Response of Cumulus Updraft and Downdraft to GATE A/B-Scale Motion Systems

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ABSTRACT

Large-scale mass, heat and moisture budgets have been computed over the GATE A/B-scale area during two priority periods in Phase 3. The computed budget results are well-correlated with the activities of the cloud clusters which develop and decay during the analyzed period. Strong upward motion exists with two maxima at 700 and 350 mb during the active period of the cloud cluster, but sinking motion appears in the middle troposphere during the dissipating period. Large amounts of apparent heat sources and moisture sinks occur in the whole troposphere and large amounts of heat energy are transported upward by cumulus clouds during the active period; but an apparent heat sink, moisture source and slight downward heat flux are observed in the middle troposphere during the dissipating period.

A diagnostic method for determination of cloud properties has been improved by including downdraft effects and tested on the mean Marshall Islands data. In this method both spectral properties of updrafts and bulk properties of downdrafts are determined without prescribing downdraft parameters. The downdraft acts as a heat sink and a moisture source and neglect of the downdraft overestimates the mass flux of shallow clouds. The downdraft originates from air of the environment somewhere in the middle troposphere and extends below the cloud base keeping temperatures cooler than the surrounding air.

This diagnostic method is applied to three different budget results in GATE which are classified according to the activities of the cloud clusters. Bimodal cloud mass flux distributions for updrafts with dominance of very shallow and very deep clouds are obtained during the period when the GATE area is not affected by cloud clusters. However, unimodal distributions are found with dominance of very deep clouds during the active stage of cloud clusters and with dominance of shallow clouds during the dissipating stage. Greater cloud downdraft mass flux is observed in GATE than in the Marshall Islands area and, specifically, net downward cloud mass flux results from the dominance of downdrafts during the dissipating period. The relation between vertical distributions of the downdraft mass flux and vertical shear of ambient winds is also discussed.

1. Introduction

The understanding of tropical meteorology has been advanced quite extensively during the past decade. Observational studies have discovered a variety of new phenomena in the tropical atmosphere and many theories have been proposed for explaining the mechanisms of these phenomena. A lot of attention has been focused on the tropics not only for individual weather systems but also with regard to the general circulation of the whole atmosphere, because the tropics contain the main heat sources which drive the general circulation of the atmosphere. However, since most of the tropical regions are covered by oceans, there are still many fewer observational networks along the tropical belt than at higher latitudes. The GARP Atlantic Tropical Experiment (GATE) was carried out during the summer season from June to September in 1974 to provide detailed observational data in the tropical oceanic area. The main objective of GATE is "to provide the data for estimating the effects of the smaller tropical weather systems on the large-scale circulations." Specifically, an interaction between large-scale motion and cumulus convection is one of the key subjects in GATE.

Observational studies for estimating cloud properties through large-scale budget analyses have been performed under various large-scale situations by many authors in parallel with the development of theories of parameterization of cumulus clouds. Yanai et al. (1973) proposed a diagnostic method for determination of bulk properties of cumulus ensembles by combining the large-scale heat and moisture budgets with an entrainment cloud model which is similar to that used in Ooyama (1971) and Arakawa and Schubert (1974) and demonstrated how the large-scale environment interacts with cumulus ensembles. Ogura and Cho (1973) and Nitta (1975) developed a refined diagnostic method using a spectral representation for clouds introduced by Arakawa and Schubert (1974) and determined the cloud mass flux distributions. This method

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was applied to data over the western Pacific ITCZ region (Ogura and Cho, 1973; Cho and Ogura, 1974; Yanai et al., 1976), to data over the western Atlantic trades (Nitta, 1975) and to a prefrontal squall line in middle latitudes (Lewis, 1975).

One of the shortcomings of the diagnostic cloud models mentioned above is the neglect of cumulus downdrafts. Byers and Braham (1949) found a downdraft circulation with a comparative order of magnitude to an updraft in their direct measurements within thunderstorm cells. Browning et al. (1976) revealed in their analyses of the recent National Hail Research Experiment (NHRE) that downdraft circulations primarily originated in the middle troposphere and descended unixed to the surface. These downdrafts were also analyzed in synoptic and mesoscale systems in the tropics by Zipser (1969) and Betts (1973). Malkus (1955) observed downdrafts at the extreme downshear edge of the visible cloud in the trade wind region. The cumulus downdrafts generated due to the drag force by the precipitation and evaporative cooling were well simulated in two- and three-dimensional numerical cloud models by Takeda (1971), Miller and Pearce (1974), Wilhelmson (1974) and Yamasaki (1975).

