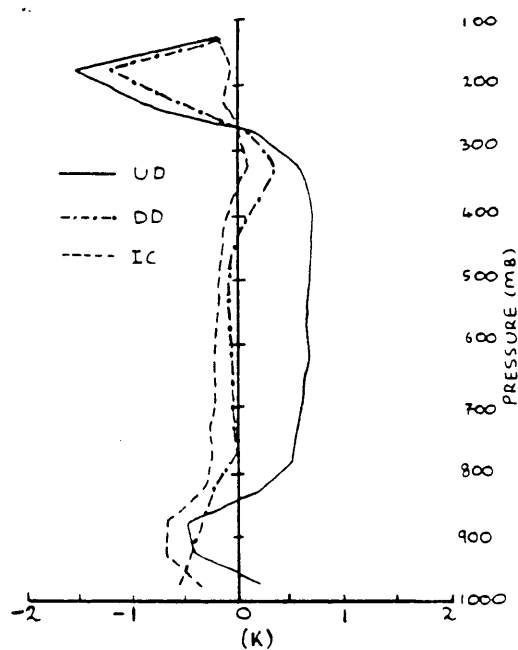
(b) Temperature and mixing ratio.

Fig 5.3 shows cloud temperature deviation from $\bar{\theta}$ for UD, DD, and IC. The UD structure is similar to that expected for deep convection. The lower levels are negatively buoyant. (The positive excess at 975mb is unrepresentative of cloud structure, as UD occurs at this level at only one grid point, and only for a few minutes of the simulation.) The profile rises through zero buoyancy around 650mb, agreeing with a parcel ascent from 975mb, with $\bar{\theta}$ and $\bar{q}$. Using a T- ϕ gram, a maximum temperature excess of 3K is expected, whereas only 1K is observed. However simple parcel ascent ignores lateral mixing, which occurs implicitly within the simulation. Averaging of time dependent effects will also reduce the maximum.

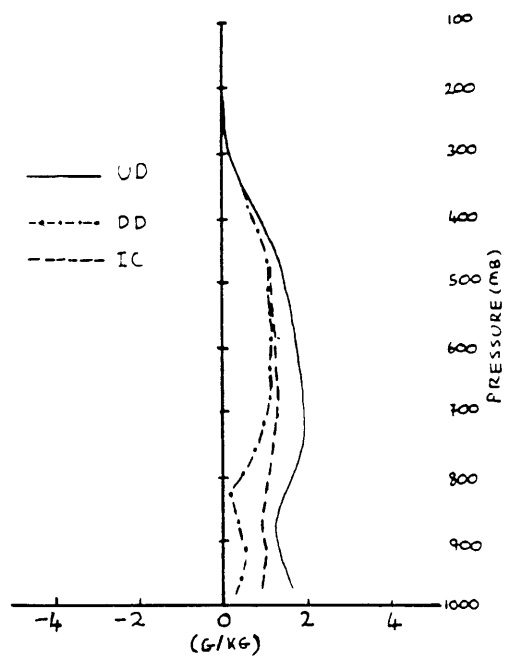
Cloud top zero buoyancy occurs at 250mb, again similar to a parcel ascent from 975mb. An overshoot of 100mb is seen. Here air is travelling upward, although it experiences negative buoyancy (deceleration). This leads to gravity wave formation in the cloud outflow region. Evidence for these is seen if the DD structure is studied. Above 250mb this is very similar to that of the UD. Mixing ratio structure is also similar (fig 5.4). Hence UD and DD in this region have a common origin, suggesting gravity wave oscillations.

As commented earlier, DD are much deeper than expected for convective DD. Air in the lowest 400mb is associated with these features, even though much of this is neutrally buoyant. Above this DD air is associated with the outflow air from various cloud populations. IC is colder than both UD and DD. In the lower layers, such air is associated with the large amount of low cloud, and inflow into UD. IC and DD have similar temperature and mixing ratio structures above 600mb, implying that this air has common origin, again indicating that DD air above this level is associated with cloud outflow layers.

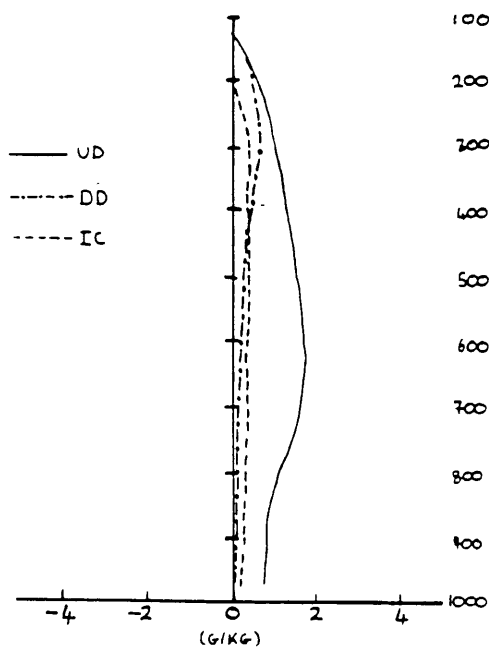
Although the averages presented are taken over many types of cloud, the temperature structure of UD and DD are very similar to those expected for deep convection. Hence deep convection within the model contributes a large signal to the average over all cloud.

Fig 5.3 $(\bar{T} - \bar{T})$ for UD, DD, IC

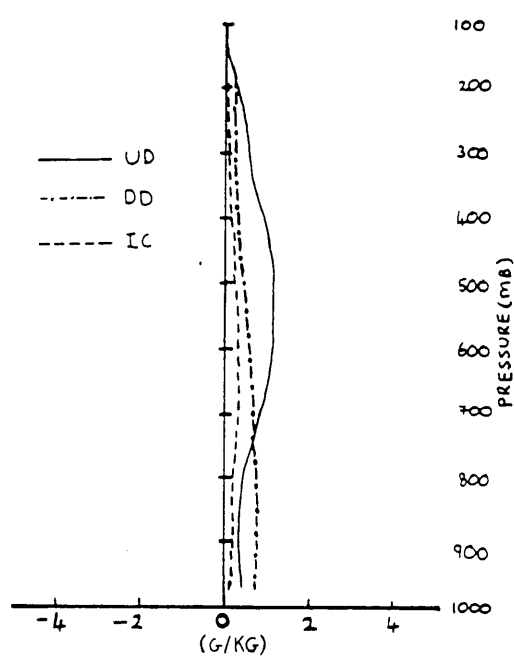
(1-9 hours).

Fig 5.4 $(\bar{q}^e - \bar{q})$ for UD, DD, IC

(1-9 hours).

Fig 5.5 $\bar{L}_c$ for UD, DD, IC

(1-9 hours).

Fig 5.6 $\bar{L}_r$ for UD, DD, IC

(1-9 hours).

(c) Cloud and rain water.

These are shown in fig 5.5 and 5.6 respectively. UD contains more liquid water than DD and IC, DD being hardly saturated in the region where convective DD occur. IC is also very tenuous.

Maximum rainwater content of the UD occurs at 500mb. $\bar{l}_r^{vp}$ follows a similar shape to $\bar{l}_c^{vp}$, due to the autoconversion process being proportional to the latter. Once formed rainwater falls, and so an increase in rainwater is expected towards the surface. Such a feature is seen in DD, not UD. $\bar{l}_c^{vp}$ is too small for autoconversion to occur, and so this increase is due to rain falling from UD.

(d) Buoyancy.

Convection is driven by buoyancy forces. In the model vertical momentum equation, considering virtual potential temperature and water loading, this is defined as,

$$g \left(\frac{\theta_v - \theta_{iv}}{\theta_{iv}} - l_c - l_r \right)$$

where $\theta_v = \theta(1 + 0.61q)$

and $\theta_{iv} = \theta_i(1 + 0.61q_i)$

Water loading may be incorporated into the temperature term by defining the liquid water potential temperature,

$$\theta_{vl} = \theta(1 + 0.61q - l_c - l_r)$$

the buoyancy term then being written with little error as,

$$g \left(\frac{\theta_{vl} - \theta_{iv}}{\theta_{iv}} \right)$$

(assuming the initial state contains no liquid water).

Above, deviations of potential temperature are from the initial profile of the model. It would seem more reasonable to define buoyancy with respect to the cloud environment temperature $\bar{\theta}$. Are the model equations consistent with such a definition?

Consider the buoyancy and pressure terms of the vertical momentum equation (water loading and virtual effects are neglected; these can be added later),

$$-g\left(\frac{\theta - \theta_r}{\theta_r}\right) + g^2 p_r \frac{\partial(h - h_r)}{\partial p}$$

Now $\phi - \phi_r = \bar{\phi} - \phi_r + \phi'$, and so these terms can be expanded to,

$$g\left(\frac{\bar{\theta} - \theta_r}{\theta_r}\right) + g\frac{\theta'}{\theta_r} + g^2 p_r \frac{\partial(\bar{h} - h_r)}{\partial p} + g^2 p_r \frac{\partial h'}{\partial p} \quad \text{eqn(5.22)}$$

Assuming that the domain averages satisfy hydrostatic equilibrium, ie.

$$\frac{\partial \bar{h}}{\partial p} = -\frac{1}{\bar{p}g}$$

then,

$$\frac{\partial h_r}{\partial p} + \frac{\partial(\bar{h} - h_r)}{\partial p} = -\frac{1}{g p_r (1 + (\bar{p} - p_r)/p_r)} = -\frac{1}{g p_r} \left(1 + \frac{\bar{\theta} - \theta_r}{\theta_r}\right)$$

The initial model state is hydrostatic and so,

$$\frac{\partial(\bar{h} - h_r)}{\partial p} = -\frac{1}{g p_r} \left(\frac{\bar{\theta} - \theta_r}{\theta_r}\right)$$

Substituting into eqn(5.22) gives,

$$g\frac{\theta'}{\theta_r} + g^2 p_r \frac{\partial h'}{\partial p}$$

Deviations are now from the domain mean state. Replacing θ_r by $\bar{\theta}$ in the temperature term introduces little error ($\lesssim 1\%$). So finally the buoyancy and pressure terms are written as,

$$g\left(\frac{\theta - \bar{\theta}}{\bar{\theta}}\right) + g^2 p_r \frac{\partial(h - \bar{h})}{\partial p}$$

The contribution to mean cloud buoyancy of $(\bar{\theta}' - \bar{\theta})$, and water loading is seen in fig 5.7a and b. Net buoyancy is shown in fig 5.7c.

The curves of fig 5.7a, and 5.7b reflect the temperature excess and liquid water content of cloud as expected. Maximum UD buoyancy is 0.03ms^{-2} . Much of the DD has slight positive buoyancy also. Inclusion of water loading has important effects, DD now being negatively buoyant throughout much of the domain (upto 400mb). UD net buoyancy is much reduced. Can such a buoyancy drive the mean convective motions within the simulation.

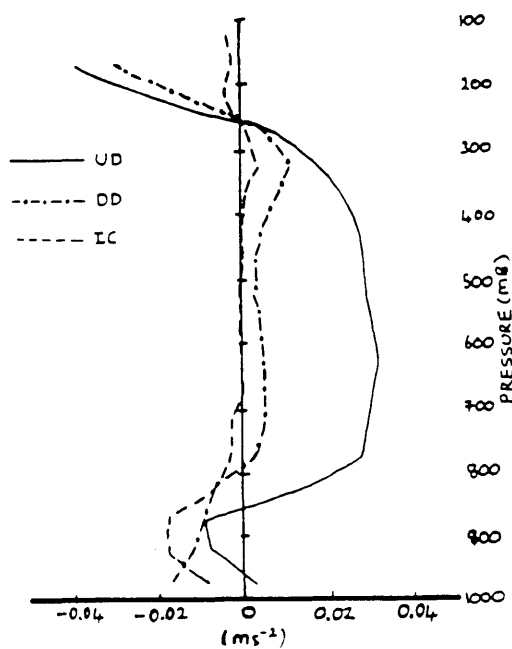


Fig 5.7(a) $g\bar{Q}'/\bar{Q}$ for
UD, DD, IC (1-9 hours).

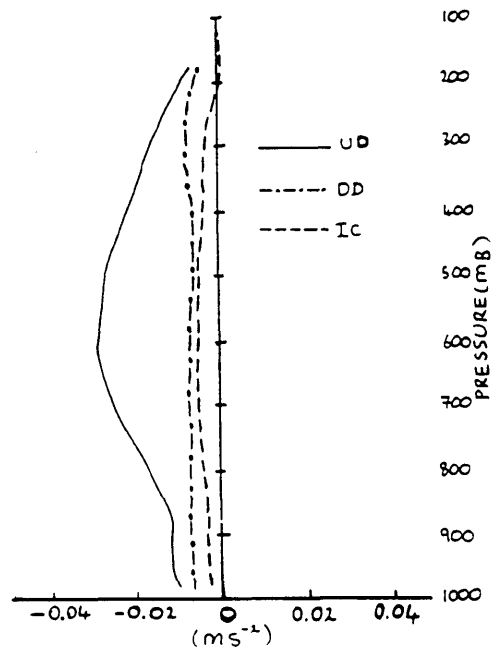


Fig 5.7(b) Water loading for
UD, DD, IC (1-9 hours).

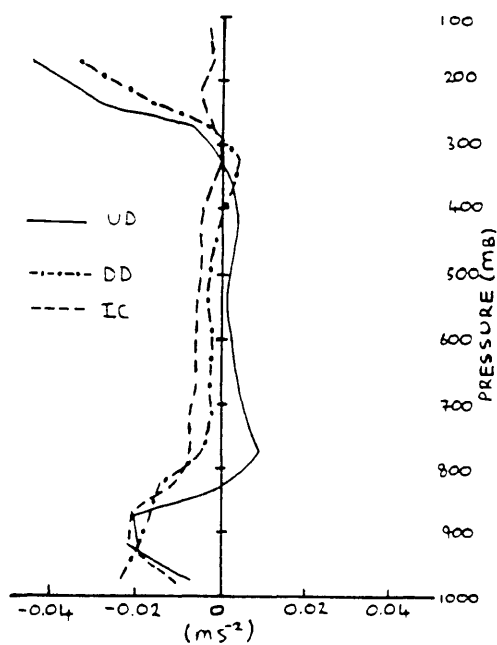


Fig 5.7(c) Net buoyancy for
UD, DD, IC (1-9 hours).

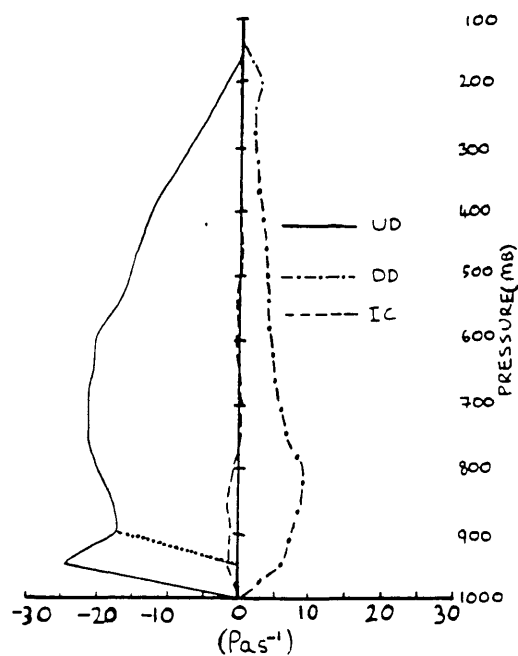


Fig 5.8 $\bar{\omega}$ for UD, DD, IC
(1-9 hours).

If the buoyancy of a parcel is integrated vertically, then the convective available potential energy (CAPE) is obtained. Assuming all this is converted into kinetic energy (simple parcel theory) then,

$$\frac{1}{2} \overline{w^c}^2 \approx \text{CAPE}$$

where $\overline{w^c}$ is the representative UD velocity.

Taking a parcel from the surface, lifting it through saturation to zero buoyancy at the top of the atmosphere, and assuming all condensed water falls out, gives a CAPE of around 1000J/kg for the tropical atmosphere, and so $\overline{w^c} \approx 40\text{ms}^{-1}$, far too large. However this value of CAPE neglects the entrainment of dry environmental air as the parcel ascends, which reduces buoyancy.

In the numerical model such a process occurs implicitly, and so values of CAPE estimated from the simulated clouds should be more realistic. From fig 5.7a, for UD air, positive buoyancy occurs over 600mb with a typical value of 0.02ms^{-2} , giving a CAPE of $\approx 120\text{J/kg}$, and a vertical velocity of $\approx 15\text{ms}^{-1}$. This is reduced further by the addition of water loading, buoyancy decreasing four fold, yielding a CAPE of $\approx 30\text{J/kg}$, and $w \approx 5\text{ms}^{-1}$. This value is of similar magnitude to the mean vertical velocity of the simulated UD, $\approx 3\text{ms}^{-1}$. Hence such low buoyancy can drive the mean cloud circulations from the simulation.

This simple calculation illustrates that water loading is important in specifying the buoyancy of cloud air. Simple analytical cloud models use some form of CAPE to calculate $\overline{w^c}$, and consideration of water loading in these is paramount.

(e) Cloud vertical velocity.

$\overline{w^c}$ is presented in fig 5.8. For UD the value at 975mb is unrepresentative of the mean cloud for the reasons given in (a) above. At 900mb $\overline{w^u}$ is 17Pa s^{-1} ($\approx 2\text{ms}^{-1}$), with only a slight increase occurring to 600mb. Above this the value decreases. This is not so for $\overline{w^u}$, as density decreases with height. $\overline{w^u}$ has a value of $\approx 2.5\text{ms}^{-1}$ between 700 and 400mb.

Maximum $\overline{\omega}^{ps}$ is lower than that of UD, as expected. Significant vertical velocities exist in DD air above 500mb. $\overline{\omega}^{ic}$ is near zero except in the lowest 200mb where the motion is upward.

Appearing in the parametric forms of $(Q1)_{conv}$ and $(Q2)_{conv}$ is the domain averaged mass flux, $\sigma \overline{\omega}^c$. This is presented in fig 5.9. Being modulated by the profile, the shape differs from that of $\overline{\omega}^c$. $\sigma \overline{\omega}^{up}$ increases rapidly up to 800mb, and decreases (almost linearly) to 200mb. Above 800mb DD mass flux is approximately 50% that of the UD, while below this it exceeds the UD value. It is seen that inactive cloud becomes important in the domain mass flux budget, especially in the lowest 200mb. Here fig 5.2 shows that σ_{xc} is much larger than UD and DD area. In the lowest 100mb $\sigma_{xc} \overline{\omega}^{ic}$ approximately balances the DD mass flux, the net cloud mass flux being close to zero (fig 5.10).

(f) En/detrainment rate.

Examining the profile of $-\partial \sigma \overline{\omega} / \partial p$ (fig 5.11), regions of cloud inflow and outflow (en/detrainment) may be identified.

UD inflow occurs between 975 and 750mb, reaching a maximum at 825mb. The broad depth of the inflow air implies that much low CAPE air is entrained into convective towers. The detrainment layer extends to 200mb, the rate at which detrainment occurs being constant throughout much of the region.

DD entrains air between 400 and 800mb, maximum detrainment occurring at the surface. Inactive cloud has a complex structure. Entrainment occurs at 975mb, and peaks of detrainment at 900, 800, 500 and 275mb. These are associated with outflow layers of different cloud types described in chapter 4 (section(4.1.2)).

5.4.4 Comparative Study Of A Single Deep Convective Cloud.

When studying the temperature structure of the mean ensemble cloud (MEC), it was commented that the structure was similar to that

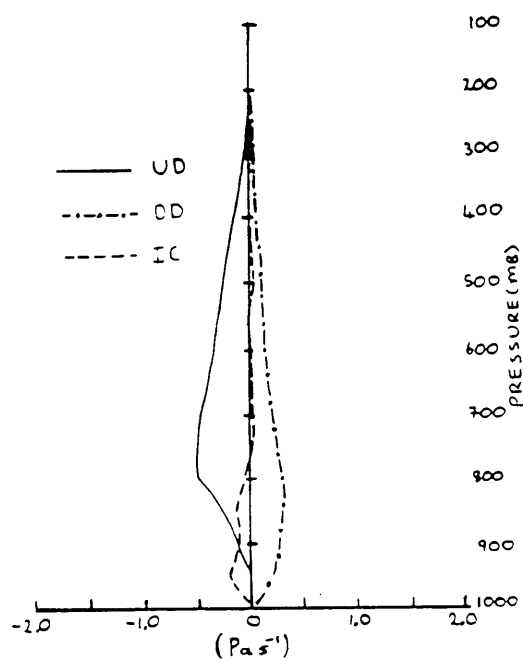


Fig 5.9 $\sigma \bar{w}$ for UD, DD, IC

(1-9 hours).

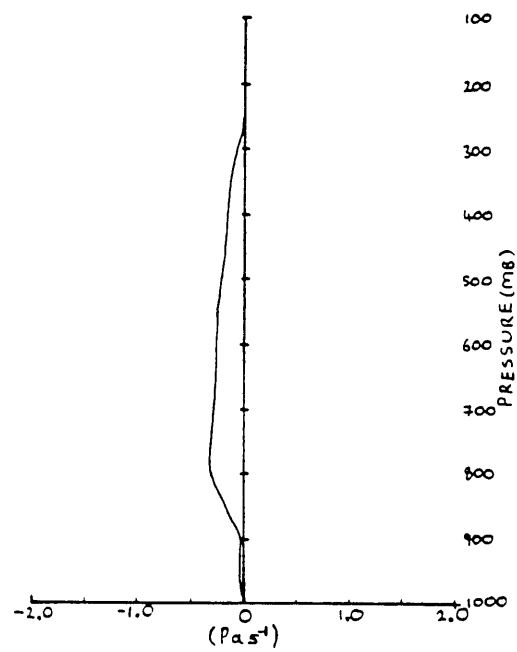


Fig 5.10 $\sigma \bar{w}$ for all cloudy air

(1-9 hours).

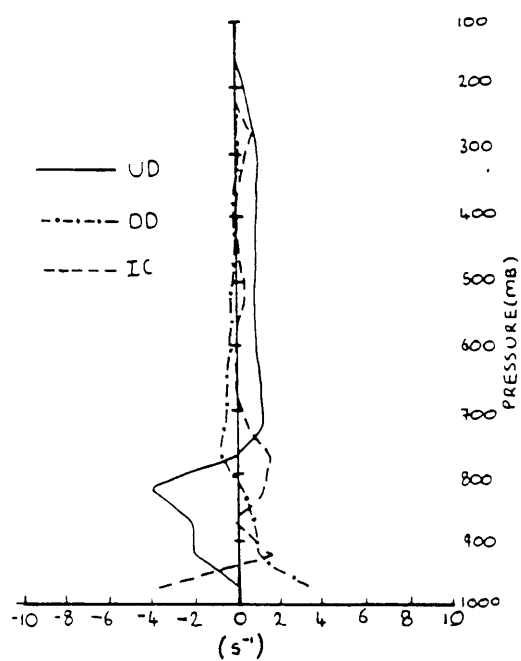


Fig 5.11 $-\partial \sigma \bar{w} / \partial p$ for UD, DD, IC (1-9 hours).

expected of deep convection. In this section the structure of a single deep convective cloud is studied for comparative purposes. The cloud developed at 270 minutes, and the average is taken over the next 80 minutes, into the decay phase. The storm produced 80mm of rain over 4km (maximum over 1km $\approx$ 30mm). Buoyancy terms and vertical velocity are presented.

(a) Updraught.

Buoyancy and water loading (fig 5.12a and b) are similar to those of the MEC, although magnitudes are greater. Net buoyancy (fig 5.12c) is negative in mid-levels, whereas only a reduction is observed in the MEC. This is a feature of this storm. Medium height cloud could not contribute significantly to buoyancy in this region as it is near their cloud top. $\bar{\omega}^p$ (fig 5.13) is more peaked than MEC below 500mb, but similar above. The difference is due to the effect of medium height cloud ($\bar{\omega}^c$ for such a cloud is also plotted in fig 5.13). $\sigma_{\omega} \bar{\omega}^p$ (fig 5.14) show significant differences to MEC, being flatter. This is caused by differences in the σ_{ω} profile (fig 5.15), maximum UD extent being at 500mb, instead of 800mb as in MEC.

(b) Downdraught.

The profiles are little different to those of MEC. Buoyancy (fig 5.12a) show no cold air at the surface, although inclusion of water loading gives negatively buoyant DD air. Study of the unaveraged fields shows that convective downdraughts, associated with large amounts of rain occur in the lowest 400mb. DD motions above this occur during the decay phase, and little rain is present.

(c) Inactive cloud.

$\bar{\omega}^c$ for IC occurs mainly in the lowest 250-300mb of the domain, and has a similar magnitude to that seen in the eight hour averages earlier (fig 5.13). Hence it would appear that much of the IC within the domain is associated with deep convection, and not with the large number of shallow clouds observed within the domain.

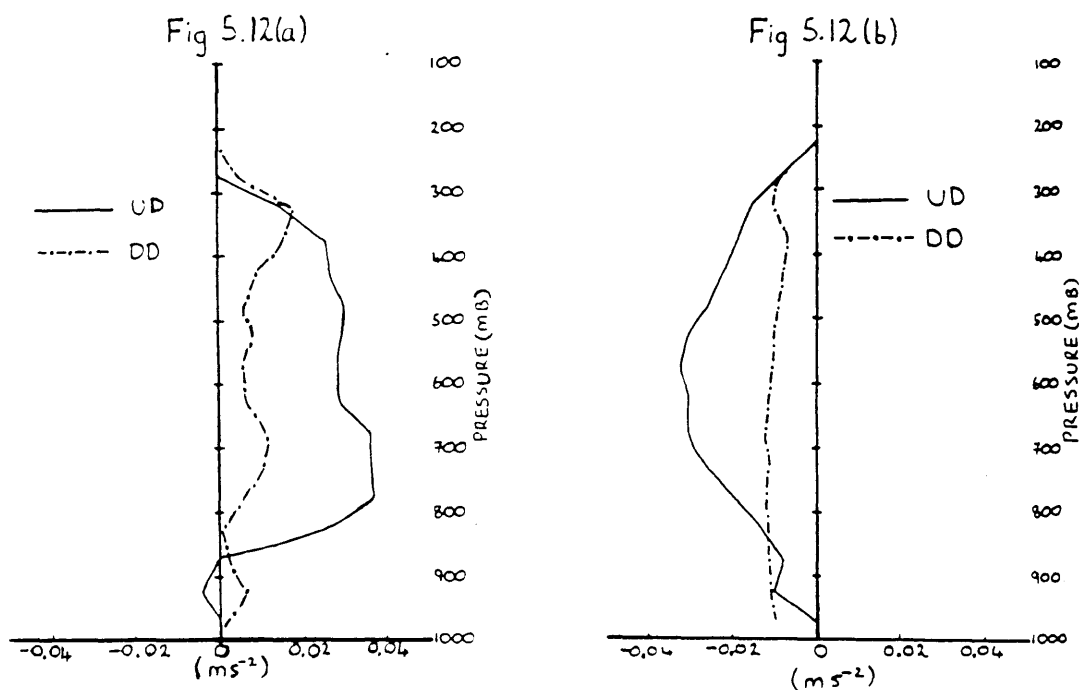


Fig 5.12(a) $g\bar{\theta}'/\bar{\theta}$ for UD and DD of a single deep convective cloud, averaged between 280 and 360 minutes.

Fig 5.12(b) Water loading for UD and DD of a single deep convective cloud, averaged between 280 and 360 minutes.

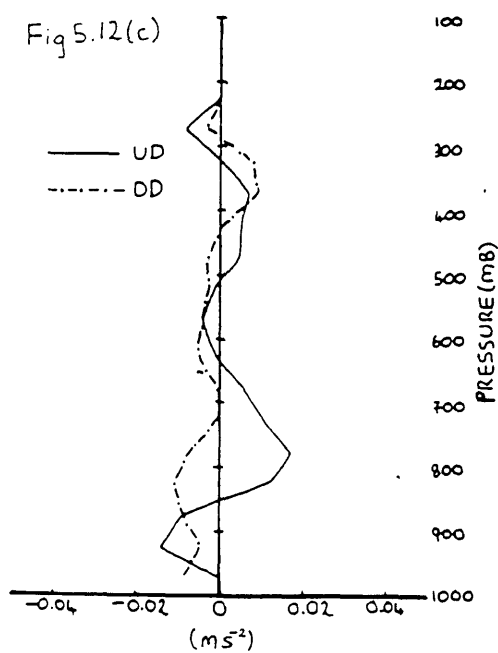


Fig 5.12(c) Net buoyancy for UD and DD of a single deep convective cloud, averaged between 280 and 360 minutes.

Fig 5.13 $\bar{\omega}^c$ for UD, DD, IC of a single deep convective cloud, averaged between 280 and 360 minutes (plus $\bar{\omega}^{ub}$ for a medium height cloud).

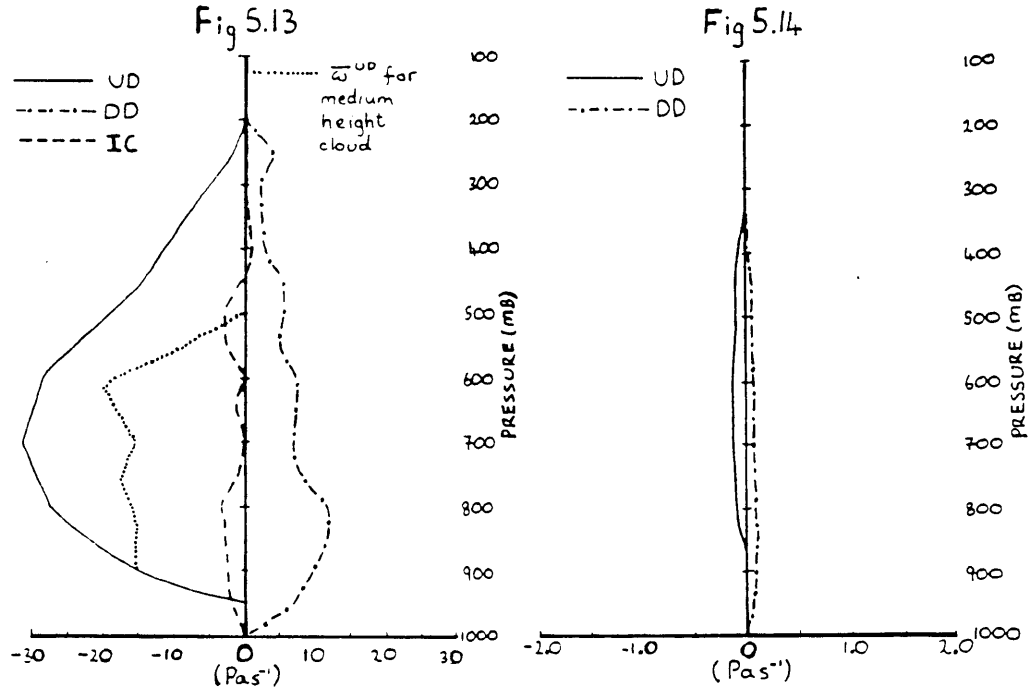


Fig 5.14 $\sigma \bar{\omega}^c$ for UD and DD of a single deep convective cloud, averaged between 280 and 360 minutes.

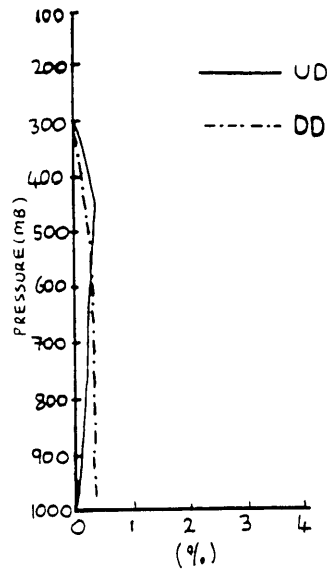


Fig 5.15 Schematic of UD and DD fractional area (σ) for a single deep convective cloud, averaged between 280 and 360 minutes.

Further evidence for this is seen if the compensating subsidence for UD and DD are studied for the whole domain, using twenty minute averages (fig 5.16). Although IC exists throughout the simulation, subsidence associated with IC is a maximum when deep convection occurs. Hence the association of IC with inflow into deep convective UD appears correct.

The above comparison indicates that the MEC cloud contains a large signal due to deep convective clouds. Smaller clouds are important in determining the mean UD structure (especially $\bar{\omega}^{up}$ and $\sigma_p \bar{\omega}^{up}$), but less so for DD. Thus the incorporation of a spectrum of cloud types within parameterization would seem important.

5.4.5 Internal Structure Of Convection-20 Minute Average.

In chapter 4, 20 minute averages of ensemble energy transports and sources were studied in relation to parameterization in meso-scale models. These cloud properties were seen to evolve with time, a typical example occurring between 280 and 380 minutes (fig 4.17). Two clouds contribute, both having similar vertical structure and time evolution. One of them is the cloud presented in the previous section, and this is again presented here. All times are measured relative to 280 minutes.

Fig 5.17 show $L\bar{Q}/c_p$ for the cloud over four consecutive 20 minute periods. As the cloud forms heating occurs in the lowest 400mb of the domain, the depth increasing with time. During 40-60 minutes, DD cooling occurs at the surface. This deepens between 60 and 80 minutes, heating only occurring in the upper half of the domain. The UD and DD structure associated with this evolution are described below.

(a) Updraught.

$(\bar{T}^{up} - \bar{T})$ (fig 5.18) has a high degree of correlation with condensational heating, as the cloud grows. When DD cooling develops, UD temperature decays and UD base moves away from the surface, implying that DD now dominates the lower domain (60-80 minutes).

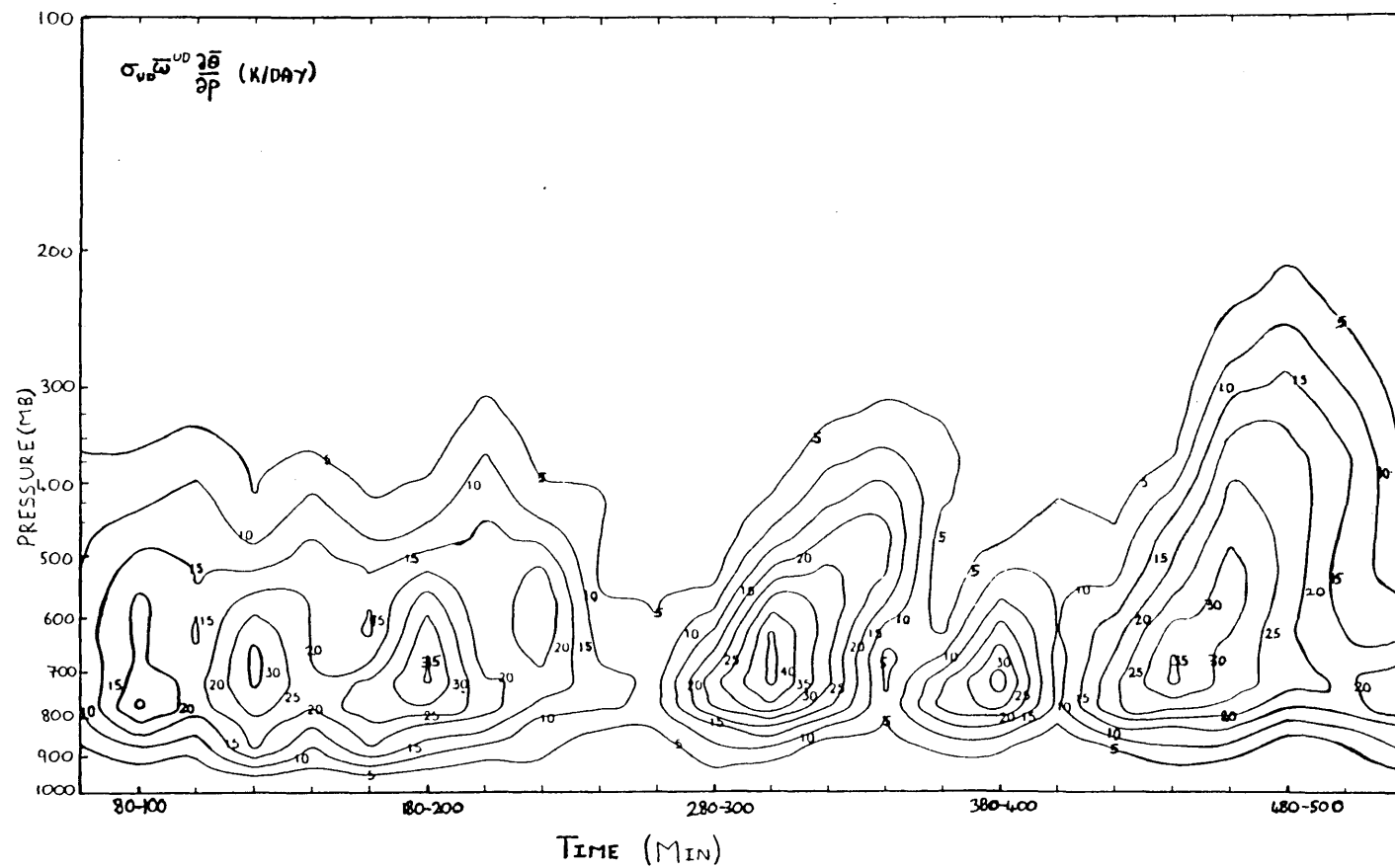


Fig 5.16 Comparison of compensating subsidence due to
UD and IC (20 minute averages).
(contour every 5K/day)

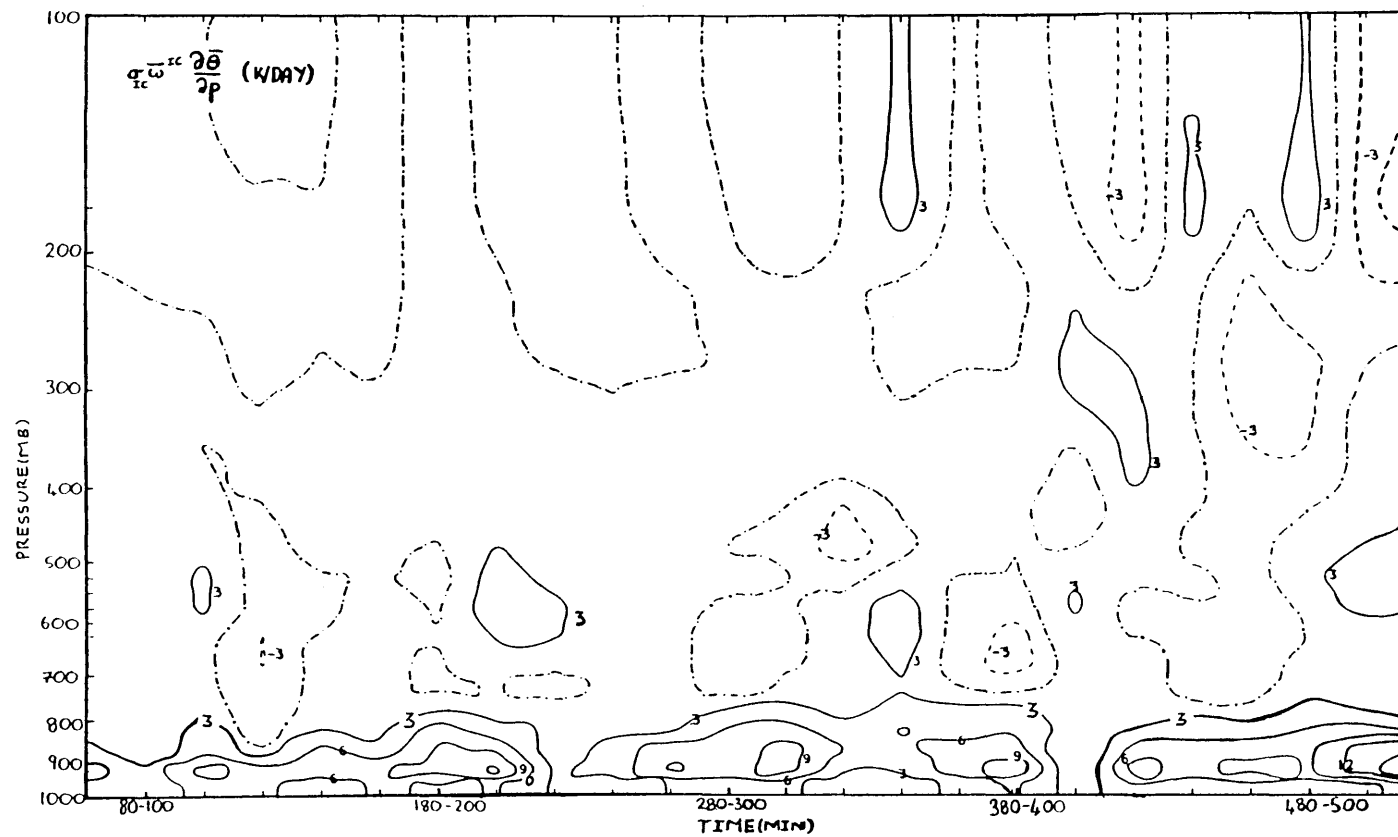


Fig 5.16 cont
 (contour every 5K/day)

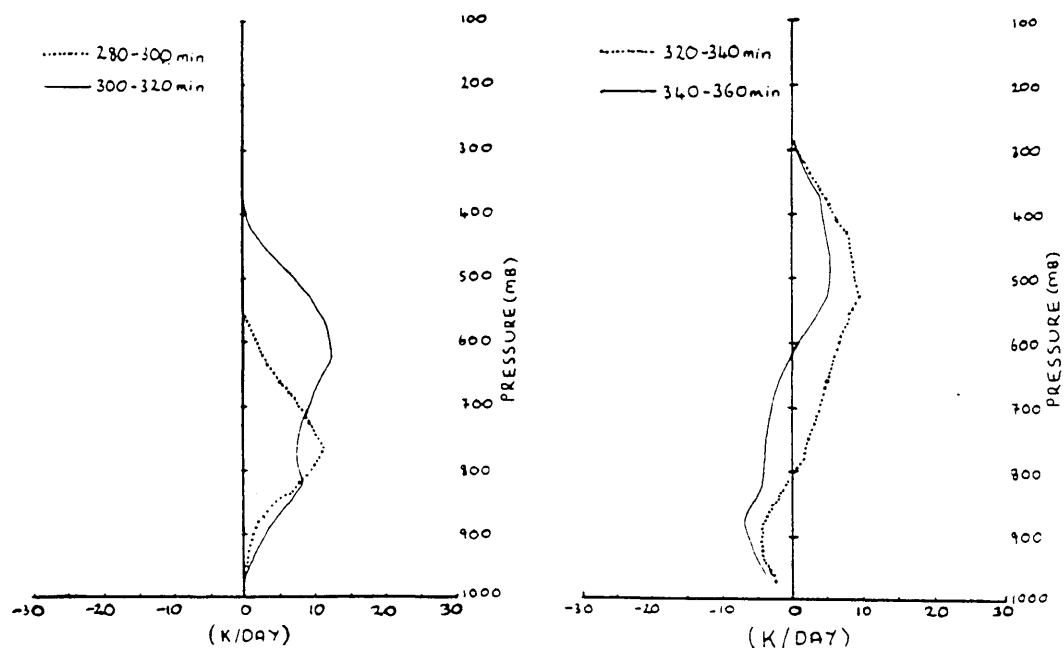


Fig 5.17 $\bar{LQ}/c_p$ for a single deep cloud (20 minute averages).

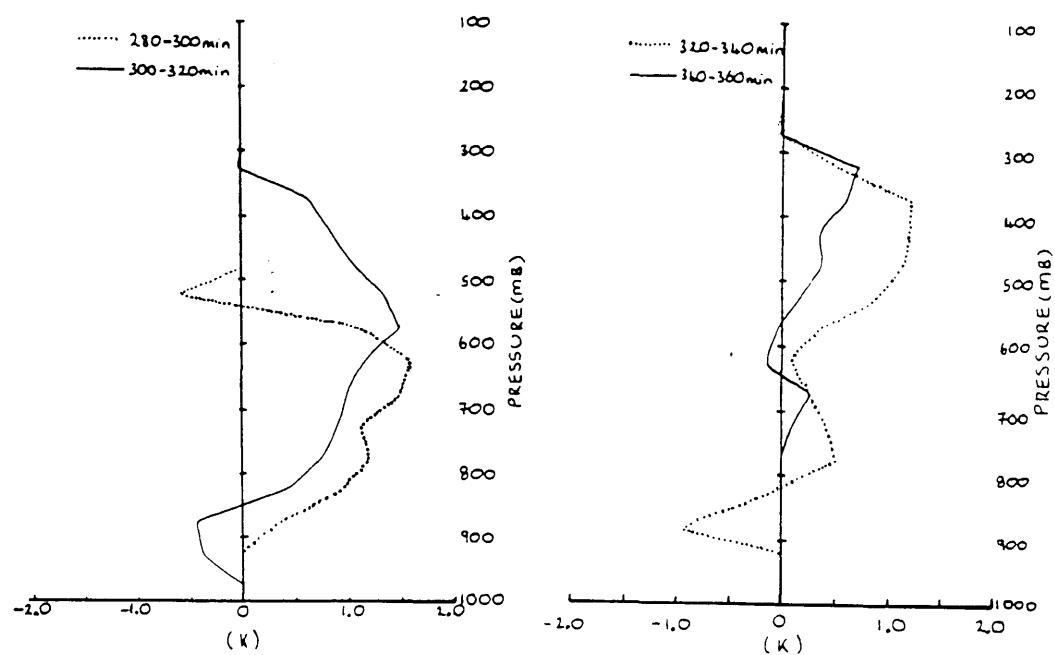


Fig 5.18 $\overline{(T-T)}$ for a single deep cloud (20 minute averages).

Buoyancy (fig 5.19) follows the temperature profile. Water loading only becomes important after 20 minutes. The structure of the mature phase is similar to that of the MEC. $\bar{\omega}^*$ (fig 5.20) peaks at maximum temperature excess for the first 40 minutes. As the cloud decays the profile becomes flatter. $\sigma \bar{\omega}^*$ (fig 5.21) exhibits a different evolution, due to the level of maximum UD evolving upward during cloud lifetime.

(b) Downdraught.