The above observational and numerical results suggest that cumulus downdrafts as well as updrafts may make a great contribution to large-scale heat and moisture budgets. Gray (1973) and Ruprecht and Gray (1976a) evaluated the downdraft mass flux in their budget analyses of cloud clusters, but those estimates were subjective. Soong and Ogura (1976) combined the large-scale heat and moisture budgets with a time-dependent two-dimensional cloud model which expresses both updrafts and downdrafts and determined a cumulus cloud population over the trade wind region, but the contributions due to downdrafts were not explicitly obtained. Betts (1976) analyzed the thermodynamic transformation of the tropical subcloud layer by downdrafts originating from low levels just above cloud base. Recently Johnson (1976) has developed a diagnostic model in which the entrainment spectral cloud model is applied to both updrafts and downdrafts. His results showed that cumulus downdrafts contribute significantly to the large-scale mass, heat and moisture fields and the neglect of cumulus downdrafts leads to overestimation of the contribution due to shallow clouds in intense convective situations. However, several important parameters of cumulus downdrafts were predetermined in his model.

The main objective of this study is to examine the response of cumulus updrafts and downdrafts to different large-scale weather systems by using the well-planned GATE data set. To achieve this purpose, large-scale mass, heat and moisture budgets are computed and classified according to the different large-scale situations. Then a diagnostic cloud model is developed by introducing downdraft effects. Both updraft and downdraft properties are directly determined from the large-scale heat and moisture budgets by this diagnostic method. This method is tested on the mean Marshall Islands data and applied to the budget results in GATE.

2. Data

The upper air observation data measured in the GATE A/B-scale area by seven USSR ships are used in this analysis. Fig. 1 shows B- and A/B-scale ship distributions during Phase 3 (30 August–19 September 1974). Six USSR ship stations compose a hexagonal A/B-scale area. The ship Prof. Vize is located near the center of the hexagon. The data used in this study consists of two priority periods, i.e., the period from 0600 GMT 5 September to 0000 GMT 7 September and the period from 0300 GMT 12 September to 1800 GMT 13 September.

Temperature, geopotential height, wind velocity, wind direction and relative humidity were processed at both standard and significant pressure levels by the USSR All-Union Scientific Research Institute of Hydrometeorological Information. About 96% of total measurements extend to 150 mb and 85% reach the 100 mb level. The vertical distributions of observation density vary in time and space, but generally indicate a high density (7–8 levels per 100 mb thickness) below 900 mb, a minimum density (3–4 levels per 100 mb) between 800 and 600 mb, and a maximum density (~10 levels per 100 mb) in the uppermost layer between 200 and 100 mb. The basic parameters are interpolated at constant pressure levels with a vertical resolution of 25 mb in the upper layer between 900 and 100 mb and a 12.5 mb resolution below 900 mb by assuming that the basic parameters vary linearly with respect to the natural logarithm of pressure between significant levels.

There are some missing data for winds near the surface, probably due to mistracking of balloons. When the missing layer of winds is confined below 950 mb, the missing wind data are filled in by extrapolation from the upper observation. If the missing layer extends to 950 mb, no vertical extrapolation is performed. The upperair observations were taken at the regular 3 h observation times in general and missing data, including those which are not filled in by vertical interpolation or extrapolation, are interpolated linearly in time. Finally we obtain 3 h basic parameters (temperature, relative humidity and winds) at constant pressure levels with a 25 mb vertical resolution above 900 mb and a 12.5 mb below.

3. Method of analysis

a. Budget computations

Before going into large-scale budget computations, any basic parameter $X = \{u\}$ (eastward wind component), $v$ (northward wind component), $\theta$ (potential tempera-
Fig. 1. Upper air observation networks for A/B- and B-scales during GATE Phase 3. Squares denote stations used in this study. The hexagonal area enclosed by solid lines is the GATE A/B-scale area used for budget computations.

ture), \( q \) (mixing ratio of water vapor)—at each level for seven ships is fitted into a quadratic surface function

\[
X = ax^2 + bxy + cy^2 + dx + ey + f,
\]

where \( x \) and \( y \) are the longitudinal and latitudinal distances relative to the center of gravity of the analyzed area, and \( a-f \) are coefficients determined by the least-square method. Then the horizontal divergence \( \nabla \cdot \mathbf{V} \), relative vorticity \( \tilde{\omega} \), vertical \( \rho \)-velocity \( \tilde{\omega} \), large-scale apparent heat source \( Q_1 \) and large-scale apparent moisture sink \( Q_2 \) averaged over each triangle subarea (Fig. 1) are computed by

\[
\nabla \cdot \mathbf{V} = \frac{1}{A} \int \nabla \cdot \mathbf{V} dxdy,
\]

\[
\tilde{\omega} = \frac{1}{A} \int \mathbf{k} \cdot \nabla \times \mathbf{V} dxdy,
\]

\[
\tilde{\omega} = \tilde{\omega}_a = \int_{p_0}^{p} \nabla \cdot \mathbf{V} dp,
\]

\[
Q_1 = \frac{c_p}{(p_0/p)^s} \left( \frac{\partial \tilde{\omega}}{\partial t} + \mathbf{V} \cdot \nabla \tilde{\omega} + \tilde{\omega} \frac{\partial \tilde{\omega}}{\partial p} \right),
\]

\[
Q_2 = -L \left( \frac{\partial q}{\partial t} + \mathbf{V} \cdot \nabla q + \tilde{\omega} \frac{\partial q}{\partial p} \right),
\]

where overbars denote area averages, \( A \) the area of each triangular area, \( \int \int (\ ) dxdy \) the horizontal space integration, \( \tilde{\omega}_a \) the vertical \( \rho \)-velocity at the surface, \( p_s \) the surface pressure, \( p_0 = 1000 \text{ mb} \), \( c_p \) the specific heat of air under constant pressure, \( \kappa = R/c_p \) and \( L \) the latent heat of condensation. \( \tilde{\omega}_a \) is computed from

\[
\tilde{\omega}_a = \frac{\partial p_s}{\partial t} + \mathbf{V}_s \cdot \nabla p_s,
\]

where \( \mathbf{V}_s \) is the surface wind. In addition, we use an upper boundary condition which is computed from the first law of thermodynamics:

\[
\tilde{\omega}_R = \tilde{\omega}(p = 112.5 \text{ mb}) = \left[ \frac{\partial \tilde{\omega}}{\partial t} + \mathbf{V} \cdot \nabla \tilde{\omega} + \frac{(p_0/p_s)^s}{c_p} \frac{\partial \tilde{\omega}}{\partial p} \right]_{p=112.5}
\]

assuming that there is no diabatic heat source or sink except radiational heating \( Q_R \) in the uppermost layer between 125 and 100 mb. The original profiles of \( \tilde{\omega} \) from (4) are corrected to satisfy \( \tilde{\omega}_R \) by assuming a linear correction of \( \nabla \cdot \mathbf{V} \) increasing with height. The corrected \( \tilde{\omega} \) is then used for computation of the vertical advection terms in (5) and (6) but no correction is applied for
computation of the horizontal advection of heat and moisture. The imposed condition (8) automatically satisfies \( Q_1 = Q_R \) in the uppermost layer. The values of \( Q_R \) estimated at 10°N in September by Dopplick (1970) are used in this study.

All terms on the right of Eqs. (2)–(8) are computed analytically by using the quadratic surface function expressed by (1). After getting the budget results averaged in each triangular subarea, the area-averaged values over the whole hexagonal A/B-scale area are obtained by weighted-averaging those of each triangle by area.

The total heat budget equation is

\[
Q_1 - Q_2 - Q_R = \frac{\partial (s' + Lq')}{\partial p} \omega' = \frac{\partial h'}{\partial p},
\]

(9)

where

\[
s' = c_T T' + g_z, \quad h' = c_T T' + g_z + Lq',
\]

(10)

and \(-\overline{h'}\) is a measure of the vertical transport of total heat due to subgrid-scale eddies. If we integrate (9) from 100 mb to \( p \) by assuming \(-\overline{h'}\omega' = 0 \) at \( p = 100 \) mb, we obtain the vertical profile of eddy total heat flux at \( p \):

\[
F(p) = -\overline{h'\omega'} = \frac{1}{g} \int_{100}^{p} (Q_1 - Q_2 - Q_R) dp.
\]

(12)

The total heat supply from the sea surface can be evaluated by taking \( p = p_s \) in (12).

b. Diagnostic cloud model with downdraft

The net large-scale apparent heat source \( Q_1 - Q_R \) and moisture source \(-Q_2 \) in the cloud layer should be balanced by effects due to cumulus clouds which interact with large-scale motion systems. The previous diagnostic studies by Yanai et al. (1973), Ogura and Cho (1973) and Nitta (1975) represented these cumulus effects by using a one-dimensional entrainment-detrainment cloud model for determination of bulk properties of cumulus ensembles. However, they constructed the diagnostic cloud model by considering only cloud updrafts.