DD forms between 20 and 40 minutes, coinciding with the growth of liquid water in the cloud. DD air is positively buoyant during this period (fig 5.22), only experiencing a downward acceleration in the lowest 300mb when water loading is considered. The depth of this layer increases, although in a complex manner. Air below 600mb is associated with convective scale DD, being nearly unsaturated but containing significant amount of rainwater (between 1 and 2g/kg). DD air above this contains much more cloud water. Several processes contribute to defining this air DD. Between 20 and 40 minutes, "vortex rings" of descending air occur next to the UD (fig 5.23). These contain both rain and cloud water, and decay with updraught, indicated by the reduction of $\bar{\omega}^*$ (fig 5.24) above 600mb between 20-40, and 40-60 minutes. During 40-80 minutes gravity waves form in the centre of the UD as it decays, and these also contribute to DD above 600mb. Below 700mb $\bar{\omega}^*$ is roughly constant with time, and similar in structure to MEC. Thus once formed convective DD change little with time.

Hence the change in cloud transport properties is linked to changes in the internal structure of UD and the formation of DD. Both are associated with the increase of liquid water within cloud.

5.4.6 Brief Comparison Of Simulated Cloud And 1D Cloud Model.

In this section comparison is made between the simulated cloud, presented in the previous two sections, and a typical one dimensional cloud model, taken from Anthes(1977). It contains consideration of liquid water in the buoyancy term (using a Kessler parameterization

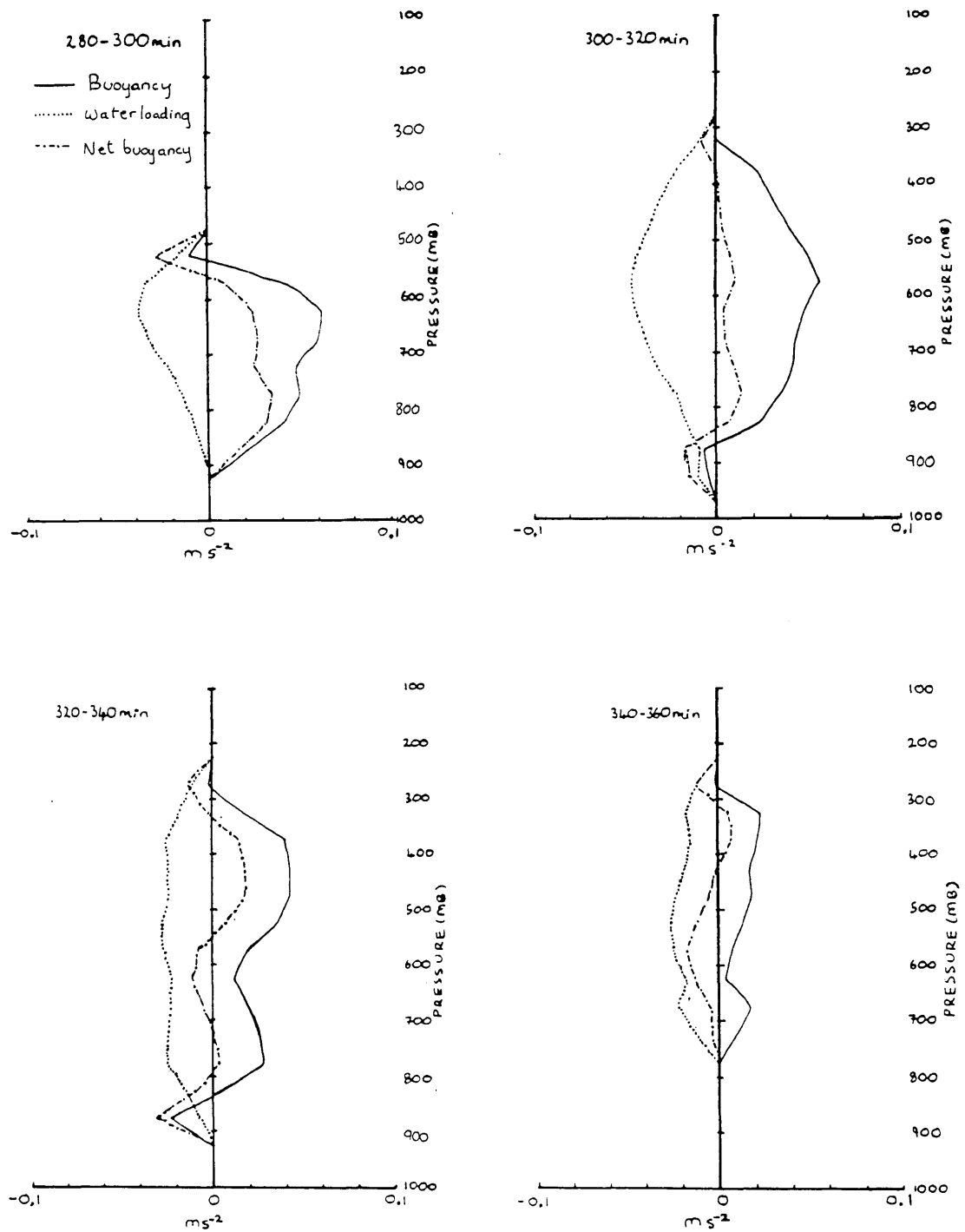


Fig 5.19 Buoyancy terms for UD of a single deep cloud (20 minute averages).

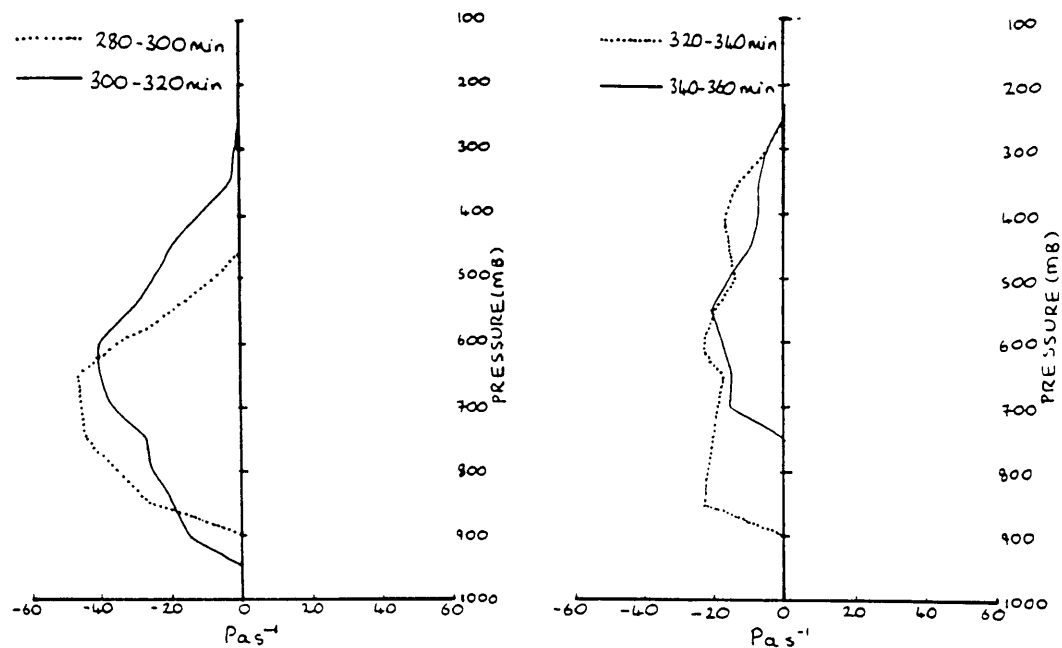


Fig 5.20 $\bar{w}^{uo}$ for a single deep cloud (20 minute averages).

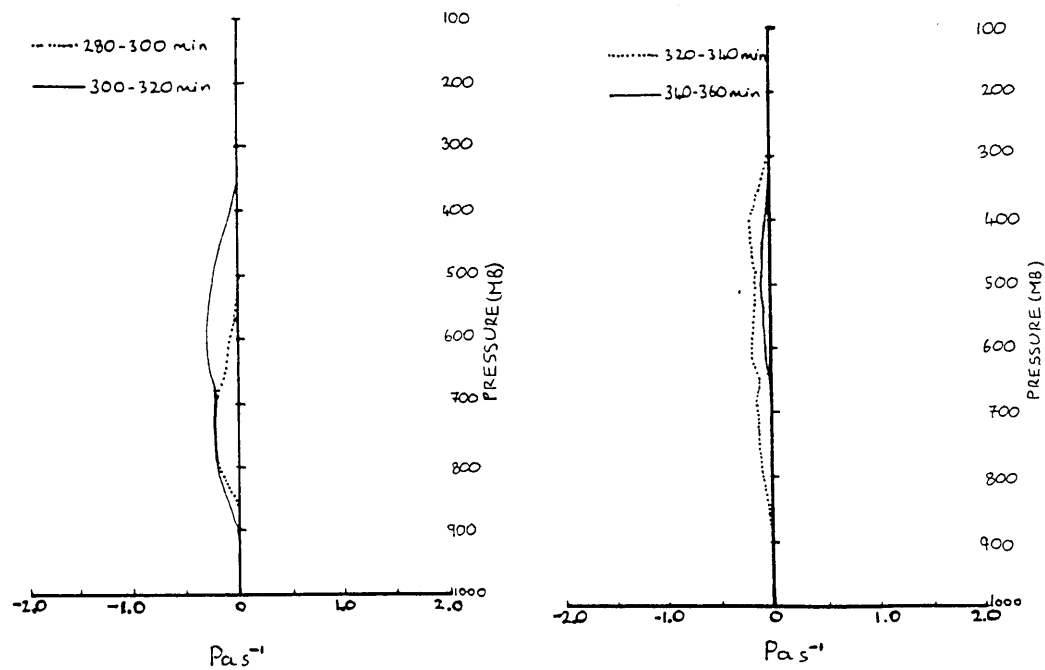


Fig 5.21 $\sigma \bar{w}^{uo}$ for a single deep cloud (20 minute averages).

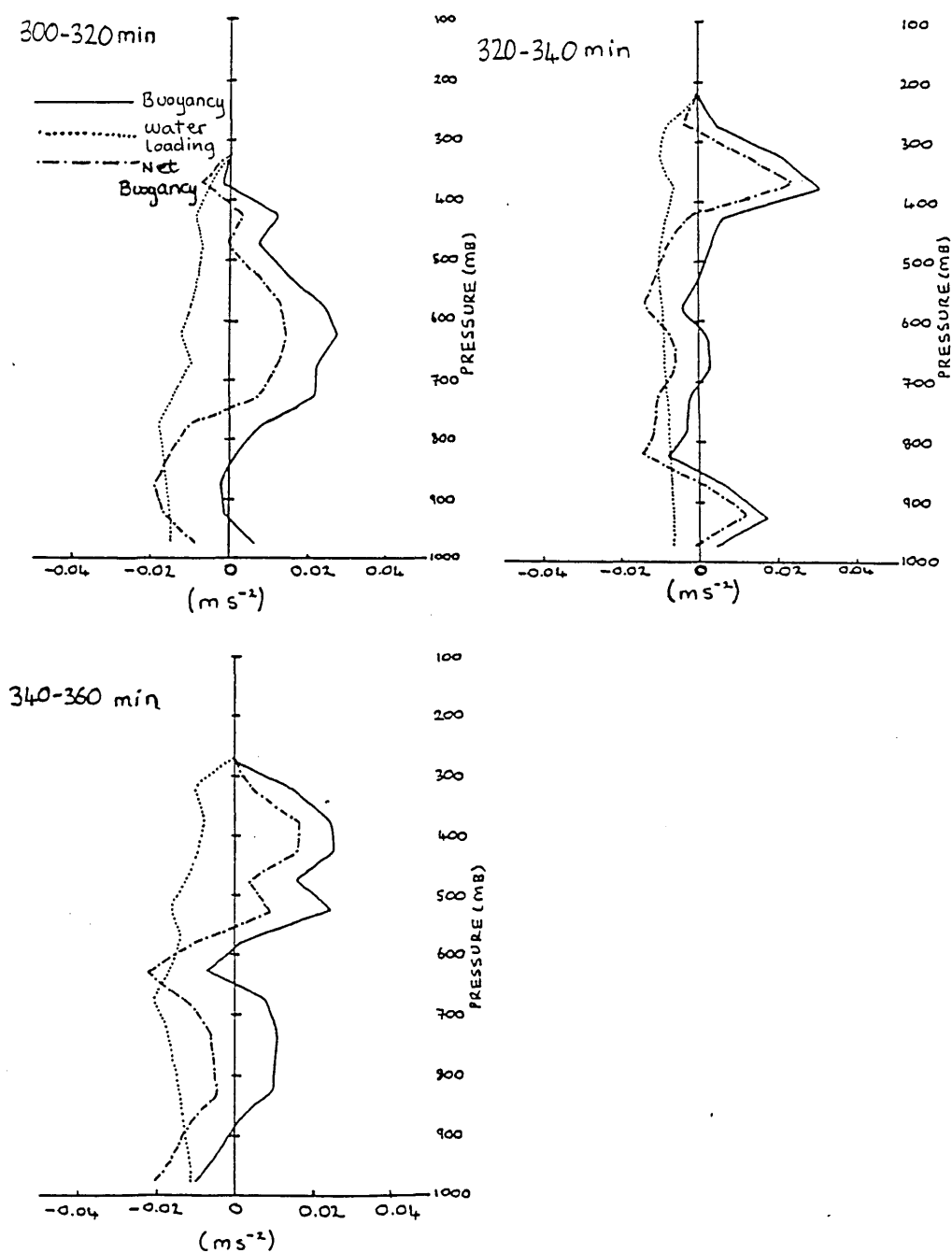


Fig 5.22 Buoyancy terms for DD of a single deep cloud (20 minute averages).

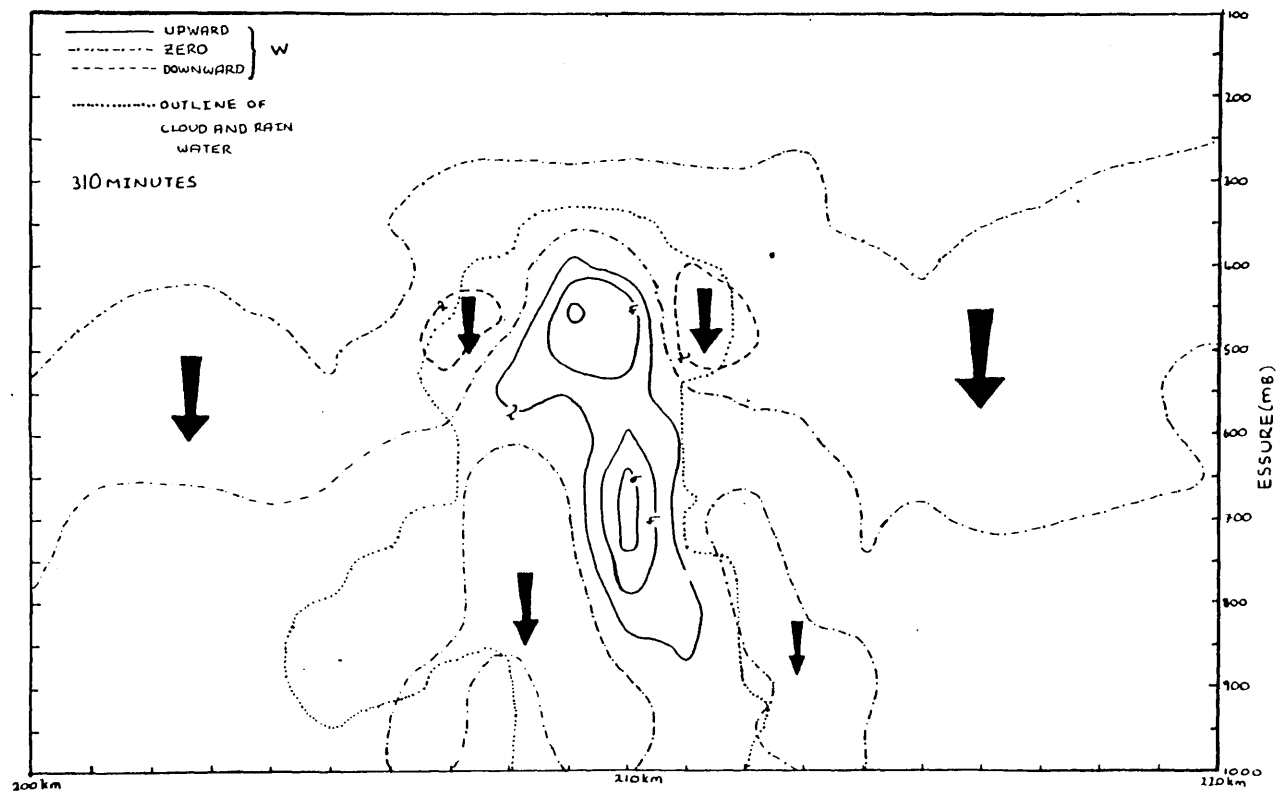


Fig 5.23 Spatial plots of a single deep cloud for 310 minutes showing vertical velocity and cloud location.

scheme), but no DD is represented. Eddy transports derived from this model were discussed in chapter 4. The simulated cloud had a width of 3000m, and so comparison is made to analytical clouds with radius between 1000 and 2000m.

$\overline{T'w}$ for the analytical model (fig 5.25a) is of larger magnitude than that averaged over the 80 minutes of the cloud life cycle. $\overline{w'w}$ (fig 5.25b) is also much too large, and peaks at a higher level. However the structure of the analytical and simulated cloud are in quite good agreement for 20-40 minutes of the life cycle, although $\overline{w'w}$ is again too large. However Kuo and Raymond(1980) considered incorporating DD in such a model, and found that $\overline{w'w}$ was halved. This would bring better agreement with the mature phase of the simulated cloud. The agreement gives some credence to the form of the entrainment rate used in the analytical model, which is proportional to w^2 .

From this brief comparison, one dimensional models are a bad representation of the mean internal structure of clouds. However comparisons with the mature phase of the simulated convection is favourable. Their performance in parameterization schemes within both large-scale and meso-scale models, perhaps, could be improved by using a simple evolution of the buoyancy profile determined from numerical studies such as the one here. The use of two dimensional analytical models (Moncrieff(1981,1985)) may also prove useful. These contain a greater consideration of dynamics, although at present the thermodynamics of such models is crude, and predicted vertical velocities too large.

5.5 SUMMARY.

In this chapter a mathematical theory of parameterization has been developed and discussed, $(Q1)_{conv}$ and $(Q2)_{conv}$ being represented in terms of cloud and large-scale parameters. Comparison has been made to approximate forms of these equations used by previous workers in many parameterization schemes.

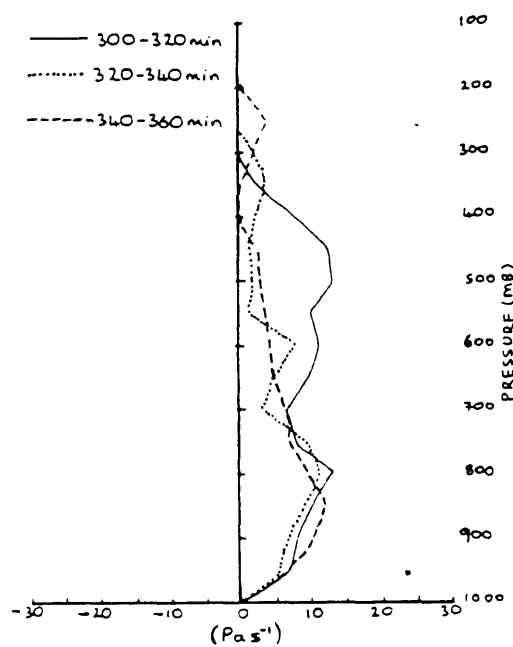


Fig 5.24 $\bar{\omega}^{sp}$ for a single deep cloud (20 minute averages).

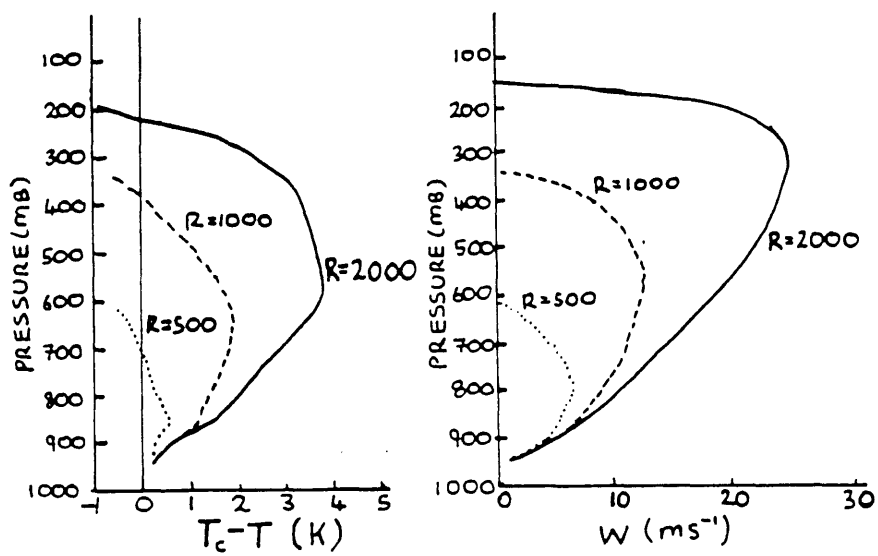


Fig 5.25 $(\bar{T}_c - \bar{T})$ and $\bar{\omega}^c$ for a typical 1D cloud model (profiles for three cloud radii, R , shown) (From Anthes(1977)).

The internal structure of cloud within the domain has been studied, separated into updraught, downdraught, and inactive cloud. In an eight hour average the mean ensemble cloud has been described. This is envisaged to be useful in studying the parameterization of precipitating convection within large-scale models, where grid sizes are on the scale of cumulus ensembles. This was compared to the mean updraught and downdraught structure of an individual cumulonimbus. Deep convection is seen to dominate the structure of the mean ensemble cloud updraught and downdraught, although the effect of weaker convection is important for updraught structure below 500mb.

The same cloud was also used to study the evolution of a single storm, on a twenty minute time scale, in relation to the evolution of cloud transport properties on this scale described in chapter 4. These are associated with changes in updraught structure, linked to the increase of rainwater within cloud and the formation of downdraughts.

Comparison of the single cloud to a one dimensional analytical cloud model was carried out. This failed to represent the mean structure of deep convection over it's life cycle, although comparison with the mature phase of convection was favourable. It was suggested that numerical models might be used to improve analytical models, by suggesting simple representations of time dependent effects, so improving their performance within parameterization schemes.

In chapter 6, the cloud model described above will be employed to understand the structure of cloud scale thermodynamic transports, and a numerical study of the parametric forms of $(Q1)_{conv}$ and $(Q2)_{conv}$ is carried out.

CHAPTER 6

THE PARAMETERIZATION OF CLOUD THERMODYNAMIC EFFECTS.

.In this chapter the parameterization of the thermodynamic effects of an ensemble of deep convective clouds is considered. The forms of $(Q1)_{para}$ and $(Q2)_{para}$ derived in chapter 5 are diagnosed using sim B, and the relative importance of each term discussed. Approximations useful for parameterization are identified and discussed. Brief comparison are made with some of the existing schemes described in chapter 5. Further study of the approximate form of the vertical eddy flux divergence is carried out, and the neglected part is found to be important.

6.1 NUMERICAL ANALYSIS OF THE PARAMETRIC FORMS OF Q1 AND Q2.

Equation (5.19) expresses $(Q1)_{conv}$ in terms of cloud and large-scale variables. Before this (and the equivalent form for $(Q2)_{conv}$) can be used in connection with the meso-scale model, a slight modification is required. Large-scale, radiational, boundary layer and cloud forcing are applied to each grid point regardless of whether it is clear or cloud air. Thus the theta equation averaged over cloud air will contain a contribution from these forcings; $\sigma \overline{(\partial \theta / \partial t)}^c_{For}$. Equation (5.19) becomes,

$$\frac{\partial \sigma \bar{\theta}^e}{\partial t} - \theta_b \frac{\partial \sigma}{\partial t} + \sigma \left(\frac{\partial v \theta}{\partial x} \right)^e + \frac{\partial \sigma \omega \bar{\theta}^e}{\partial p} - (\omega \theta)_b \frac{\partial \sigma}{\partial p} = \sigma \frac{L \bar{Q}}{c_p \pi} + \sigma \left(\frac{\partial \theta}{\partial t} \right)_{\text{FOR}}^e + \sigma \bar{S}_\theta^e \quad \text{eqn(6.1)}$$

Working through to obtain the form of $(Q1)_{\text{para}}$ using eqn(6.1) as in chapter 5 gives,

$$\begin{aligned} (Q1)_{\text{para}} = & (1-\sigma) \frac{L \bar{Q}}{c_p \pi} + (\bar{\theta} - \theta_b) \frac{\partial \sigma}{\partial t} + \frac{\partial \sigma \bar{\theta}^e}{\partial t} + (\omega_b \bar{\theta} - (\omega \theta)_b) \frac{\partial \sigma}{\partial p} \\ & + \sigma \left(\frac{\partial v \theta}{\partial x} \right)^e + \sigma \left(\frac{\partial v}{\partial x} (\theta - \bar{\theta}) \right)^e + \sigma \bar{\omega}^e \frac{\partial \bar{\theta}}{\partial p} - \sigma \bar{S}_\theta^e - \sigma \left(\frac{\partial \theta}{\partial t} \right)_{\text{FOR}}^e \end{aligned} \quad \text{eqn(6.2)}$$

An equivalent term appears in the parametric form of $(Q2)_{\text{conv}}$. As σ is small for a large depth of the atmosphere, the contribution of these forcing terms is expected to be small.

Most of the terms in eqn(6.2) are easily calculated from the model grid. Care was taken to ensure that similar space differencing schemes were employed in diagnosing the parametric form of $(Q1)_{\text{conv}}$ (and $(Q2)_{\text{conv}}$), as were used in the model, and also the diagnostic studies of chapters 3 and 4. (Success in this matter is seen later in the agreement of the value of $(Q1)_{\text{conv}}$ calculated from eqn(6.2) and the lhs of the domain averaged theta equation.)

The term representing moist processes within the environment, $((1-\sigma)L\bar{Q}/c_p\pi)$ cannot be calculated independently, as it requires knowledge of θ_b , which is difficult to specify from the numerical results. The same is also true of $(\bar{\theta} - \theta_b)\partial\sigma/\partial t$. However their joint value may be obtained. Consider the theta equation averaged over environmental air,

$$\begin{aligned} \frac{\partial (1-\sigma) \bar{\theta}^e}{\partial t} - \theta_b \frac{\partial (1-\sigma)}{\partial t} + (1-\sigma) \left(\frac{\partial v \theta}{\partial x} \right)^e + (1-\sigma) \left(\frac{\partial \omega \theta}{\partial p} \right)^e \\ = (1-\sigma) \frac{L \bar{Q}}{c_p \pi} + (1-\sigma) \bar{S}_\theta^e + (1-\sigma) \left(\frac{\partial \theta}{\partial t} \right)_{\text{FOR}}^e \end{aligned} \quad \text{eqn(6.3)}$$

Using eqn(5.3) and adding $\bar{\theta} \partial \sigma / \partial t$ to both rhs and lhs of eqn(6.3), after some rearrangement eqn(6.4) is obtained:-

$$\begin{aligned} \left\{ \frac{\partial \bar{\theta}}{\partial t} - \frac{\partial \sigma \bar{\theta}^e}{\partial t} \right\} + \bar{\theta} \frac{\partial \sigma}{\partial t} + (1-\sigma) \left(\frac{\partial v \theta}{\partial x} \right)^e + (1-\sigma) \left(\frac{\partial \omega \theta}{\partial p} \right)^e - (\bar{S}_\theta - \sigma \bar{S}_\theta^e) \\ - \left(\left(\frac{\partial \theta}{\partial t} \right)_{\text{FOR}} - \sigma \left(\frac{\partial \theta}{\partial t} \right)^e \right) = (1-\sigma) \frac{L \bar{Q}}{c_p \pi} + (\bar{\theta} - \theta_b) \frac{\partial \sigma}{\partial t} \end{aligned} \quad \text{eqn(6.4)}$$

The lhs of eqn(6.4) consists of terms known from the meso-scale model. Hence the rhs, which forms part of the rhs of eqn(6.2) are calculated without any knowledge of θ_b .

The other two terms involve values upon the cloud boundary. These are calculated as residuals. One from the cloud averaged continuity equation (multiplying by $\bar{\Theta}$),

$$\omega_b \bar{\Theta} \frac{\partial \sigma}{\partial p} = \bar{\Theta} \sigma \left(\frac{\partial \omega}{\partial x} \right)^c + \bar{\Theta} \frac{\partial \sigma \bar{\omega}^c}{\partial p} - \bar{\Theta} \frac{\partial \sigma}{\partial t} \quad \text{eqn(6.5)}$$

the other from the rule for integrating over variable boundaries,

$$-(\omega \Theta)_b \frac{\partial \sigma}{\partial p} = \sigma \left(\frac{\partial \omega \Theta}{\partial p} \right)^c - \frac{\partial \sigma \omega \Theta^c}{\partial p} \quad \text{eqn(6.6)}$$

The same methods are used in diagnosing the parametric form of $(Q2)_{\text{conv}}$.

6.2 THE PARAMETERIZATION OF Q1.

In this section, budgets and vertical profiles of the parametric form of $(Q1)_{\text{conv}}$ are used to study how the effects of convection upon the large-scale temperature field may be simply represented. This is carried out using data averaged over eight hours. Comments are made concerning twenty minute averages later (section 6.5).

The budget of the rhs of eqn(6.2) is presented in table 6.1 for all cloud air, and also for the components defined in chapter 5. This table will be referred to extensively throughout the following sections, the terms being referred to by their abbreviated names.

(a) "BAL" terms.

Study of table 6.1 indicates that the three terms labelled "BAL" are at least one order of magnitude larger than $(Q1)_{\text{para}}$. However on addition a large degree of cancellation occurs. For all cloud, the net value of the three terms is only -43 Wm^{-2} , less than 1% of the value of the smallest term, and only 8% of $(Q1)_{\text{para}}$. Similar results are obtained for UD, DD, and IC.

Fig(6.1) shows the vertical profile of $(\text{BAL})_{\text{T}}$ again for all cloud. The balance is apparent at each level of the model. Maximum values are around 2K/day, whereas individual terms may exhibit maxima of 400K/day.

Term	Abbreviation	All cloud (Wm^{-2})	Updraught (Wm^{-2})	Downdraught (Wm^{-2})	Inactive cloud (Wm^{-2})
$\pi\sigma\omega\frac{\partial\bar{\theta}}{\partial p}$	CS	692	1026	-508	174
$(1-\sigma)\frac{L\bar{Q}}{c_p} + \pi(\bar{\theta}-\theta_v)\frac{\partial\sigma}{\partial t}$	EE	-197	-	-	-
$\pi\sigma\left(\frac{\partial u}{\partial x}(\bar{\theta}-\bar{\theta})\right)$	ED	52	39	4	9
$\pi\sigma\left(u\frac{\partial\bar{\theta}}{\partial x}\right)^c$	ADV	-58	-17	-22	-19
$\pi\frac{\partial\sigma\bar{\theta}^c}{\partial t}$	BAL	11669	-104	1444	10329
$\pi\omega_b\bar{\theta}\frac{\partial\sigma}{\partial p}$		-19139	3659	-5250	-17548
$-\pi(\omega\bar{\theta})_b\frac{\partial\sigma}{\partial p}$		7427	-3537	3791	7173
$-\pi\sigma\bar{S}_\theta^c$	TURB	-17	24	-7	-34
$-\pi\sigma\left(\frac{\partial\bar{\theta}}{\partial t}\right)_{\text{Forc}}$	FORC	98	3	28	67
$\pi(Q1)_{\text{para}}$		527	1093	-520	151
$\pi(Q1)_{\text{conv}}$		527	-	-	-

Table 6.1 Budgets of terms on rhs of the parametric form of $(Q1)_{\text{conv}}$, for all cloud, updraught, downdraught, and inactive cloud (averaged between 1 and 9 hours).

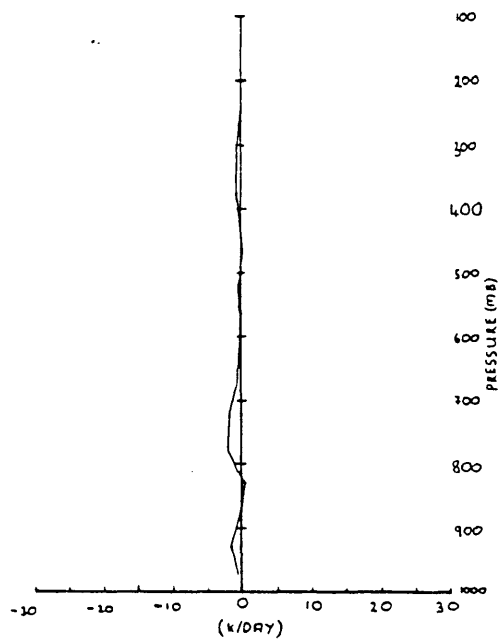


Fig 6.1 $(BAL)_t$, all cloud (1-9 hours).

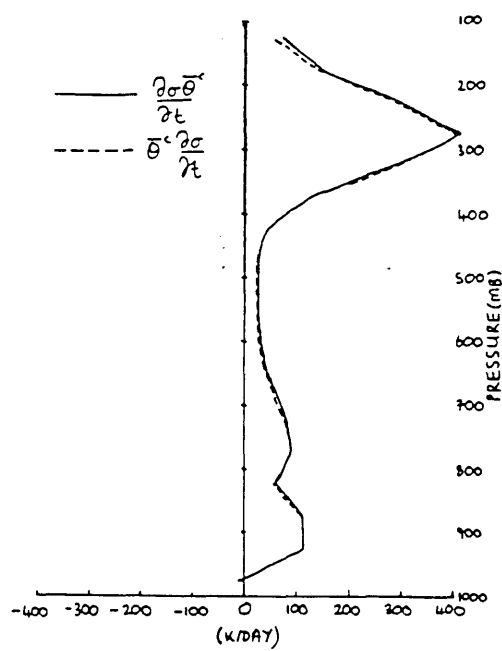


Fig 6.2 $\partial\sigma\bar{\theta}/\partial t, \bar{\theta}\partial\sigma/\partial t$, all cloud (1-9 hours).

The nature of the balance is not the same for each component of cloud. Compared to IC, UD is in a near steady state. Indeed IC contributes 90% of all the cloud value of $\partial\sigma\bar{\theta}/\partial t$. Much of this variation is due to the change in cloud area with time, as seen in fig 6.2 which compares $\partial\sigma\bar{\theta}/\partial t$ to $\bar{\theta}^c\partial\sigma/\partial t$, again for all cloud. Agreement is good. The large change at high levels is due to the increase in "anvil" cloud as the simulation progresses. The internal thermodynamic structure of cloud air, especially of UD and DD, is approximately steady state in nature. Hence in parameterization schemes the use of some form of steady state cloud model may be justified.

Why do these three terms balance to such a large extent at all levels of the domain? Consider the cloud averaged continuity equation, (eqn(5.15)). Multiplying by $\bar{\theta}$ and rearranging,

$$\omega_b \bar{\theta} \frac{\partial \sigma}{\partial p} = \sigma \bar{\theta} \left(\frac{\partial \bar{\omega}}{\partial x} \right)^c + \bar{\theta} \frac{\partial \sigma \bar{\omega}^c}{\partial p} - \bar{\theta} \frac{\partial \sigma}{\partial t}$$

The difference between $\bar{\theta}$ and $\bar{\theta}^c$ is only a few degrees in 300K (see chapter 5), ($\approx 0.5\%$), and so using the result of fig 6.2,

$$\bar{\theta} \frac{\partial \sigma}{\partial t} \approx \bar{\theta}^c \frac{\partial \sigma}{\partial t} \approx \frac{\partial \sigma \bar{\theta}^c}{\partial t}$$

with an error of $<1\%$ in the budget.

Thus it can be written,

$$\omega_b \bar{\theta} \frac{\partial \sigma}{\partial p} \approx \sigma \bar{\theta} \left(\frac{\partial \bar{\omega}}{\partial x} \right)^c + \bar{\theta} \frac{\partial \sigma \bar{\omega}^c}{\partial p} - \frac{\partial \sigma \bar{\theta}^c}{\partial t} \quad \text{eqn(6.7)}$$

Consider now the definition of $(\omega\theta)_b \partial\sigma/\partial p$,

$$(\omega\theta)_b \frac{\partial \sigma}{\partial p} = \frac{\partial \sigma \bar{\omega} \bar{\theta}^c}{\partial p} - \sigma \left(\frac{\partial \bar{\omega} \bar{\theta}}{\partial p} \right)^c \quad \text{eqn(6.8)}$$

Expanding the second term on the rhs,

$$(\omega\theta)_b \frac{\partial \sigma}{\partial p} = \frac{\partial \sigma \bar{\omega} \bar{\theta}^c}{\partial p} - \sigma \left\{ \left(\bar{\omega} \frac{\partial \bar{\theta}}{\partial p} \right)^c + \left(\bar{\theta} \frac{\partial \bar{\omega}}{\partial p} \right)^c \right\}$$

Suppose the terms on the rhs were written uncorrelated, ie,

$$(\omega\theta)_b \frac{\partial \sigma}{\partial p} \approx \frac{\partial \sigma \bar{\omega}^c \bar{\theta}^c}{\partial p} - \sigma \left(\bar{\omega}^c \left(\frac{\partial \bar{\theta}}{\partial p} \right)^c + \bar{\theta}^c \left(\frac{\partial \bar{\omega}}{\partial p} \right)^c \right) \quad \text{eqn(6.9)}$$

This is not usually thought to be a good approximation, especially for cloud air. However budgets of the model indicate that using this form an error of only 10% is induced, mostly arising from

the second term in the rhs of eqn(6.9).

Using the model continuity equation (eqn(3.3)), after some rearrangement,

$$(\omega\theta)_b \frac{\partial \sigma}{\partial p} \approx \bar{\omega}(\theta_b - \bar{\theta}) \frac{\partial \sigma}{\partial p} + \bar{\theta} \frac{\partial \sigma \bar{\omega}}{\partial p} + \sigma \bar{\theta} \left(\frac{\partial \bar{\omega}}{\partial x} \right)$$

It would seem reasonable to assume that $\theta_b \approx \bar{\theta}$, and so the first term may be neglected when compared to the second. Also using $\bar{\theta} \approx \bar{\theta}$,

$$(\omega\theta)_b \frac{\partial \sigma}{\partial p} \approx \bar{\theta} \frac{\partial \sigma \bar{\omega}}{\partial p} + \sigma \bar{\theta} \left(\frac{\partial \bar{\omega}}{\partial x} \right)$$

with an error of less than 1% (using model data).

Substituting into eqn(6.7) and rearranging,

$$\frac{\partial \sigma \bar{\theta}}{\partial t} + (\omega_b \bar{\theta} - (\omega\theta)_b) \frac{\partial \sigma}{\partial p} \approx 0 \quad \text{eqn(6.10)}$$

Thus from this argument, the three terms are seen to balance (approximately) as a consequence of the continuity equation. Indeed eqn(6.10) is an approximate form of the cloud averaged continuity equation. No budget assumptions have been made in deriving eqn(6.10), and so the balance exists at all level of the atmosphere as shown in fig 6.1.

What consequence does this balance have for parameterization? Because the $(BAL)_T$ terms only contribute a small amount to $(Q1)_{para}$ (48%), it might be expected that these terms could be neglected in calculating an approximate value of $(Q1)_{conv}$. Parameterization schemes have done so in the past by assuming to be constant wrt pressure and time. This has been shown not to be the case (fig 5.2 and 6.2). However because of the balance described above, parameterization schemes have suffered little through this assumption.

(b) Advection and en/detrainment (ADV,ED).

Previous workers (eg Yanai et al(1973)) assumed that convective towers entrained air at the environmental temperature. Hence entrainment caused no exchange of energy between the two. Detrainment was assumed to occur in a thin layer at cloud top, the outflow air being warmer than cloud environment. Budget studies employing

observational data indicated that this term was small. The budgets of the model indicate that this is so, $(ED)_T$ being $\simeq 10\%$ of $(Q1)_{para}$. However a second term exists involving the lateral transport of energy across cloud boundary, $(ADV)_T$. This term has not been studied by previous workers, being incorporated into their detrainment term. Budgets show that this term opposes $(ED)_T$, their net value accounting for $\simeq 1\%$ of $(Q1)_{para}$. Thus the effect of lateral transports across cloud boundaries is small.

Fig 6.3a shows vertical profiles of $(ED)_T$ and $(ADV)_T$ averaged over all cloud air. The balance indicated by the budgets exists at all levels of the model.

The balance can be understood by considering a simple symmetric cloud model described in fig 6.4. In this case air is being entrained into cloud, although the argument below also applies for a detraining cloud. $(ADV)_T$ may be approximated by,

$$\overline{\left(u \frac{\partial \theta}{\partial x}\right)^c} \simeq \frac{1}{2} \left[u \frac{(\bar{\theta}^c - \bar{\theta})}{L} + (-u) \frac{(\bar{\theta}^c - \bar{\theta})}{L} \right] \simeq u \frac{(\bar{\theta}^c - \bar{\theta})}{L}$$

In a similar manner $(ED)_T$ is given by,

$$\overline{\left(\frac{\partial u}{\partial x} (\theta - \bar{\theta})\right)^c} \simeq \frac{1}{2} \left[\frac{0-u}{L} (\bar{\theta}^c - \bar{\theta}) + \frac{(-u-0)}{L} (\bar{\theta}^c - \bar{\theta}) \right] \simeq -u \frac{(\bar{\theta}^c - \bar{\theta})}{L}$$

The approximate expressions of $(ADV)_T$ and $(ED)_T$ are seen to be equal and opposite, at each level of the cloud. The slight imbalance observed in the value obtained from the model is a consequence of the simulated clouds lacking horizontal symmetry.

Although the terms oppose each other for each component of cloud air, the balance is not as good in magnitude when compared to that averaged over all cloud. This is due to asymmetry caused by the arrangement of UD, DD, and IC wrt each other, and cloud environment. Such asymmetry is mostly averaged out when all cloudy air is considered.

Vertical profiles of $(ED)_T$ and $(ADV)_T$ for each cloud component are shown in fig 6.3b. $(ED)_T$ for UD is larger than for the other two components, as indicated in table 6.1. This is due to the larger cloud large-scale temperature difference for UD (up to 1K when averaged over 8 hours). The detrainment of warm cloud air from UD

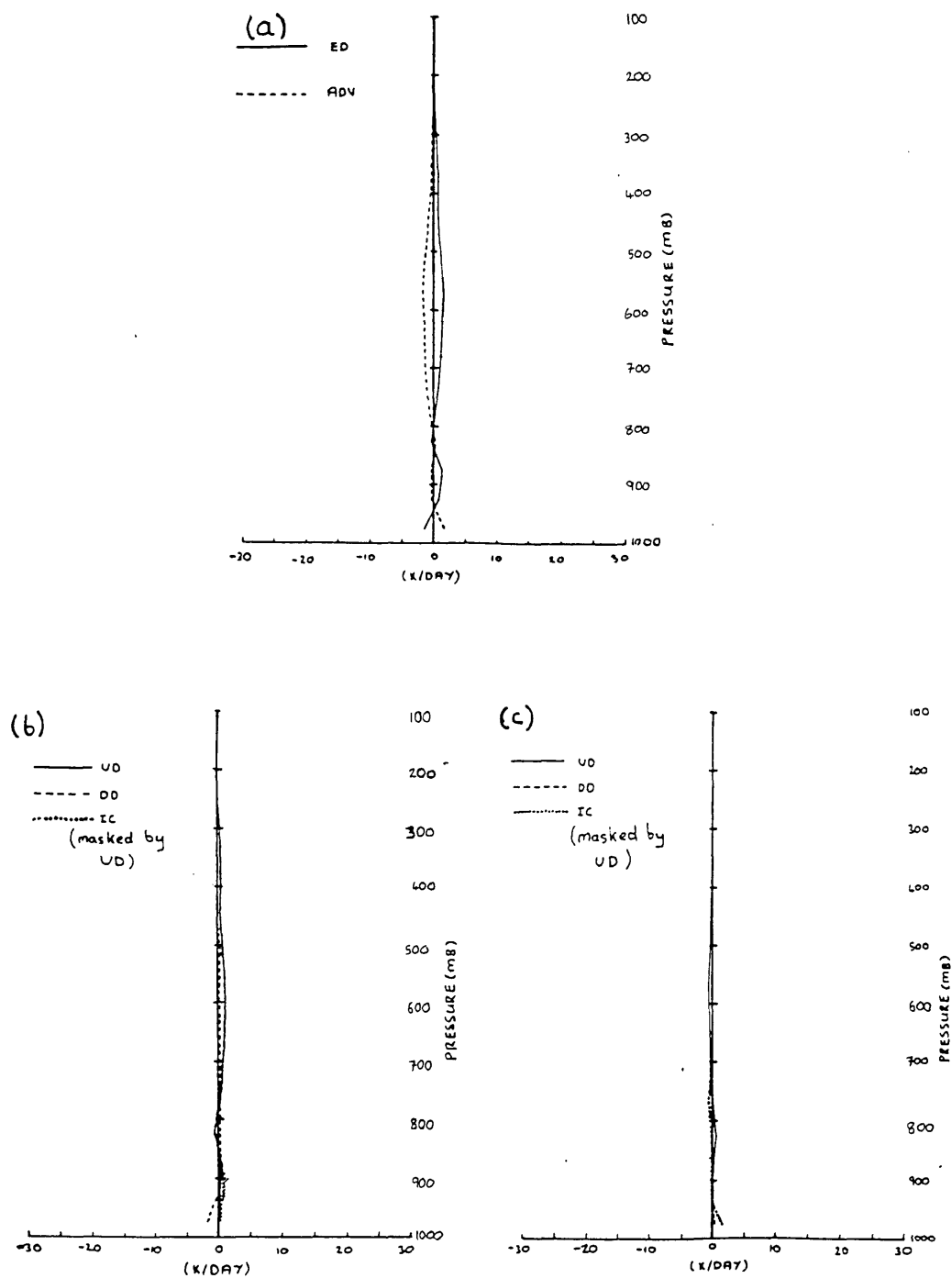


Fig 6.3(a) $(ED)_T$, $(ADV)_T$, all cloud (1-9 hours).
 (b) $(ED)_T$ for UD, DD, IC (1-9 hours).
 (c) $(ADV)_T$ for UD, DD, IC (1-9 hours).

above 800mb causes the environment to warm, although the evaporation of detrained liquid water can cause a net cooling to be attributed to detrainment. Study of fig 6.6 shows that the cooling due to $(EE)_T$ is larger than warming due to $(ED)_T$. Below 850mb warming is also seen. In this region UD air is colder than the environment. Hence $(ED)_T$ represents cold air entrained from cloud environment, implying a warming of the large-scale. The cooling around 800mb is due to the entrainment of warm air as UD temperature changes from negative to positive with height. Below 900mb DD cools the large-scale through detrainment of cold air as density currents travel outward from decaying storms (maximum cooling is $\approx 2K/day$ at 975mb). Inactive cloud warms the large-scale through $(ED)_T$ in the lowest 150mb of the domain ($\approx 0.5K/day$). This may be associated with entraining cloud air around the base of UD, not incorporated into the definition of UD.