If we take into account both cloud updrafts and downdrafts, the equations of the large-scale heat and moisture budgets are

\[
Q_1 - Q_R = \delta_u (\delta_u - \delta) - L \delta_u l_u - M_u \frac{\partial \delta}{\partial p} + \delta_d (\delta_d - \delta)
\]

\[
- L \delta_d \delta_d - M_d \frac{\partial \delta}{\partial p},
\]

(13)

\[
- Q_2 = L \delta_u (\delta_u - \delta + l_u) - L M_u \frac{\partial \delta}{\partial p} + L \delta_d (\delta_d - \delta + l_d)
\]

\[
- L M_d \frac{\partial \delta}{\partial p},
\]

(14)

where \( M \) is the total vertical cloud mass flux multiplied by \( g \) (positive upward), \( \delta \) the rate of mass detrainment per unit pressure interval, \( \delta, q \) and \( P \) are the dry static energy, mixing ratios of water vapor and liquid water in the form of cloud droplets at the cloud top, and the subscripts \( u \) and \( d \) denote properties of updrafts and downdrafts, respectively. Here a single bulk model for the downdraft is assumed.

If we add (13) to (14), the total heat budget equation

\[
Q_1 - Q_2 - Q_R = \delta_u (\delta_u - \delta) - M_u \frac{\partial \delta}{\partial p} + \delta_d (\delta_d - \delta) - M_d \frac{\partial \delta}{\partial p}
\]

(15)

is obtained.

Ogura and Cho (1973) used (14) and (15) separately to determine the cloud mass flux and compared both results. They found that a scheme using (15) is most convenient, because this scheme does not depend on parameterization of a conversion process from cloud droplets to raindrops. Nitta (1975) used (15) for computations of the cloud mass flux. Johnson (1976) has recently determined both updraft and downdraft mass fluxes by solving (15), but his method was principally the same as Ogura and Cho (1973) and Nitta (1975), because he specified the relative magnitude of each downdraft mass flux to each updraft mass flux from the beginning and solved (15) for updraft cloud-base mass fluxes. These authors did not use both equations of (15) and (13) [or (14)] simultaneously for the cloud mass flux computations. However, if we introduce a parameterization of conversion from cloud droplets to raindrops as proposed by Arakawa and Schubert (1974), we can obtain \( l_u \) from a simple entrainment relation of the cloud liquid water and a lower boundary condition \( l_u (p_b) = 0 \) at the cloud base \( p_b \). Then (13) [or (14)] can be used for determination of another unknown parameter. In this study, both updraft and downdraft cloud mass fluxes are determined by solving (15) and (13) simultaneously.

If we assume that there is no mutual interaction of mass, heat, moisture and cloud liquid water between updrafts and downdrafts, the same spectral cloud representation as that adopted in Nitta (1975) is applicable to the updraft. The total mass flux and the mass detrainment rate for the updraft are written as

\[
M_u (p) = \int_0^{h_u (p)} m_B (\lambda) \eta (p, \lambda) d \lambda,
\]

(16)

\[
\delta_u (p) = m_B [\Delta_D (p)] \eta (p, \Delta_D (p)) \frac{\partial \Delta_D (p)}{\partial p},
\]

(17)

\( l_u \) is not necessarily equal to the mixing ratio of the liquid water which spreads into the environment. It is assumed that all of the detrained liquid water is evaporated at the same level where the detrainment occurs. More details were discussed in Arakawa and Schubert (1974).
where \( \lambda \) is the fractional rate of entrainment which classifies updraft cloud types, \( m_B(\lambda) \) the mass flux distribution function at the cloud base of the cloud type which has entrainment rate of \( \lambda \), and \( \lambda_B(p) \) the \( \lambda \) of clouds which detrains at pressure level \( p \). The budgets of mass and total heat for each updraft give the normalized mass flux and moist static energy

\[
\eta(p, \lambda) = \begin{cases} 
\frac{\lambda}{1 + \lambda}, & p_B < p < p_D \\
0, & p < p_D
\end{cases}
\]  
(18)

\[
h_u(p, \lambda) = \frac{1}{\eta(p, \lambda)} \int_{s(p_B)}^{s(p)} \eta(p', \lambda) h(p') dz(p')
\]  
(19)

where \( s(p) \) is the height at pressure level \( p \) and \( h_u \) the moist static energy of the updraft at the cloud base \( p_B \). From the assumption that the air is saturated in the updraft, we have

\[
s_u(p, \lambda) = \frac{1}{1 + \gamma(p)} \left[ h_u(p, \lambda) - h^*(p) \right]
\]  
(20)

\[
q_u(p, \lambda) = q^*(p) + \frac{\gamma(p)}{1 + \gamma(p)} \left[ h_u(p, \lambda) - h^*(p) \right]
\]  
(21)

where \( q^* \) is the saturation mixing ratio of water vapor in the environment and

\[
\gamma(p) = \frac{L}{T}
\]  
(22)

\[
\bar{h}^* = c_p T + \bar{q} + L q^*.
\]  
(23)