$(ADV)_T$ is more difficult to interpret simply. For all cloud components it shows a similar structure and magnitude. Much of the depth of the domain is cooled by this term, the maximum rate being $\approx 1K/day$.

(c) Turbulence and forcing ($TURBC$, $FORC$).

$(FORC)_T$ contributes $\approx 20\%$ of the magnitude of $(Q1)_{para}$. However this is somewhat balanced by a combination of $(ED)_T$, $(ADV)_T$, $(BAL)_T$ and $(TURBC)_T$. Adding these to $(FORC)_T$, the residual is only $\approx 5\%$ of $(Q1)_{para}$. Fig 6.5 shows the vertical structure of $(FORC)_T$ for all cloud. Maxima occur at high and low level, due to the large amounts of "anvil" and low cloud, the main contributor to the term being IC. UD and DD are little effected by $(FORC)_T$.

$(TURBC)_T$ plays little part in $(Q1)_{para}$ (fig 6.5). Maximum magnitudes of $\approx 2K/day$ are seen, again mostly due to IC. For this the term is a maximum in the lowest 200mb.

(d) Moist processes within cloud environment (EE).

$(EE)_T$ contributes $\approx 15\%$ to $(Q1)_{para}$. Previous workers have associated this term with the evaporation of detrained liquid water, which seems reasonable. The negative value of $(EE)_T$ when integrated throughout the domain ($\approx 200Wm^{-2}$) would seem to agree with this

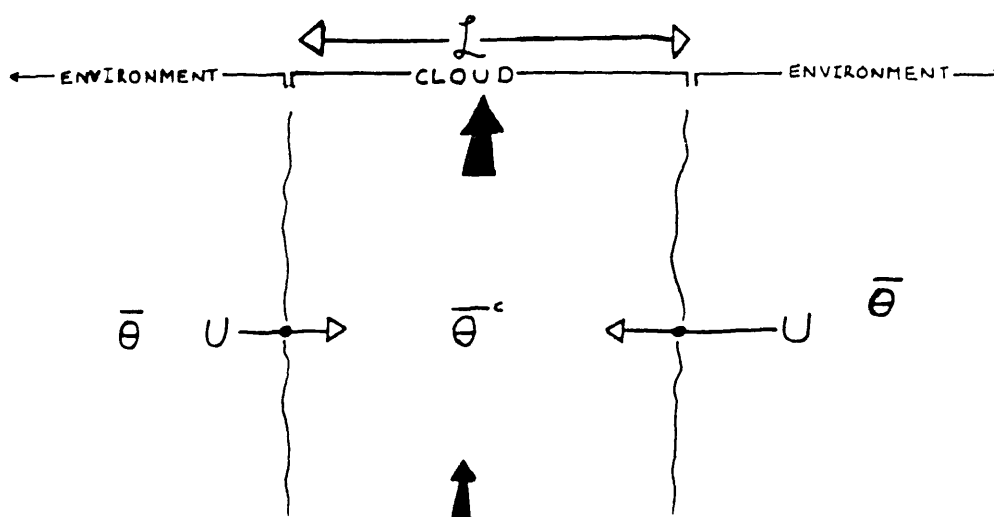


Fig 6.4 Simple symmetric cloud model used to explain balance between ED and ADV.

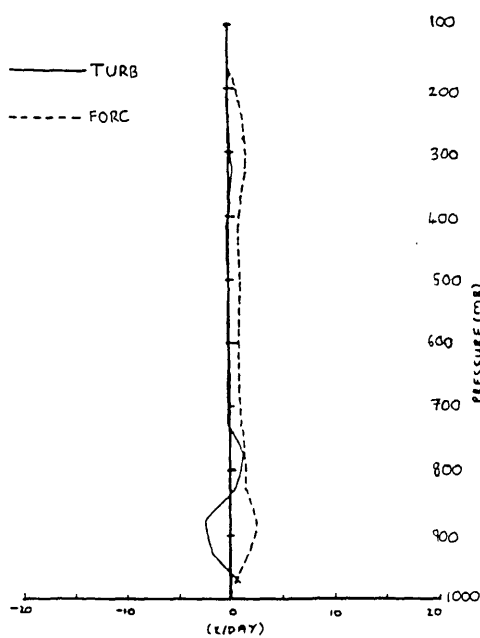


Fig 6.5 $(FORC)_T$, $(TURB)_T$,
all cloud (1-9 hours).

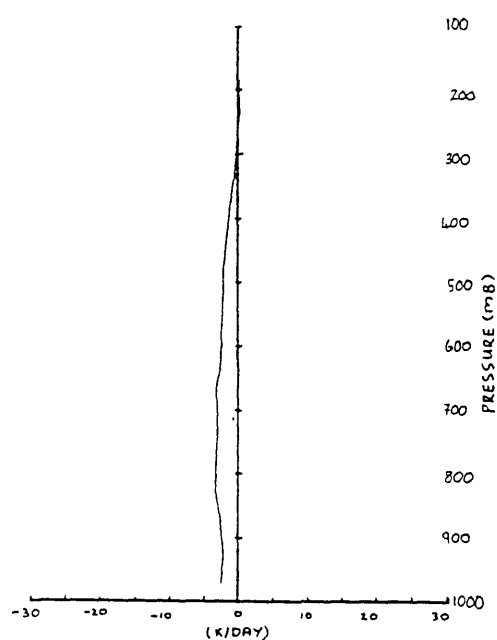


Fig 6.6 $(EE)_T$, all cloud
(1-9 hours).

assumption. Fig 6.6 shows the vertical profile of $(EE)_\tau$, which is negative for all levels of the domain, strengthening the argument. However it must be remembered that $(EE)_\tau$ as presented here contains a contribution from $(\bar{\theta} - \theta) \partial \sigma / \partial t$. As seen previously $\partial \sigma / \partial t$ is large at high levels of the domain. In the discussion of the $(BAL)_\tau$ terms, the approximation $(\bar{\theta} - \theta) \ll \bar{\theta}$ was used successfully. Therefore it is reasonable to assume that the contribution from the time dependent term is small. Agreement of the magnitude and structure of this term with that calculated from mixing ratio, would bring further confirmation, and this is studied later (section(6.2)).

(e) Compensating subsidence (CS) .

This term represents the response of cloud environment to the motions within cloud. Previous workers using observational data and analytical cloud models have found this to be the largest contributor to $(Q1)_{para}$ (eg Yanai et al(1973)). This is confirmed by the budget results. For all cloud, $(CS)_\tau$ over estimates $(Q1)_{para}$ by $\approx 30\%$. $(CS)_\tau$ is also the major term in the budgets of UD, DD, and IC. Fig 6.7a shows the vertical structure for all cloud, while 6.7b that of the three cloud components. $(CS)_\tau$ resembles $(Q1)_{conv}$. However it is an overestimate by $\approx 3K/day$ between 800 and 600mb, and no cooling is seen in the lowest 100mb. $(CS)_\tau$ follows the shape of $\sigma \bar{\omega}^*$ described in chapter 5, $\partial \bar{\theta} / \partial p$ being roughly constant with height.

Budgets show that $\overline{(CS)}_\tau^{ud}$ is approximately twice $(Q1)_{para}$. The profile peaks around 750mb with a maximum of $\approx 20K/day$. $\overline{(CS)}_\tau^{db}$ is of opposite sign to $(Q1)_{para}$, but magnitudes are similar. This shows a lower maxima than $\overline{(CS)}_\tau^{ud}$, and is relatively flat between 700 and 900mb. $\overline{(CS)}_\tau^{ic}$ is small above 700mb. However below this it acts in opposition to the cooling caused by DD. Thus no cooling is seen in the net $(CS)_\tau$ profile, as observed in $(Q1)_{conv}$.

(f) The importance of downdraught in determining Q1.

Many parameterization schemes do not include the effects of downdraughts (eg Kuo(1974), Anthes(1977), Arakawa-Schubert(1974)). This is not because they are considered unimportant, but that it is difficult to determine their structure in a simple manner. However these model results show how crucial DD are in determining the value

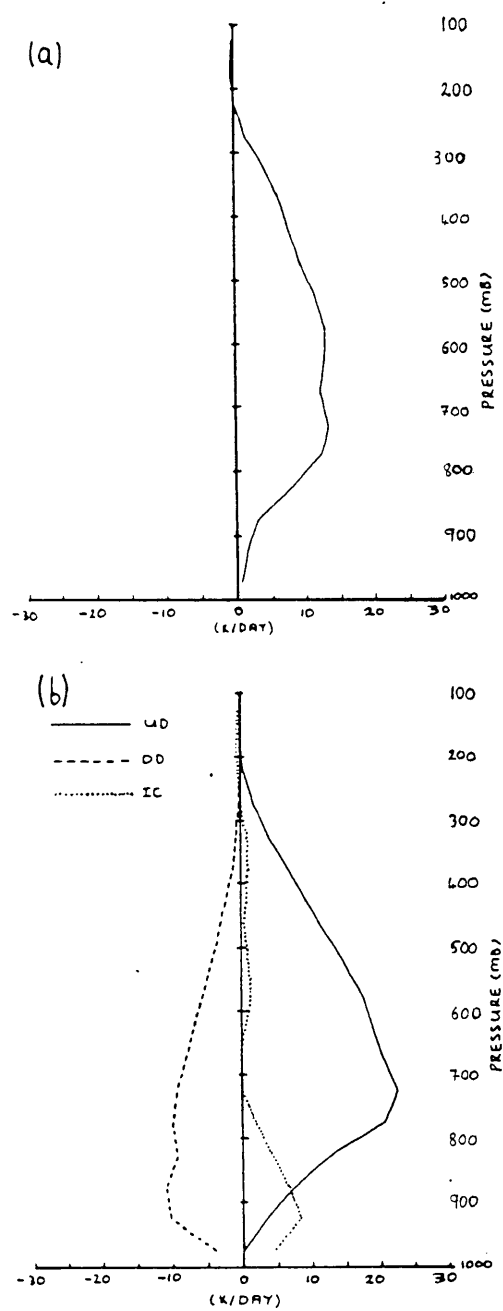


Fig 6.7(a) $(CS)_T$, all cloud (1-9 hours).

(b) $(CS)_T$ for UD, DD, IC (1-9 hours).

of $(Q1)_{conv}$.

The contribution of UD to $(Q1)_{conv}$ is roughly twice $|(Q1)_{conv}|$. Thus if a parameterization scheme considers only the effects of UD, it will greatly overestimate the heating due to convection upon the large-scale atmosphere. DD opposes the heating due to UD with a magnitude similar to $(Q1)_{conv}$. Hence large-scale processes plus DD oppose the effects of UD upon the atmosphere. The relationship between UD, DD and large-scale forcing will be discussed further in the next chapter.

(g) Brief comparison with alternative simulation (Sim F).

In section(4.2.1) of chapter 4 budgets of $(Q1)_{conv}$ and $(Q2)_{conv}$ were presented for a simulation identical to sim B, except that an alternative random number forcing was used to specify the cloud forcing. Budgets from both simulations compared well. A similar result is found if parametric budgets are considered. $(CS)_T$ differs by less than 10% between the two simulations (for eight hour averages). Other terms also compare well in magnitude. Thus the results presented in this section are independent of the temporal structure of the convection. As stated in chapter 3, the magnitude of convection would appear to be dependent upon the thermodynamic forcing supplied by large-scale processes. This will be investigated in the next chapter.

(h) Approximations to Q1 useful in parameterization.

Here simple approximations are identified by which $(Q1)_{conv}$ may be parameterized. The terms involved will need to be related to large-scale variables. This has been achieved by several methods in the past, the use of analytical cloud models being the most widely used. This subject is not dealt with here, rather the aim is to identify which terms need to be calculated.

The main term in the all cloud budgets is $(CS)_T$. Other large terms were $(EE)_T$ ($\approx 15\% (Q1)_{para}$), and $(FORC)_T$ ($\approx 10\% (Q1)_{para}$). This last term can be viewed as being balanced by the remaining terms of the budget. Adding $(FORC)_T$ to these gives a residual which is only $\approx 5\%$ of $(Q1)_{para}$.

Adding $(CS)_T$ and $(EE)_T$ gives a value of 495 Wm^{-2} , only 5% less than $(Q1)_{\text{conv}}$. Fig 6.8 shows the vertical structure of this combination, compared to the structure of $(Q1)_{\text{conv}}$. Above 800mb good agreement is seen. Between 800 and 900mb, $(Q1)_{\text{conv}}$ is underestimated by as much as 3K/day. The cooling of the lowest 100mb is reasonably represented, although at 925mb it is underestimated. This surface cooling is not caused by "up-sidence", compensating DD motion (which is masked by inactive cloud induced subsidence), or by the detrainment of cold air from downdraughts (which as seen earlier, does occur within the simulation). Rather the cooling is caused by evaporation of water within cloud environment.

In this approximation, $(CS)_T$ includes contributions from UD, DD and IC. UD is associated with positively buoyant air, while DD with negatively buoyant air. Such features have been modelled using simple analytical cloud models. Thus the mass flux for this cloud air, and so the compensating subsidence could perhaps be calculated. This is not so clear for the inactive cloud within the domain. The internal structure of such cloud is irregular (see chapter 5). Also turbulence plays a much larger role in the sensible heat budget compared to UD and DD. Obtaining a representation of such cloud using a simple model would seem difficult. As seen earlier (section 5.4.4), there is some indication that such cloud air is associated with UD inflow in the lower domain where it's effects are important, in which case they might be included in a simple model of the UD.

Some parameterization schemes do not consider $(EE)_T$ when approximating $(Q1)_{\text{conv}}$. However as seen from fig 6.8 this proves to be a bad approximation. From table 6.1, it is seen that $(EE)_T$ and $(\overline{CS})_T^{xc}$ oppose each other, with roughly equal magnitudes. Thus a possible method of not considering $(EE)_T$ in parameterization schemes might be to use $(\overline{CS})_T^{uo}$ and $(\overline{CS})_T^{oo}$. The net value of these two terms is 518 Wm^{-2} , closer to $(Q1)_{\text{conv}}$ than the method suggested previously. However comparison of the vertical structure of this combination to $(Q1)_{\text{conv}}$ is disappointing (fig 6.9). Above 800mb $(Q1)_{\text{conv}}$ is over-estimated by up to 4K/day, while below this it is under-estimated, and the depth of the cooling increased.

Fig 6.8 $(CS)_T + (EE)_T$, all cloud compared to $(Q1)_{conv}$ (1-9 hours).

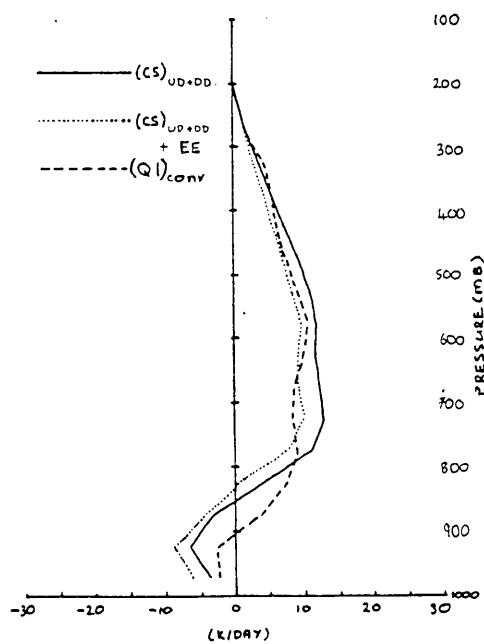
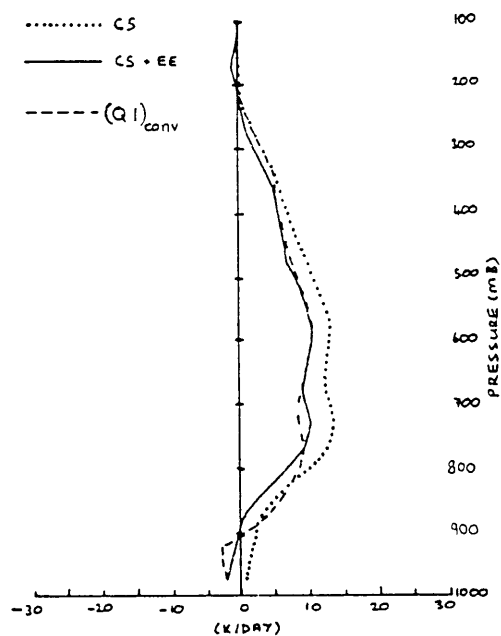


Fig 6.9 $(CS)_T^{UD} + (CS)_T^{DD}$ (1-9 hours).

Thus a combination of $(CS)_T$ and $(EE)_T$ best represents $(Q1)_{conv}$. Two other terms which might be simply represented using simple cloud models are $(ADV)_T$ and $(ED)_T$. However addition of both, or one of these terms brings little improvement.

These results agree with the results of previous workers. Yanai et al(1973) using Marshall Island data and a bulk one dimensional cloud model obtained a similar result (fig 6.10) (detrainment being only important at cloud top). Arakawa and Schubert(1974) used a combination of $(CS)_T$ and $(EE)_T$ as the basis for their parameterization scheme (again together with a detrainment term). This scheme has proved successful in large-scale models.

(i) "Errors" in the parametric representation of $Q1$.

Budget studies indicate that the value of $(Q1)_{conv}$ calculated from eqn(2.26) and eqn(6.2) (the parametric form of $(Q1)_{conv}$) are identical. However when the vertical profile of this residual is studied (fig 6.11), it is non-zero, although because of the budget result it must integrate to zero in the vertical.

Checks were carried out on the time and space differencing schemes used in the diagnosis. Equation(2.26) contains the same terms as the model thermodynamic equation, and the space differencing scheme (although not the time differencing scheme-see chapter 3) is identical. Many terms in the parametric form of $(Q1)_{conv}$ have no equivalent within the model equations. However the space scheme used in computing eqn(6.2) is identical to that used for eqn(2.26).

A term which has been neglected in obtaining eqn(6.2) is the contribution to $-\partial\overline{\omega\theta}/\partial p$ due to correlations between ω and θ within cloud environment;

$$-\frac{\partial((1-\sigma)\overline{\omega\theta^c} - (1-\sigma)\overline{\omega^c\theta})}{\partial p}$$

This integrates to zero over the entire depth of the domain as required by the residual of fig 6.11. A study of this term will be carried out after the parameterization of $(Q2)_{conv}$ has been considered.

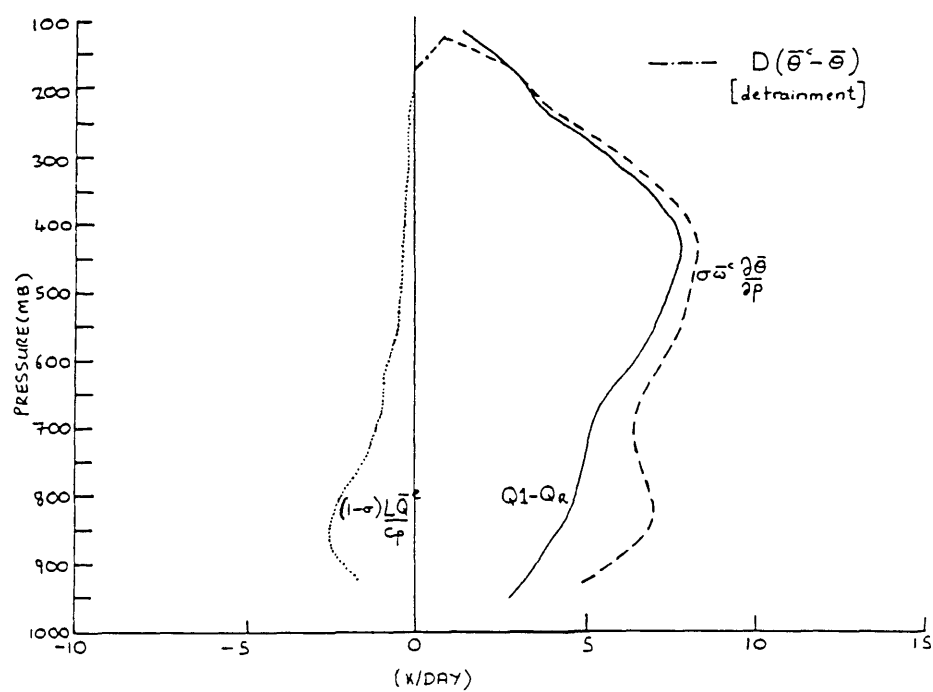


Fig 6.10 $(CS)_T + (EE)_T$ from an observational study
(from Yanai et al(1973)).

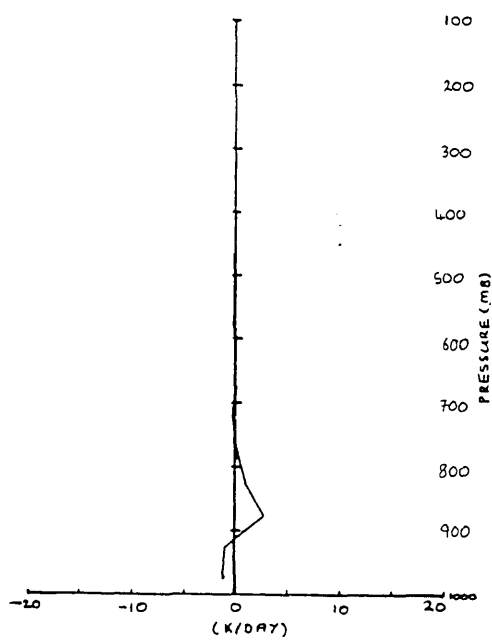


Fig 6.11 $(Q1)_{conv} - (Q1)_{para}$ (1-9 hours).

6.3 THE PARAMETERIZATION OF Q2.

This section has similar aims to the previous one, except the parametric form of $(Q2)_{\text{conv}}$ is studied. Budgets are presented in table 6.2 for all cloud and its three components. Several differences are apparent to the budgets of $(Q1)_{\text{para}}$, and these are discussed as each term is considered.

(a) "BAL" terms

The magnitude of these terms is of the same order as $(Q2)_{\text{para}}$, unlike the $(Q1)_{\text{para}}$ budget. The balance is also not as good. For all cloud, on addition, the residual is $\approx 20\%$ of the smallest term and $\approx 15\%$ of $(Q2)_{\text{para}}$. This is also exhibited in the vertical profiles, with a maximum magnitude of 3g/kg/day (fig 6.12) (cf. an average value of $(Q2)_{\text{para}}$ of $\approx 2\text{g/kg/day}$).

The lack of balance implies that the approximations made in the arguments of the previous section for $(\omega\phi)_q \partial\sigma/\partial p$ and $\partial\sigma\bar{\phi}/\partial t$ are worse for moisture than temperature. Errors of 5% are found in each of these terms for q , compared to $\approx 1\%$ for θ , due to larger errors in neglecting correlations, and also in the assumption that $(\bar{q}-\bar{q}) \ll \bar{q}$.

UD is seen to be in a steady state whereas DD and IC exhibit time dependent behaviour. Again this is due to changes in cloud area rather than in cloud internal structure. The main contributor comes from the lower domain, the large upper level increase in having little effect, as q is small in this region.

(b) En/detrainment and advection (ED,ADV).

Adding these two terms together for all cloud, the residual is 10% of $(Q2)_{\text{para}}$, much larger than similar terms in the $(Q1)_{\text{para}}$ budget. Although the terms oppose each other, there is a large difference in their magnitude. For DD and IC these terms are also in opposition, but not so for UD. Profiles of $(ED)_q$ and $(ADV)_q$ are presented in fig 6.13.

Term	Abbreviation	All cloud (Wm ⁻²)	Updraught (Wm ⁻²)	Downdraught (Wm ⁻²)	Inactive cloud (Wm ⁻²)
$\sigma \bar{\omega}^c \frac{\partial \bar{q}}{\partial p}$	CS	-801	-1209	661	-252
$-(1-\sigma) \bar{Q}^c + (\bar{q}-q_b) \frac{\partial \sigma}{\partial t}$	EE	202	-	-	-
$\sigma \left(\frac{\partial v}{\partial x} (\bar{q}-\bar{q}) \right)^c$	ED	-163	27	-92	-97
$\sigma \left(v \frac{\partial q}{\partial x} \right)^c$	ADV	105	45	41	19
$\frac{\partial \sigma \bar{q}^c}{\partial t}$	BAL	364	<0	115	249
$\omega_b q \frac{\partial \sigma}{\partial p}$		-859	654	-525	-988
$-(\omega q)_b \frac{\partial \sigma}{\partial p}$		571	-744	421	894
$-\sigma \bar{S}_2^c$	TURB	138	55	<0	83
$-\sigma \left(\frac{\partial q}{\partial t} \right)_{\text{forc}}^c$	FORC	-84	-10	-20	-56
$(Q2)_{\text{para}}$		-527	-1182	601	-148
$(Q2)_{\text{conv}}$		-527	-	-	-

Table 6.2 Budgets of terms on rhs of the parametric form of (Q2) ,
for all cloud, updraught, downdraught, and inactive cloud
(averaged between 1 and 9 hours).

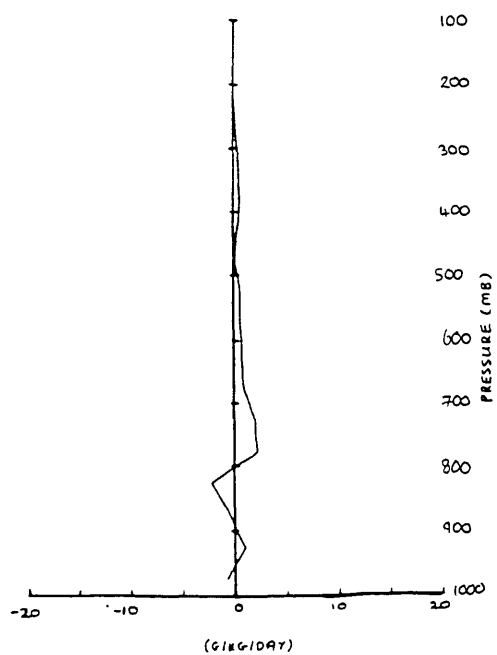


Fig 6.12 $(BAL)_q$, all cloud (1-9 hours).

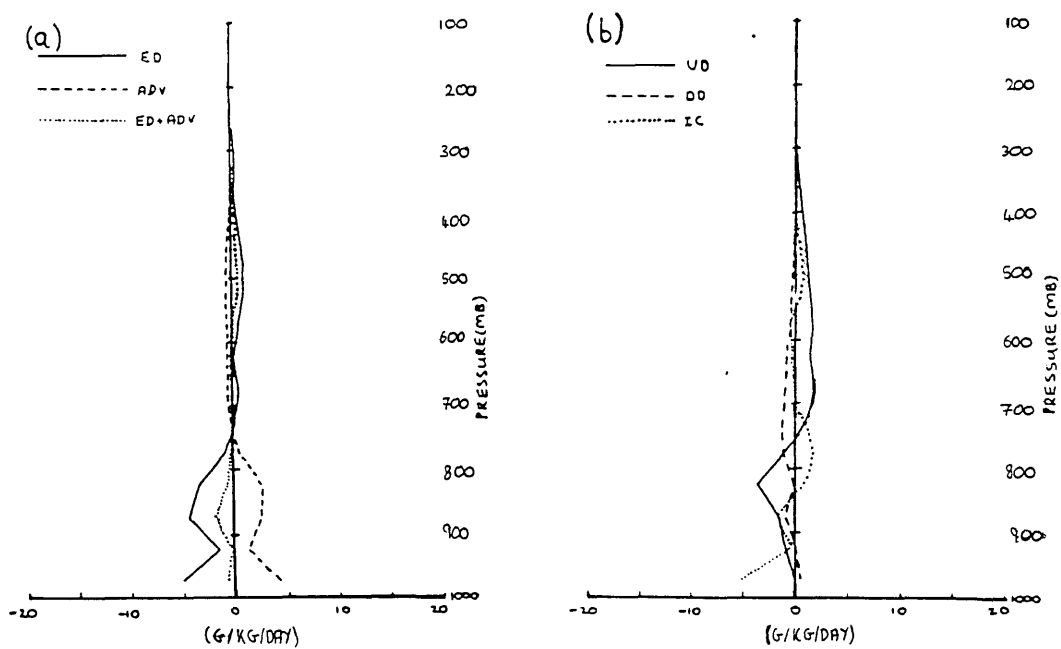


Fig 6.13(a) $(ED)_q$, $(ADV)_q$, all cloud (1-9 hours).
 (b) $(ED)_q$ for UD, DD, IC (1-9 hours).

The net value of these terms is negative. Hence convection dries the domain through lateral flow across cloud boundary. In simple analytical models, clouds only moisten the large-scale through detrainment in cloud outflow layers. Air is assumed to be entrained at the large-scale thermodynamic values, causing no change in the temperature and moisture profile of the cloud environment in the entrainment layer. The profile of $(ED+ADV)_q$ (fig 6.13(a)) does show slight moistening around 500mb. However drying in the lowest 250mb dominates. This is associated with entrainment into UD and IC. Within the simulation air in the UD inflow region is wetter than the domain average by up to 1g/kg, caused by convergence. Hence the process of entrainment dries the atmosphere.

UD moistens the domain by detrainment above 750mb (fig 6.13(b)). However this is opposed by entrainment into DD. DD moistens the lowest 100mb of the domain, mainly through advection, although detrainment contributes at the lowest model level. Above 950mb, IC mass detrainment layers are associated with slight moistening of the domain.

(c) Turbulence and forcing (TURB, FORC).

$(FORC)_q$ contributes 15% to $(Q2)_{para}$, mainly from IC in the lower domain. Magnitudes increase from UD to IC due to increasing cloud area. $(TURB)_q$ acts in opposition to $(FORC)_q$, and is larger than in the $(Q1)_{para}$ budget. DD contributes little to this term.

(d) Moist processes within cloud environment (EE) .

$(EE)_q$ is shown in fig 6.14(a). Budgets show that this term contributes $\approx 15\%$ of $(Q2)_{para}$, similar to $(Q1)_{para}$ budget, suggesting that the time dependent effects of this term are small. However the profile of $(EE)_q$ shows large differences from $(EE)_T$ (fig 6.14(b)). If the time dependent term were zero then no differences would be seen.

It is difficult to ascribe to which quantity the time dependent term is of greater importance. Fig 6.14(c) shows twenty minute average budgets of $(EE)_q$, $(EE)_T$, and the contribution of DD to $\langle c_p \pi (Q1)_{conv} \rangle$. Large variations of $(EE)_q$ are seen when DD forms, whereas $(EE)_T$ shows much steadier changes. As commented earlier, the approximation $(\bar{q} - q_k) \approx 0$ appears worse for q than θ . This suggests that

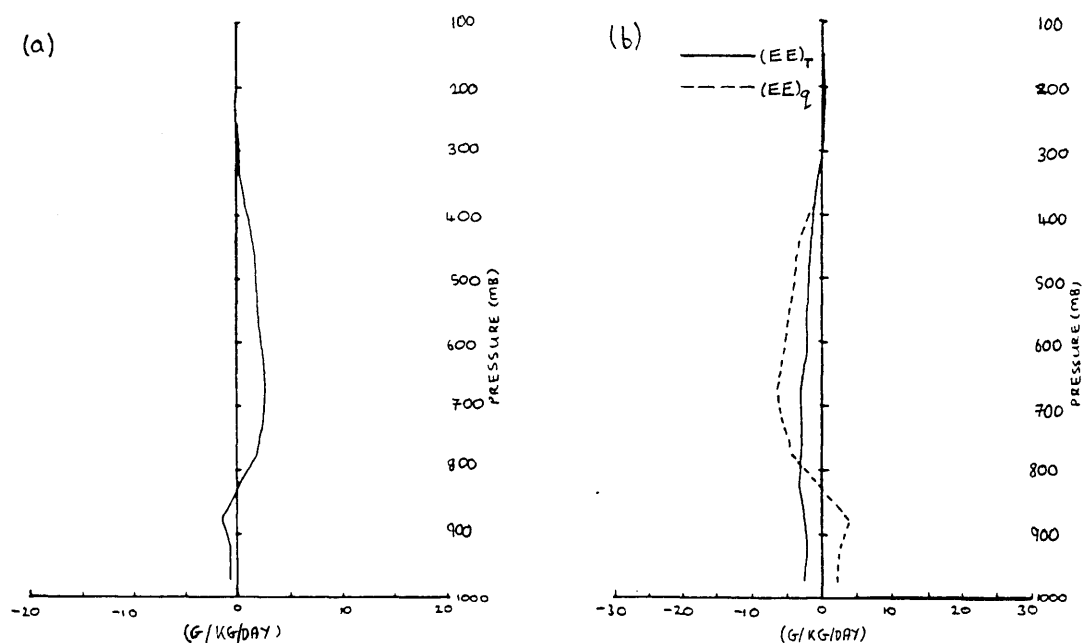


Fig 6.14(a) $(EE)_q$ (1-9 hours).

(b) Comparison of $(EE)_q$ and $(EE)_T$ (1-9 hours).

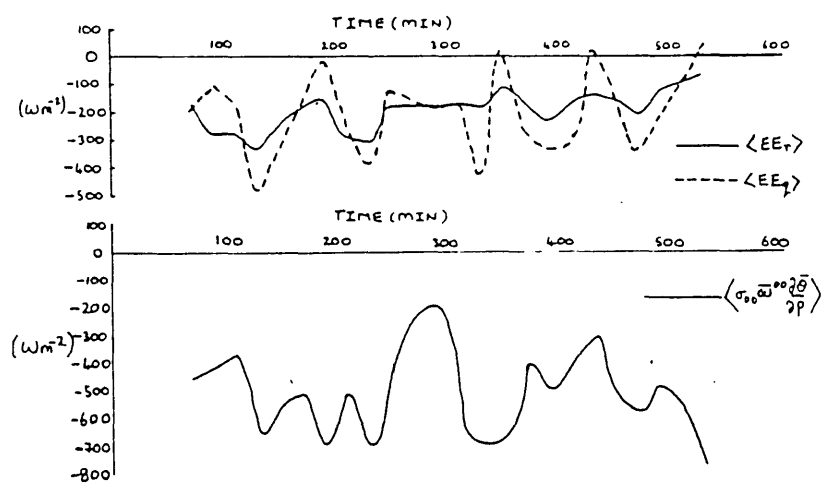


Fig 6.14(c) 20 minute budgets of $(EE)_q$, $(EE)_T$, and $(Q1)_{conv}^{DD}$ (1-9 hours).

greater time dependent effects are recorded in $(EE)_q$. This will prove a great difficulty for parameterization schemes!

(e) Compensating subsidence (CS).

All cloud $(CS)_q$ over estimates $(Q2)_{para}$ by $\approx 300W/m^2$ ie. $\approx 50\%$ (cf. $\approx 30\%$ for the $(Q1)_{para}$ budget). It is also evident that $(CS)_q$ does not follow the shape of $(Q2)_{para}$ (fig 6.15(a)), peaking at 775mb instead of 925mb. $(CS)_q$ follows $\sigma\bar{\omega}^c$, although $(\partial q/\partial p)$ does apply a modulating influence. $(\partial q/\partial p)$ decreases with height, and so $(CS)_q$ falls off more rapidly above the maximum for moisture than for temperature. The increase in $(\partial q/\partial p)$ towards the surface is not large enough to bring the maximum drying down to the surface. The same is true for each cloud component. $\overline{(CS)}_q^{ud}$ is approximately twice $(Q2)_{para}$, balanced by $\overline{(CS)}_q^{dp}$, as for the $(Q1)_{para}$ budget. $\overline{(CS)}_q^{xc}$ is negligible above 700mb, opposing the "up-sidence" due to DD (fig 6.15(b)).

(f) Approximations to Q2 useful in parameterization.

For temperature, $(Q1)_{conv}$ was well approximated by $(CS)_T + (EE)_T$, with an error of only 5%. Addition of other simple terms eg $(ED+ADV)_T$ brought no improvement. Adding $(CS)_q$ and $(EE)_q$ give a value of $\approx -600Wm^2$, an error of $\approx 30\%$ compared to $(Q2)_{conv}$. The profile of this quantity does not have the shape of $(Q2)_{conv}$ (fig 6.16(a)), drying the domain most around 800mb instead of the surface. The use of $(EE)_T$ converted to a drying rate (assuming time dependent effects are small), brings little improvement, although slightly more drying occurs at the surface.

Are any other approximations useful? Addition of $(CS)_q$ due to UD and DD gives a similar budget approximation to $(Q2)_{conv}$. However the profile shape is totally incorrect, with moistening at the surface.

Going back to $(CS)_q + (EE)_q$, are there any other simple terms which can be used to improve the representation of $(Q2)_{conv}$? Again the terms need to be simple, so that they can perhaps be calculated from simple cloud models. Detrainment terms have been used together with $(CS)_q$ and $(EE)_q$ in the past. However as seen earlier $(ED)_q$ and $(ADV)_q$ are very different from the "classical" detrainment term, causing much drying below 800mb. Addition of these terms to $(CS)_q$ and $(EE)_q$

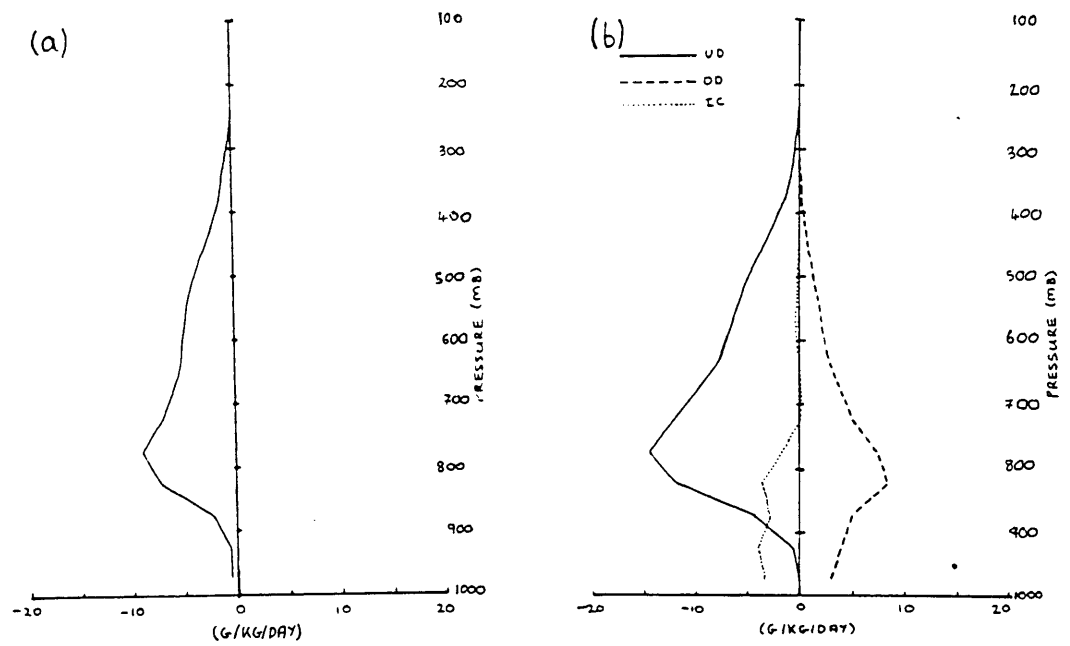


Fig 6.15(a) $(CS)_q$, all cloud (1-9 hours).
 (b) $(CS)_q$ for UD, DD, IC (1-9 hours).

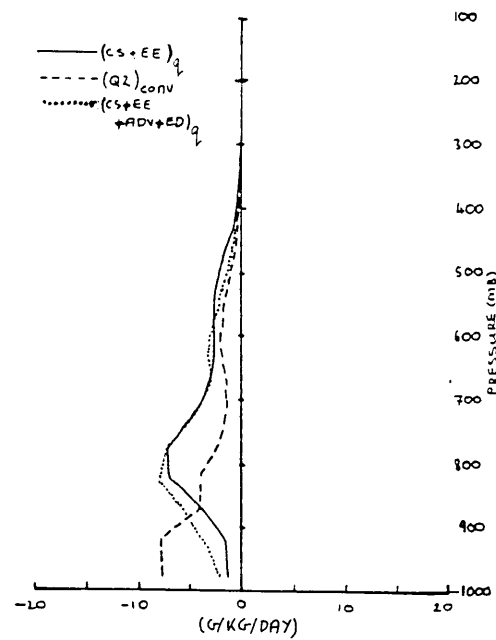


Fig 6.16 Approximations to $(Q2)_{\text{conv}}$ (1-9 hours).

(fig 6.16(b)) brings greater drying below 800mb. However little is done to the maxima at 800mb, and the drying above this. Hence $(Q2)_{conv}$ is not well represented by these simple terms. Addition of time dependent terms, turbulence, and vertical fluxes through cloud boundaries are also needed.

The over-estimation of $(Q2)_{conv}$ above 800mb, and under-estimation below this is a common complaint of some parameterization schemes. Kuo and Anthes(1984c) used a simple cloud model and $(Q2)_{conv}$ in the form of eqn(5.12). They found that the maximum drying was higher than observed (fig 6.17). They suggested the difference was due to the transport properties of shallow convection, not represented in their deep convective scheme. The Arakawa-Schubert scheme does predict the maxima of $(Q2)_{conv}$ in the lower atmosphere as observed (fig 6.18). This scheme considers a spectrum of clouds, including shallow convection. Hence Kuo and Anthes suggestion may be correct. However the simulation contains many small clouds in the lowest 200mb, and studies of the "clear" regions observed during the integration (section(4.2.2) of chapter 4) suggest that their transport properties were reasonably represented. This question of shallow convection will be returned to shortly in section(6.4).

(g) "Errors" in the parametric form of $Q2$.

As with $(Q1)_{conv}$, if all the parametric terms are added, and subtracted from $(Q2)_{conv}$ the budgets is zero. However the profile of this quantity (fig 6.19), indicates that a residual does exist, which appears more important than for the $(Q1)_{conv}$ budget ($\approx 10\text{g/kg/day}$ compared to $\approx 5\text{g/kg/day}$ for $(Q2)_{conv}$, and $\approx 2\text{K/day}$ compared to $\approx 10\text{K/day}$ for $(Q1)_{conv}$). Again the residual is not due to errors in the finite difference equations used in the analysis. As indicated earlier, the answer is arrived at by studying the approximation to the vertical eddy flux used in deriving the parametric form of $(Q1)_{conv}$ and $(Q2)_{conv}$. This is studied in the next section.

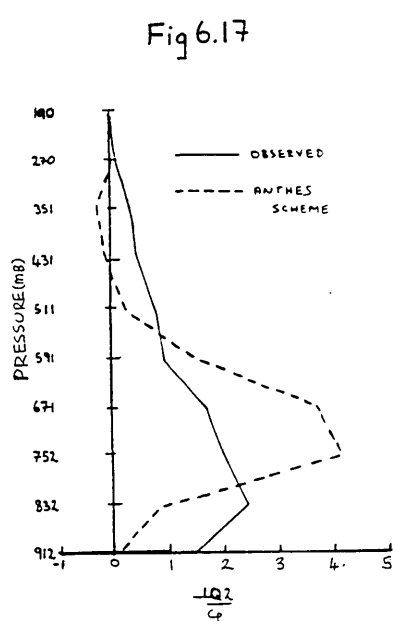


Fig 6.17 Q_2 for a Kuo type scheme compared to observations (from Kuo and Anthes(1984)).

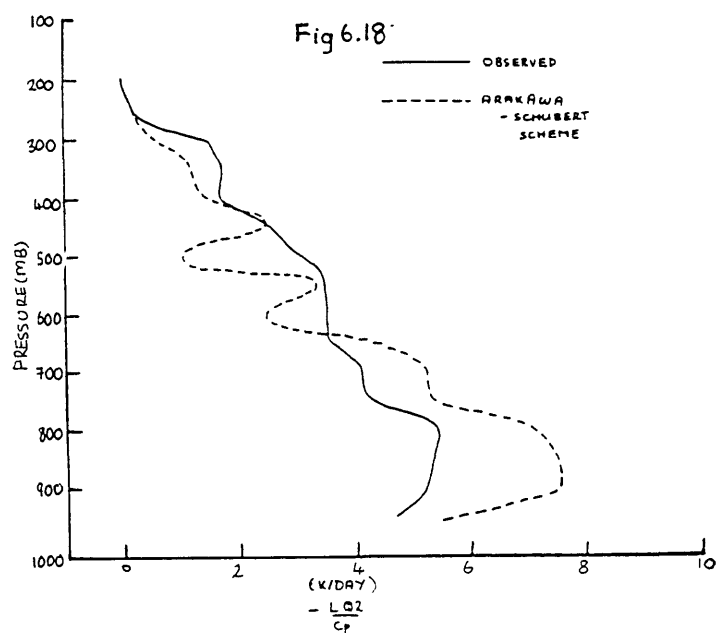


Fig 6.18 Q_2 for the Arakawa-Schubert scheme compared to observations (from Lord and Arakawa(1980)).

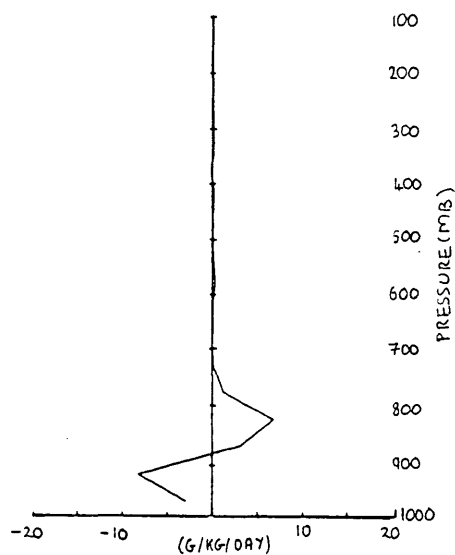


Fig 6.19 $(Q_2)_{\text{conv}} - (Q_2)_{\text{para}}$ (1-9 hours).

6.4 FURTHER STUDY OF VERTICAL EDDY FLUX DIVERGENCE.

(a) The eddy flux due to cloud.

In chapter 5 an approximation for $\overline{\omega'\phi'}$ was developed, in terms of cloud and large-scale variables;

$$\overline{\omega'\phi'} = \sigma(\overline{\omega\phi^c} - \overline{\omega^c\phi}) \quad \text{eqn(6.11)}$$

This approximation assumes that motion within the environment is uncorrelated with the thermodynamic fields, and so transports no energy.

With the numerical model eqn(6.11) can be evaluated for each component of cloud. These fluxes are presented for θ and q in fig 6.20. The shape of the profile can be roughly understood using the internal cloud properties of chapter 5, and the approximation to eqn(6.11), $\sigma\overline{\omega'(\bar{\theta}^c - \bar{\theta})}$.

Above 800mb, UD dominates the flux of temperature and moisture. Both are positive, peaking around 800mb for θ , and 750mb for q , lower due to the increase of $\partial\bar{q}/\partial p$ towards the surface. For temperature DD contributes little to the total flux above 800mb, IC being $\approx 5\%$ of the UD. DD plays a more important role in the total moisture flux, being $\approx 20\%$ of the UD in magnitude ($(\bar{q}^{\text{mo}} - \bar{q})$ cf. $\bar{q} > (\bar{\theta}^{\text{mo}} - \bar{\theta})$ cf. $\bar{\theta}$). IC flux is again $\approx 5\%$ UD magnitude. Below 800mb, DD and IC are more important. DD temperature flux is positive, peaking at 950mb. This is opposed by IC. The IC flux is approximately twice that due to UD. For moisture UD plays a more important role, exceeding the flux of IC in the top part of the layer. DD contribute less than for temperature as their is little increase in DD moisture content over that of the domain average at these pressures.

The gradient of these flux curves gives the heating and moistening rates due to eddy transports within cloud (vertical eddy flux divergence). These are presented in fig 6.21.

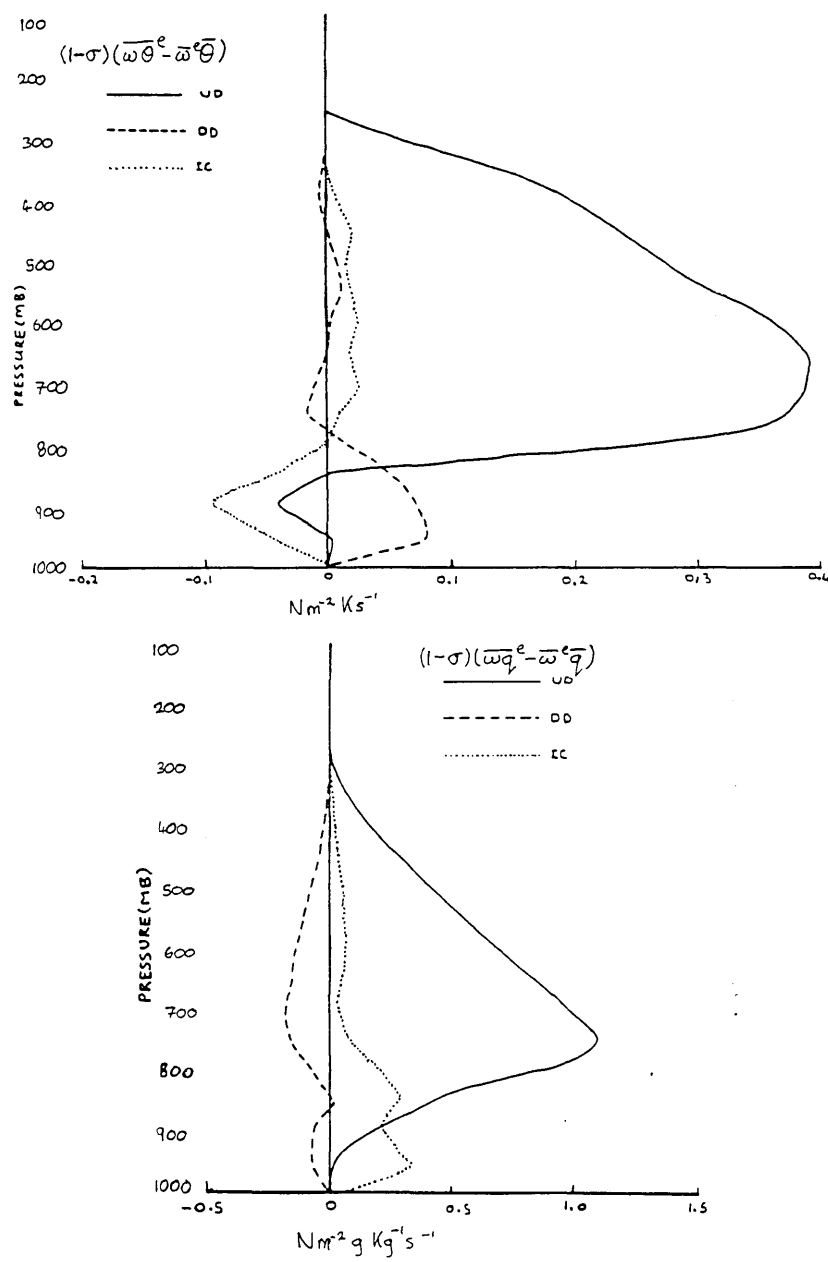


Fig 6.20 $\overline{\omega'\theta}$, $\overline{\omega'q}$ for UD, DD, IC (1-9 hours).

Fig 6.21 $-\pi \frac{\partial \bar{\omega}^c}{\partial p}$, $-\frac{\partial \bar{q}^c}{\partial p}$ for UD, DD, IC
(1-9 hours).

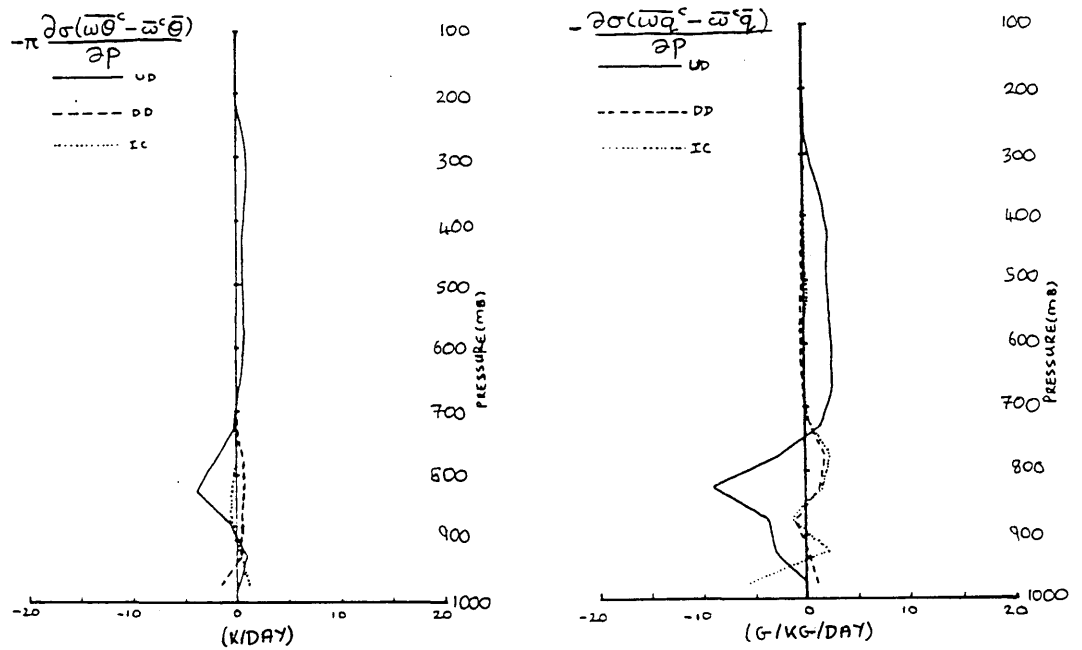
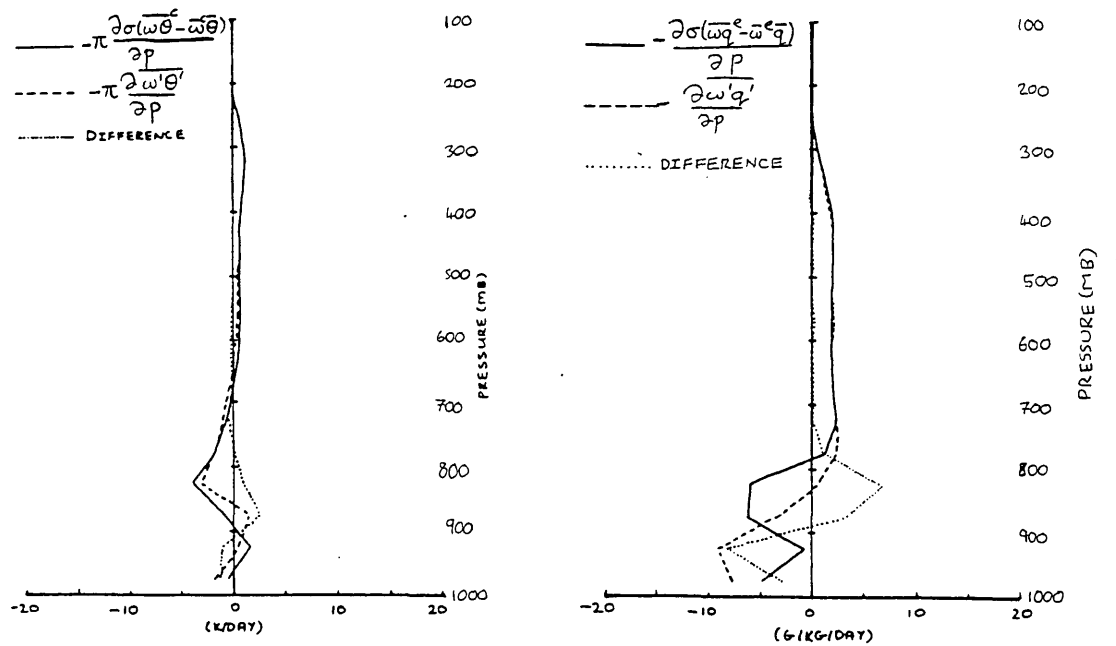


Fig 6.22 Comparison of exact profiles of $-\pi \frac{\partial \bar{\omega}^c}{\partial p}$, $-\frac{\partial \bar{q}^c}{\partial p}$ and profiles calculated by the approximate method (1-9 hours).



UD vertical eddy flux divergence basically follows the mass en/detrainment profile presented in chapter 5. Temperature and moisture are both transported upward. The heating above 700mb is small compared to $(Q1)_{conv}$, but moisture transports are larger compared to $(Q2)_{conv}$. Maximum drying of -10g/kg/day occurs at 825mb (the level of maximum entrainment), the moistening between 700 and 400mb being constant.

DD transports little energy compared to UD. Temperature transports are negligible above 700mb. Cooling occurs only in the lowest 50mb ($\approx 2\text{K/day}$), with slight warming above. The transport of moisture extends higher into the atmosphere. Drying occurs between 300 and 700mb ($\approx 0.5\text{g/kg/day}$). Below the structure is complicated, with alternative layers of moistening and drying.

Inactive cloud heat transports are small, with cooling between 700 and 900mb, and heating below. Effects above 700mb are negligible. The structure of the profile is opposite to that of DD. For moisture, detrainment layers indicated by mass en/detrainment profile are linked to layers of moistening due to eddy transports. Considerable drying occurs at 975mb ($\approx 6\text{g/kg/day}$). Again this is linked to entrainment into the deep convective updraughts.

Addition of the profile for UD, DD and IC gives the net vertical eddy transport due to convection. If the formula used for is correct, then agreement should be found with the $-\partial\overline{\omega\phi}/\partial p$ calculated exactly from the simulation, and presented in chapter 4. Comparison is made in fig 6.22. Differences are seen, mainly below 800mb, and especially for moisture. For temperature, the minimum at 725mb is over-estimated by $\approx 40\%$, and the maximum at 925mb is too low by 50mb. Insufficient cooling occurs at the lowest model level. For moisture, the drying at the surface is under-estimated by $\approx 30\%$. Too much drying occurs between 900 and 800mb, and the cross over is too high.

If the approximate and true profiles are subtracted, the residual shown in fig 6.22 results. This is the same as the "error" in the parametric profiles of $(Q1)_{conv}$ and $(Q2)_{conv}$ presented earlier. The cause of these differences must arise in the neglect of the environmental eddy flux in forming an approximate form of $\overline{\omega'\phi'}$ (eqn(6.11)).

(b) The environmental vertical eddy flux (eef).

Calculating $(1-\sigma)(\overline{\omega\phi^e} - \overline{\omega}\phi^e)$ by averaging over clear air within the domain for θ and q , gives the profiles shown in fig 6.23(a)). Taking the vertical derivative of these quantities gives the environmental vertical eddy flux divergence (fig 6.23(b)). This is in agreement with the residuals of fig 6.19. The divergence of eef warms and moistens the atmosphere between 800 and 900mb, and cools and dries below. Thus it appears that eef cannot be neglected when calculating $\overline{\omega'\phi'}$.

What is the cause of this component of $\overline{\omega'\phi'}$? The response of cloud environment to convection is gravity wave motion, and both upward and downward motion are observed (fig 6.24). Areas of ascent and decent within clear air are roughly equal, as expected for gravity waves. The net motion of the environment is downward. This motion is dry adiabatic, and so warming and moistening is caused. This can be seen when $\overline{\theta}$ and $\overline{q}$ are compared to their domain averaged values (fig 6.25). Differences are small confirming that $\overline{\phi} = \overline{\phi^e}$ for most purposes is a good approximation.

Fig 6.25 also shows the thermodynamic fields associated with upward and downward motion within the environment. Above 825mb the correlations are as expected. However for temperature below this, they are reversed. Moisture deviations are the same in this region as above but increased in magnitude. The reversal of temperature correlations occurs because of the cloud forcing. Not all of the perturbations are strong enough to force saturated ascent. The above result indicates that a weak dry circulation is induced in the lower layers of the domain, within cloud environment. The warming and cooling of the air causes air to ascend and descend. Hence upward moving air is warmer than $\overline{\theta^e}$, and downward moving air cooler. This is illustrated in fig 6.26.

Such a dry circulation is observed over the clear regions of the domain. The motions are small, at maximum $20\text{--}30\text{cm s}^{-1}$. They do not penetrate far above the cloud forcing region (950–850mb). This is indicated in the temperature structure of these motions (fig 6.25). Above 850mb, the correlations are as expected for unforced ascent and decent, while below this they are reversed as explained above. Pressure time plots of the vertical divergence of eef and average

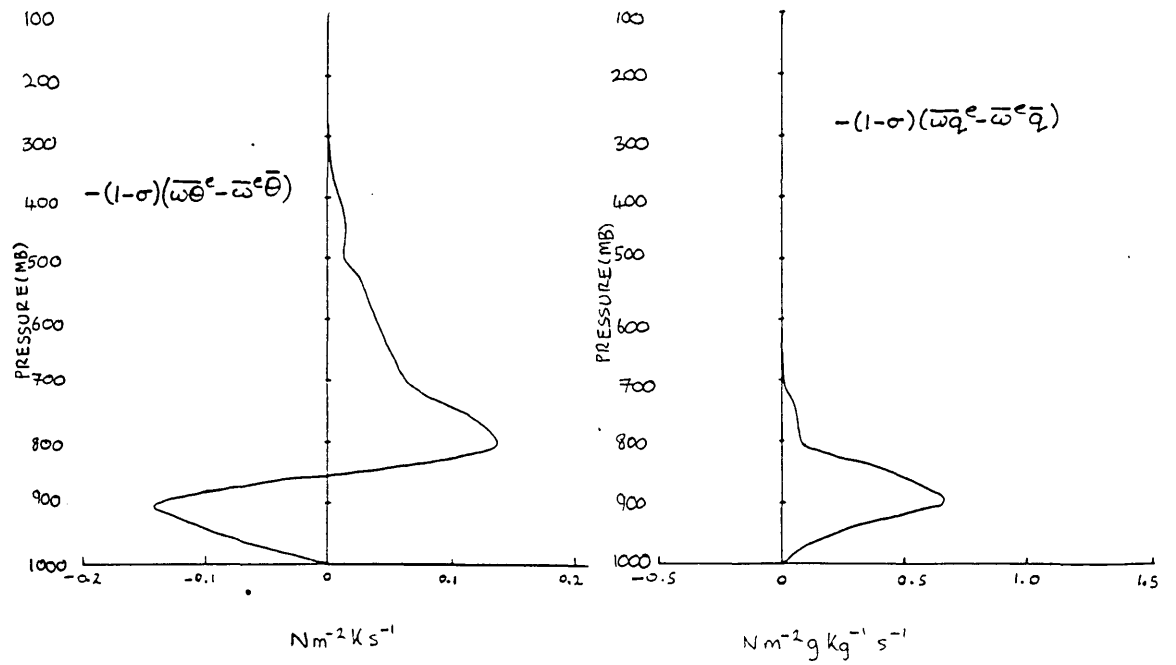


Fig 6.23(a) Environmental eddy flux $((1-\sigma)(\overline{w\Theta^e} - \overline{w^e\Theta}))$ for Θ and q (1-9 hours).

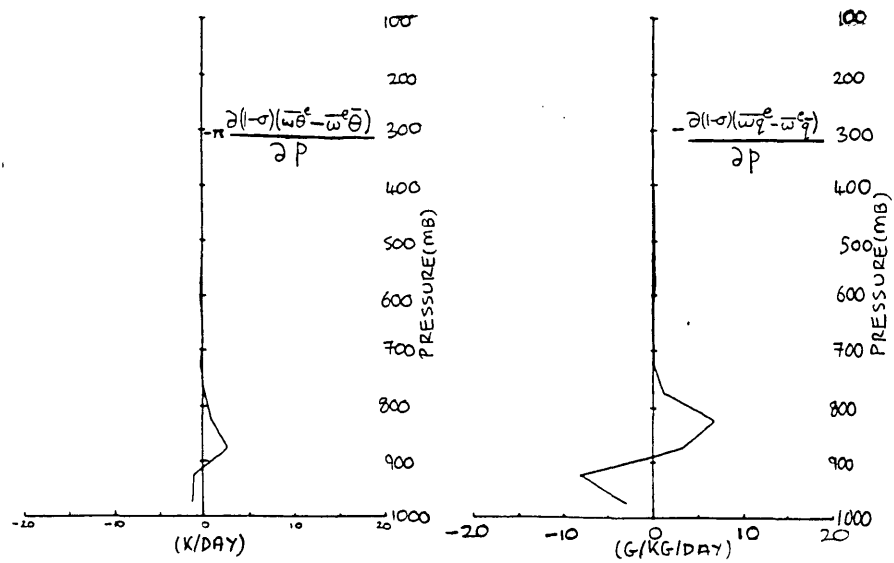


Fig 6.23(b) $-\partial(1-\sigma)(\overline{w\Theta^e} - \overline{w^e\Theta})/\partial p$ for Θ and q (1-9 hours).

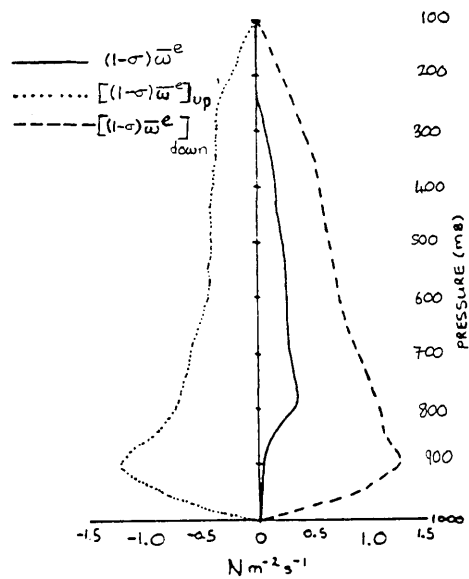


Fig 6.24 Vertical motion within cloud environment (1-9 hours).

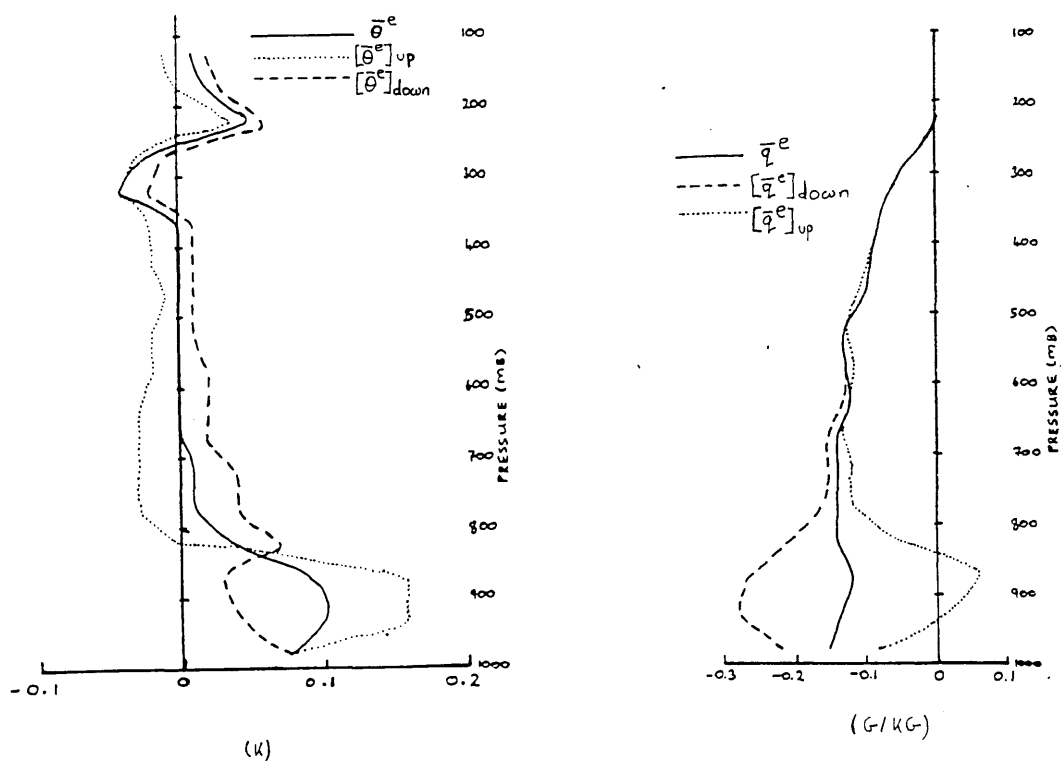


Fig 6.25 Environmental thermodynamic profiles for upward and downward moving air (1-9 hours).

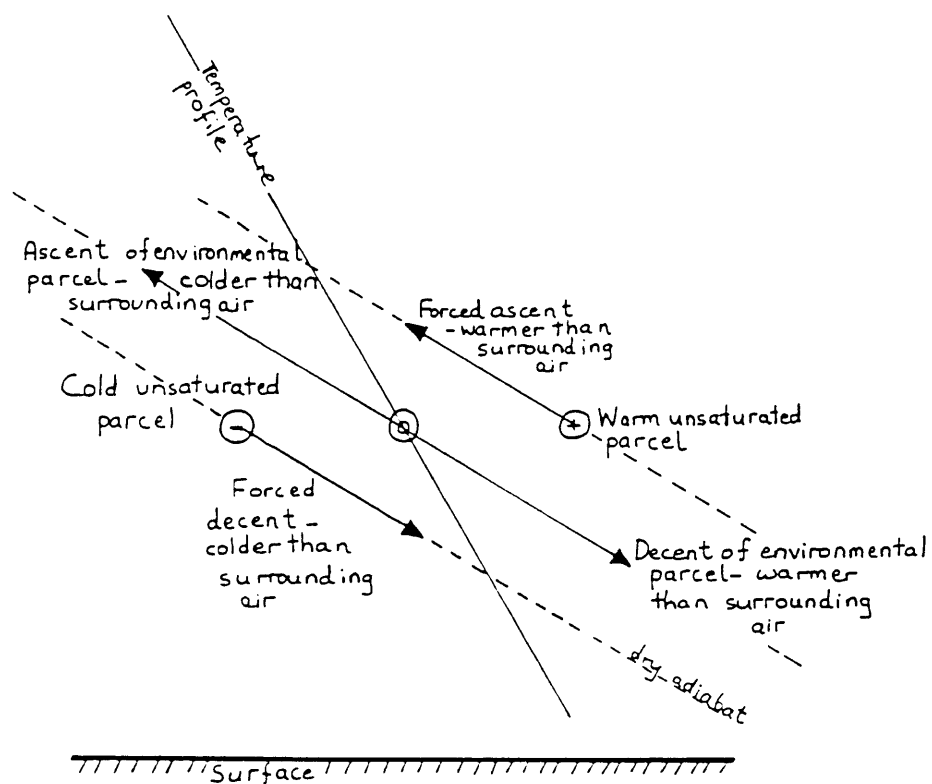


Fig 6.26 Representation of correlated motions on a $T-\theta$ gram.

cloud forcing across the domain show that the two processes are related; maxima of the eddy flux occurring when strong forcing exists (fig 6.27).

The environmental mass flux profile also provide evidence that the circulations do not penetrate above 850mb. Above this the ratio between the net subsidence, $(1-\sigma)\bar{\omega}^e$, and the upward, $((1-\sigma)\bar{\omega}^e)_{up}$, and downward, $((1-\sigma)\bar{\omega}^e)_{down}$, mass flux are observed to be,

$$\left. \begin{aligned} \frac{((1-\sigma)\bar{\omega}^e)_{up}}{(1-\sigma)\bar{\omega}^e} &\approx -2 \\ \frac{((1-\sigma)\bar{\omega}^e)_{down}}{(1-\sigma)\bar{\omega}^e} &\approx 3 \end{aligned} \right\} \quad \text{eqn(6.12)}$$

Hence the ratio of down to upward motion within the environment above 850mb is 3/2. Below this the above ratios increase towards the surface, as indicated in table 6.3.

It is difficult to see how such small net motion can give rise to large oscillations, when above 850mb much larger cloud motions induce smaller ones. Hence additional forcing would seem to be required, in this case, the cloud forcing. The net forcing between one and eight hours is approximately zero. The balance between upward and downward motion below 850mb reflects this. The residual left would seem to be the subsidence induced by the cloud motions upon the environment.

If the cloud forcing were not present, what would be the effect of the convective induced motion? Above 800mb fig 6.23(b) indicates that the effect is negligible. Can the contribution below 800mb be estimated? Here the moisture flux is considered. This is much larger than that of temperature, and so any contribution due to cloud induced motion will be larger.

From the analysis of the model, it is found that,

$$[(1-\sigma)\bar{\omega}\bar{\phi}^e]_{up} \approx [(1-\sigma)\bar{\omega}^e]_{up}[\bar{\phi}^e]_{up}$$

and similarly for downward motion.

Assume that the correlations below 850mb can be estimated (q'). Hence $(1-\sigma)\bar{\omega}\bar{q}^e$ can be approximated by,

$$\begin{aligned} (1-\sigma)\bar{\omega}\bar{q}^e &\approx [(1-\sigma)\bar{\omega}^e]_{up}[\bar{q}^e]_{up} + [(1-\sigma)\bar{\omega}^e]_{down}[\bar{q}^e]_{down} \\ &\approx -2(1-\sigma)\bar{\omega}^e(\bar{q}^e - q') + 3(1-\sigma)\bar{\omega}^e(\bar{q}^e - q') \\ &\approx (1-\sigma)\bar{\omega}^e\bar{q}^e - 5(1-\sigma)\bar{\omega}^eq' \end{aligned}$$

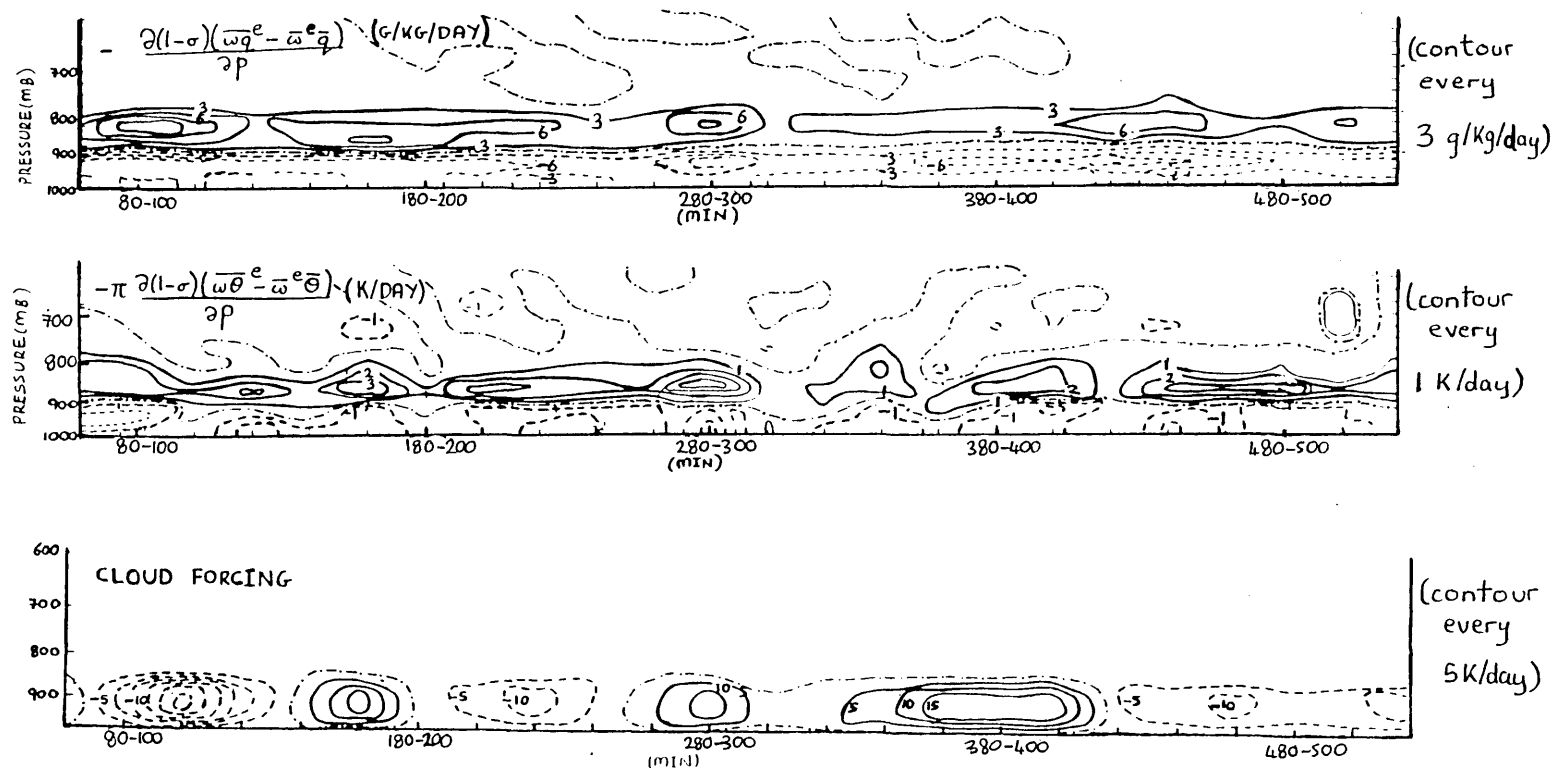


Fig 6.27 Comparison of cloud forcing and $-\partial(1-\sigma)(\bar{\omega}\phi^e - \bar{\omega}^e \bar{\phi})/\partial p$ for θ and q (20 minute averages).

Pressure (mb)	$(1-\sigma)\bar{\omega}^e$	$\frac{[(1-\sigma)\bar{\omega}^e]_{up}}{(1-\sigma)\bar{\omega}^e}$	$\frac{[(1-\sigma)\bar{\omega}^e]_{down}}{(1-\sigma)\bar{\omega}^e}$	$\frac{[(1-\sigma)\bar{\omega}^e]_{up}}{[(1-\sigma)\bar{\omega}^e]_{down}}$
850	0.149	-7	8	1.1
900	0.032	-39	40	1.0
950	0.040	-21	22	1.1

Table 6.3 Ratio of upward and downward environmental motion to the net environmental motion (averaged between 1 and 9 hours).

Pressure (mb)	$(1-\sigma)\bar{\omega}^e$ (Nm s)	$(\bar{q}^e - \bar{q})$ (g/kg)	q' (g/kg)	$5(1-\sigma)\bar{\omega}^e$ (Nm s)	$(eef)_{conv}$ (Nm s gkg)	$-\frac{\partial(eef)}{\partial p}$ (g/kg/day)
700	0.27	-0.14	0.02	1.37	-0.07	0.2
725						
750	0.32	-0.14	0.02	1.58	-0.08	
775						0.0
800	0.32	-0.15	0.02	1.61	-0.08	-0.9
825						
850	0.15	-0.13	0.01	0.75	-0.03	
875						-0.4
900	0.03	-0.12	0.00	0.16	-0.01	0.0
925						
950	0.04	-0.14	0.00	0.19	-0.01	
975						-0.2
1000	0.00	-	-	-	0.00	

Table 6.4 Estimation of eef due to deep convection (averaged between 1 and 9 hours).

using the ratios of eqn(6.12), and assuming that the correlations are equal in the upward and downward moving air.

Thus eef can be estimated by,

$$(1-\sigma)\overline{\omega q^e} - (1-\sigma)\overline{\omega^e q} \simeq (1-\sigma)\overline{\omega^e}(\overline{q^e} - \overline{q}) - 5(1-\sigma)\overline{\omega^e q'} [= (eef)_{conv}]$$

Assuming that the correlation in q at 825mb is unaffected by the cloud forcing, and that correlations below this due to the convective circulation are linearly proportional to this, then q' may be estimated. The calculation of eef, due cloud forced motion, is presented in table 6.4.

Values of $-\partial(eef)_{conv}/\partial p$ for moisture are much smaller than those analysed from the simulation, showing that cloud forcing dominates. The profile shape is also different. Hence convection plays little part in contributing to this term.

Is this redistribution of energy by cloud forcing realistic in any manner? The forcing was introduced to represent (in a crude manner) the effect of small scale convection observed in the lower atmosphere. Deep convection is observed to grow from fields of these shallow clouds. The vertical divergence of eef for both temperature and moisture are similar to the eddy transports of such clouds. Fig 6.28 shows idealised profiles of these quantities based upon the modelling work of Sommeria and Le Mone(1978), using BOMEX data. Magnitudes within the meso-scale model are also similar to those of the BOMEX clouds.

However in the model these shallow circulations are dry, whereas in the real atmosphere such transports are due to saturated ascent. Hence transport of liquid water also occurs, and this can effect the net energy source due to shallow convection. Assuming that the liquid water transports are similar to those of water vapour, the liquid water is removed from below 900mb (where it has condensed), and redistributed above this (where it evaporates). This causes a heat source (moisture sink) below 900mb, and sink (source) above. Hence transports of liquid water oppose those of sensible heat. Over the BOMEX area, Holland and Rasmusson(1973) found that shallow convection warmed the atmosphere below 900mb, and cooled it above, opposite to

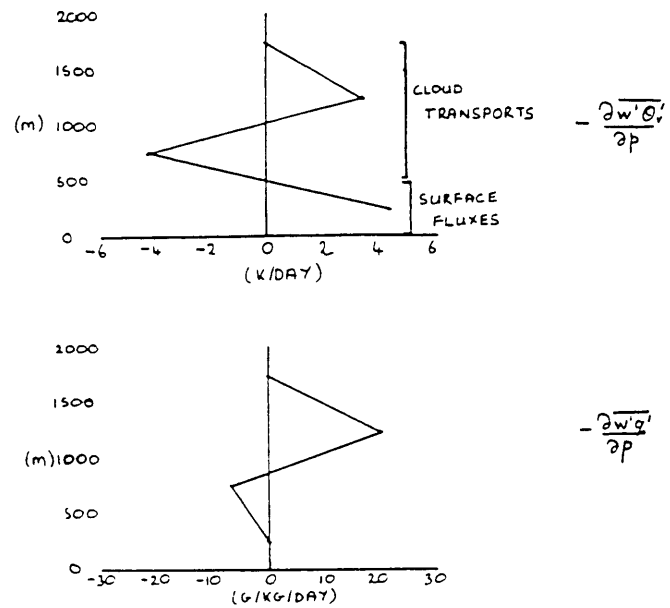


Fig 6.28 Transport of θ and q by shallow convection (from Sommeria and LeMone(1978)).

the mechanical transport of sensible heat. For the moisture field, transports of water in both vapour and liquid phases act together causing drying below 900mb, and moistening above. Work by Soong and Ogura(1980) indicate that moist processes contribute $\approx 25\%$ of the total effect.

Hence for the moisture field, the effects of cloud forcing simulate shallow convection reasonably well. However for temperature, the effects is opposite to that observed, due to the neglect of moist processes. This has little effect on the $(Q1)_{conv}$ profile as the contribution of eef divergence is small.

(c) Further comments on eddy transports in "clear" regions (shallow clouds).

The identification of cloud forcing contributing a large part of $-\partial\overline{\omega\theta}/\partial p$ in the lowest 200mb, must bring reinterpretation of the eddy transports diagnosed for "clear" areas of the domain, and presented in section(4.2.2). During the first 6 hours of the simulation, little convection occurs in these regions, while during the last 3 hours, cloud with tops up to 600mb develop. Before 6 hours, the eddy transport properties of the region are similar to those of shallow convection (see 1 to 3 hour profile of fig 4.24), and it was suggested that the model had sufficient vertical resolution to represent such features correctly. However these eddy transports, for both θ and q , are of a similar shape to the environmental eddy flux divergence presented above. Hence it would appear that these features are due to cloud forcing not shallow clouds, ie a dry circulation not a moist one. This argument is further strengthened by the observation that $L\overline{Q}/c_p$ for the clear regions during this period is negative (fig 4.22, 1 to 3 hour average). This is due to evaporation of liquid water transported into the "clear" regions from the "anvil" areas.

The large number of clouds with tops in the lowest 200mb of the domain are negligible when the model energy budget is considered. As stated previously, these exist on the grid scale. Their transport properties are not as expected for shallow convection, much of the motion within them being downward (at a few tens of centimeters per second). This suggests that although upward motion must cause the initial saturation, subsidence due to large clouds dominates, causing

the clouds to eventually dissipate.

(d) Consequences for parameterization.

Adding the vertical divergence of eef to $(CS)_q$, $(EE)_q$, $(ADV)_q$, and $(ED)_q$ brings large changes to the approximation of $(Q2)_{conv}$ below 800mb. More drying occurs at the surface, while the maxima at 800mb is much reduced (fig 6.29). Addition of this term also improves the approximation of $(Q1)_{conv}$. Hence the effects of shallow convection are important in predicting the correct shape of $(Q2)_{conv}$. Consideration of the effects of deep convection is insufficient. Such a situation was suggested by Kuo and Anthes(1984c) as reported earlier. The Arakawa-Schubert scheme would seem to represent these effects sufficiently well, as the predicted $(Q2)_{conv}$ peaks at the correct level (fig 6.18).

However the addition of shallow convective effects is still insufficient to bring agreement between $(Q2)_{conv}$ and its approximate parametric form. Large differences are still apparent at all levels of the domain. Closer agreement is only achieved with the addition of $(BAL)_q$ and $(TURB)_q$ terms (fig 6.29). Fine adjustment is achieved through the addition of $(FORC)_q$. It has already been stated that such terms cannot be calculated simply. Hence it would seem that no simple approximation to $(Q2)_{conv}$ based upon eqn(5.20), for use in parameterization schemes exists.

6.5 FURTHER STUDY OF 20 MINUTE AVERAGES.

These have been studied in chapters 4 and 5 with relevance to parameterization within meso-scale models. Here the study is briefly continued, and the question of how $(Q1)$ can be represented on this short time scale considered.

(a) Eddy transports on short time scales.

In chapter 4, $-\partial\overline{\omega\theta}/\partial p$ for Θ and q were seen to evolve with time as cloud grows and decays. In the last chapter the evolution of internal cloud structure associated with these features was presented.

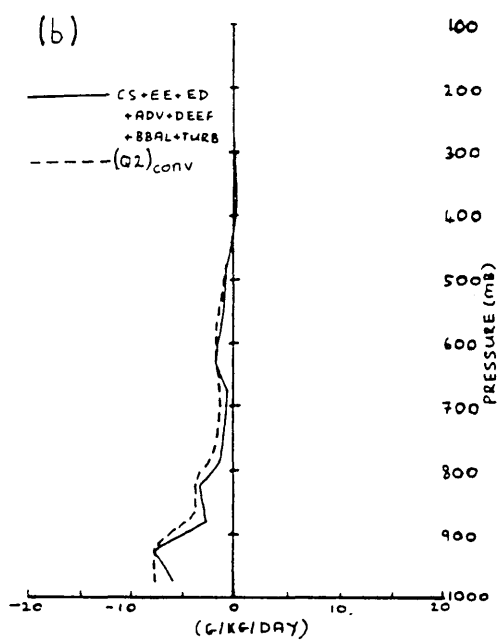
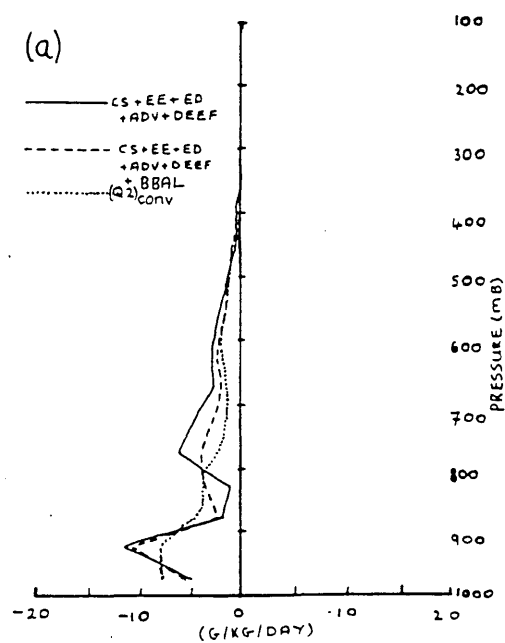


Fig 6.29(a) Addition of the vertical divergence of eef to the approximation to $(Q2)$ (fig 6.16) (1-9 hours).

(b) Addition of further terms to the approximation to $(Q2)$ (1-9 hours).

UD show a characteristic evolution, whereas DD show roughly constant structure once formed. The association of this change in UD structure with those of eddy transports is clearly seen in fig 6.30.

Here pressure-time plots of $-\partial\sigma(\overline{\omega\theta^c}-\overline{\omega^c\theta})/\partial p$ are shown for UD and DD. Comparison with fig 4.19 shows that the majority of changes in $-\partial\overline{\omega\theta}/\partial p$ are associated with the UD, although some changes due to DD are seen in the lower domain. Other differences are due to inactive cloud, and cloud forcing.

The evolution can be explained by using a simple evolution of cloud structure based upon the results of chapter 5. This is illustrated in fig 6.31. Cloud evolution is characterised by a mature and decay phase, and eddy transports calculated using the approximation to eqn(6.11), $\sigma\overline{\omega^c}(\overline{\theta^c}-\overline{\theta})$. The deepening of the cooling layer due to eddy transports is associated with the decay of upward mass flux in the lower levels, and the maxima moving upward (hence increasing the depth of the entrainment layer).

Similar arguments apply to moisture. Here the eddy flux divergence due to cloud forcing (shallow convection) is more important in the lower levels. However this feature appears reasonably constant throughout the simulation (fig 6.27).

Hence if some simple evolution in UD structure could be developed, then the change in cloud eddy transports properties could be taken into account in meso-scale models.

(b) Representation of Q1 on short time scales.

Here the simple representation of $(Q1)_{conv}$ on short time scales is considered. $(Q2)_{conv}$ is not considered as no simple approximation to it has been found on longer temporal time scales.