Arakawa and Schubert (1974) parameterized the conversion from cloud droplets to raindrops by

\[
\sum_{\lambda(\lambda_B + \Delta \lambda)} R_i = c_0 \eta(p, \lambda) m_B(\lambda) \frac{\partial M_d}{\partial \lambda} d\lambda
\]  
(24)

where \( R_i \) is the rate of conversion of the liquid water to precipitation per unit height due to clouds which have entrainment rate \( \lambda_i \) and \( c_0 \) is a constant parameter.

\( \lambda \) is chosen as \( 2 \times 10^{-3} \) m\(^{-1}\) so that the calculated values of liquid cloud water agree approximately with the observed values. If we adopt the same parameterization as (24), the mixing ratio of cloud liquid water is

\[
l_u(p, \lambda) = \frac{1}{\eta(p, \lambda)} \int_{s(p_B)}^{s(p)} \exp \left\{ -c_0 [z(p) - z(p')] \right\}
\]  
\[
\times \left\{ -\frac{\partial}{\partial p} \left[ \eta(p', \lambda) q_u(p', \lambda) \right] - \frac{\lambda}{\rho g} \right\} dp',
\]  
(25)

where \( \rho \) is the density of air.

It is assumed that the clouds lose their buoyancy at the cloud top and this vanishing buoyancy condition gives

\[
h_u(p, \lambda_B(p)) = \frac{(1 + \gamma)}{1 + 0.608(\beta p/L)}
\]  
\[
\times \left\{ 0.608(q^* - q) - L \left[ h_u(p, \lambda_B(p)) \right] \right\}
\]  
(26)

The same cloud-base condition for \( h_u \) as that in Nitta (1975), i.e.,

\[
h_u = h^*(p_B) - \frac{(1 + \gamma) c_p T}{1 + 0.608(\beta p/L)} \times 0.608(q^* - q) p_B \]  
(27)

is used, where it is assumed that all updrafts share the same cloud-base height and the same moist static energy at the cloud base. The height of the cloud base is estimated from the lifting condensation level which lies at \( \sim 950 \) mb in the GATE area. Using these cloud-top and cloud-base conditions with (18), (19) and (25), \( \lambda_B(p) \) is calculated and then all parameters for the updraft except \( m_B(\lambda) d\lambda \) are obtained under the given environmental profiles of \( \bar{s}, \bar{h} \) and \( \bar{h}^* \).

As for the downdraft, we do not specify any spectral cloud model like that for updrafts, because we do not have enough information with regard to properties of the downdraft. Rather we deal with bulk properties of the downdraft in our diagnostic cloud model. To close the system, we use the following assumptions for the downdraft:

1. The entrainment and detrainment of mass are given by

\[
\epsilon_d = 0 \quad \text{and} \quad \delta_d = \frac{\partial M_d}{\partial p}, \quad \text{when} \quad \frac{\partial M_d}{\partial p} > 0,
\]  
(28)

\[
\epsilon_d = -\frac{\partial M_d}{\partial p} \quad \text{and} \quad \delta_d = 0, \quad \text{when} \quad \frac{\partial M_d}{\partial p} \leq 0.
\]  
(29)

2. The air inside the downdraft is saturated. Relations similar to (20) and (21) are obtained, viz.,

\[
s_d(p) = \bar{s}(p) + \frac{1}{1 + \gamma(p)} \left[ h_d(p) - h^*(p) \right]
\]  
(30)

\[
q_d(p) = q^*(p) + \frac{\gamma(p)}{1 + \gamma(p)} \left[ h_d(p) - h^*(p) \right]
\]  
(31)

3. There are no cloud droplets inside downdraft, i.e.,

\[
l_d(p) = 0.
\]  
(32)

4. The thermodynamic parameters of the detrained air from the downdraft, \( s_d, q_d \) and \( h_d \) have the same values as the mean values inside the downdraft, i.e.,

\[
\bar{s}_d = s_d,
\]  
(33)

\[
\bar{q}_d = q_d,
\]  
(34)

\[
\bar{h}_d = h_d.
\]  
(35)
The first of these is an assumption of mass continuity for the downdraft. A similar bulk assumption for the mass balance was also used by Reed and Johnson (1974) in their vorticity analysis. From the budget of moist static energy in the downdraft, we obtain

\[ (h' - h_d) \frac{\partial M_d}{\partial p} - M_d \frac{\partial h_d}{\partial p} = 0, \tag{36} \]

where

\[ h' = \begin{cases} h_d, & \frac{\partial M_d}{\partial p} > 0 \quad \text{(detrainment)} \\ h, & \frac{\partial M_d}{\partial p} < 0 \quad \text{(entrainment).} \end{cases} \tag{37} \]

Since \( M_d(p_0) = 0 \) and \( \frac{\partial M_d}{\partial p} \big|_{p=p_0} < 0 \) at the downdraft originating level \( p = p_0 \), Eq. (36) gives

\[ h_d(p_0) = h(p_0). \tag{38} \]

This is a natural result from the assumption that the air entrained into the downdraft comes only from the environment. The saturation condition in the downdraft may contradict observations in the Line Islands Experiment analyzed by Zipser (1969) who found an unsaturated downdraft. The difference between results obtained from the saturation and unsaturation assumptions will be discussed in Section 6. There were few direct measurements of cloud liquid water in the downdraft, but Warner (1969) found that zero or low cloud droplet counts occur generally in downdraft regions, and aircraft measurements during penetrations of hailstorms by Browning et al. (1976) also revealed little water content of cloud droplets near the downdraft originating level. Numerical results of a precipitating convective cloud by Takeda (1971) indicated that downdraft regions have a relatively low concentration of small size drops. It seems that the third assumption is an acceptable first approximation. The fourth assumption is needed to close the system. Since the dominant terms due to the downdraft in (13) and (15) are \(-M_d(\partial \delta/\partial p)\) and \(-M_d(\partial h/\partial p)\), as shown in the following sections, the bulk assumptions adopted in this study would not significantly affect the general results.

Using these assumptions for downdrafts with the spectral cloud model for updrafts, \( m_d(\lambda) d\lambda \) and \( M_d(p) \) are determined in a coupled way by solving (13) and (15). In the actual computations of \( m_d(\lambda) d\lambda \) and \( M_d(p) \), Eqs. (13) and (15) are solved iteratively, and \( m_d(\lambda) d\lambda \) and \( M_d(p) \) are determined by neglecting the detrainment terms due to downdrafts at the beginning. Then \( h_d(p) \) is computed from (36). If \( \partial M_d(p)/\partial p > 0 \), the detrainment terms are evaluated by (28), (30), (32), (33) and (35), and new values of \( m_d(\lambda) d\lambda \) and \( M_d(p) \) are calculated by including the detrainment terms. In the application of the cloud diagnostic method to the discrete observation data, finite-difference forms should be carefully designed to conserve mass, heat and moisture budgets for both large-scale systems and cumulus clouds as shown by Arakawa and Mintz (1974) and Nitta (1975).

The method for determination of the cloud downdraft in this study is quite different from that proposed by Johnson (1976) even though both use the same spectral cloud model for updrafts. As stated briefly in a previous paragraph, he introduced the same spectral representation for downdrafts as updrafts assuming that each downdraft possesses the same entrainment rate as each updraft. The term \( m_d(\lambda) d\lambda \) was determined from (15) with the assumption \( \delta_d = 0 \) by specifying the relative magnitude of the downdraft mass flux to that of the updraft through a parameter \( \epsilon \) and the downdraft depth through a parameter \( \beta \); selection of optimum values of \( \epsilon \) and \( \beta \) was made based on large-scale budget considerations. However, a single bulk cloud model for the downdraft is introduced in this study and bulk properties of the downdraft are determined without specifying any parameters of the downdraft.

4. An evolution of the cloud cluster and its associated mass circulation

During both priority periods, cloud clusters develop and decay over the A/B-scale area. Three hourly IR satellite pictures shown in Fig. 2 reveal the great evolution of the cloud clusters during the whole period. On 5 September, two systems of cloud clusters interact with each other, become intense over the northern part of the A/B-scale area and then decay. Several smaller cloud systems develop on 6 September, but the area is relatively less disturbed as a whole than the previous day. During the period from 0600 to 1300 GMT 12 September two systems of cloud clusters are coupled tightly and develop along a line from southwest to northeast. This system gradually dissipates from 1730 GMT 12 September to 0400 GMT 13 September. Afterward, the area is again disturbed due to an approaching new cloud cluster.