Many terms in the parametric $(Q1)_{conv}$ budget are noisy on the 20 minute time scale, eg $(BAL)_T$ terms. However these terms are never locally large (maxima $\approx 5K/day$), and when added together some degree of cancellation occurs. $(ED)_T$ and $(ADV)_T$ follow a well ordered evolution, the cooling due to $(ED)_T$ and warming due to $(ADV)_T$ in the lower atmosphere, increasing in depth as the UD entrainment layer

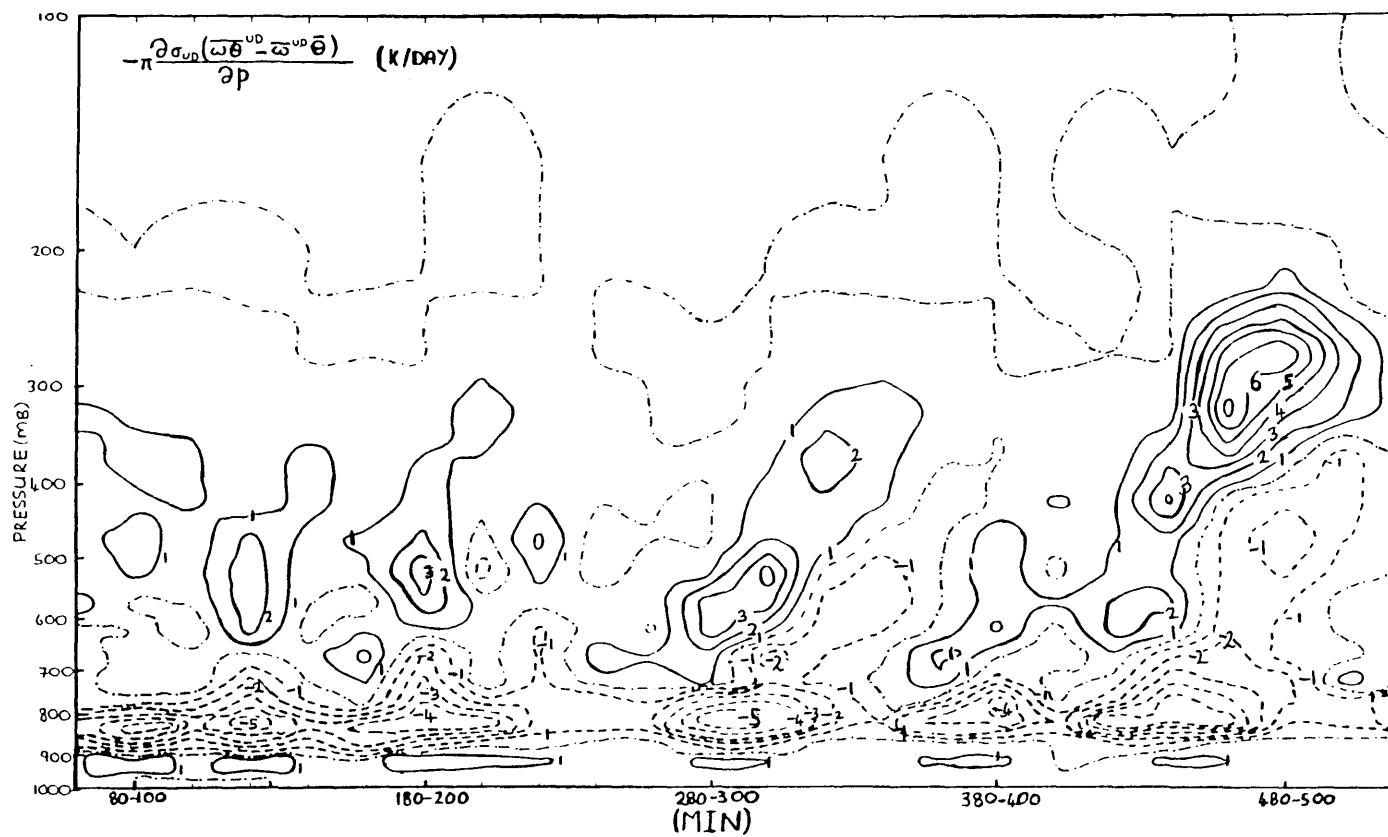


Fig 6.30 $-\pi \frac{\partial \bar{\omega} \bar{\theta}}{\partial p}$ for UD and DD (20 minute averages).

(contour every 1K/day)

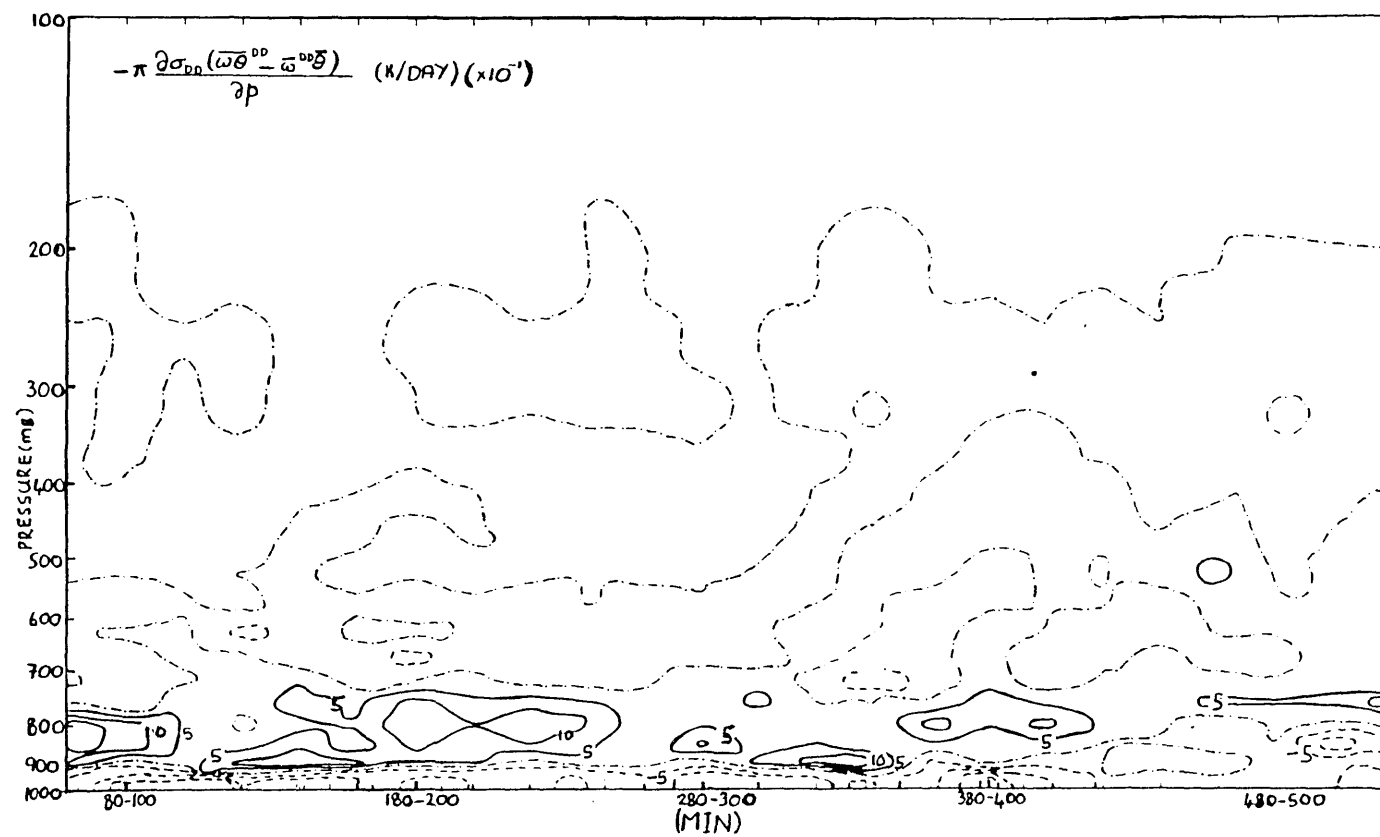


Fig 6.30
(contour every 0.5 K/day)

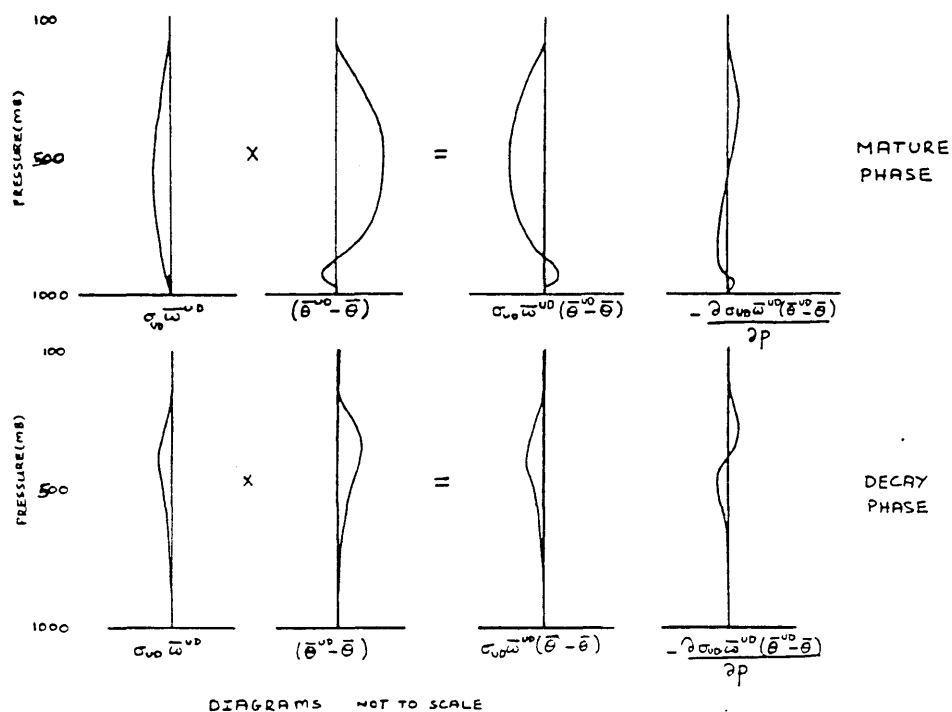


Fig 6.31 Simplified evolution of UD and DD to explain change in $-\pi \partial \bar{\omega} / \partial p$.

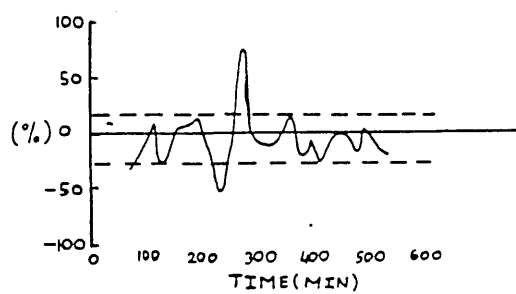


Fig 6.32 Percentage difference of $\langle CS+EE \rangle_r$ from $(Q1)_{conv}$ (20 minute averages).

deepens, during cloud decay.

Can $(Q1)_{conv}$ be approximated by $(CS+EE)_T$, as over longer time periods? Fig 6.32 shows the quantity,

$$\frac{(Q1)_{conv} - \left(\sigma \bar{\omega}^c \frac{\partial \bar{\theta}}{\partial p} + (1-\sigma) \frac{L \bar{Q}}{c_p \pi} + (\bar{\theta} - \theta_b) \frac{\partial \sigma}{\partial t} \right)}{(Q1)_{conv}} \times 100\%$$

budgeted over the domain for twenty minute time periods. The averaged error is $\approx -6\%$, with a standard deviation of $\approx 20\%$. Most of the errors are below 25%, and so most points lie within one standard deviation of the mean. Two points indicate a large errors; one is associated with strong cloud forcing occurring within cloudy air (220-240 minute), the other with little convection (260-280 minutes).

Comparison of profiles on these time scales is made in fig 6.33. $(CS)_T$ due to all cloud looks like $(Q1)_{conv}$, without the addition of $(EE)_T$. DD events are seen (eg. 220-240, 320-340 minutes), although they are under-estimated (no DD cooling was seen in the eight hour average of $(CS)_T$). However there is a lack of cooling below some convective events (eg. 280-300 minutes). Addition of $(EE)_T$ brings improvement. $(EE)_T$ is noisy and time dependent effects are apparent (eg. 180-200 minutes) below 800mb (the region where they were important in $(EE)_q$). There is some indication that DD cooling is associated with low level increases in $(EE)_T$ (eg. 80-100, and 180-200 minutes), but this is not always the case (eg. 400-420 minutes). This low level cooling can reach -9K/day. Above 800mb, maxima in cooling (up to -5K/day) are associated with periods of convective heat release.

$(CS)_T + (EE)_T$ gives a better representation of $(Q1)_{conv}$. DD and decay coolings are slightly over-estimated. More cooling is seen under convective heating events as indicated by $(Q1)_{conv}$. The maxima of $(Q1)_{conv}$ are slightly under-estimated. However generally agreement is quite good.

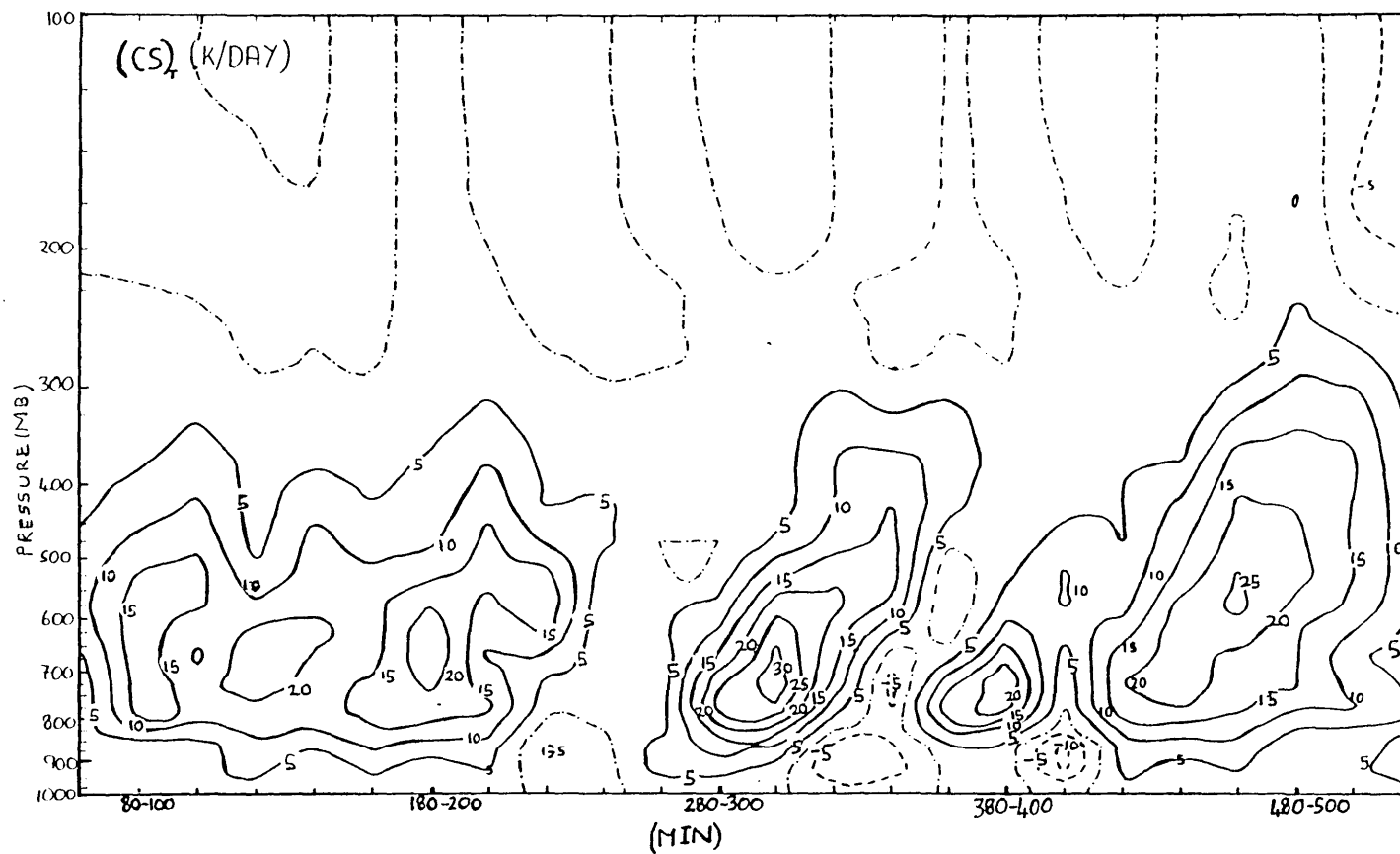
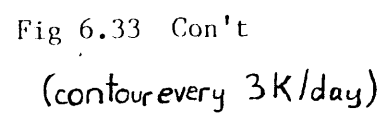


Fig 6.33 P-T plots of $(CS)_T$, $(EE)_T$, $(CS+EE)_T$, and $(Q1)_{conv}$
(20 minute averages).

(contour every 5 K/day)



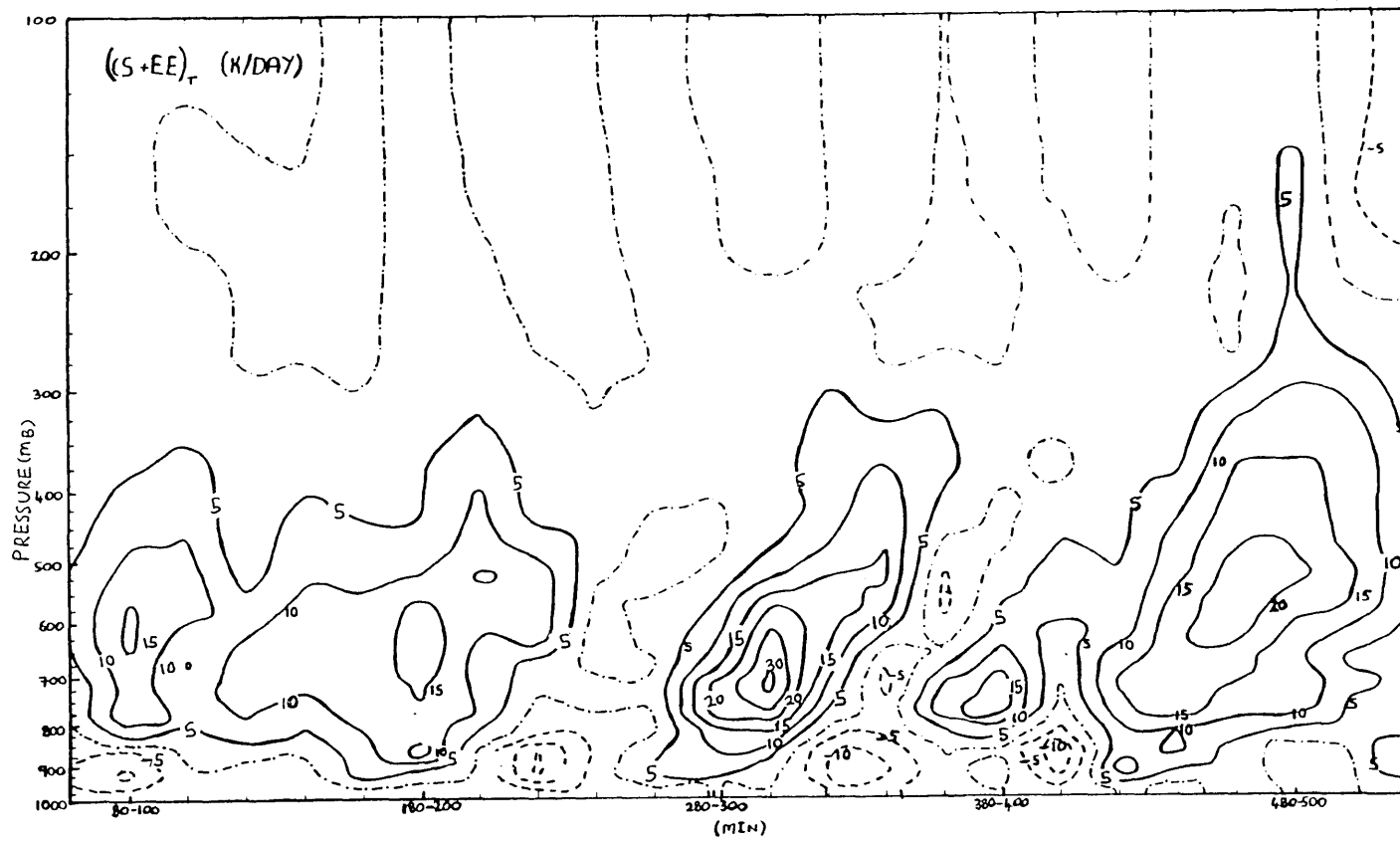


Fig 6.33 con't
 (contour every 5 K/day)

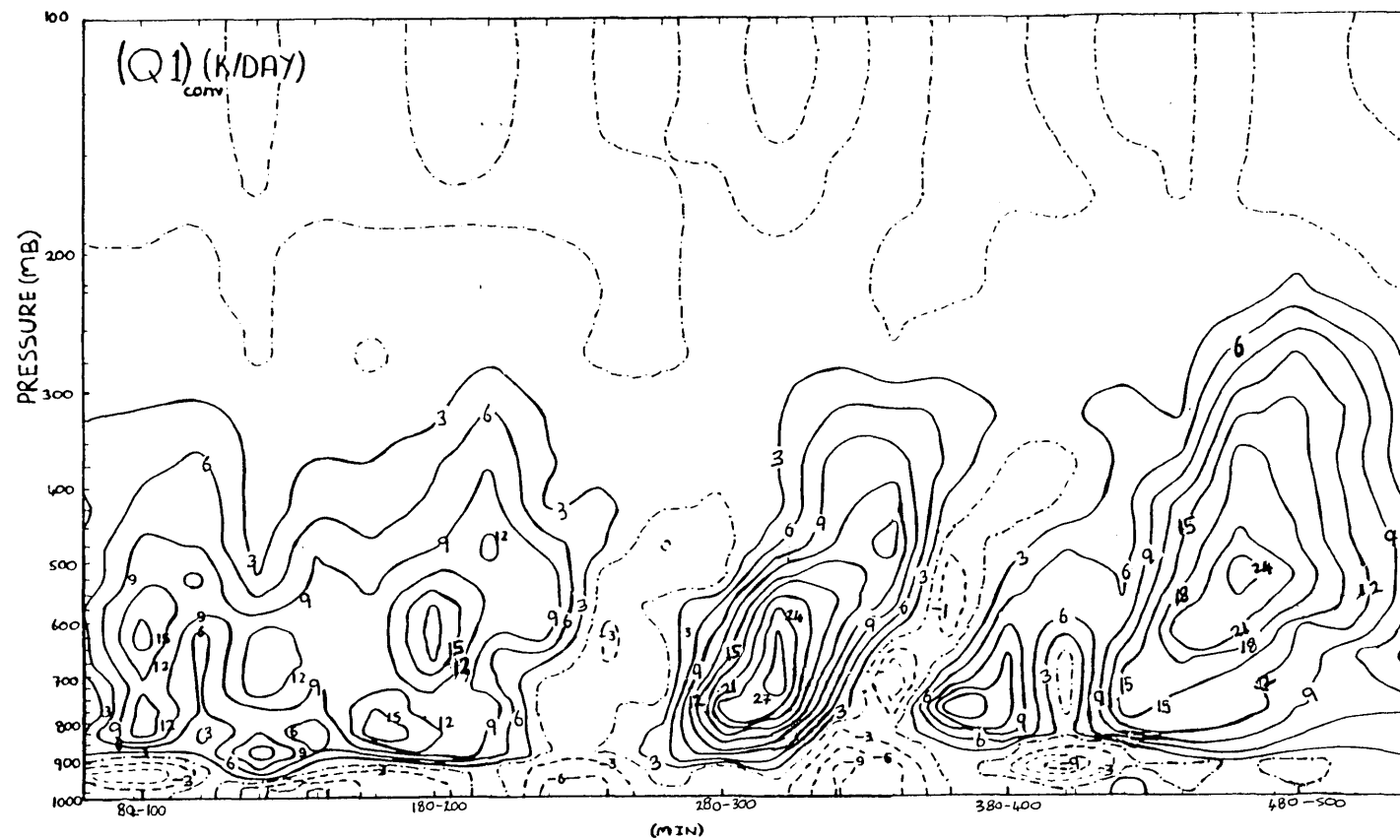


Fig 6.33 con't

(contour every 3 K/day)

6.6 SUMMARY.

In this chapter part of the parameterization problem has been addressed. The parametric budgets of $(Q1)_{conv}$ and $(Q2)_{conv}$ have been analysed using an eight hour averaging period, and approximations useful for parameterization studied. $(Q1)_{conv}$ is well represented by compensating subsidence (due to all cloud), and moist processes within the environment. Other terms in the budget balance, indicating that approximations used by previous workers are correct, although for the wrong reasons.

Although the budgeted value of $(Q2)_{conv}$ is reasonably represented by compensating subsidence, and environmental moist processes, the profile is not. This peaks at 800mb, whereas $(Q2)_{conv}$ peaks in the lowest 100mb of the domain. Addition of time dependent and more complex terms brings good agreement above 800mb, but not below. Study of $\overline{\omega'\phi'}$ for Θ and q indicate that this discrepancy is due to a contribution to this term from the cloud environment. This is due to dry thermals induced by the cloud forcing. The vertical divergence of the "environmental eddy flux" acts in a similar manner to the transports of small scale convection (at least for moisture), and brings the maxima of $(Q2)_{conv}$ down to the surface. Thus small scale convection would seem to be important in specifying the $(Q2)_{conv}$ profile, deep convection alone is insufficient.

Studies were briefly presented using 20 minute averages. Time variation of $-\partial\overline{\omega'\Theta'}/\partial p$ is seen to be caused mainly by UD evolution with time. The representation of $(Q1)_{conv}$ by subsidence and moist environmental process, although not as accurate as over eight hours is reasonable on this shorter time scale.

In the following chapter, the relationship between convection and large-scale structure and processes is studied, with reference to the closure problem of parameterization schemes.

CHAPTER 7

THE RELATIONSHIP BETWEEN DEEP CONVECTION AND LARGE-SCALE PROCESSES.

In this chapter, the relationship between Q1 and Q2 and large-scale structure and processes is studied with regard to the closure problem of convective parameterization schemes. The Kuo and Arakawa-Schubert schemes are concentrated upon, mainly using model data averaged over eight hours, although brief comments are made concerning short time scales. Several simulations with varying initial profiles and large-scale forcings are described. The storage of water vapour within the atmosphere before the onset of deep convection is discussed. Comments are also made concerning CAPE and the new adjustment scheme of Betts and Miller.

7.1 THE KUO CLOSURE SCHEME-ACTIVE CONVECTION CASE (SIM B).

The basis of the Kuo scheme is that the convection rains out a fraction (1-b) of the moisture convergence, M_t , the rest going to moisten the atmosphere. Using the eight hour average of the meso-scale model it is observed that,

$$\langle c_p \pi(Q1)_{\text{conv}} \rangle = 0.78 M_t$$

$$\text{where } M_t = L \left\langle \left(\frac{\partial q}{\partial t} \right)_{L_s} + \left(\frac{\partial q}{\partial t} \right) \right\rangle$$

Hence convection rains out a large portion of the applied moisture forcing. This agrees well with the observational results of the tropical atmosphere. Using GATE phase III data, Krishnamurti et al(1980) suggested that convection rains out $0.8M_t$. Thus the result confirms that the meso-scale model is able to represent the balance between convection and large-scale forcing in a realistic manner, and also gives support for the Kuo closure hypothesis. Many Kuo schemes do not consider the effects of convective scale downdraughts. $(Q1)_{conv}$ is assumed to be that associated with the energy release in updraughts. However in section(6.2) this was seen to be a very poor representation of $(Q1)_{conv}$. However using the meso-scale model, a Kuo type closure scheme can be specified including downdraughts, inactive cloud and evaporation with cloud environment. This is:-

$$\langle c_p \pi(Q1)_{conv, UD} \rangle = 1.63M_t$$

$$\langle c_p \pi(Q1)_{conv, DD} \rangle = -0.77M_t$$

$$\langle c_p \pi(Q1)_{conv, IC} \rangle = 0.22M_t$$

$$\langle c_p \pi(Q1)_{conv, EE} \rangle = -0.29M_t$$

It has been seen that much of the effect of IC occurs within the lowest 200mb of the domain, associated with the occurrence of deep convection. Hence it would seem reasonable to include IC with UD to give a modified closure relation,

$$\langle c_p \pi(Q1)_{conv, UD} \rangle = 1.85M_t$$

If the vertical distribution of $(Q1)_{conv}$ by UD, DD and EE can be determined then the above relations form the basis for a Kuo type scheme. Previously $(Q1)_{conv}$ has been shown to be well approximated by $(CS)_T$ plus $(EE)_T$, and so it is more reasonable to define the closure relations using $(CS)_T$ instead of $(Q1)_{conv}$. Again combining UD and IC, this gives:-

$$\langle c_p \pi(CS)_T \rangle_{UD} = 1.79M_t$$

$$\langle c_p \pi(CS)_T \rangle_{DD} = -0.76M_t$$

This type of closure scheme is of little use on short time scales. Chapter 4 showed that $(Q1)_{conv}$ undergoes rapid variations on the twenty minute time scale, whereas the moisture forcing over the simulation is roughly constant (table 4.1). Hence there is no simple definition of the constant (1-b).

The scheme does seem reasonable for large ensembles of cloud. However there is problem. As seen in chapter 2 convection does not always balance the large-scale moisture convergence, as is the case of category 3 (20-30hours) of the composite easterly wave. At times the atmosphere stores significant amounts of water vapour. Hence the simple Kuo scheme demonstrated above breaks down. This question will be returned to later.

7.2 ARAKAWA-SCHUBERT CLOSURE-STRONG CONVECTIVE CASE (SIM B).

This scheme was described briefly in chapter 5. Here the closure hypothesis is presented in more detail. Arakawa and Schubert (A-S) considered a spectrum of clouds which comprised of updraughts only. Each cloud type is specified by the entrainment rate, λ . Closure is achieved by assuming a balance exists between the generation of moist convective instability, and it's destruction by convection.

Consider the vertical kinetic energy budget of a cloud type λ ,

$$\frac{dK(\lambda)}{dt} = A(\lambda) - D(\lambda) \quad \text{eqn(7.1)}$$

where $A(\lambda)$ is the generation rate of
cloud scale KE,

and $D(\lambda)$ is the dissipation rate of
cloud scale KE.

Eqn(7.1) is normalised with respect to the vertical mass flux at some reference level, chosen by A-S to be the top of the sub-cloud layer, $m_B(\lambda)$. Hence,

$$\frac{dK(\lambda)}{dt} = \left(\frac{A(\lambda)}{m_B(\lambda)} - \frac{D(\lambda)}{m_B(\lambda)} \right) m_B(\lambda) = (A_n(\lambda) - D_n(\lambda)) m_B(\lambda)$$

Let τ_{dis} be the dissipation time scale, usually less than the cloud lifetime ie. $\sim 10^2$ to 10^3 seconds, and τ be the time scale on which eqn(7.1) applies. This is usually assumed to be the time scale due to large-scale processes ie. $\sim 10^5$ seconds. Hence an order of magnitude approximation to the kinetic budget is,

$$\frac{K(\lambda)}{\tau} \sim A_N(\lambda) m_b(\lambda) - \frac{K(\lambda)}{\tau_{dis}}$$

As $\tau_{dis} \ll \tau$, the lhs may be neglected compared to the rhs provided that $m_b(\lambda)$ is positive. Hence,

$$A_N(\lambda) \simeq D_N(\lambda) \quad \text{eqn(7.2)}$$

A-S chose $A_N(\lambda)$ to be the generation of KE by buoyancy (termed the cloud work function), defined as

$$A_N(\lambda) = \int_{z_{bottom}}^{z_{top}} \frac{g \gamma(z, \lambda)}{\bar{T}_v(z)} [\bar{T}_{ve}(z, \lambda) - \bar{T}_v(z)] dz \quad \text{eqn(7.3)}$$

where $\bar{T}_{ve}(z, \lambda)$, and $\bar{T}_v(z)$ are the cloud and large-scale virtual potential temperatures (including consideration of liquid water), and $\gamma(z, \lambda)$ is the normalised convective mass flux. $A_N(z, \lambda)$ is the generation of KE by buoyancy per unit cloud base mass flux.

$D(\lambda)$ is assumed to be proportional to $m_b(\lambda)$, and so is a constant. Hence on differentiating eqn(7.2) wrt time,

$$\frac{\partial A_N(\lambda)}{\partial t} \simeq 0$$

The change of $A(\lambda)$ wrt time is split into two parts:- one due to cloud effects upon the large-scale, the other due to large-scale processes, ie.

$$\frac{\partial A_N(\lambda)}{\partial t} = \left[\frac{\partial A_N(\lambda)}{\partial t} \right]_{cum} + \left[\frac{\partial A_N(\lambda)}{\partial t} \right]_{LS} \quad \text{eqn(7.4)}$$

By assuming $\partial A_N(\lambda)/\partial t \approx 0$, a closure relationship for $m_b(\lambda)$ is obtained, details of which are not presented here.

To check if the closure assumption is valid, Lord and Arakawa(1980) used observational data to estimate $[\partial A_N(\lambda)/\partial t]_{LS}$ and $\partial A_N(\lambda)/\partial t$. If

$$\frac{\partial A_N(\lambda)}{\partial t} \ll \left[\frac{\partial A_N(\lambda)}{\partial t} \right]_{LS}$$

then the equilibrium assumption on A_{λ} is reasonable.

Similar calculations can be made using the simulated cloud with the meso-scale model. However as stated previously it is difficult to specify the effects due to individual cloud types. Hence the KE budget over the entire ensemble is used. In this case the generation of KE due to buoyancy is:-

$$A = \int_{1000\text{mb}}^{100\text{mb}} \left(\frac{\sigma \bar{\omega} \bar{\Theta}_v^c - \sigma \bar{\omega}^c \bar{\Theta}_v}{\bar{\Theta}_v} \right) \frac{dp}{\rho g} \quad \text{eqn(7.5)}$$

Again this is defined for all clouds within the ensemble. Because this is calculated averaged over cloud lifetime, it may be a more reasonable estimate than that calculated from observational data and analytical cloud models.

Both UD and DD are present within the simulation, both of which create KE. Thus it seems reasonable to normalise the rhs of the KE budget wrt

$$|m_B| = \frac{1}{g} |\sigma_{\nu} \bar{\omega}^{\nu}|_B + \frac{1}{g} |\sigma_{\nu} \bar{\omega}^{\nu}|_B (= \rho |W_B|)$$

at some reference level, specified by the sub-script B. The model contains no specific representation of the sub-cloud layer, and so the top of this layer cannot be used as a reference level as in the original scheme. For the purposes of this study p_B is defined at 825mb, because UD and DD exist at this level throughout the simulation.

Briefly reviewing the analysis, the KE budget for all of the ensemble is,

$$\frac{dK}{dt} = (A_N - D_N) |m_B|$$

Following the same line of argument, as $|m_B| > 0$,

$$A_N \approx D_N$$

Again D_N is assumed to be proportional to $m_B(\lambda)$, and so

$$\frac{\partial A_N}{\partial t} \approx 0$$

and

$$\frac{\partial A_N}{\partial t} = \left[\frac{\partial A_N}{\partial t} \right]_{\text{cum}} + \left[\frac{\partial A_N}{\partial t} \right]_{\text{LS}}$$

The inclusion of DD effects into the closure argument does not effect the derivation of a practical closure assumption. Test of closure is carried out as previously. An expression for $[\partial A_N / \partial t]_{\text{LS}}$ is formulated

below. Differentiating eqn(7.5) wrt time, and collecting cloud and large-scale terms,

$$\frac{\partial A_N}{\partial t} = - \int_{1000 \text{ mb}}^{100 \text{ mb}} \frac{1}{|m_B|} \left(\frac{1}{\bar{\theta}_v} \frac{\partial \sigma \bar{\omega} \bar{\theta}_{vL}^c}{\partial t} - \frac{\partial \sigma \bar{\omega}^c}{\partial t} \right) \frac{dp}{p g} + \int_{1000 \text{ mb}}^{100 \text{ mb}} \frac{1}{|m_B|} \frac{\sigma \bar{\omega} \bar{\theta}_{vL}^c}{\bar{\theta}_v^2} \frac{\partial \bar{\theta}_v}{\partial t} \frac{dp}{p g} \quad \text{eqn(7.6)}$$

Now for the model,

$$\frac{\partial \bar{\theta}}{\partial t} = - \frac{\partial u \bar{\theta}}{\partial x} - \frac{\partial \bar{\omega} \bar{\theta}}{\partial p} + (Q1)_{\text{conv}} + \bar{S}_\theta + [F_\theta]$$

$$\frac{\partial \bar{q}}{\partial t} = - \frac{\partial u \bar{q}}{\partial x} - \frac{\partial \bar{\omega} \bar{q}}{\partial p} + (Q2)_{\text{conv}} + \bar{S}_q + [F_q]$$

where $[F_\theta]$ and $[F_q]$ are the applied forcing.

The mean advective terms are zero for a cyclic numerical model, and so combining to form $\partial \bar{\theta}_v / \partial t$

$$\frac{\partial \bar{\theta}_v}{\partial t} = ((Q1)_{\text{conv}} + 0.61(Q2)_{\text{conv}}) + (\bar{S}_\theta + 0.61\bar{S}_q) + ([F_\theta] + 0.61[F_q])$$

Substituting into eqn(7.6) above and collecting cloud and large-scale terms gives

$$\begin{aligned} \frac{\partial A_N}{\partial t} &= \left[\frac{\partial A_N}{\partial t} \right]_{\text{cum}} + \int_{1000 \text{ mb}}^{100 \text{ mb}} \frac{1}{|m_B|} \frac{\sigma \bar{\omega} \bar{\theta}_{vL}^c}{\bar{\theta}_v^2} ([F_\theta] + 0.61[F_q]) \frac{dp}{p g} \\ &= \left[\frac{\partial A_N}{\partial t} \right]_{\text{cum}} + \left[\frac{\partial A_N}{\partial t} \right]_{\text{LS}} \end{aligned}$$

Hence in evaluating the integral on the rhs of the last equation and $\partial A_N / \partial t$ direct from eqn(7.5), the A-S closure hypothesis can be tested using the meso-scale model.

Fig 7.1 shows $[\partial A_N / \partial t]_{\text{LS}}$ and $\partial A_N / \partial t$ for all cloudy air, averaged over twenty minute intervals between 1 and 9 hours of sim B. $[\partial A_N / \partial t]_{\text{LS}}$ is always positive $\partial A_N / \partial t$ oscillating around this. Averaging over the entire of this period gives,

$$\overline{\left[\frac{\partial A_N}{\partial t} \right]_{\text{LS}}}^{1 \rightarrow 9 \text{ hr}} = -390 \text{ J/kg/day} \quad \overline{\left[\frac{\partial A_N}{\partial t} \right]}^{1 \rightarrow 9 \text{ hr}} = -60 \text{ J/kg/day}$$

Hence averaging over many cloud lifetimes, the A-S closure hypothesis would seem to be valid.

In calculating A_N above $\bar{\theta}_{vL}$ has been used, water loading having a great effect upon cloud buoyancy. Calculations using $\bar{\theta}_v$ and $\bar{\theta}$ in the buoyancy term indicate that A-S closure is not valid if the water loading is neglected, $\partial A_N / \partial t$ being of similar magnitude to $[\partial A_N / \partial t]_{\text{LS}}$.

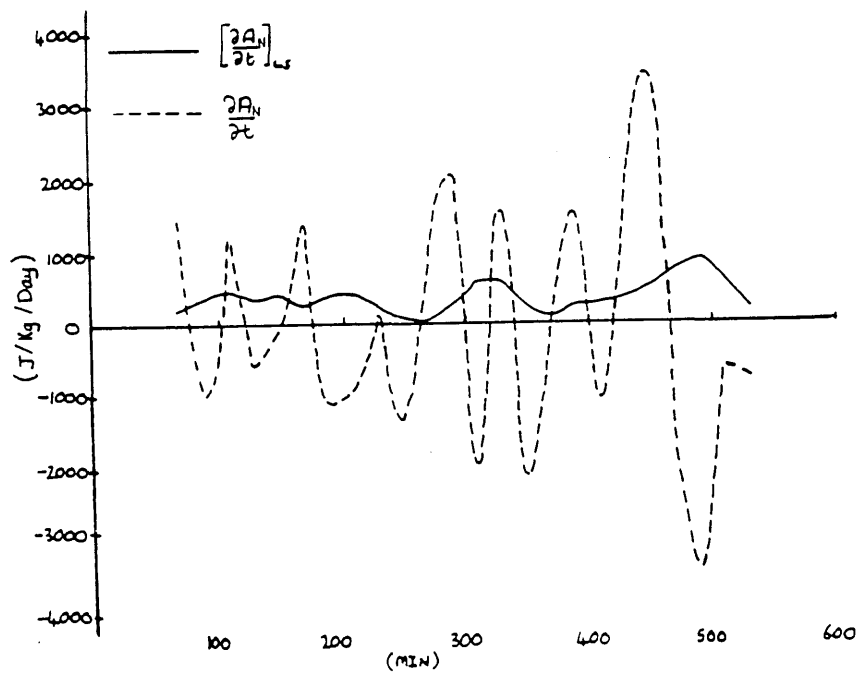


Fig 7.1 $[\frac{\partial A_N}{\partial t}]_{L5}$ and $\frac{\partial A_N}{\partial t}$ for sim B, 20 minute averages.

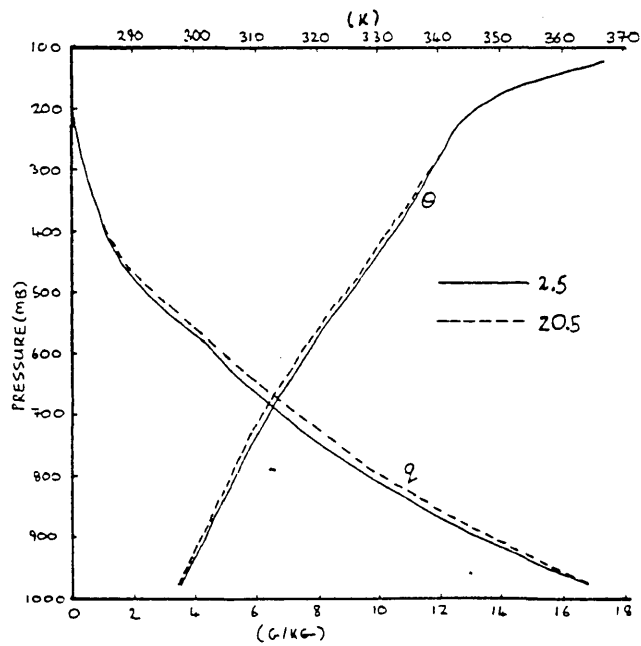


Fig 7.2 Comparison of 2.5 and 20.5 hour thermodynamic profiles of the GATE composite easterly wave.

The meso-scale model may be used to calculate $\partial A/\partial t$ due to upward moving air only (but in this case normalised to the modulus of upward and downward moving air). This gives

$$\overline{\left[\frac{\partial A_u}{\partial t}\right]}_{1-9hrs} = -860 \text{ J/Kg/day} \quad \overline{\left[\frac{\partial A_d}{\partial t}\right]}_{1-9hrs} = -70 \text{ J/Kg/day}$$

Hence A-S closure seems valid for a system of cloud containing UD only. $[\partial A/\partial t]_{1-9hrs}$ for upward and downward moving air have a ratio of 1.8. This appears to be a statement of the ratio of upward and downward moving mass fluxes within cloud, which when integrated over the entire cloud layer have a ratio of 1.9. It is noted that at 800mb, the ratio of mass fluxes is 1.8. Such a fact would be useful in deriving a closure assumption to obtain $|m_0|$ in the A-S scheme, for a cloud ensemble containing both UD and DD.

Is the A-S scheme valid on shorter time-scales than eight hours, or for systems with fewer clouds? Averaging over one hour periods brings little better results, $\partial A/\partial t$ still oscillating around $[\partial A/\partial t]_{1-9hrs}$. Within the simulation 3 periods of convective activity can be seen; 1-4, 4-7, and 7-9 hours. These are separated by periods when little convective heating occurs, the atmosphere being destabilized by large-scale forcing after being stabilized by convection. Integrating A over these periods gives the results shown in table 7.1.

Over the first two events, A-S closure seems reasonable. The third event would appear to be incomplete by the end of the simulation (the atmosphere not having sufficient time to destabilize again). Hence the A-S closure hypothesis would seem to be valid on the scale of 3 hours, if this includes the period of convective activity, and also the period in which the atmosphere "recovers" from the stabilization due to convection.

7.3 SIMULATION OF WEAKER CONVECTIVE CASES (STORAGE PROBLEM).

The model has been seen to provide a reasonable representation of category 3 (20-30 hours of the TPRR data set) of an easterly wave, in which deep convection is occurring (sim B). Is it able to represent other wave categories, and the response of convection to weaker

Event (hours)	$[\partial A/\partial t]_{L_s}$ (J/kg/day)	$\partial A/\partial t$ (J/kg/day)	$\frac{[\partial A/\partial t]_{L_s}}{\partial A/\partial t} \times 100\%$
1-4	353	-49	14%
4-7	294	-29	9%
7-9	589	-274	46%

Table 7.1 Comparison of $[\partial A/\partial t]_{L_s}$ and $\partial A/\partial t$ calculated from the model, over separate convective events.

Pressure (mb)	Profiles	
	2.5 hours (J/kg)	20.5 hours (J/kg)
975	1110	1260
925	310	550
875	-30	124

Table 7.2 CAPE for the lowest three model layers of the 2.5 and 20.5 profiles.

Forcing	Temperature (Wm^{-2})	Moisture (Wm^{-2})
Large-scale	-288	322
Boundary	5	89
layer		
Radiation	-128	-

Table 7.3 Forcings applied to sim G and H averaged between 1 and 9 hours of the simulation (based upon observed values for 2-10 hours of the TPRR data set)

[Note:-radiational cooling was kept constant between sim B, G and H.]

large-scale forcing? From the TPRR data this appears to be non-linear, convection not always raining out a large proportion of the moisture convergence. This is especially seen in category 1 of the easterly wave (fig2.5(a) of chapter 2). Here the sum of the large-scale and boundary layer moisture forcing is 400Wm^{-2} , whereas the observed rainfall is 100Wm^{-2} . Hence $3/4$ of the forcing goes into moistening the atmosphere.

Fig 7.2 compares the 2.5 hour temperature and moisture profile to that of 20.5 hours used in sim B (the 2.5 hour profile has been filtered in a similar manner to remove any wave structure; this does not greatly effect it's shape). The 2.5 hour profile is slightly drier and warmer than the 20.5 hour profile, by up to -0.5g/kg , and 0.8K at maximum. The CAPE of the lowest three model layers is presented in table 7.2. The 2.5 hour profile is less unstable to moist convective instability, and the depth of the unstable layer is less.

Two experiments were carried out. In the first (sim G), the model was initialized using the 20.5 hour profile, the large-scale and boundary layer forcing being provided by the 2 to 10 hour observed values (comparison between the 2-10 and 20-30 hour forcing is made in table 7.3). The second used the 2.5 hour profile and the 2-10 hour forcing (sim H).

(a) 20.5 hour profile, 2-10 forcing (sim G).

The simulation was continued for nine hours, the convection being initialized using the same random number forcing as in sim B. The budget value of $(Q1)_{\text{conv}}$ over the last eight hours of the simulation is 300Wm^{-2} . The moisture forcing during this time was 411Wm^{-2} . Hence the convection rains out 72% of the moisture convergence, a similar figure as seen in sim B. The contributions to $(Q1)_{\text{conv}}$ from UD, DD, IC, and EE in terms of M_t are

$$\langle c_p \pi (Q1)_{\text{conv UD}} \rangle = 1.73 M_t$$

$$\langle c_p \pi (Q1)_{\text{conv DD}} \rangle = -0.81 M_t$$

$$\langle c_p \pi (Q1)_{\text{conv IC}} \rangle = 0.14 M_t$$

$$\langle c_p \pi (Q1)_{\text{conv EE}} \rangle = -0.33 M_t$$

These are also similar to the values obtained for sim B in section(7.1). The noticeable exception is the contribution due to IC which is reduced by $\approx 35\%$ from sim B. This seems closely related to the reduction in boundary layer moisture forcing, $\approx 30\%$ between 20-30 hours and 2-10 hours. IC mainly has an effect in the lowest 200mb. An experiment was carried out using the 2-10 large-scale forcing with the 20-30 hour boundary layer forcing. Here IC contributed $\approx 20\%$ to $(Q1)_{conv}$, similar to sim B. Hence convection in the lowest layers of the model seem sensitive to this parameter.

Profiles of $(Q1)_{conv}$ and $(Q2)_{conv}$ for sim G are shown in fig 7.3, compared with those of sim B. $(Q1)_{conv}$ has an identical shape to that of sim B, although magnitudes are less in response to the reduction in large-scale forcing. $(Q2)_{conv}$ is also of similar shape to sim B, although again a reduction in magnitude is seen. Above 800mb much of these changes are due to reductions in condensational effects, the eddy flux divergence being of similar magnitude (fig 7.4). However below this, especially for moisture, the reduction in transport properties is important.

Although the net convective effect is reduced, individual convective storms appear to be little different from those of sim B, when internal cloud structure is considered. Fig 7.5 shows updraught buoyancy, and vertical velocity for sim G and sim B. Little difference is seen, and indeed both quantities are slightly larger in sim G between 700 and 500mb.

Difference is seen when UD area is considered. Sim G shows approximately a 30% reduction compared to sim B (fig 7.6). The cloud height spectrum also shows less deep convection (fig 7.7). This should lead to less overlap in cloud lifetimes, and hence more noise in the budget of $\langle C_p \pi (Q1)_{conv} \rangle$ with time, compared to sim B (fig 7.8). This is seen during the first four hours of the simulation. Sim B exhibits three peaks during this period, while sim G only two, with a deeper minimum between. The event at 300 minutes is lost in sim G, while both simulations have large energy release around 450 minutes. In sim G the event at 300 minutes is suppressed, the large-scale forcing not having sufficient time to de-stabilize the domain, after the initial convection. Approximately twice as long is required, reflecting that the applied forcing is approximately half that of

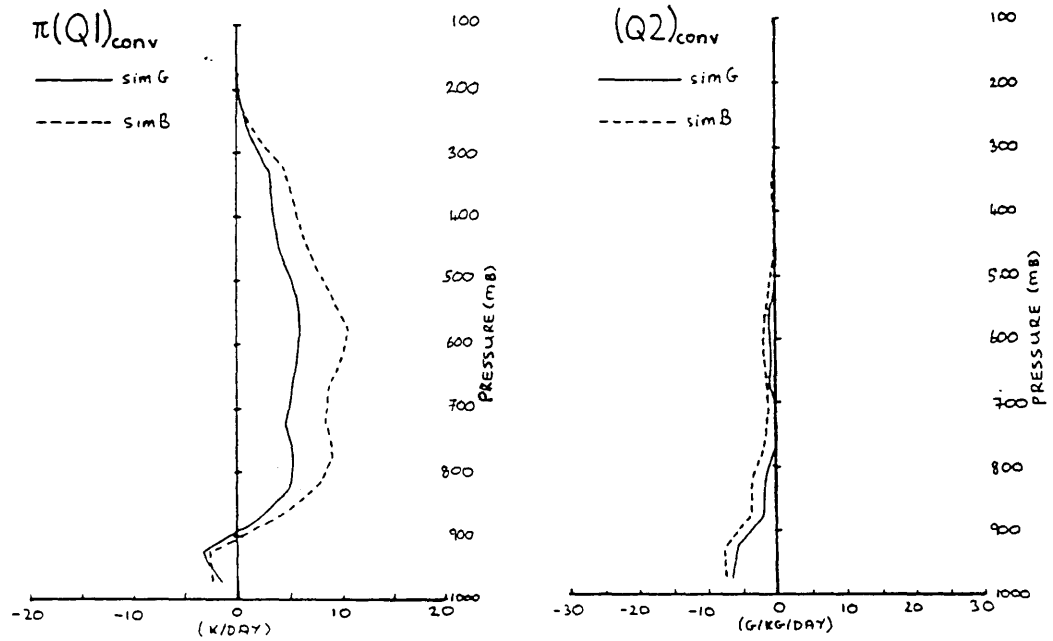


Fig 7.3 Comparison of $(Q1)_{conv}$ and $(Q2)_{conv}$ for sim B and G, averaged between 1 and 9 hours.

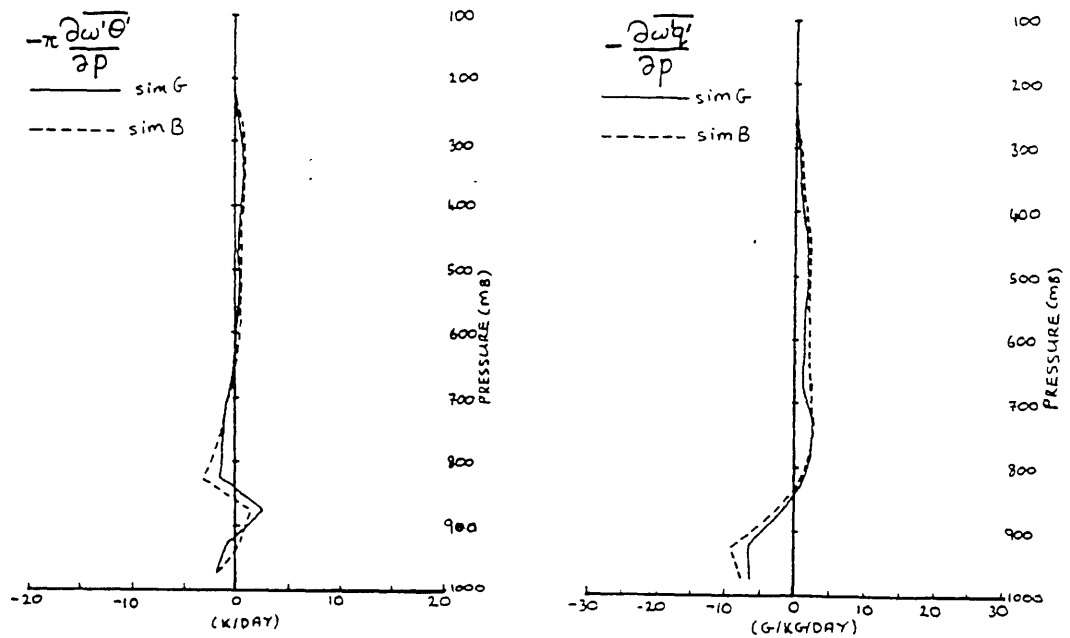


Fig 7.4 Comparison of vertical eddy transports for θ and q , for sim B and G, averaged between 1 and 9 hours.

Fig 7.5 Comparison of $g \frac{\overline{\theta'_v}}{\overline{\theta'_v}}$, and $\overline{\omega}^{\text{ob}}$ for sim B and G, averaged between 1 and 9 hours.

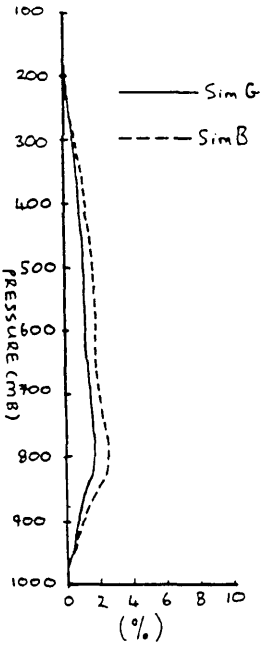
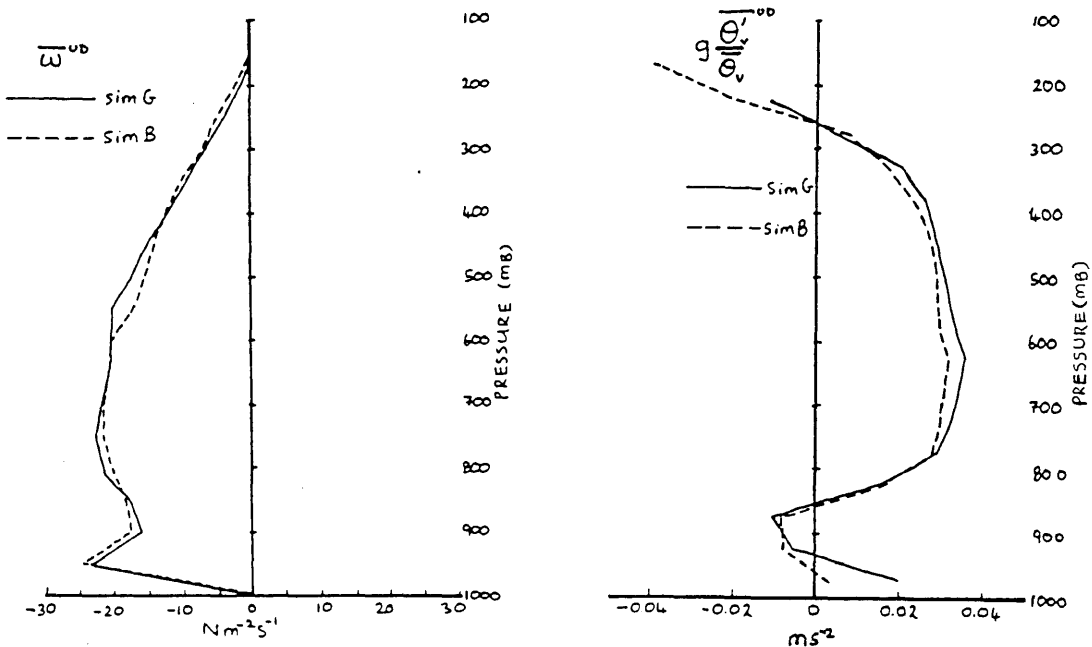


Fig 7.6

Comparison of σ for sim B and G, averaged between 1 and 9 hours.

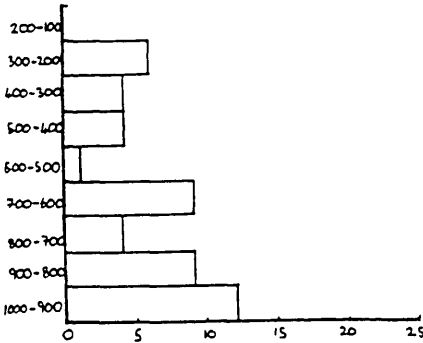


Fig 7.7

Cloud height spectrum for sim G.

sim B.

Hence if the initial profile is kept fixed, the meso-scale model indicates that the simulated convection reacts linearly to a change in large-scale forcing. However the experiment was an attempt to simulate wave category one of the TPRR data set. In this it has failed. The simulated $(Q1)_{conv}$ is approximately 3 times larger than observations suggest. A reasonable low level shear did exist during this period ($10ms^{-1}$ in 200mb). It is known that strong shear can inhibit convection (Kuo(1965)), and so an experiment was carried out identical to sim G, except the model was also initialized with the 2.5 hour shear, in attempt to reduce the activity of the simulated convection. This did not occur, and the eight hour average of $(Q1)_{conv}$ was little different from that of sim G, although the onset of convection was delayed by for 3 hours due to the shear. (The effect of shear, and momentum transports by convection will be briefly considered in the next chapter.)

Because this set of experiments failed to represent the 2-10 hour period, a second experiment was initiated, using the same cloud forcing, and initialized with the 2.5 hour thermodynamic profiles.

(b) 2.5 hour profile, 2-10 forcing (sim H).

As explained above the simulation was initialized with the 2.5 hour profile. The integration was continued for nine hours, as previously. Over the last eight hours $\langle c_p(Q1)_{conv} \rangle$ has a value of $150Wm^{-2}$, with a surface rainfall of $110Wm^{-2}$. This agrees well with the observed precipitation rate of $100Wm^{-2}$. Fig 7.9 compares $(Q1)_{conv}$ and $(Q2)_{conv}$ for sim G and H. The profiles are nearly identical in the lowest 150mb of the domain. Below 500mb $(Q1)_{conv}$ follows a similar shape to sim G. However above this a large reduction is seen. This implies that much of the convection is limited to below this level. Above 800mb, $(Q2)_{conv}$ shows little drying, being dominated by two layers of moistening, between 700 and 800mb, and 400 and 600mb. These are caused by moistening due to eddy transports rather than evaporation, and are linked to cloud detrainment layers. Eddy transports (fig 7.10) of temperature are similar to sim G below 700mb, and warm the atmosphere above. However maximum heating is at 500mb, compared to 350mb in sim G. Moisture transports also do not extend to such

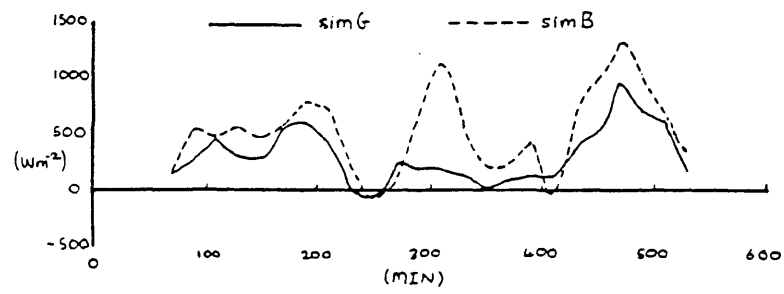


Fig 7.8 Comparison of $\langle c_p \pi(Q1)_{conv} \rangle$ for sim B and G, 20 minute averages.

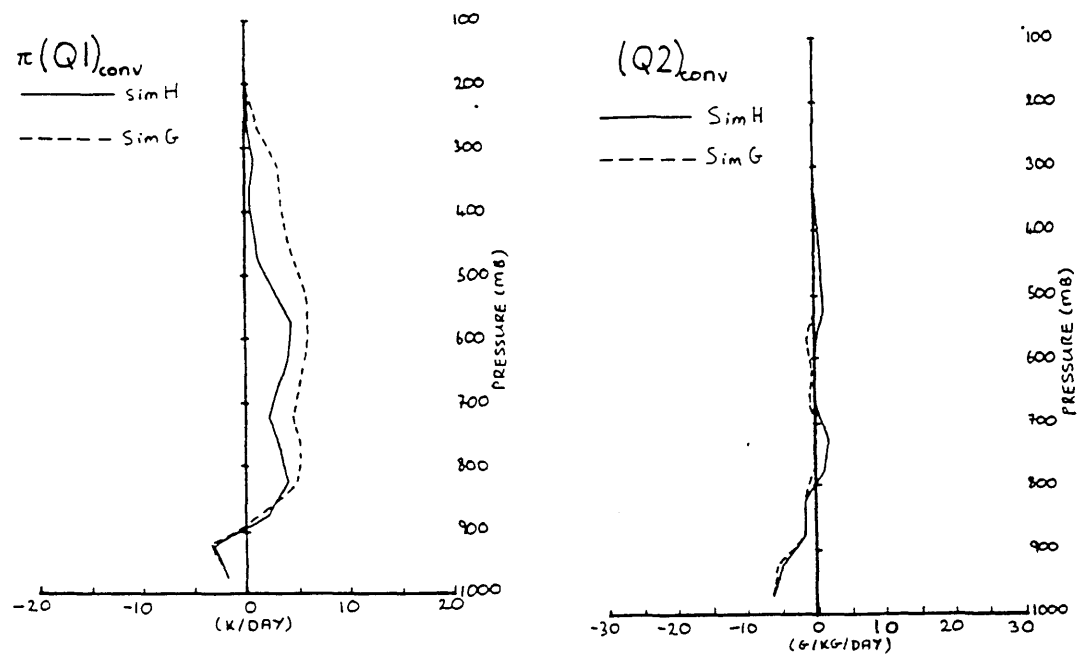


Fig 7.9 Comparison of $(Q1)_{conv}$ and $(Q2)_{conv}$ for sim G and H, averaged between 1 and 9 hours.

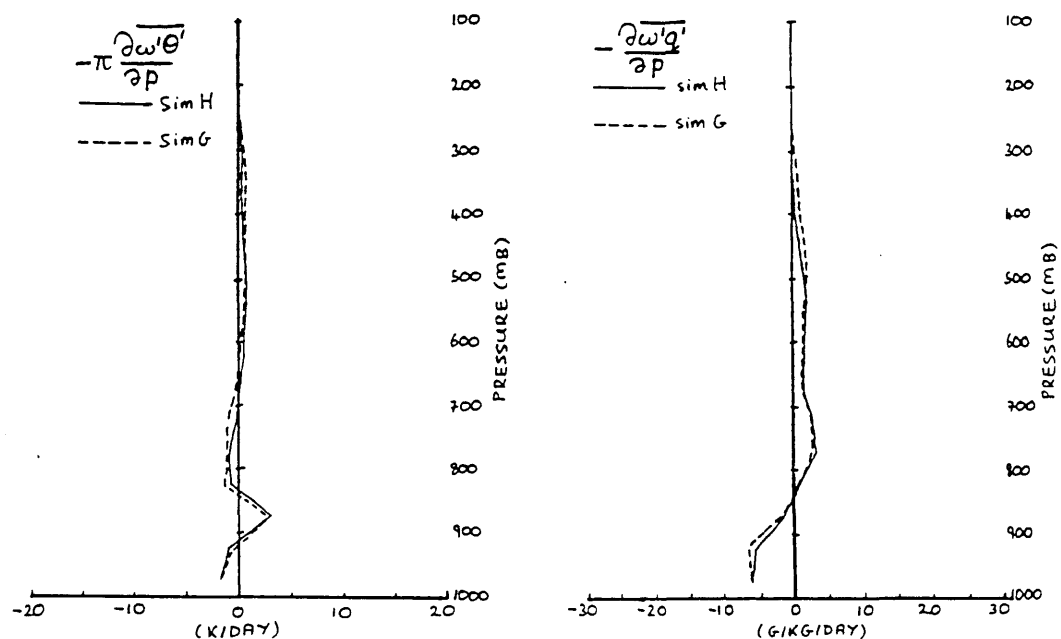


Fig 7.10 Comparison of vertical eddy transports for Θ and q , for sim G and H, averaged between 1 and 9 hours.

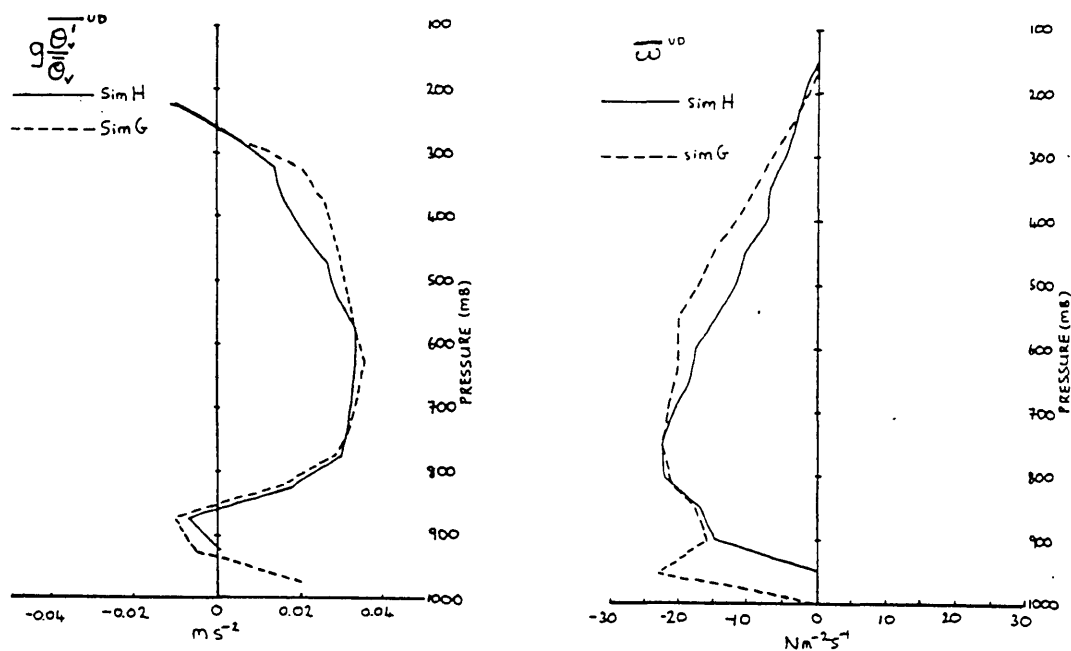


Fig 7.11 Comparison of $g\overline{\Theta_v'}/\overline{\Theta_v}$, and $\overline{\omega^{vp}}$ for sim G and H, averaged between 1 and 9 hours.

high levels as in previous simulations.

Contributions to $(Q1)_{conv}$ from UD, DD, IC and EE in terms of M_t are

$$\langle c_p \pi(Q1)_{conv} \rangle_{UD} = 1.08 M_t$$

$$\langle c_p \pi(Q1)_{conv} \rangle_{DD} = -0.54 M_t$$

$$\langle c_p \pi(Q1)_{conv} \rangle_{IC} = 0.12 M_t$$

$$\langle c_p \pi(Q1)_{conv} \rangle_{EE} = -0.30 M_t$$

These ratios are different to those seen in sim G and B, for UD and DD. EE is similar to values reported previously. The contribution due to IC is similar to that for sim G. IC mainly occurs in the lowest 200mb where sim H and G are similar.

Although there appears to be less deep convection in sim H, when it occurs it has a similar structure to deep convection seen earlier. Fig 7.11 shows UD buoyancy, and vertical velocity for sim H and G. Below 700mb UD structure is similar, but above this the sim H profile are less in magnitude.

Comparison of UD area between the two simulations confirms that less deep convection occurs in sim H (fig 7.12). The reduction is most marked above 500mb, agreeing with the decrease in condensational heating above this level. A plot $\langle c_p \pi(Q1)_{conv} \rangle$ against time (fig 7.13) is more noisy than sim G, showing that less overlap in convective lifetime occurs. The initial convection is $\approx 50\%$ in magnitude of that observed in sim G. However by 5 hours convective events are similar to those seen initially in sim G. Further increase occurs toward the end of the simulation.

Why is Sim H less active than sim G? The only difference lies in the initial profiles. Fig 7.14 compares the evolution of $\langle Lq \rangle$ for both simulations. The straight line indicates the moistening rate due to the applied forcing. $\langle Lq \rangle$ increases more rapidly in sim H, indicating that greater storage of water vapour occurs. Although $\langle L(l_c + l_r) \rangle$ is only $\approx 1\%$ of $\langle Lq \rangle$, greater amounts of liquid water are also stored in sim H, compared to sim B and G. For sim H

$$\left\langle L \left(\frac{\partial l_c}{\partial t} + \frac{\partial l_r}{\partial t} \right) \right\rangle \xrightarrow{1 \rightarrow 9 \text{ hrs}} \approx 30 \text{ W m}^{-2} \quad \left\langle L \frac{\partial q}{\partial t} \right\rangle \approx 300 \text{ W m}^{-2}$$

ie. liquid water storage is $\approx 10\%$ of the water vapour storage (cf.

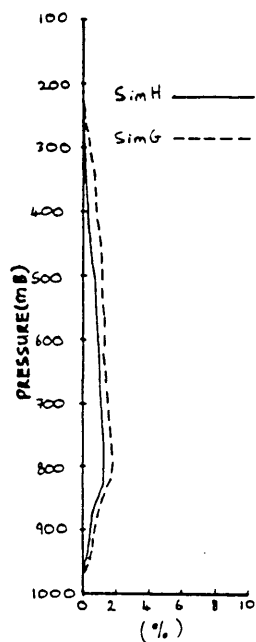


Fig 7.12 Comparison of σ for sim G and H, averaged between 1 and 9 hours.

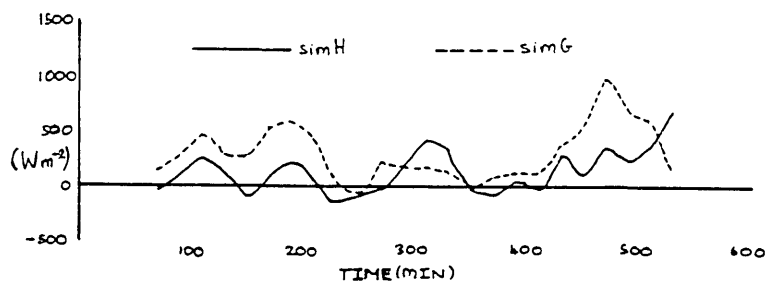


Fig 7.13 Comparison of $\langle c_p \pi(Q1)_{conv} \rangle$ for sim G and H, 20 minute averages

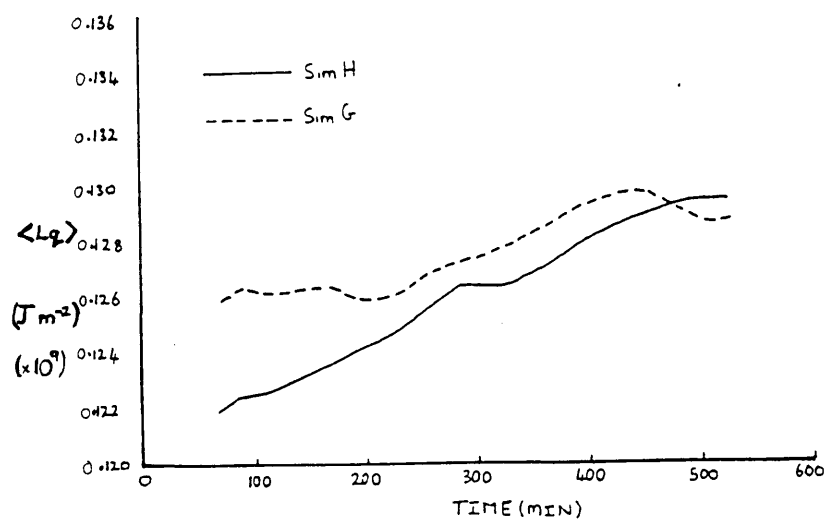


Fig 7.14 Evolution of $\langle L_q \rangle$ for sim G and H, 20 minute averages.

$\approx 3\%$ in sim B and G). This is due to an increase in the amount of deep convection over the simulation, and is similar to the figure calculated by Johnson(1980) for category one of the TPRR easterly wave. Study of the twenty minute averages (fig 7.15) shows that during convective events, liquid water storage can exceed that of water vapour. However between convective events, some of this liquid water evaporates, water vapour storage then exceeding that of the large-scale and boundary layer forcing. The remainder of the reduction in liquid water is due to fall out of rain water.

The reason for the increased storage of water vapour can be understood if the domain between 1000 and 200mb is partitioned into four, 200mb thick layers. Fig 7.16 shows $L\bar{Q}/c_p$ and $\langle Lq \rangle$ for each of these layers over twenty minute averaging periods for sim H and G.

Layer 1 (1000-800mb) shows a similar evolution in both simulations. $\langle Lq \rangle$ is approximately constant when compared to the input of moisture by the external forcing. This is transported upward by convection. Although not identical, $L\bar{Q}/c_p$ have similar structures.

The evolution of layer 2 (800-600mb) is different between the two simulations. This layer is moistened by large-scale forcing and cloud transports from below. However for sim G, the rate of moistening is less than that due to large-scale forcing alone, indicating that some of the moisture input is rained out, a few percent being stored as liquid water. For sim H the rate of moistening is greater than that due to large-scale forcing. Hence convection transports water vapour into this region without raining it out of the atmosphere. This agrees with the layer of moistening seen in the $(Q_2)_{conv}$ profile (fig 7.9). During the first 5 hours of the simulation, $L\bar{Q}/c_p$ is $\approx 50\%$ less in peak magnitude than in sim G, and also noisier. Hence the drier profile would seem to suppress convection. Little convection occurs in layers 3 and 4 during the first 5 hours of the integration.

By 5 hours, layer 2 contains as much water vapour as the initial profile of sim G (and sim B), and appears to be able to support convection as active as initially in sim G. This convection extends into layer 3 also, with a similar magnitude as in sim G.

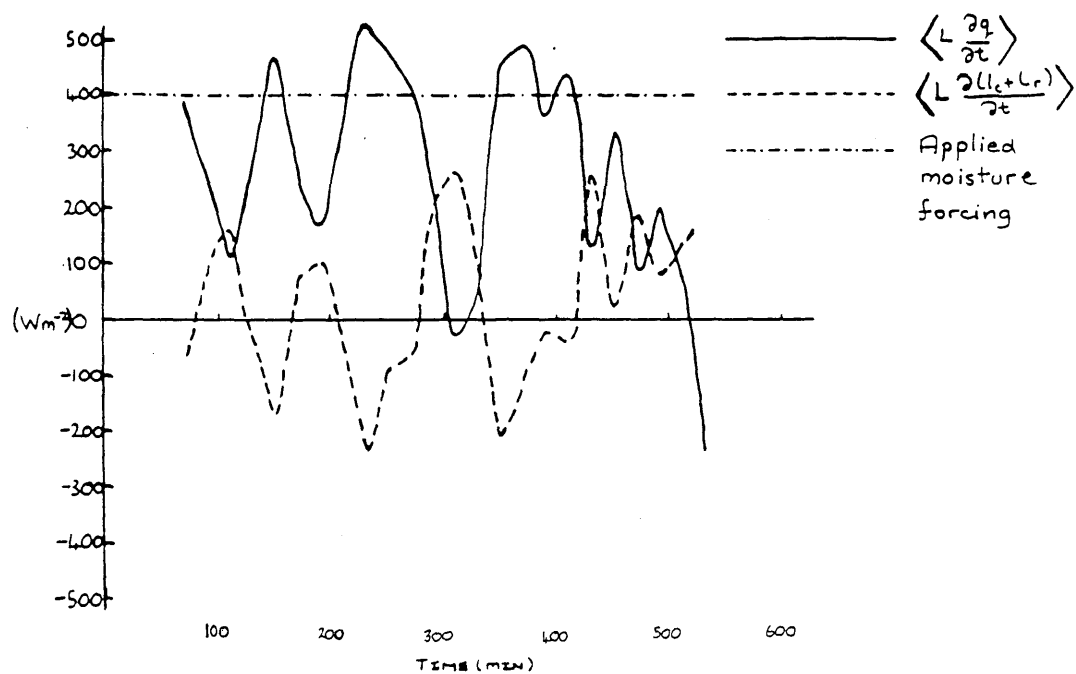


Fig 7.15 Storage of liquid water and water vapour, for sim H, 20 minute averages.

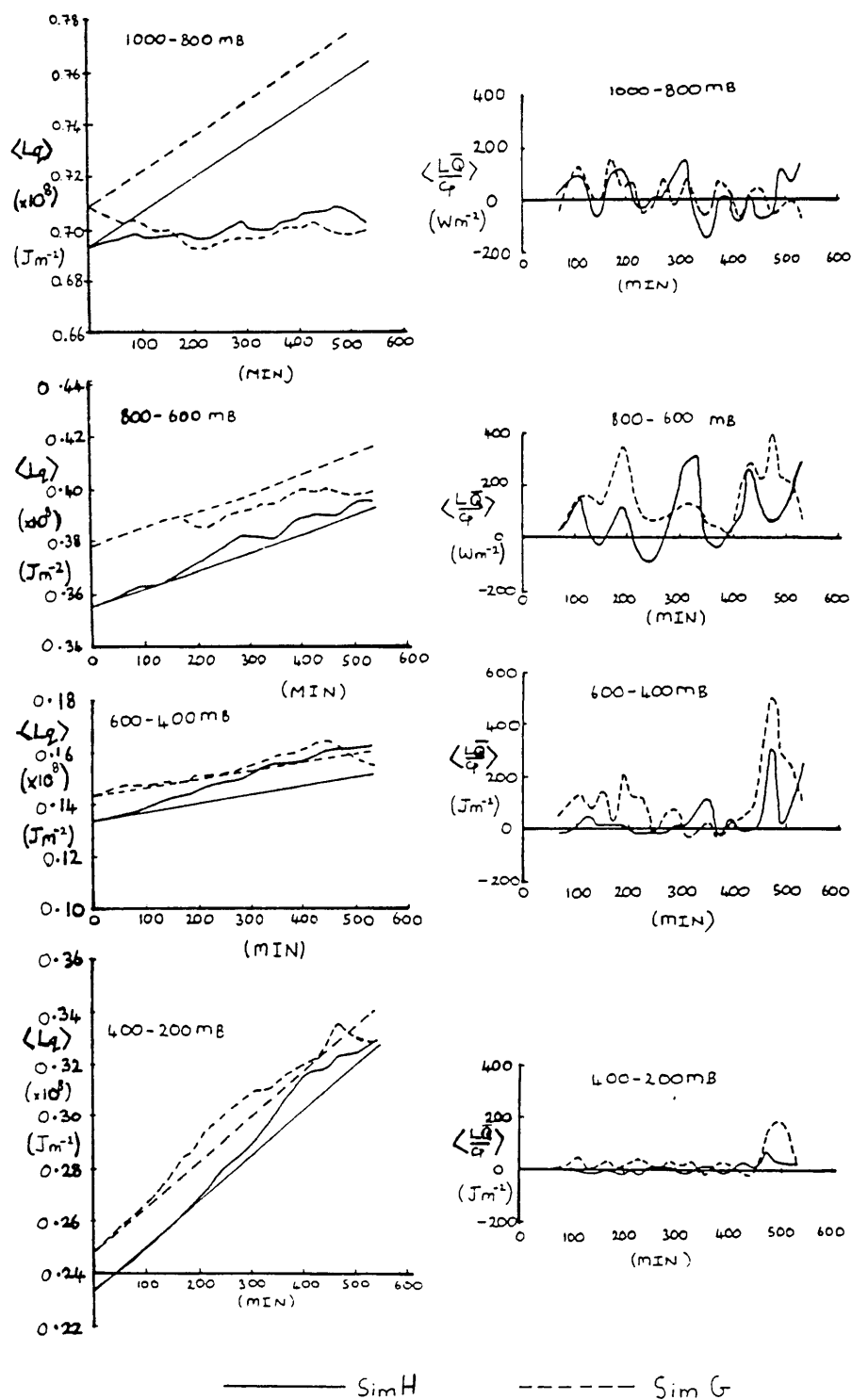


Fig 7.16 Evolution of $\overline{LQ}/c_p$ and $\langle Lq \rangle$, for 200mb thick layers, for sim H and G, 20 minute averages.

Hence a dry atmosphere suppresses convection. The mechanism for this is entrainment of environmental air into growing cells. As cloud grows through a dry layer, entrainment will cause evaporation of liquid water, and so give a reduction in buoyancy. If the layer is dry enough, then the cloud is unable to grow through the layer, detraining, and so causing moistening. Once the layer is sufficiently moist, the cloud can grow through it to higher levels of the atmosphere.

(c) Further discussion; consequence for parameterization.

These simulations indicate that the model is able to simulate periods of the easterly wave when large amounts of water vapour are stored in the atmosphere. Hence numerical models are useful in studying such events, which many parameterization schemes fail to represent (eg the original Kuo scheme).

The lowest model layer of the 2.5 hour profile has a large CAPE. However it is not able to support deep convection. UD inflow occurs in the lowest 150 to 200mb, and so the CAPE of these layers must be considered. These are considerably lower than those of the 20.5 hour profile. Also the dryness of the atmosphere above 800mb inhibits convection through entrainment as explained above. Thus the atmospheric profile modulates the convective response of the atmosphere to a change in large-scale forcing, by changing the population of convective clouds. If the atmospheric profile allows deep convection to exist, then the response of such convection to a change in large-scale forcing is approximately linear. Individual convective storms appear to be no weaker in this case, but their occurrence is less frequent ie σ is reduced.

What is the consequence of these results for parameterization? The Kuo scheme is inadequate if the parameter "b" is kept constant. Some dependence upon atmospheric structure, perhaps moisture content, is required. A scheme based upon moisture content was proposed by Anthes(1977). This is

$$b = \begin{cases} \left[\frac{(1-\langle RH \rangle)}{(1-RH_c)} \right]^n & \text{if } \langle RH \rangle \leq RH_c \\ 1 & \text{if } \langle RH \rangle > RH_c \end{cases} \quad \text{eqn(7.7)}$$

where $\langle RH \rangle$ is the mean relative humidity of the column,
 RH_c is a critical value of relative humidity,
 n is a positive exponent of order 1.

Kuo and Anthes (1984(c)), in a semi-prognostic test, using mid-latitude observational data, found best agreement between observed and diagnosed rainfall rates when n is between 2 and 3, and RH_c is between 0.25 and 0.5.

n and RH_c can be estimated using the values of b calculated from sim B and sim H (0.2 and 0.7 respectively), and $\langle RH \rangle$, averaged over the cloud layer (1000–250mb) of the initial profiles (0.71 and 0.66 respectively). These give values for n and RH_c of 8 and 0.64, very different from those predicted by Kuo and Anthes. This may indicate that the parameters n and RH_c need to vary between the tropics and mid-latitudes for the Anthes expression to work.

This calculation indicates that meso-scale models can be used to investigate the inclusion of storage into the Kuo scheme, by improving empirical relations for b . However linked to this, the cloud model used within the scheme would have to be able to model the variation of cloud top height caused by the entrainment process, as the atmospheric moisture content changes.

What about the A-S closure hypothesis? Averaging between 1 and 9 hours, for sim G,

$$\overline{\left[\frac{\partial A_N}{\partial t} \right]_{LS}}^{1 \rightarrow 9 \text{ hrs}} = -240 \text{ J/Kg/day} \quad \overline{\left[\frac{\partial A_N}{\partial t} \right]}^{1 \rightarrow 9 \text{ hrs}} = -90 \text{ J/Kg/day}$$

and for sim H,

$$\overline{\left[\frac{\partial A_N}{\partial t} \right]_{LS}}^{1 \rightarrow 9 \text{ hrs}} = -210 \text{ J/Kg/day} \quad \overline{\left[\frac{\partial A_N}{\partial t} \right]}^{1 \rightarrow 9 \text{ hrs}} = -40 \text{ J/Kg/day}$$

Hence the closure assumption seems reasonable for sim H, but not for sim G. However a great degree of cancellation does occur between $[\partial A_N / \partial t]_{LS}$ and $[\partial A_N / \partial t]_{cum} (\approx 60\%)$. If the average is taken over all but the last twenty minutes of the simulation (starting at 1 hour), then for sim G,

$$\overline{\left[\frac{\partial A_N}{\partial t} \right]}^{1 \rightarrow 9 \text{ hrs}} = -240 \text{ J/Kg/day} \quad \overline{\left[\frac{\partial A_N}{\partial t} \right]}^{1 \rightarrow 9 \text{ hrs}} = -20 \text{ J/Kg/day}$$

Hence the imbalance would seem to result in averaging over a period in which the atmosphere has not had time to recover from convective effects.

The results generally indicate that the A-S closure hypothesis is valid for weaker convective cases. However the scheme is insensitive to storage, and its accuracy will depend on whether the analytical cloud ensemble used is able to represent the variation of cloud population with atmospheric structure.

7.4 SIMULATION OF STRONGER CONVECTIVE CASES (SIM I).

The 20-30 hour period of the easterly wave is not the most convectively active. Between 30 and 50 hours surface precipitation reaches a maximum (fig 2.5). During wave category 5 (40-50 hours), surface precipitation is 600Wm^{-2} , compared to moisture supply by large-scale and boundary layer forcing of 350Wm^{-2} . Hence the convection removes moisture from the atmosphere stored during the first half of the wave cycle. The Kuo parameter, b , is 1.7.

An experiment was carried out with the meso-scale model in an attempt to simulate the 40-50 hour period. The model was initialized using the 20.5 hour profile. Comparison of the 20.5 and 40.5 profiles is made in fig 7.17. The 40.5 hour profile is drier at the surface and moisture above 600mb. The CAPE of the lowest model level (975mb) is 950Jkg , compared to 1300Jkg for the 20.5 hour profile. Hence less convective activity might be expected for the same large-scale forcing. The large-scale moisture forcing is based upon that observed for this period, decreasing from 275Wm^{-2} (41.5 to 43.5 hours), to 154Wm^{-2} (47.5 to 49.5 hours) over the simulation. Averaged between one and nine hours of a nine hour integration, its value is 176Wm^{-2} . Little difference is seen in the boundary layer forcing between wave categories three and five. Thus the values used in sim B are used (10Wm^{-2} for sensible heat, 127Wm^{-2} for latent heat). The cloud forcing is identical to that used in sim B, G and H.

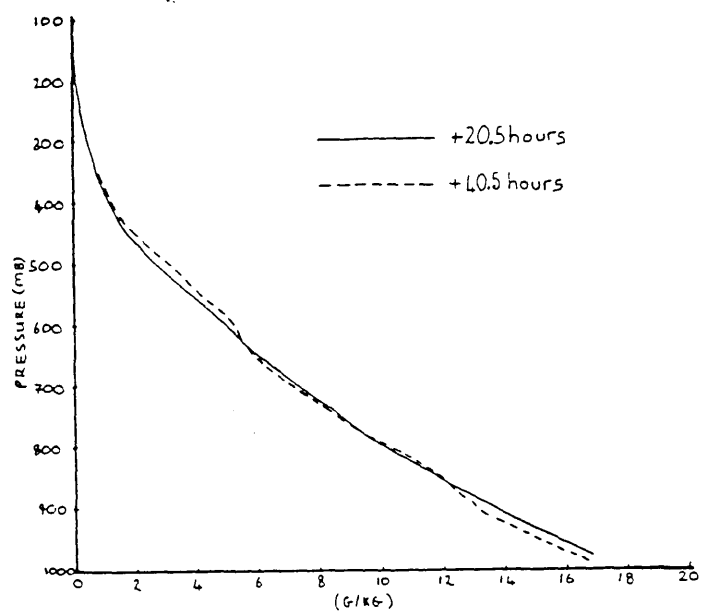


Fig 7.17 Comparison of 20.5 and 40.5 moisture profiles of the GATE composite easterly wave.

The simulation failed to produce the observed convective energy release. The ratio of $\langle c_p \pi(Q1)_{conv} \rangle$ to M_t is 0.71 (averaged between 1 and 9 hours), similar to that observed in sim B and G. Hence the convection reacts linearly to the decrease in forcing.

The failure would seem to be due to an initialization problem. The simulated convection grows to maximum amplitude as the moisture forcing decays. In reality deep active convection exists before the decay occurs. Once the decay takes place, deep convection is able to feed off the moisture stored in the atmosphere in the first half of the wave cycle. Decay of convection finally occurs between 60 and 70 hours.

Initializing the model with the 40.5 hour profile would not bring better results. It would seem necessary to initialize the convective circulations before the integration is started. This could perhaps be achieved by using a large external forcing until deep convection occurs, and then changing to the forcing observed between 40 and 50 hours. However changes in the domain averaged thermodynamic profiles (especially moisture) will have occurred by this time due to the 2D nature of the ensemble, and so it is unclear how the simulation would respond.

7.5 BRIEF COMMENTS ON CAPE.

Fritsch and Chappell(1980a), Kreitzberg and Perkey(1976), and Miller and Moncrieff(1983) have developed parameterization schemes using CAPE to predict convective activity. The first two were developed for use within coarse meso-scale models, while the last was tested at ECMWF. Numerical cloud models, such as the one used in this study can be used to test if CAPE is a sensitive parameter for use in convection schemes.

Here the results from sim B are presented. CAPE is calculated from the domain averaged thermodynamic profiles. However in the lower layer of the model these undergo large modification due to the 2D nature of the model, especially moisture. The lowest 150mb of the initial profile (20.5 hours) exhibits positive CAPE. Which model

layer is least (abnormally) effected by the models two dimensionality? Fig 7.18(a) show q averaged over twenty minute intervals between 1 and 9 hours of sim B, for the lowest three model levels. After initial modification by transport properties of dry thermals in the cloud forcing region during the first hour, 875mb seems to be the least effected, staying in a quasi-steady state. This is shown in more detail in fig 7.18(b). Thus it would seem reasonable to use CAPE calculated from 875mb (level 16) here $((\text{CAPE})_{875})$.

Fig 7.18(c) shows $(\text{CAPE})_{875}$ together with $\langle c_p \pi(Q1)_{\text{conv}} \rangle$. CAPE reduces when convection is active and increases when convection decays, as expected. There is some correlation between peaks in the CAPE sequence and those of $\langle c_p \pi(Q1)_{\text{conv}} \rangle$, a lag of up to 40 minutes being apparent. Some correlation between the magnitudes of the peaks also exists. However CAPE does not predict convective activity on a point to point basis. Also no critical value of CAPE above which convection occurs is apparent.

The sequence brings some understanding as to why the observational data in the tropics shows CAPE to be a minimum when convective activity is a maximum (fig 2.6 of chapter 2). CAPE is observed to decrease before the peak of active convective events, due to modification of the domain averaged thermodynamic structure by convection. In the real atmosphere, when convection is active, radio sonde observations will be of the convectively modified atmosphere. Hence CAPE calculated from large-scale observations are likely to be less than those seen by convection as it grows.

These results indicate that CAPE shows variation with convective activity. However the relationship appears to be complex. Meso-scale models (perhaps of a single cloud) will probably be of use in determining the nature of this relationship more fully. Also the CAPE calculated here does not include water loading effects. Again meso-scale models will prove useful in specifying these.