Fig. 3 shows the time-height distribution of the vertical velocity averaged over the whole hexagonal A/B-scale area. In Fig. 3, no time smoothing has been applied. Although the computed profiles of the vertical velocity may include some uncertainties caused by observational errors and the simple method for the correction, they agree quite well with the visual appearance of the cloud clusters revealed by the satellite. During the first priority period on 5 September when the cloud cluster develops, the magnitude of the upward motion increases and the level of the maximum rising motion shifts from the lower level (~700 mb) to the upper level (~300 mb), indicating strong enhancement of deep clouds. During the rest of the first priority period, the rising motion still dominates the whole tro-
posphere but is confined mainly in the lower troposphere.

General features during the developing stage of the cloud cluster for the second priority period are similar to those for the first period, but strong downward motion replaces the upward motion in the middle troposphere during the decaying period. Afterward, rising motion again predominates due to the approach of the next cloud cluster.

The other budget results (not shown) also indicate large variations associated with the evolution of the cloud clusters. The budget results for the individual triangle subareas reveal the existence of regional differences, but only mean budgets in the whole A/B-scale area are discussed in this study; because the A/B-scale networks used in this study are not sufficient to examine regional differences accurately.

5. Mass, heat and moisture budgets for different A/B-scale fields

a. Classification of the A/B-scale fields

Though the analyzed area is highly disturbed by cloud clusters during the whole of both priority periods in general, stages of the development of the cloud clusters and their related budget results vary strongly in time as revealed in the previous section. Our main interest is to examine the effects of cumulus ensemble on the large-scale heat and moisture budgets under the different large-scale weather situations. Thus the combination of the two priority periods is subdivided into the following three different subperiods based upon both results of 3 h IR satellite cloud pictures and the A/B-scale area-averaged vertical velocity:

Period 1. The cloud cluster develops and reaches the
mature stage (0900–1800 GMT 5 September and 0600–
1500 GMT 12 September).

Period 2. The cloud cluster decays with the strong
downward motion (1800 GMT 12 September–0300
GMT 13 September).

Period 3. The A/B-scale area is not affected by the
organized cloud cluster (0000 GMT 6 September–0000
GMT 7 September).

The GATE area is also highly disturbed by the ap-
proaching cloud cluster during the period from 0600 to
1800 GMT 13 September, but this period is not in-
cluded in Period 1, because the center of the cloud
cluster system was located outside of the A/B-scale area
during this period. These three subperiods generally
correspond to the developing and mature stage of the
cloud cluster (Period 1), the dissipating stage of the
cloud cluster (Period 2) and the relatively undisturbed
period (Period 3). Budget computations were carried
out for each 3 h observation time, but only mean budget
results for each subperiod will be discussed in this study.

b. Horizontal divergence, relative vorticity and vertical
velocity

Fig. 4 shows the vertical profiles of the horizontal
divergence for the three subperiods. During Period 1 a
large convergence occurs in the lower layer below 700
mb but large divergence in the upper layer around 250
mb. There also exist weak divergence and convergence
in the middle layer. During Period 2 the amplitude of
the low-level convergence decreases and its vertical
depth is much limited to below 900 mb. On the contrary,
the amplitudes and the vertical thickness of the diver-
gence and the convergence in the middle troposphere
increase extremely. During Period 3 low-level conver-
gence takes place below 700 mb and divergence above,
but there is no strong divergence in the upper troposphere as that during Period 1.

Fig. 5 illustrates the vertical profiles of the relative vorticity. There exist deep cyclonic circulations with maxima near 650 mb and upper level anticyclonic circulations for all subperiods. The large relative vorticity near 650 mb is mainly due to the strong horizontal wind shear near the easterly jets. A secondary maximum of
the cyclonic circulation appears at about 900 mb for all periods. The upper level anticyclonic circulation becomes intense during the active period of the cloud cluster but weakened during the dissipating period. The value of the maximum relative vorticity in the middle layer increases during the dissipating period. The relative vorticity in the lower layer in the relatively undisturbed period is larger than those in the other super-periods. Though vorticity budgets will not be examined in this study, the results of the relative vorticity for different periods suggest that cumulus clouds may make large contributions to large-scale vorticity fields.