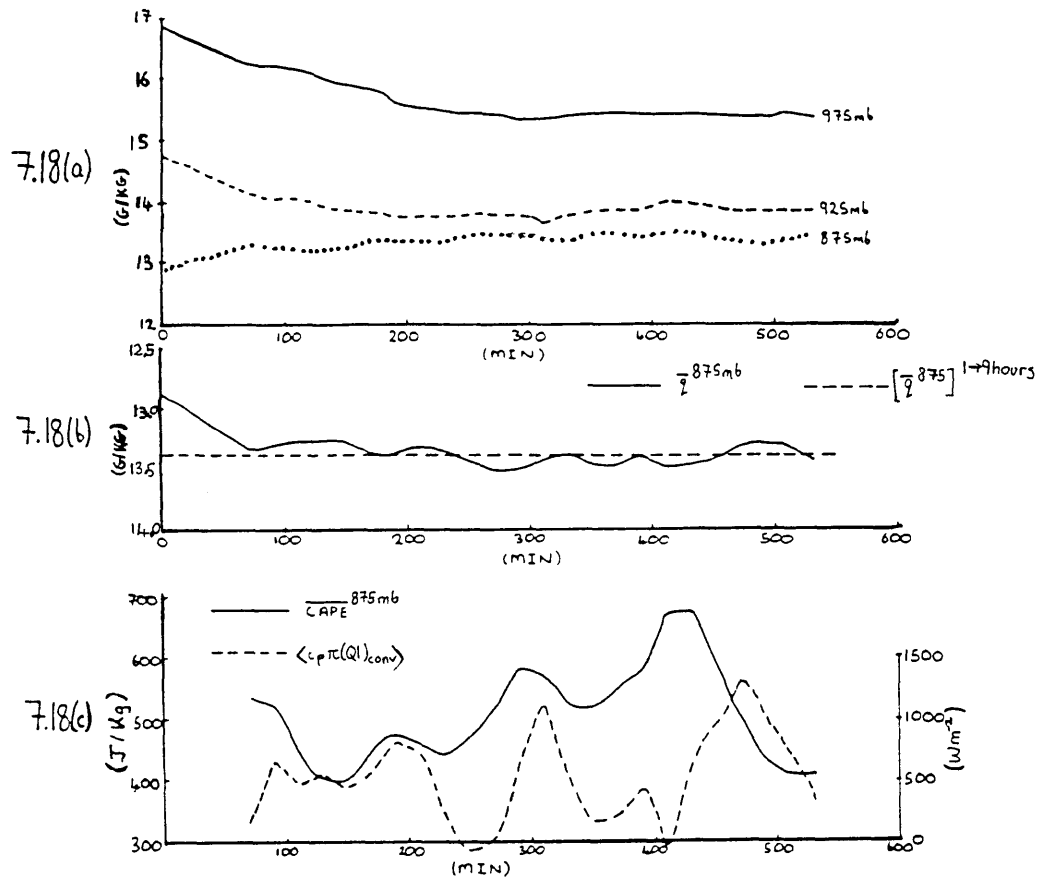


Fig 7.18(a) $\bar{q}$ for 975, 925, and 875mb, for sim B, 20 minute averages.

Fig 7.18(b) $\bar{q}$ for 875mb, for sim B, 20 minute averages (expanded scale).

Fig 7.18(c) CAPE for 875mb, and $\langle c_p \pi(Q1)_{conv} \rangle$ for sim B, 20 minute averages.

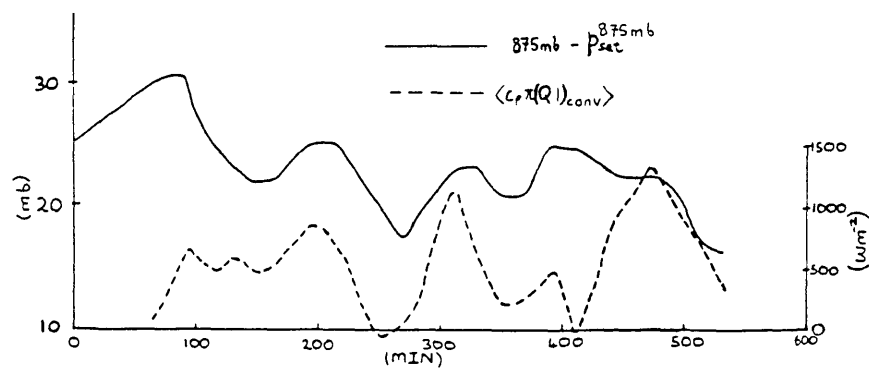


Fig 7.19 $P (=p_{saturation} - p_{initial})$ for 875mb, for sim B, 20 minute averages.

7.6 BRIEF COMMENTS ON THE BETTS-MILLER SCHEME.

Betts and Miller(1984) have developed an adjustment scheme using in part the saturation point parameter. Is such a parameter sensitive to convective activity? Fig 7.19 shows $P(=p_{init}-p_{sat})$ for 875mb (level 16) for sim B, averaged over twenty minute intervals. P increases as convection stabilizes the atmosphere, and increases as the large-scale forcing destabilizes the atmosphere. Hence the saturation point parameter would seem a useful indicator of convective activity.

The concept behind the Betts-Miller scheme is similar to moist adiabatic adjustment. The atmosphere is adjusted back to some characteristic structure when convection occurs. The failure of moist adiabatic adjustment comes because a Θ_{es} curve is not characteristic of a convective atmosphere. However Betts(1982a) observed that convective atmospheric profiles closely follow a virtual saturated equivalent potential temperature curve (Θ_{esv}) below the freezing level. Above this the profile diverged back to the Θ_{es} curve from cloud base, near cloud top.

Θ_{esv} differs from the definition of Θ_{es} , because the condensed water is retained as the parcel ascends, rather than falling out. The gradients of the two curves are related by

$$\left(\frac{\partial \Theta}{\partial p}\right)_{\Theta_{esv}} \simeq 0.9 \left(\frac{\partial \Theta}{\partial p}\right)_{\Theta_{es}}$$

From this Θ_{esv} can be drawn on a T- θ gram.

Does the model atmosphere have the structure Betts and Miller suggest? Fig 7.20 shows the model temperature profile for sim B averaged between 1 and 9 hours, together with a Θ_{es} and Θ_{esv} curve drawn through the saturation point of the lowest level. Above 800mb the model profile parallels the Θ_{esv} curve. Above freezing level they diverge as suggested by Betts and Miller. Below 800mb the model profile does not follow Θ_{esv} , probably due to the cloud forcing applied in this region.

The model profile lies to the right of the Θ_{esv} curve. This is probably due to the drying of the lowest model layer, causing the saturation point to be higher than observed, and Θ_{es} to be lower for moist ascent. A 3D model might give a better fit.

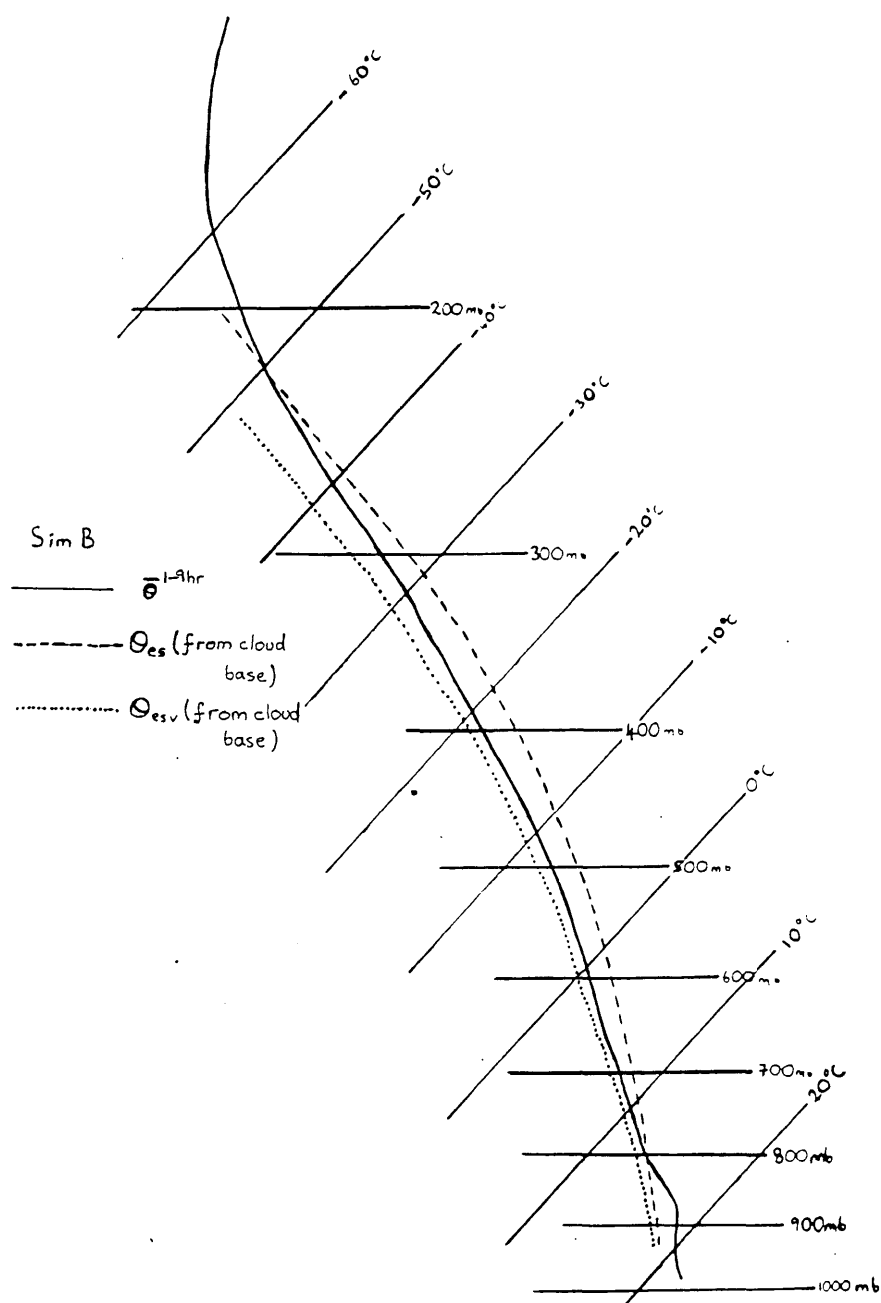


Fig 7.20 Temperature structure of sim B, averaged between 1 and 9 hours, compared to Θ_{es} and Θ_{esv} curves through the saturation point of 975mb.

Thus there is reasonable evidence that the model is able to simulate the observed thermodynamic structure of the atmosphere reasonably well.

Many features of the scheme are not well understood. It is unclear why the atmosphere should retain such a structure. Also the effects of shear and organization may require different adjustment profiles to be used. The present scheme cannot represent storage processes. Numerical cloud models could be used to study these questions, perhaps with more ease than using observational data. The question over what space and time-scales the scheme is valid could also be addressed.

7.7 SUMMARY.

The relationship between convection and large-scale processes has been considered, in relation to the closure problem of cumulus parameterization. If the initial thermodynamic profiles are able to support deep convection, the response of such convection to a change in large-scale forcing is linear. The clouds rain out approximately 80% of the total moisture forcing applied to the model.

The model is able to simulate periods when large amounts of water vapour are stored within the atmosphere. This is caused by deep convection being inhibited by the atmosphere being too dry, causing increased evaporation in growing towers. Both large-scale forcing, and detrainment from convection moisten the atmosphere until deep convection can exist.

Difficulties have been encountered when trying to model phases of the wave where convection removes more moisture than provided by the large-scale and boundary layer forcing. This seems to be due to problems in initializing convection within the model, as explained earlier.

Comments have been made concerning both the Kuo and A-S closure schemes. The Kuo parameter "b" calculated from the model, when deep convection occurs, is similar to that derived from observational

studies. However the scheme is not able to deal with periods when storage of moisture occurs. The form of b due to Anthes(1977) has been studied, values for the exponent n , and RH_c calculated from model data not agreeing with those suggested by Kuo and Anthes(1984c).

The A-S closure hypothesis appears more general, and the model confirms previous work using observational data. However here the calculations include the effects of downdraughts, not included in the original A-S scheme.

Brief comments have also been made on the use of CAPE and saturation point as indicators of convective activity. Both are sensitive to convective activity. The domain averaged thermodynamic structure is similar to the equilibrium structure used in the Betts-Miller scheme when active convection occurs.

In the next chapter momentum transports by cloud in sheared flows are briefly considered, and aspects of their parameterization studied.

CHAPTER 8

PARAMETERIZATION OF MOMENTUM TRANSPORTS AND THE EFFECTS OF SHEAR.

Previous simulations presented in this study have been unsheared. In this chapter the effects of shear are considered. Several simulations are presented using initial thermodynamic conditions of sim B, but with differing wind shears derived from the TPRR data set. Thermodynamic properties are compared to those of unsheared ensembles. Convective scale momentum transports are discussed, and their parameterization considered. The effects of organization upon the ensemble is also briefly discussed.

8.1 THEORY AND REVIEW OF PREVIOUS WORK.8.1.1 Observational And Parametric Theory.

The area averaged horizontal momentum equation (in 2D) (eqn(2.9)) is

$$\frac{\partial \bar{u}}{\partial t} + \frac{\partial \bar{u}^2}{\partial x} + \frac{\partial \bar{u} \bar{v}}{\partial p} + g \frac{\partial (\bar{h} - \bar{h}_x)}{\partial x} = - \frac{\partial \bar{\omega' u'}}{\partial p} + \bar{S}_u \quad \text{eqn(8.1)}$$

neglecting horizontal eddy transports, and the coriolis force which is not included in the two dimensional model.

The average horizontal pressure gradient contains both mean and eddy terms. Measurements of pressure is difficult in the tropics, due to non-geostrophic motion. Thus the importance of the eddy pressure term is unclear, and is usually neglected when using observational data to diagnose the rhs of eqn(8.1). With a cyclic model such horizontal terms are zero when averaged over the entire domain, and so their neglect causes no error in the diagnosis of the model.

The eddy transport term on the rhs of eqn(8.1) is usually termed cumulus friction (Schneider and Lindzen(1976)), and is here denoted by Q3. As with thermodynamic eddy transports it is usually associated with cumulus convection. To parameterize this term it must be expressed in terms of cloud and large-scale variables. Thus following the method of chapter 5

$$Q3 = -\frac{\partial \sigma (\overline{\omega \bar{u}^c} - \bar{\omega}^c \bar{u})}{\partial p} - \frac{\partial (1-\sigma)(\overline{\omega \bar{u}^e} - \bar{\omega}^e \bar{u})}{\partial p} \quad \text{eqn(8.2)}$$

The second term on the rhs is usually neglected. In the meso-scale model the equivalent term for θ and q is associated with cloud forcing, and above 800mb has negligible effect. Hence for deep convection,

$$(Q3)_{conv} = -\frac{\partial \sigma (\overline{\omega \bar{u}^c} - \bar{\omega}^c \bar{u})}{\partial p} \quad \text{eqn(8.3)}$$

which gives on expansion:-

$$(Q3)_{conv} = -\frac{\partial \sigma \bar{\omega}^c \bar{u}^c}{\partial p} + \sigma \bar{\omega}^c \frac{\partial \bar{u}}{\partial p} + \bar{u} \frac{\partial \sigma \bar{\omega}^c}{\partial p} \quad \text{eqn(8.4)}$$

The cloud averaged horizontal momentum equation is:-

$$\frac{\partial \sigma \bar{u}^c}{\partial t} - u_b \frac{\partial \sigma}{\partial t} + \sigma \left(\overline{\frac{\partial \bar{u}^2}{\partial x}} \right)^c + \frac{\partial \sigma \bar{\omega}^c \bar{u}^c}{\partial p} - (\omega u)_b \frac{\partial \sigma}{\partial p} + g \sigma \left(\overline{\frac{\partial (h-h_1)}{\partial x}} \right)^c = \sigma \bar{S}_u^c$$

Substituting into eqn(8.4) for $\partial \sigma \bar{\omega}^c / \partial p$, and using the cloud averaged continuity equation (eqn(5.15)) in a similar manner as when deriving $(Q1)_{para}$, after some manipulation $(Q3)_{para}$ may be written as:-

$$\begin{aligned} (Q3)_{para} = & \frac{\partial \sigma \bar{u}^c}{\partial t} + (\bar{v} - u_b) \frac{\partial \sigma}{\partial t} + [\omega_b \bar{v} - (\omega u)_b] \frac{\partial \sigma}{\partial p} + \sigma \left(\overline{v \frac{\partial \bar{u}}{\partial x}} \right)^c \\ & + \sigma \left(\overline{\frac{\partial \bar{u}}{\partial x} (\bar{v} - \bar{u})} \right)^c + g \sigma \left(\overline{\frac{\partial (h-h_1)}{\partial x}} \right)^c + \sigma \bar{\omega}^c \frac{\partial \bar{u}}{\partial p} - \sigma \bar{S}_u^c \end{aligned} \quad \text{eqn(8.5)}$$

This is equivalent to $(Q1)_{para}$ and $(Q2)_{para}$ derived in chapter 5, and has formed the basis for diagnostic studies of cumulus momentum transports using observational data, and steady state analytical cloud models. Later in this chapter eqn(8.5) will be diagnosed using the

results of the meso-scale model.

8.1.2 Review Of Previous Work.

There are few observational studies of large-scale momentum budgets. A great difficulty in calculating the lhs of eqn(8.1) is obtaining an accurate value for the pressure gradient term. This is especially difficult within the tropics, where pressure gradients are small. Most work on dynamical transports have been in terms of vorticity. Reed and Johnson(1974) found a substantial residual in the vorticity budget of a composite easterly wave, which they attributed to convection. Using the TPRR data set, Stevens(1979) found that for the composite GATE phase III easterly wave, a vorticity sink existed in the lower levels of the atmosphere, with a source above. Cho et al(1979) attempted to model the large-scale budget residuals using simple cloud models, and found that this could be done using a compensating subsidence and twisting term.

The GATE data set was of sufficient quality to estimate the geopotential gradient. Stevens(1979) also calculated the large-scale momentum budget for the TPRR data set, and found that the rhs of the large-scale horizontal momentum equation was approximately 50% of the other terms.

Shapiro and Stevens(1980) used these observed budgets to test the assumption that the residual was due to cumulus friction (originally proposed by Schneider and Lindzen(1976)). They used an approximate form of eqn(8.5), equivalent to eqns(5.17) and (5.18) for $(Q1)_{para}$ and $(Q2)_{para}$, for use with analytical cloud models, ie.

$$(Q3)_{para} = \sum_{\text{detraining cloud}} D_i (\bar{u}_i^c - \bar{u}) + \sigma \bar{\omega}^c \frac{\partial \bar{u}}{\partial p} + g \sigma \frac{\partial \bar{h}}{\partial x}^c$$

They assumed $\bar{u}_i^c$ to be constant (at the inflow speed) and neglected pressure gradients. They found that cumulus friction explained the bulk of the residual, although agreement was poor above 500mb. This was attributed to the neglect of pressure gradients.

Lee(1984) used $(Q3)_{conv}$ in the form of eqn(8.3) to compare cumulus friction derived using simple cloud models, and residuals of hurricane momentum budgets. He found better agreement than Shaperio and Stevens, attributable to not having to include any representation of cloud pressure gradients in calculating $(Q3)_{conv}$.

However such results are disputed. Thompson and Hartmann(1979) used tropical data to study the Hadley circulation and found that cumulus friction was unimportant. Sardeshmukh and Held(1984), studying the vorticity budget of the monsoon within a GCM came to a similar conclusion. Hence observational and modelling studies would seem to be ambiguous to the role of cumulus momentum transports in large-scale dynamics.

The above studies of cumulus friction used simple one dimensional cloud models with a simple constraint on $\bar{u}^2$. This may be inadequate in studying momentum transports. Shear is known to produce organization within convective systems. Several distinct types of convection occur within the atmosphere, each having a different flow configuration and momentum transports, both up and down gradient (Moncrieff(1981)). Hence to diagnose and study convective momentum transports it may be necessary to use dynamical cloud models, such as those presented in Moncrieff(1981). In some cloud types, pressure gradients across the system are important in driving the storm (eg the propagating model of Moncrieff(1981)). How the magnitude of these is predicted from large-scale flow is unknown.

Studies of numerical models of individual storms may prove useful in determining momentum transports for various types of cloud, and also the magnitude of the cloud pressure gradient. Moncrieff and Miller(1976) discusses a model of a tropical squall line, while Soong and Tao(1984) have studied tropical line convection in relation to cloud momentum transports.

Further complications arise because of meso-scale pressure gradients observed to exist in the tropics, associated with organized tropical systems (Zipser(1977)). The effect of these upon the large-scale flow is, at present, unclear.

8.2 THE EFFECTS OF WEAK SHEAR UPON SIM B (SIM J).

The simulation presented here is identical to sim B, except that the model is also initialized with the 20.5 hour zonal wind shear (fig 8.1). This has westerly flow in the lowest 200mb, with easterlies above. Two jets are seen, one at 700mb, the other at 200mb. The low level jet maxima is 3.5ms^{-1} , and the shear in the lowest 300mb is 7ms^{-1} . The profile is typical of that associated with the development of squall lines, although here the low level shear is too weak for this. Shears of between 15 and 25ms^{-1} in the lowest 300mb are usually associated with squall lines.

No large-scale momentum forcing is applied to the simulations presented in this chapter. The accuracy of this quantity from observations is unsure. Also the contribution of the large-scale pressure gradients is also not well known. It is unclear whether convective momentum transports are balanced by those of the large-scale. For thermodynamic variables, such a balance exists because clouds stabilize the atmosphere, while large-scale motions destabilize it.

The same random forcing as in sim B was used, the model being integrated for nine hours. Active convection develops by 100 minutes, slightly longer than in sim B, due to the inhibiting effects of the low level shear. The clouds are transient in nature, with a life span typical to those observed in sim B (1 to 1.5 hours). However below 500mb, the clouds travel to the west at approximately -2ms^{-1} . This is the velocity of the wind shear at 650mb. Such motion is typical of organized tropical systems (squall lines), which are much longer lasting (several days), and travel at much higher speeds. Thus the presence of even weak shear seems to have some organizational effect. However above 500mb, the clouds move at the mean wind speed.

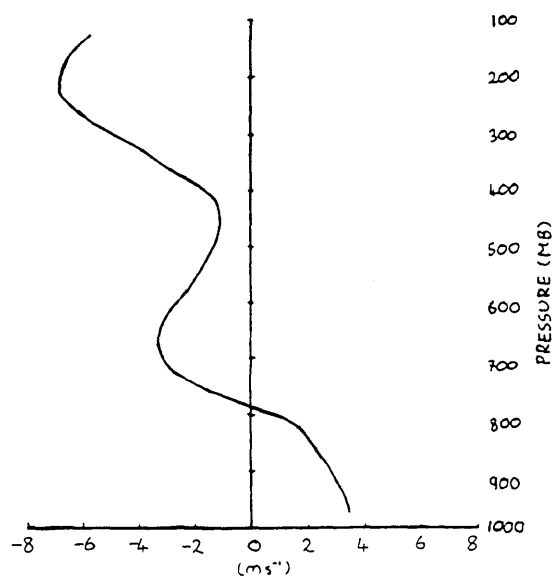


Fig 8.1 20.5 hour zonal velocity profile of the GATE composite easterly wave.

8.2.1 Thermodynamic Budgets And Cloud Structure.

In this section the thermodynamic budgets and cloud structure of sim J and sim B are compared.

$\langle c_p \pi(Q1)_{conv} \rangle$ averaged between 1 and 9 hours is 495 Wm^{-2} , slightly less than in sim B. The time structure is similar (fig 8.2), although direct correlation between events is not seen. However the frequency and duration of events is comparable. The vertical structure of $(Q1)_{conv}$ and $(Q2)_{conv}$ are also similar (fig 8.3) to sim B, although values are slightly less above 500mb. Vertical transports have an identical structure, although do not extend to the same altitude as in sim B.

The contribution of UD and DD to $(Q1)_{conv}$ and $(Q2)_{conv}$ are similar to sim B. UD structure is compared in fig 8.4, where buoyancy and vertical velocity are shown. The convection of sim J is slightly weaker above 500mb. Cloud area is also slightly less above this pressure, probably due to the low level shear weakening the convection as it grows.

From these results, it appears that the weak shear of 20.5 hours has little effect upon the thermodynamic budget, and structure of the cloud ensemble.

8.2.2 Convective Momentum Transports.

In this section the vertical transport of horizontal momentum by convection is presented and discussed. Parametric interpretation is considered in the next section. Fig 8.5 shows the internal UD and DD zonal velocity. These largely follow the domain averaged mean wind profile (except UD in the lowest 100mb). Such a feature was not observed when cloud thermodynamic profiles were considered, and so the cause would not seem to be simple mixing or time dependent effects. This suggests that the cloud scale pressure gradient plays an important role in determining the cloud internal velocities. LeMone(1983) in an observational study of a GATE convective line found the effects of cloud scale pressure gradients to be important in determining momentum transports, and internal cloud velocity. Hence

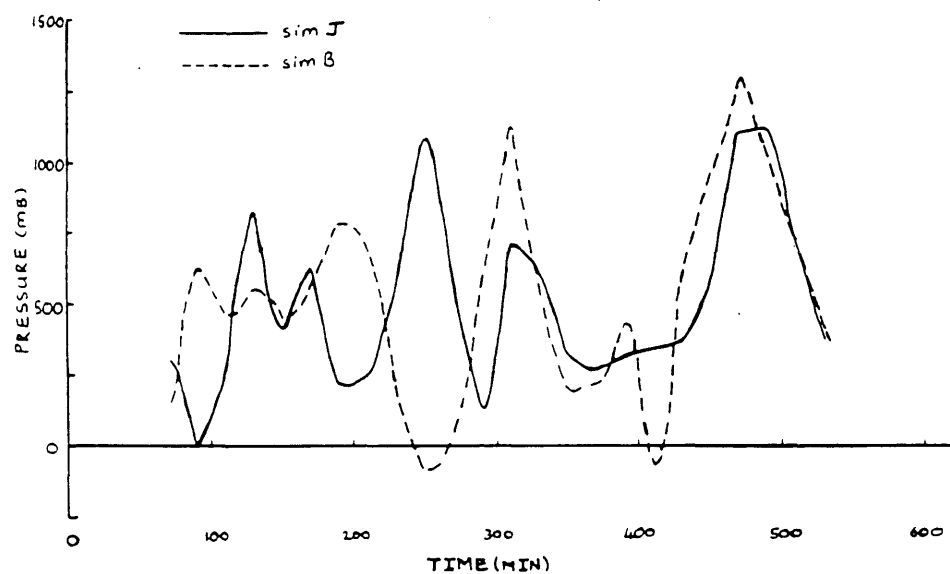


Fig 8.2 Comparison of $\langle c_p \pi(Q1)_{conv} \rangle$ for sim B and J, 20 minute averages.

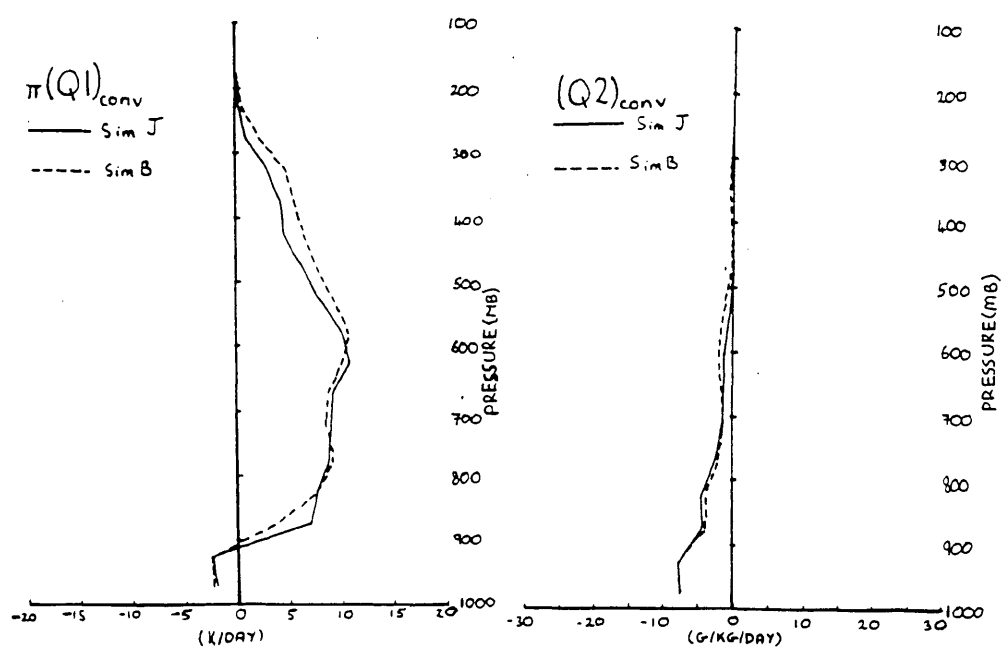


Fig 8.3 Comparison of $(Q1)_{conv}$ and $(Q2)_{conv}$ for sim B and J, averaged between 1 and 9 hours.

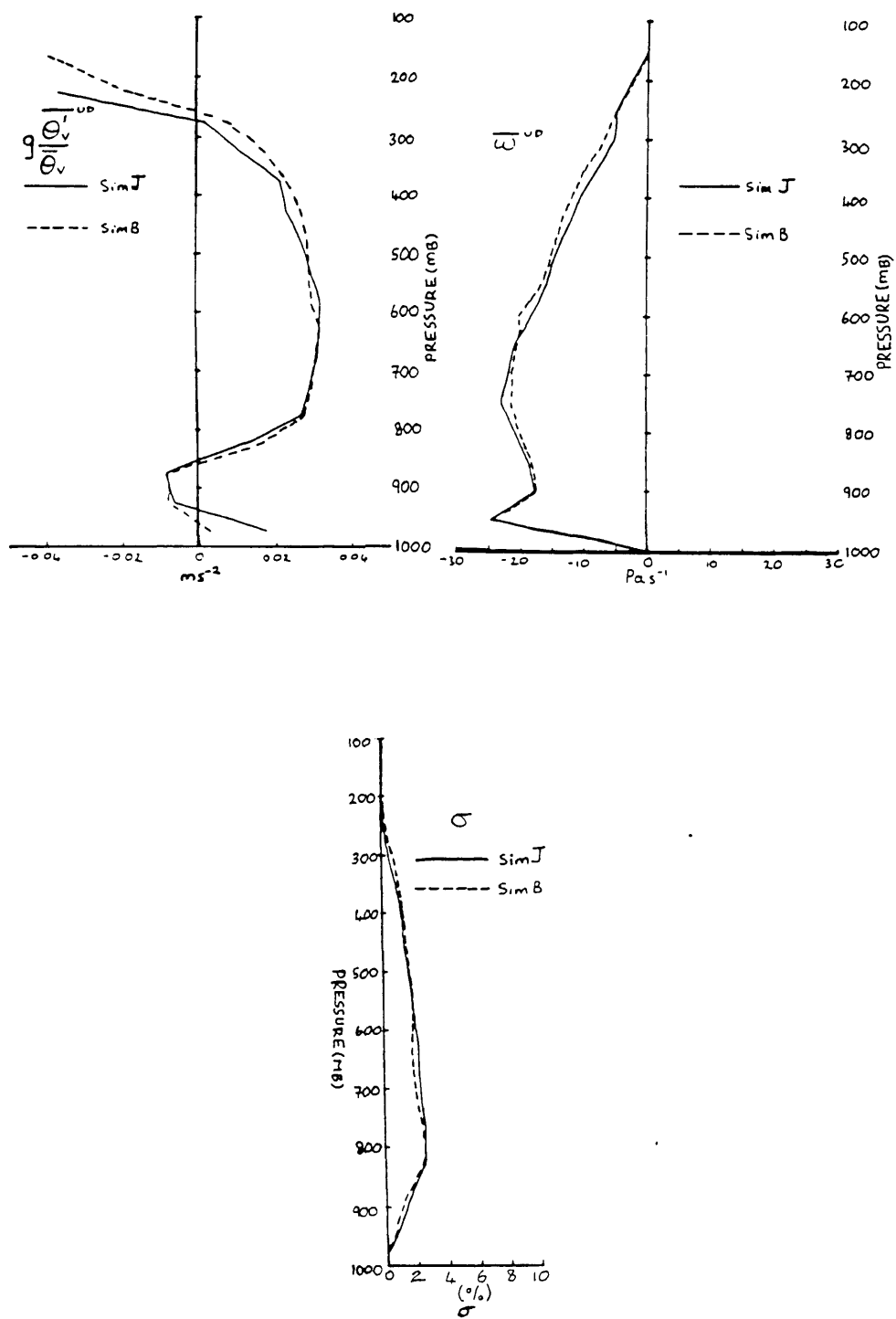


Fig 8.4 Comparison of $\frac{g \overline{\theta'_v}}{\overline{\theta_v}}^{up}$, $\overline{\omega}^{up}$, and σ_{vo} for sim B and J, averaged between 1 and 9 hours.

the assumption used by Schneider and Lindzen(1976), and other workers, that $\bar{u}$ is conserved, would seem to be incorrect.

The vertical transport of horizontal momentum by eddies is shown in fig 8.6(a), together with contributions from convective and environmental motions, all averaged between 1 and 9 hours. The convection accelerates the lower part of the 700mb jet to the west and the surface flow to the east ie. a flux of easterly momentum upwards. Hence the clouds increase the low level shear. Although this inhibits cloud growth, larger shears lead to greater organization and storm efficiency. The jet maxima is accelerated to the east. Above 650mb the flow is accelerated to the west and to the east above this. These transports above the low level jet may be due to a mixing length process, proportional to $-\partial\bar{u}/\partial p$ (compare the shape of $-\partial\bar{u}/\partial p$ above 650mb to that of $(CS)_0$ in fig 8.7(a)), suggesting that in the upper half of the domain, UD slope with the wind shear ie. an unorganized structure.

The majority of eddy transports occur within cloudy air. However environmental transports play a larger role above 800mb than for thermodynamic fields.

Fig 8.6(b) compares the contributions of UD, DD and IC to $-\partial\bar{u}/\partial p$. As for thermodynamic fields UD dominates, providing a reasonable estimate of $-\partial\bar{u}/\partial p$. DD effects oppose UD in the lowest 200mb. However DD also acts to increase the low level shear accelerating the flow of the jet maxima to the west (whereas UD brings the jet maxima towards the surface). IC contributes little above 800mb, and below this opposes DD. Below 700mb neither upward or downward moving cloud mix momentum down gradient, implying that both have some degree of organization. These effects are similar to those found associated with tropical squall lines and line convection, which increase the low level shear normal to the system.

However the TPRR data set shows that the low level shear decreases when convection is active. The difference could be due to several factors. The model here is two dimensional. Three dimensional effects may be important. In a 3D atmosphere environmental air may flow around UD air, changing cloud pressure

Fig 8.5 $\bar{u}'$, $\bar{u}''$, and $\bar{u}$ for sim J, averaged between 1 and 9 hours.

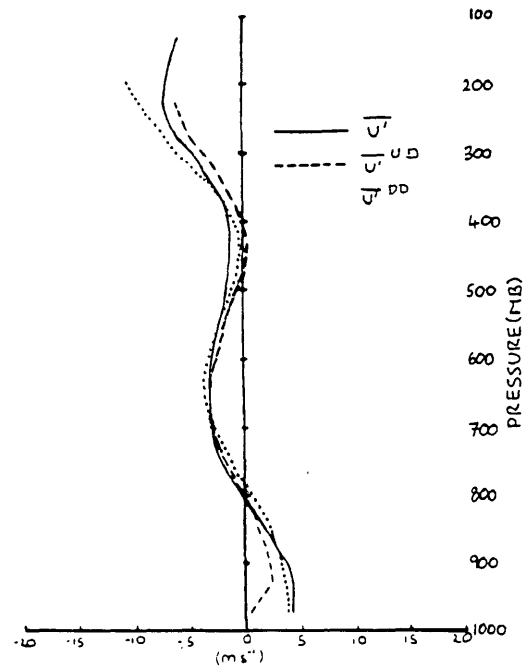


Fig 8.6(a) $-\partial \bar{\omega' u'} / \partial p$, together with cloud and environmental contributions, for sim J, averaged between 1 and 9 hours.

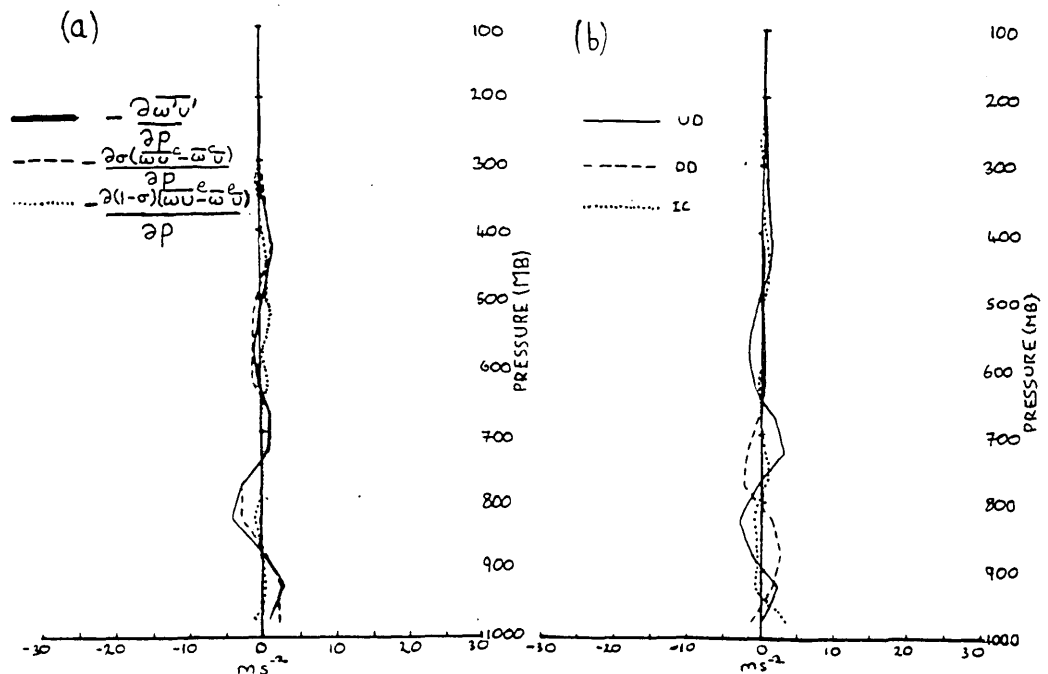


Fig 8.6(b) Contributions of UD, DD, and IC to $-\partial \bar{\omega' u'} / \partial p$ for sim J, averaged between 1 and 9 hours.

gradients, and momentum transports. However observations of line convection (LeMone(1983)), and 3D modelling of tropical squall lines (Miller and Betts(1977)) shows that the shear normal to such systems is increased by cloud scale momentum transports. Perhaps more importantly the large-scale momentum forcing, and pressure gradients may dominate cloud transports, so causing a decrease in the wind shear observed.

8.2.3 Parametric Interpretation Of Momentum Budgets.

The study of parametric momentum budgets presented here is brief, concentrating upon compensating subsidence, advection, en/detrainment and cloud scale pressure gradients. Averages between 1 and 9 hours are presented.

Fig 8.7(a) shows $(CS)_u$ due to all cloud. As for thermodynamic fields the profile shape is dominated by the UD contribution. $(CS)_u$ follows $\partial\bar{u}/\partial p$ and is a poor representation of $-\partial\sigma(\bar{w} - \bar{w}^c)/\partial p$. Above 650mb $(CS)_u$ is opposite to the cloud eddy transport. This gives further indication that cloud scale momentum transports in the upper half of the domain occur by a mixing length process proportional to $-\partial\bar{u}/\partial p$.

The cloud scale pressure gradient $(CPG)_u$ is a large-term, as suggested earlier, and opposes $(CS)_u$ (fig 8.7(a)). Their combination brings worse agreement with cloud eddy transports below 650mb, but better agreement above.

Like $(CS)_u$ cloud scale pressure gradients are dominated by UD above 900mb, DD being important below this. Fig 8.7(b) shows $(CPG)_u$ for UD and DD, and $(CS)_u$ for UD. For UD $(CPG)_u$ is roughly proportional to $-\partial\bar{u}/\partial p$. This observation may provide a simple method of parameterizing the effects of cloud scale pressure gradients in simple cloud models. The same is not true for DD. Here $(CPG)_u$ is only important below 600mb (the region associated with convective scale DD from thermodynamic studies), increasing roughly linearly to the surface.

Fig 8.7(a) $(CS)_u$, $(CPG)_u$, $(CS+CPG)_u$, for sim J,
averaged between 1 and 9 hours.

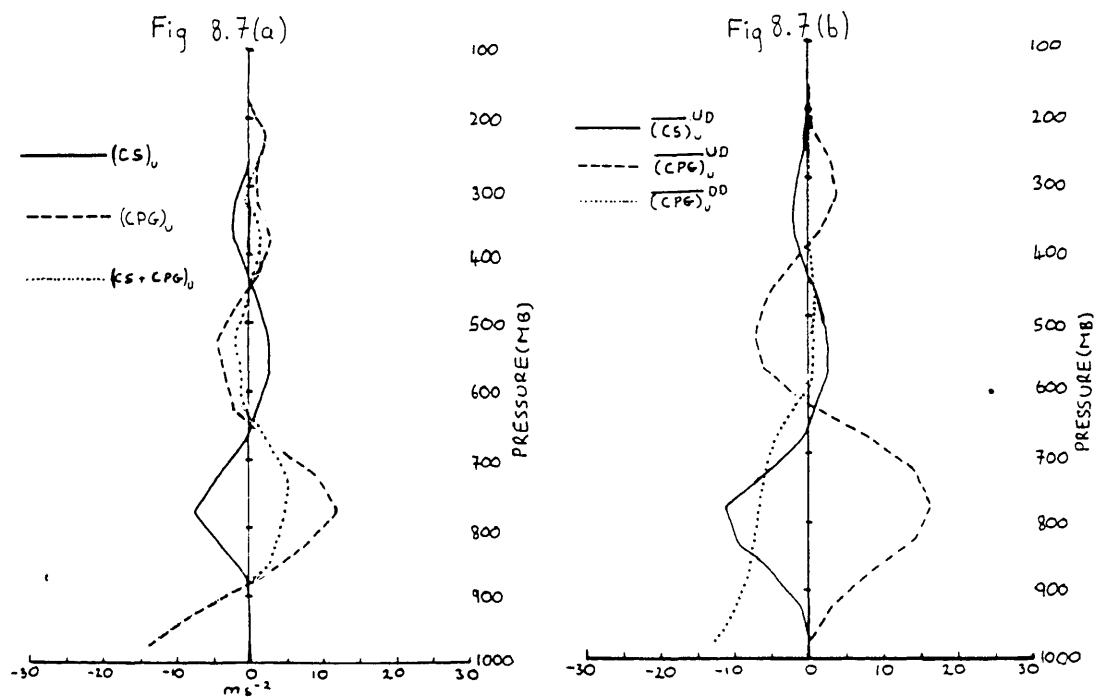


Fig 8.7(b) $(CS)_u^{UD}$, $(CPG)_u^{UD}$, $(CPG)_u^{DD}$, for sim J,
averaged between 1 and 9 hours.

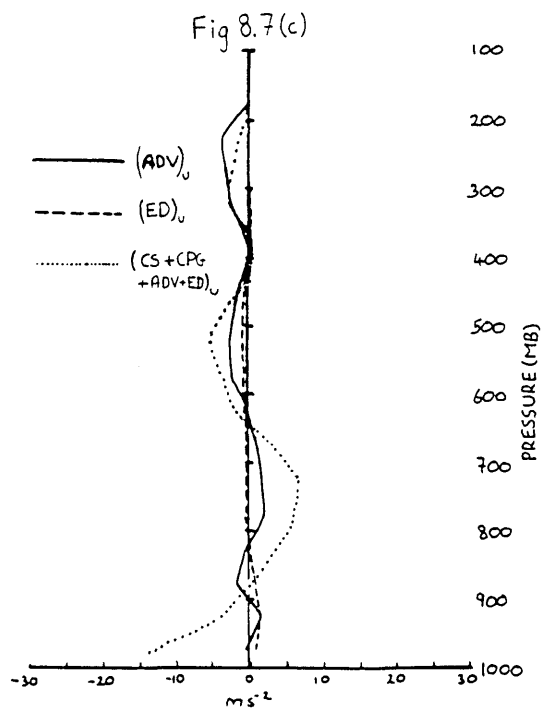


Fig 8.7(c) $(ADV)_u$, $(ED)_u$, $(CS+CPG+ADV+ED)_u$, for sim J,
averaged between 1 and 9 hours.

Inclusion of horizontal terms $((ADV)_v, (ED)_v)$ brings no improvement to the simple representation of $(Q3)_{conv}$, and makes it worse above 650mb (fig 8.7(c)). Improvement requires the inclusion of time dependent and boundary terms. As stated in chapter 6 these cannot be obtained from simple cloud models. Thus the parameterization of cloud momentum transports appears difficult.

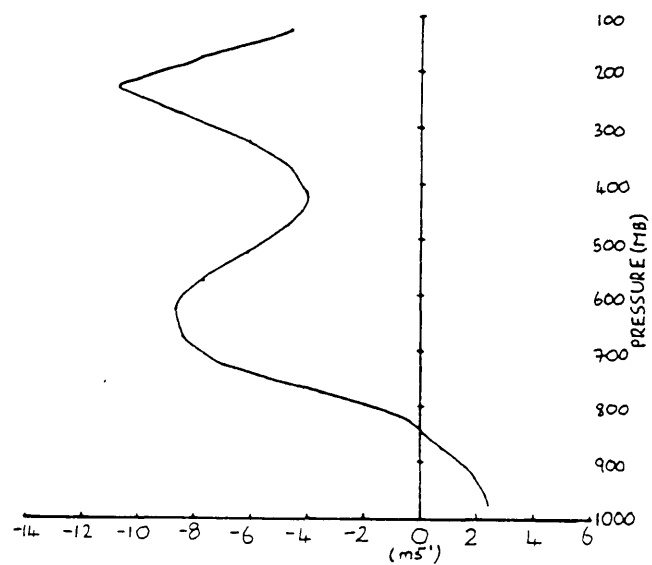
8.3 THE EFFECTS OF STRONGER SHEAR UPON CONVECTION (SIM K).

As stated earlier the 20.5 hour low level shear is weak. Stronger shears do exist within the TPRR data set, associated with periods of weak convection. However these are only $\approx 50\%$ of those associated with large organized tropical systems. This is probably due to averaging over the entire GATE area.

An experiment was carried out equivalent to sim H, but with the 2.5 hour initial wind shear (fig 8.8). This inhibited the formation of deep convection until 420 minutes. The surface rainfall between 1 and 9 hours was $220Wm^{-2}$, double that of sim H, even though convection takes place for a much shorter time. This is due to the surface rainfall in the last 3 hours of the integration being $780Wm^{-2}$, compared to $30Wm^{-2}$ between 1 and 6 hours. The convection appeared more organized than in previous simulations, new cells forming to the west of existing ones. Hence the increased shear aids the storage of water vapour in the atmosphere during the 2-10 hour period, and also produces organization in the convection once formed.

To study the organizing effect of the stronger shear further, a final experiment was carried out. Sim K had the initial thermodynamic conditions and forcing as sim B (known to allow deep convection), but was also initialized with the 2.5 hour zonal wind shear. The presence of the increased shear delays the onset of convection until 200 minutes, growth being gradual. Once formed the domain is dominated by one large convective system, which propagates from east to west. Fig 8.9 is a plot of cloud area with time showing the motion. Few other clouds penetrate 500mb.

Fig 8.8 2.5 hour zonal velocity profile of the GATE composite easterly wave.



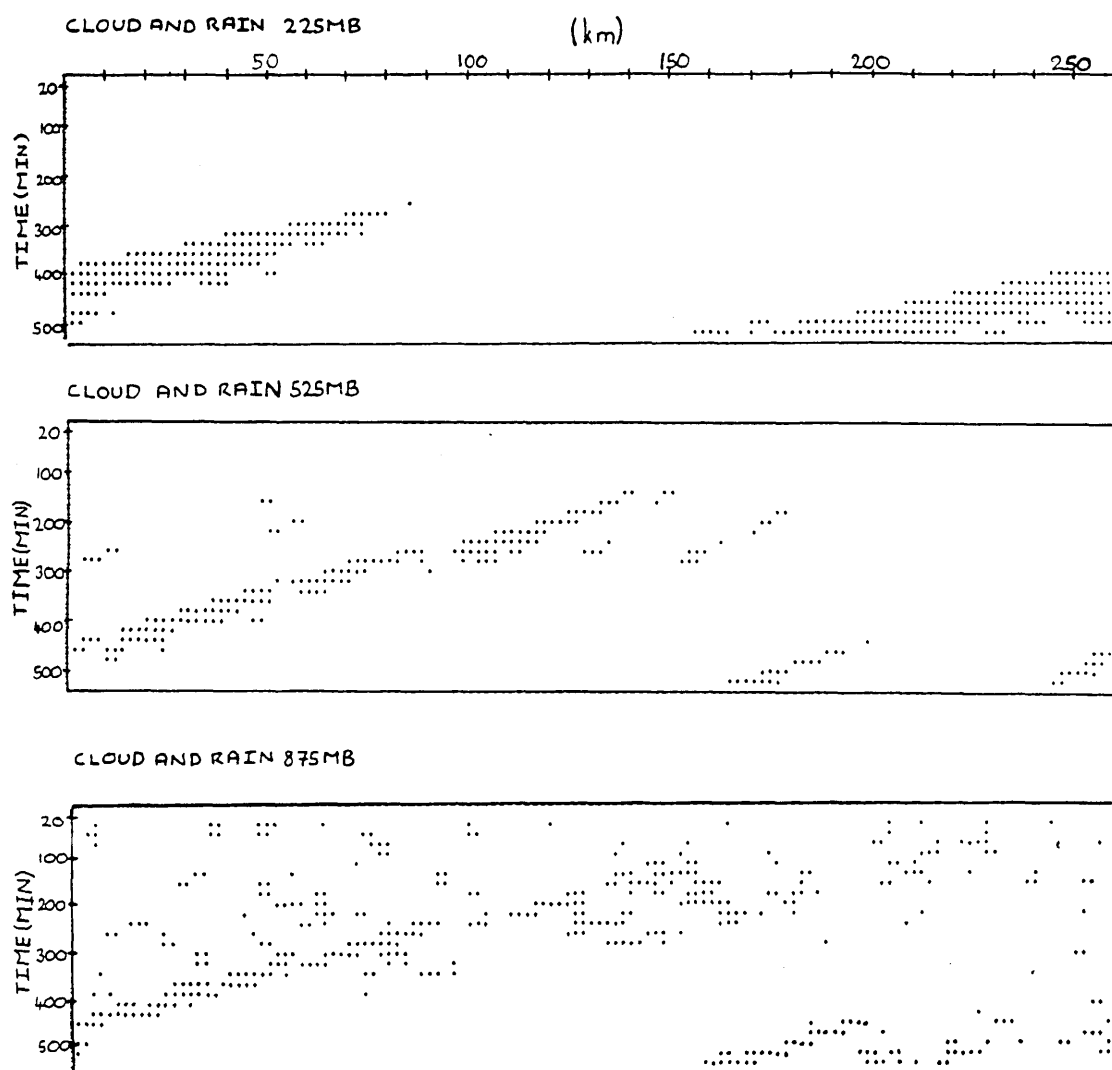


Fig 8.9 Plot of cloud area against time for 225, 525, and 875mb, for sim K, plotted every 20 minutes.

The system has squall line characteristics, with an UD sloping against the mean wind shear, and outflow ahead and behind the storm. The DD forms a density current. New cells form to the west of the density current, which is between -1 and $-1.5K$ colder than air upstream of the system and approximately 100mb deep. This implies a propagation speed of 6 to $7ms^{-1}$. The system is observed to move at 7 to $8ms^{-1}$ to the west, agreeing approximately with the propagation speed of the density current. This is typical of squall lines. Fig 8.10 illustrates this motion.

The propagation velocity of the system is the velocity of the mean flow at 700mb. This is similar to the transient clouds of sim J (weaker shear). This is further evidence that the transient convection which develops in the case of weak low level shear has some squall line characteristics.

8.3.1 Thermodynamic Properties And Cloud Structure.

$\langle c_p \pi(Q1)_{conv} \rangle$ between 1 and 9 hours is $650Wm^{-2}$, removing 92% of the applied moisture forcing. Thus the system is more efficient than either the unsheared or weakly sheared simulations (sim B and J). Twenty minute averages of $\langle c_p \pi(Q1)_{conv} \rangle$ (fig 8.11) show that the initial development is more gradual than sim B or J. Several cells are seen to form, but greater overlap in cloud life span is seen, giving more continuity in convective activity.

The vertical structure of $(Q1)_{conv}$ and $(Q2)_{conv}$ is shown in fig 8.12. Both are larger throughout the domain than sim B, especially between 800 and 500mb. Eddy transports are very similar, sim K being slightly stronger than sim B between 800 and 300mb, for both θ and q .

Contributions of UD and DD to $(Q1)_{conv}$ are slightly larger than seen in sim B, $\approx 180\%$, and $\approx -90\%$ respectively. However the ratio between the two contributions is roughly the same.

Fig 8.13 compares the UD structure of sim K and B. Buoyancy and $\overline{\omega''\theta''}$ are similar, buoyancy being slightly larger in mid-levels. Difference is seen when the cloud area is considered, UD in sim K

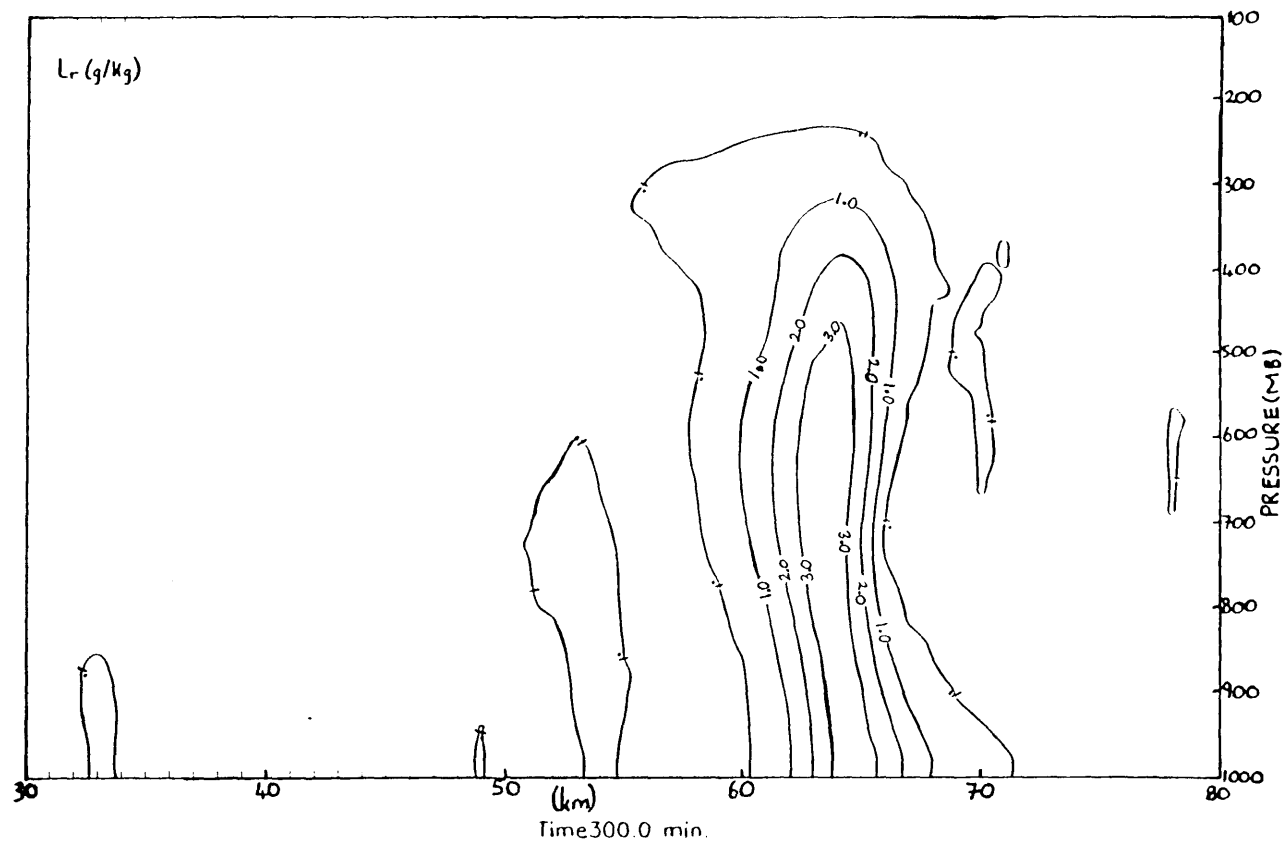


Fig 8.10 Cross-sections of squall line system of sim K, for 300 and 340 minutes, showing the east-west propagation.

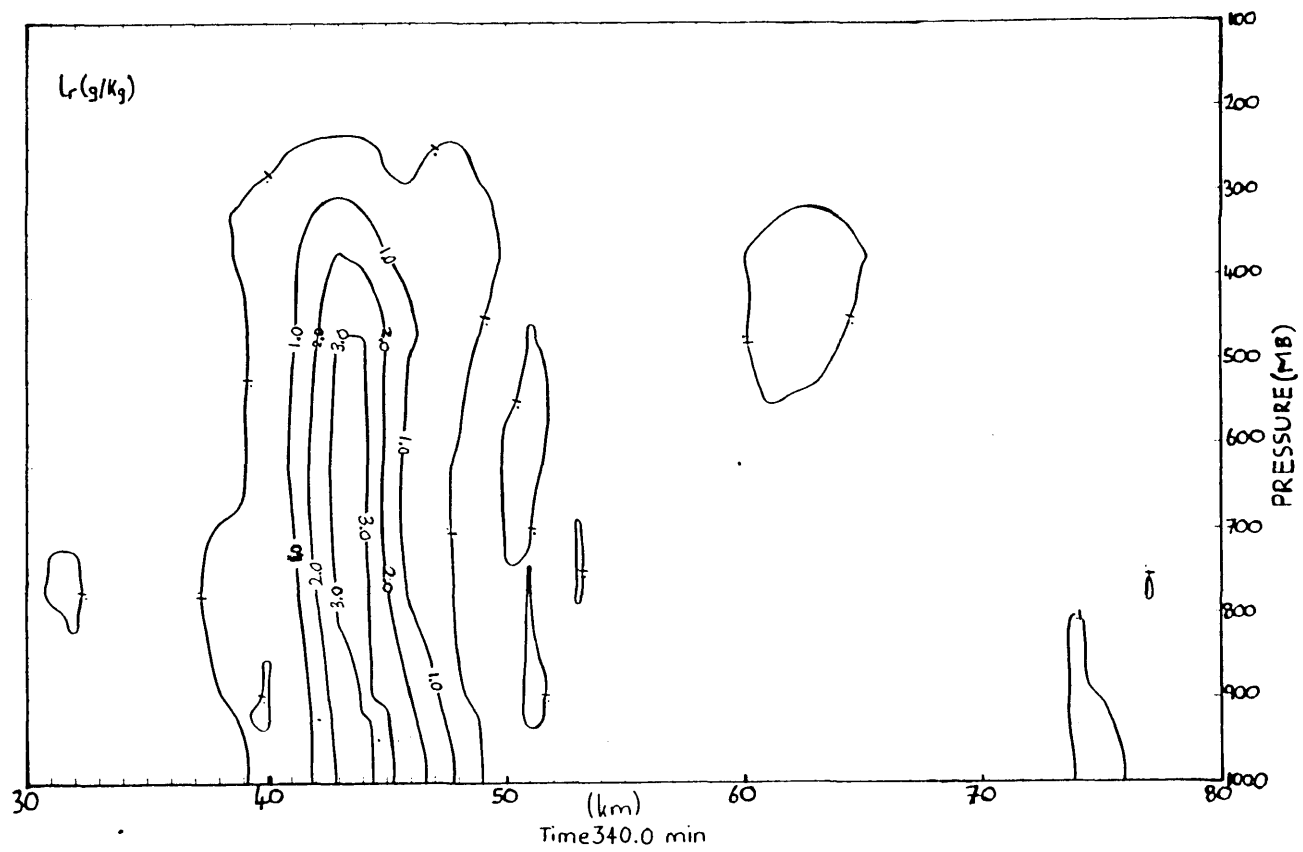


Fig 8.10 cont

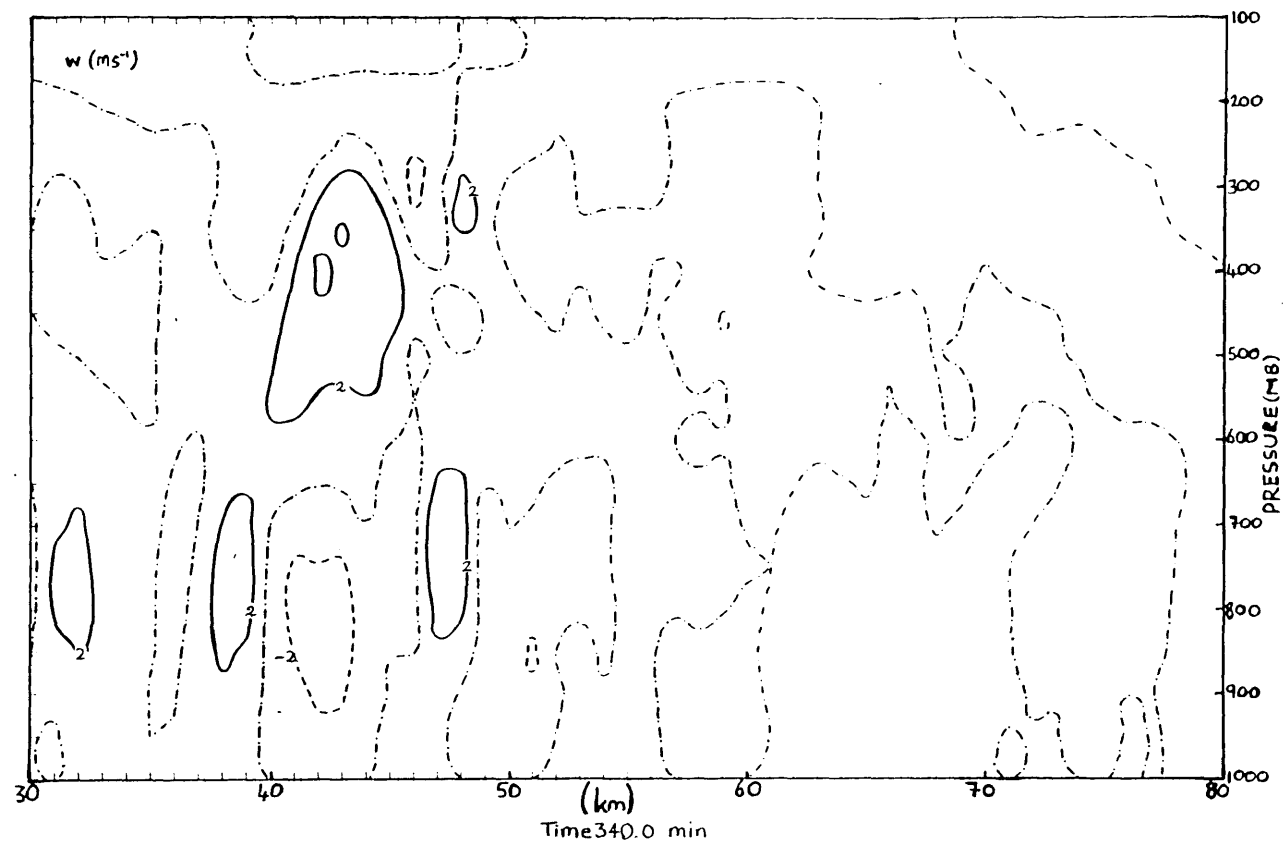


Fig 8.10 cont.

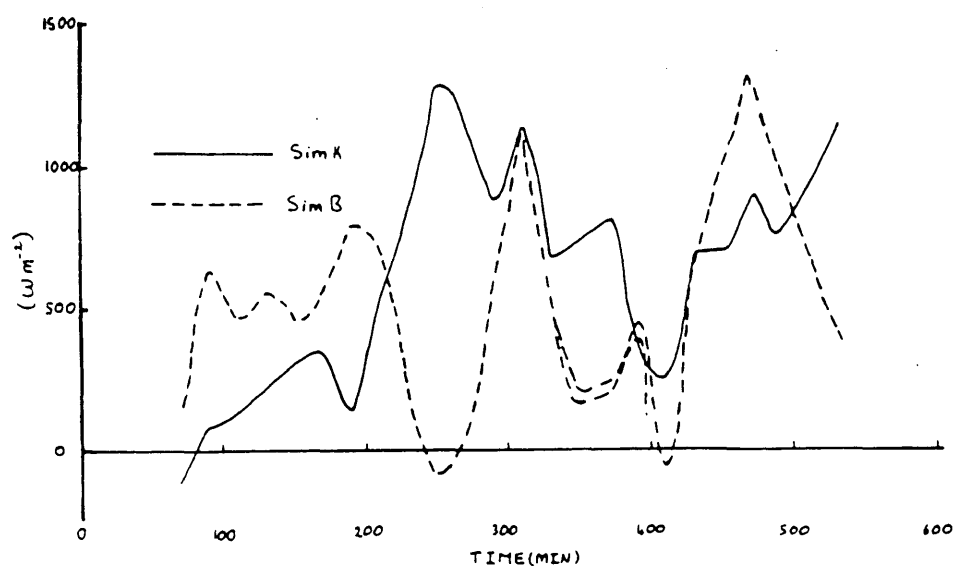


Fig 8.11 Comparison of $\langle c_p \pi(Q1)_{\text{conv}} \rangle$ for sim B and K, 20 minute averages.

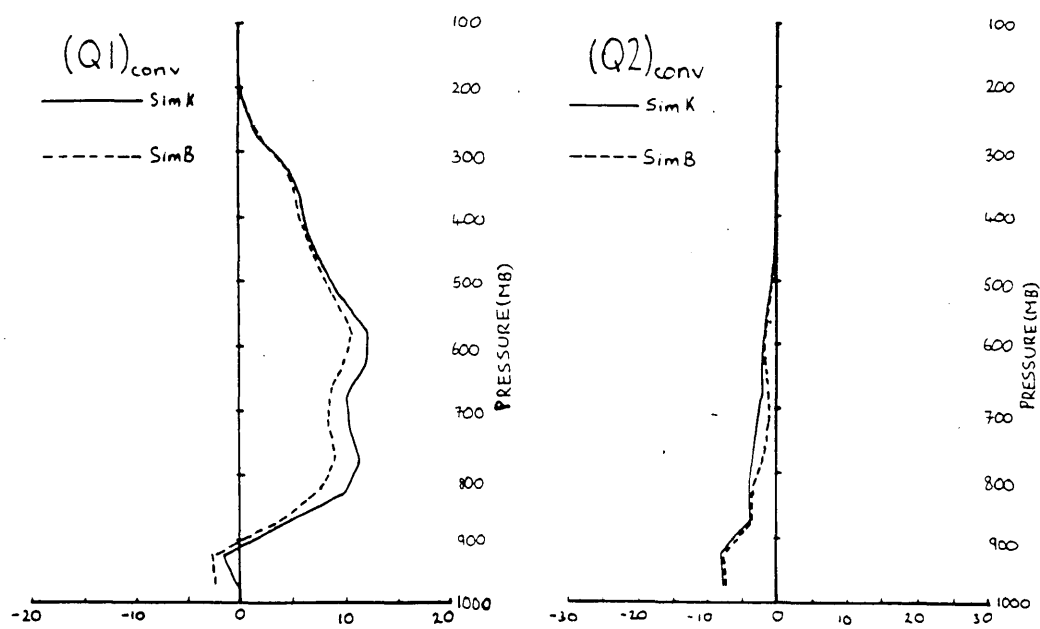


Fig 8.12 Comparison of $(Q1)_{\text{conv}}$ and $(Q2)_{\text{conv}}$ for sim B and K, averaged between 1 and 9 hours.

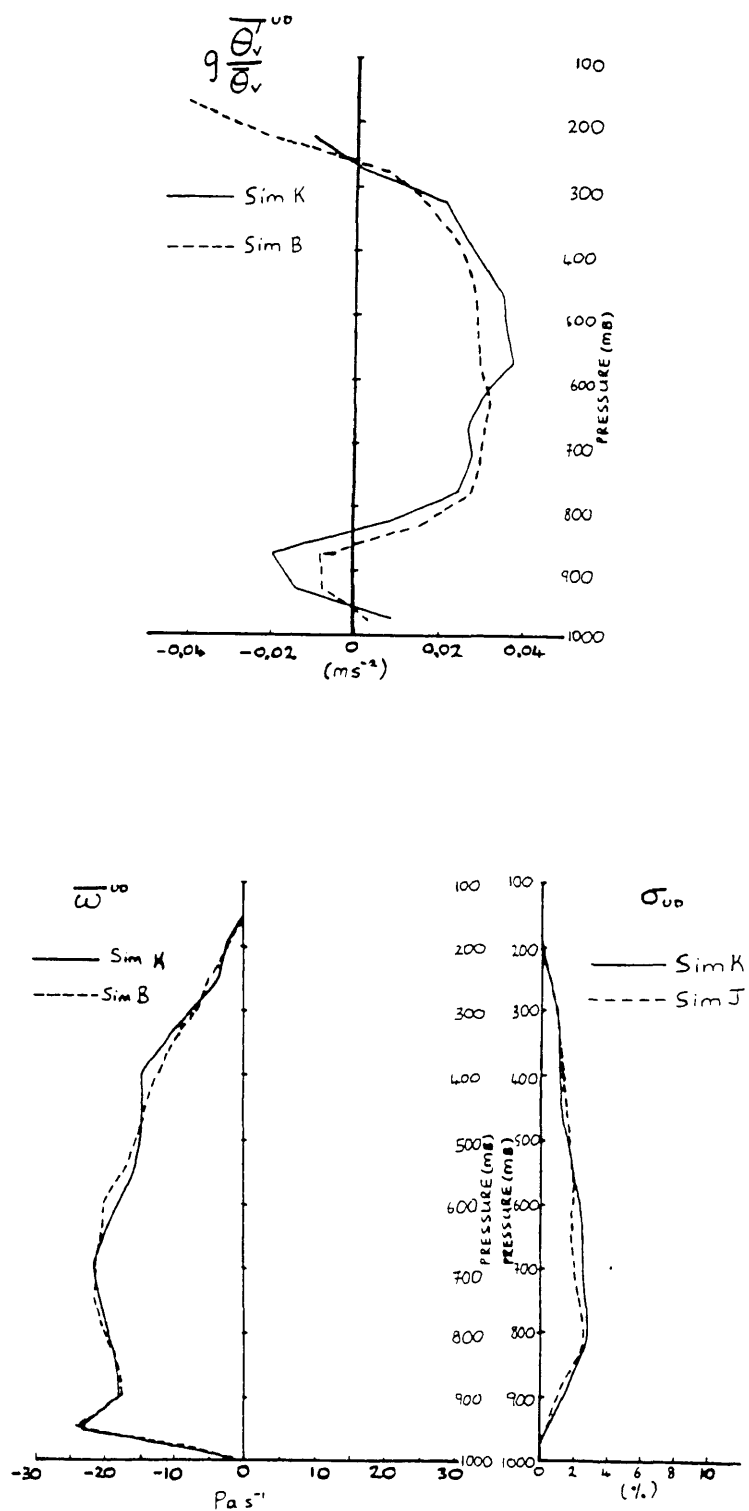


Fig 8.13 Comparison of $g \frac{\overline{\theta_v'}^{\text{up}}}{\overline{\theta_v}}$, $\overline{\omega}^{\text{up}}$, and σ_{up} for sim B and K, averaged between 1 and 9 hours.

being broader than in sim B, implying a greater efficiency in removing water vapour from the atmosphere.

Thus although the cloud system is more organized due to the large initial shear, little difference is seen in the thermodynamic features of the storm.

8.3.2 Momentum Transports.

Fig 8.14 shows internal UD and DD horizontal momentum. Differences are apparent with those of sim J. The deviation of $\bar{u}''$ from $\bar{u}$ is larger, being approximately constant between 700 and 400mb. Eddy momentum transports are also different. This is shown in fig 8.15(a), together with contributions from cloudy and clear air. Below 650mb, the structure is similar to that observed in sim J. Above this differences are seen. Eddy transports accelerate the atmosphere to the east between 650 and 200mb. Such momentum transports have been observed to be characteristic of tropical squall lines (Miller and Betts(1977)). This would imply that clouds are more organized above 650mb than in the weak shear case. Further evidence for this is seen in fig 8.10 where the UD is seen to slope against the shear up to 300mb. The upper level cloud is also observed to propagate with the speed of the low level jet maxima, whereas in sim J it moved with the local wind speed.

Again UD dominates the structure and is seen to be responsible for the differing transport properties above 650mb (fig 8.15(b)). DD transports have a similar structure as sim J. They are more important in the upper half of the domain than previously. Again DD effects are opposed by IC.

The parametric budgets are similar to those of the weak shear case. Again pressure gradients are important, and interpretation complex.

Hence the organization of convective motions into a propagating system does effect the structure of cloud momentum transports, and should be considered when parameterizing such quantities.

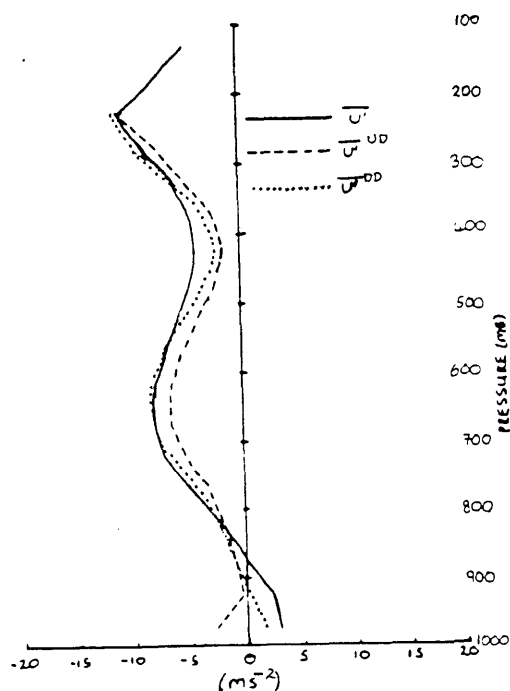


Fig 8.14 $\bar{u}^{up}, \bar{u}^{dp}, \bar{u}$ for sim K, averaged between 1 and 9 hours.

Fig 8.15(a) $-\partial \bar{\omega} \bar{u} / \partial p$, together with cloud and environmental contributions for sim K, averaged between 1 and 9 hours.

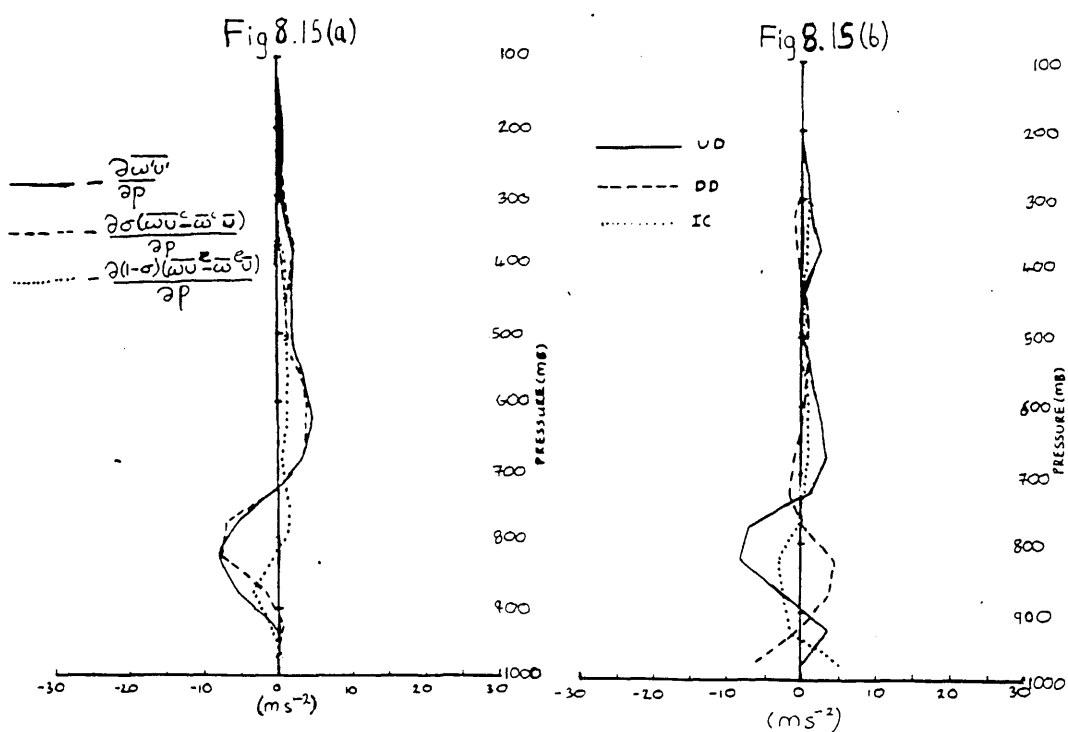


Fig 8.15(b) Contributions of UD, DD, and IC to $-\partial \bar{\omega} \bar{u} / \partial p$, for sim K, averaged between 1 and 9 hours.

8.4 SUMMARY.

A brief study of the effects of shear upon tropical convection, and convective scale momentum transports has been presented. Weak shears have little effect upon cloud thermodynamics. They do cause some organization in the lower part of the atmosphere, causing momentum transports to have a similar structure to more organized tropical systems. Stronger shears produce greater organization, one propagating system (with squall line structure) dominating the domain. Thermodynamic efficiency is increased by $\approx 10\%$, and momentum transports in the upper troposphere differ from the weak shear case.

Interpretation of parametric momentum budgets is complex, and a simple representation of cumulus friction does not appear possible. Cloud scale pressure gradients cannot be neglected, and the assumption of Schneider and Lindzen(1976), that $\bar{u}$ is approximately conserved during ascent has been seen to be incorrect.

This study is very limited. Different shears are known to give different organization, and momentum transports. Zonal and meridional transports are observed to differ for some clouds, and three dimensional effects may be important. Much more work is required if momentum transports are to be parameterized correctly.

CHAPTER 9

SUMMARY AND CONCLUDING REMARKS.

9.1 THE MODELLING OF A CLOUD ENSEMBLE.

In this study the meso-scale model first described by Miller and Pearce(1974), in a two dimensional form, has been used to model an ensemble of tropical deep convection, and study aspects of convective parameterization. Data from the TPRR easterly wave composite data set, analysed from GATE phase III data, was used to initialize the model, and also provide values for large-scale forcing and surface fluxes. Radiative cooling was provided by the profile calculated by Cox et al(1979) for the GATE area. Cloud forcing is provided by a sequence of random numbers applied at 925 and 875mb to the theta field. Further details of the methods used can be found in chapter 3. For most integrations no initial wind field was communicated to the model.

9.1.1 Comparison With Observations.

Category 3 (20-30 hours) of the easterly wave has been concentrated upon. During this period, deep convection was present, growing to maximum amplitude during category 4. Although the

representation of large-scale, radiative, boundary layer, and cloud forcings are crude, comparison of the simulated convection with observations, shows that the model is able to represent the convective response of the atmosphere during this period of the wave reasonably well.

The convection warms and dries the atmosphere, in opposition to the effects of large-scale forcing. Quantities which cannot be directly obtained from observational data have been calculated (condensational heating, eddy transports, liquid water budget). Because agreement with the observations is good, some confidence can be placed in such quantities.

9.1.2 Short Time Scale Structure.

The model allows the effects of cloud evolution upon cloud transport properties to be investigated. On a twenty minute time-scale, time dependent effects are seen, and these are perhaps important for the parameterization of convection within coarse meso-scale models. As well as warming and drying the atmosphere, convection is seen to cool and moisten a large depth of the domain (up to 500mb). This is associated with the decay of convection, evaporation of residual cloud occurring, together with a change in vertical transport structure.

9.1.3 Inadequacies Of A Two Dimensional Model.

Several inadequacies in the use of a two dimensional model to study an ensemble of clouds are apparent from this study. The moisture field is not well represented, convection causing too much drying in the lowest 100mb, and too much moistening above. Also the response of the atmosphere on a two hour time-scale is not steady, as indicated by the TPRR data set. This because too few clouds exist within the domain, and insufficient overlap in cloud life span occurs. However the eight hour average is in good agreement with observations, and is taken to be representative of a three dimensional ensemble on a

shorter time-scale.

The use of a three dimensional domain would prevent these problems from occurring.

9.1.4 Cloud Structure And Analytical Cloud Models.

A realistic spectrum of clouds exists within the simulation. However the "mean ensemble cloud" is dominated by deep convection. Weaker convection is important in specifying the shape of σ_w below 500mb. The large number of shallow clouds within the domain contribute little to the mean cloud structure, and the energy budget of the model. For deep convection, water loading is seen to be important in specifying the net cloud buoyancy.

Brief comparison has been made with one dimensional cloud models, used in many parameterization schemes. These are seen to be a poor representation of the mean life cycle of deep convective cloud. They appear to be similar to the mature structure of deep clouds. It may be possible to improve analytical cloud models by using a simple evolution of the buoyancy profile, specified by a numerical cloud model.

No meso-scale convective feature was seen in any of the simulations, either un-sheared or sheared, and so the physics and parameterization of such features has not been commented upon. Inclusion of an ice phase into the model cloud microphysics scheme is probably necessary to represent such features.

9.2 THE PARAMETERIZATION OF CONVECTION.

Having seen that the model adequately represents a tropical cloud ensemble, it has been used to study the parameterization of deep convection. Fuller parametric forms of $(Q1)_{conv}$ and $(Q2)_{conv}$ have been derived, than those of previous workers. Several new terms have been introduced, although under the assumptions of steady state, and

constant cloud area with height, the formulations are similar. These equations have been diagnosed using data from the meso-scale model to determine "simple" representations of $(Q1)_{conv}$ and $(Q2)_{conv}$. The model has also been used to investigate the closure problem of cumulus convection.

9.2.1 The Representation Of $Q1$.

$(Q1)_{conv}$ is well represented by compensating subsidence, plus evaporation within cloud environment. Horizontal transports across cloud boundary (advection and en/detrainment) are unimportant. Time dependent effects and those involving a change in cloud area with height balance, and so approximations of a steady state ensemble, and $\sigma \approx \text{constant}$ are reasonable. Hence the results of previous workers analysing observational data, and the methods used to represent $(Q1)_{conv}$ in parameterization schemes are vindicated.

Many parameterization schemes consider updraught effects only. The model results show that both updraught and downdraughts are necessary to specify $(Q1)_{conv}$ well. Alone updraughts effects are twice $|(Q1)_{conv}|$, downdraughts effects having a value of approximately $|(Q1)_{conv}|$, opposing the updraught. The effect of inactive cloud is important in the lowest 200mb of the domain, associated with updraught inflow.

9.2.2 The Representation Of $Q2$.

The budget value of $(Q2)_{conv}$ is reasonably represented by compensating subsidence, and the evaporation of liquid water within cloud environment. However this is not true when the profile of these quantities is considered. The net of these two terms peaks around 700mb, instead of near the surface. Addition of other terms is necessary for better agreement.

Horizontal transports across cloud boundary are more important in the parametric moisture budget. It is found that these dry the atmosphere in the updraught inflow layer. This effect is not usually incorporated into parameterization schemes, cloud inflow air being assumed to have the thermodynamic values of the large-scale.

Addition of other terms to the approximation to $(Q2)_{conv}$ brings better agreement above 800mb. However even if all terms are included comparison of $(Q2)_{conv}$ and $(Q2)_{para}$ shows large differences below 800mb.

9.2.3 Environmental Eddy Flux And Shallow Convection.

The difference between $(Q2)_{conv}$ and $(Q2)_{para}$ is due to the neglect of the contribution to $\overline{\omega'q'}$ from environmental air, when deriving expressions for $(Q1)_{para}$ and $(Q2)_{para}$. Calculations have shown that this is only important in the lowest 200mb of the domain. It is caused by weak dry circulations driven by the random cloud forcing. The effect is also present in the sensible heat budget, but is weaker than in the moisture budget.

Although the circulations are dry, moisture transports are similar to those due to shallow convection observed over the tropical oceans. Doubts must be expressed for the temperature field as the effects of liquid water transports are important. Hence although the cloud forcing technique is crude, the statement that it is introduced to represent the forcing of deep convection by shallow convection seems reasonable in part.

Thus the effects of shallow convection are important in specifying the $(Q2)_{conv}$ profile. This has been suggested by previous workers (Kuo and Anthes(1984c)). Hence the inclusion of several cloud types in convective parameterization schemes would seem necessary.

However as stated above, time dependent and cloud boundary terms are also needed to provide a good representation of $(Q2)_{conv}$. These terms cannot be simply calculated from analytical cloud models. Hence using mass flux techniques, together with simple cloud models, a simple representation of $(Q2)_{conv}$ does not exist (unlike $(Q1)_{conv}$). Other

techniques need to be developed.

9.2.4 The Closure Problem.

The relationship between convective magnitude, large-scale structure and processes has also been studied in relation to the closure problem of cumulus parameterization. Several simulations with differing initial profiles and large-scale forcing have been presented. The Kuo and Arakawa-Schubert closure schemes have been considered in detail, with brief comments being made on CAPE and the new Betts-Miller adjustment scheme.

(a) Rainfall, moisture forcing, and storage.

The basis of the Kuo closure scheme is that convection rains out a fraction (1-b) of the large-scale and boundary layer moisture forcing. For category 3 of the composite easterly wave the model indicates that convection rains out $\approx 80\%$ of the moisture convergence, in good agreement with observed values. Hence the model responds in a realistic manner to the external forcing supplied (even though methods used are crude).

The model was used to investigate the response of convection to changes in the magnitude of the moisture forcing. This was found to be linear if deep convection is possible. However the composite easterly wave data set indicates that the response is non-linear. For category 1 the moisture forcing is $\approx 50\%$ of category 3, the convection only raining out $\approx 30\%$ of this, the remainder being stored by the atmosphere.

The model has been used to investigate this phenomenon. The storage is caused by the inability of the thermodynamic profile to support deep convection due to dryness above the lowest 200mb of the domain. This causes increased entrainment effects in growing clouds, decreasing their maximum height. Clouds actively aid the large-scale forcing in moistening the atmosphere through detrainment and evaporation.

These results suggest that the Kuo parameter, b , needs to vary with atmospheric structure. This has been suggested by Anthes(1977), and tested by Kuo and Anthes(1984c). Using the model, convection appears to be very sensitive to a small change in the relative humidity of the atmosphere, much more so than suggested by Kuo and Anthes. However they used mid-latitude data for their calculations. This may indicate that a further variation of b is required between mid-latitude and the tropics.

Hence the control of convective magnitude in the tropics is complex, involving both large-scale dynamical control (moisture forcing), and thermodynamical control through the atmospheric profile.

(b) Cases when $b < 1$.

It is sometimes observed that convection rains out more moisture than the large-forcing supplies. Kuo type schemes cannot represent such periods. An attempt was made to simulate category 5 of the TPRR data set, where such a situation exists. However the simulation failed, the convection decaying with the large-scale forcing, an almost linear response being seen.

The failure seems to be due to an initialization problem. If convective circulations had been present at the start of the decay of large-scale forcing, convective activity would have been maintained. This suggests that a time delay of several hours should be incorporated into Kuo type schemes, when moisture forcing decays, if the atmospheric profile can still maintain deep convection.

(c) The Arakawa-Schubert closure hypothesis.

The A-S closure hypothesis is more general than that of the Kuo scheme. It has been tested using observational data, considering updraughts only, and found to be valid. Calculations have been carried out using model clouds, which contain representations of both updraughts and downdraughts, and also a reasonable representation of water loading. If water loading effects are considered, then the closure hypothesis is valid. If they are neglected then it is not.

The model also indicates that the scheme is valid when large amounts of water vapour are being stored in the atmosphere. However in such a situation the scheme will only work if the cloud model used is able to simulate the variation of cloud top height with changes in the thermodynamic profile.

(d) Kuo and A-S closure hypothesis on short time-scales.

Both Kuo and A-S closure methods do not work on short time-scales (on the order of cloud lifetime). Hence they are unsuitable for use with coarse meso-scale models. However there is evidence that on a time-scale of a few hours they are valid, if the atmosphere has had enough time to de-stabilize after the effects of convection.

(e) CAPE.

CAPE has been used to determine convective activity in several schemes, mainly for use in coarse meso-scale models. The model shows that CAPE is sensitive to convective activity, but does not predict its magnitude on a point to point basis. CAPE reduces rapidly when convection occurs, and grows when convection decays. No critical value of CAPE, above which convection occurs, is apparent.

(f) The Betts-Miller adjustment scheme.

The Betts-Miller adjustment scheme uses the saturation point parameter to define a characteristic thermodynamic profile, to which the large-scale model thermodynamic profiles are adjusted under the action of deep convection. The model indicates that the saturation point parameter is sensitive to convective activity, and so is suitable for parameterization. The model thermodynamic profile has a similar structure to that suggested by Betts and Miller for a convective atmosphere, and so the scheme would seem reasonable.

9.3 THE EFFECTS OF SHEAR.

A brief study of the effects of wind shear upon convection was made using the zonal wind profiles observed over the GATE area, obtained from the TPRR data set. Weak low level shear has little effect upon the thermodynamic properties of the ensemble. With stronger shear, a Squall line type system dominates the domain, with increased thermodynamic efficiency ($\approx 90\%$ of the moisture forcing is rained out). Momentum transports are stronger for the Squall line system, although both cases have similar structure below 500mb, increasing the low level shear. This indicates that even transient clouds have some organization. The increase in low level shear is opposite to that observed. This could be caused by the lack of large-scale momentum forcing, and pressure gradients in the model, which may exceed cloud effects.

Parametric interpretation of momentum transports is difficult. Cloud horizontal momentum is not conserved during ascent, as assumed in some momentum parameterization schemes. A contributing factor is the large cloud pressure gradients seen. Compensating subsidence is a bad representation of $(Q3)_{\text{conv}}$, all other terms being required for improvement. It is clear that much more work is required before an effective momentum parameterization scheme is developed.

9.4 CONCLUDING REMARKS AND FURTHER WORK.

From this study a numerical cloud model is shown to be a useful tool for studying the parameterization of convection, and the interaction of cloud and large-scale processes. The response of the simulated convection to variations in thermodynamic profiles and large-scale forcing agrees well with observations. Data can be obtained on a much finer spatial and temporal scale than from any observational network. Cloud structure is of much higher quality than that obtained from observations and analytical cloud models. Hence more confidence is placed in the diagnosis of mechanisms by which clouds effect their environment. However this study of parameterization is not extensive and much further work can be contemplated.

Further work on the cloud initialization scheme would prove useful, improving the "parameterization" of shallow convection within the domain. This, perhaps, could be linked with the inclusion of a boundary layer representation within the model.

The study of cloud structure diagnosed from the model could be used to improve analytical representations of convection, including the dynamical models of Moncrieff. Thus both parametric techniques, and the methods used for diagnosing the nature and effects of convection from observations could be improved.

Comparison of an ensemble model with single column tests of parameterization schemes is also possible. Further work, regarding the response of convection to changes in atmospheric thermodynamic structure, and large-scale forcing might lead to better empirical relationships for use with the Kuo scheme, so that storage effects are represented. Improvements in the Betts-Miller adjustment scheme could be achieved through the use of numerical models.

These studies would be improved by the use of a three dimensional model. However due to computational limitations it may be necessary to use coarser resolution than here (perhaps 2000m, and 100mb). Comparison of two dimensional models with this grid spacing with the model presented in this study, would show if the coarser resolution is able to represent the mean effects of an ensemble. A three dimensional model would provide a better representation of the moisture field, and also a greater steady state response. More definite statements concerning the spatial and temporal scales over which parameterization schemes are valid could be made.

A numerical model of a single cloud would provide information on the scale of the cloud subsidence region, important when considering parameterization in coarse meso-scale models. Using open lateral boundaries, horizontal eddy transports could be derived, and their parameterization correctly considered.

Much more can be achieved concerning the parameterization of momentum transports, and the effect of organization upon cloud thermodynamic effects. The response of convection to differing zonal and meridional shears can be diagnosed. The use of Moncrieff's

dynamical models in parallel to these studies may bring greater understanding of how to parameterize cloud momentum transports.

No meso-scale convective feature was seen in the simulations presented in this thesis, perhaps because of the lack of an ice phase in the cloud microphysics scheme. With a domain size of several hundred kilometers, as in this study, such features could perhaps be modelled to high resolution. Using such a model their structure and parameterization could be investigated.

This study has been carried out using tropical data. The use of mid-latitude data is also possible, although incorporation of cloud microphysics concerning the ice phase may be necessary.

By using a meso-scale model the interaction of convection with larger scales has been seen to be more complicated than previously thought. The use of such models will prove to be a powerful tool for studying the parameterization of convection, and developing new schemes in the future.

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Abbreviations.

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 JAS Journal of Atmospheric Sciences
 MWR Monthly Weather Review

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