The profiles of the vertical \( p \)-velocity are shown in Fig. 6. The upward motion dominates the whole troposphere during both Periods 1 and 3 with a maximum value at 700 mb. There exists a secondary maximum near 300 mb in Period 1, probably due to the enhanced deep clouds. On the other hand, the downward motion predominates the middle layer during the dissipating period.

c. Apparent heat source, moisture sink and vertical eddy heat flux

The large-scale apparent heat source \( Q_1 \), apparent moisture sink \( Q_2 \) and the radiational heating \( Q_R \) for the three subperiods are illustrated in Figs. 7a–7c. The vertical profile of \( Q_R \) obtained by Doplick (1970) is used for all periods. During the active period of the cloud cluster (Fig. 7a), there exist a large apparent heat source and a moisture sink throughout the whole troposphere, but the former has its maximum at 400 mb and the latter at 700 mb. The large value of \( Q_2 \) near 1000 mb mainly results from large positive values of local time change and vertical advection terms in Eq. (6). This large apparent moisture sink suggests the existence of strong compensating subsidence induced by cumulus clouds in the subcloud layer, but more detailed studies of heat and moisture budgets in the subcloud layer is required to confirm this. During the dissipating period (Fig. 7b), a large-scale apparent heat sink (\( Q_1 < 0 \)) and a moisture source (\( Q_2 < 0 \)) appear in the middle layer where the strong subsidence occurs, while a heat source and a moisture sink occur below 700 mb. General profiles of \( Q_1 \) and \( Q_2 \) in Period 3 (Fig. 7c) are similar to those in Period 1, but are mostly confined in the middle and lower troposphere and their amplitudes are much reduced.

Fig. 8 shows the distributions of the vertical transports of the total heat energy due to subgrid-scale eddies. There are significant differences of heat transports by clouds for the different A/B-scale fields. Large amounts of heat energy are transported upward by the clouds during the active stage of the cloud cluster, while the weak downward heat flux occur above 700 mb during the dissipating stage. The heat transport by clouds in Period 3 is nearly half of that in Period 1 probably due to fewer deep clouds.

The supply of the total heat energy from the ocean surface evaluated by using the bulk-aerodynamic formula can be compared with that given by the budget computations. A variety of measurements have been taken near the ocean surface during GATE, but most of
these data are still being processed and reliable ranges of the transfer coefficients of heat and moisture have not yet been established. Thus the same values established in BOMEX are used in this study in setting $e_v$ (transfer coefficient of water vapor) = $1.2 \times 10^{-3}$ and B (Bowen ratio) = 0.1. Three hourly surface meteorological observations (standard WMO) for seven A/B-scale ships are used for the computation. Sea surface temperature is assumed to be constant during the whole analyzed period and monthly mean temperature for September for 30 years reported by Burpee and Dugdale (1976) is used. The heat supplied estimated from the bulk aerodynamic relation are 117 W m$^{-2}$ (Period 1), 95 W m$^{-2}$ (Period 2) and 109 W m$^{-2}$ (Period 3), respectively. The budget results for the three periods are more scattered than those by the aerodynamic method as shown in Fig. 8. As described in Eq. (12), the total heat supply from the ocean surface is calculated from the vertical integration of $Q_1 - Q_2 - Q_R$ in the whole troposphere and hence the errors involved in the computations of $Q_1$, $Q_2$ and $Q_R$, such as those due to using the climatological values of $Q_R$, the correction of $\bar{e}$, etc., would affect the estimate of the heat flux at the sea surface. It should be also noted that the total length of each subperiod is too short for random-type errors to cancel with each other. The mean value of the total heat flux during the whole period given by the budget computation is 86 W m$^{-2}$ which is $\sim 80\%$ of that estimated by the aerodynamic method.

Fig. 7. Mean large-scale apparent heat source $Q_1$, apparent moisture sink $Q_2$ and radiation heating $Q_R$ for (a) Period 1, (b) Period 2 and (c) Period 3.

Fig. 8. Mean vertical total heat flux due to subgrid-scale eddies for Period 1 (solid), Period 2 (dashed) and Period 3 (dash-dotted).
d. Regional difference between the GATE area and the western Pacific ITCZ

Most of the previous studies of the large-scale budgets (Reed and Recker, 1971; Nitta, 1972; Yanai et al., 1973; Williams and Gray, 1973) used the data taken in the western Pacific region, because of the denser upper air network there than in other tropical regions. Ruprecht and Gray (1976a,b) recently analyzed internal structures of cloud clusters observed both in the western Pacific (equator to 18°N) and in the west Indies (15–30°N). They found significant differences between the two regions which appeared to be due to location at different latitudes. However, both GATE and western Pacific regions are generally located over the active convective regions at similar latitudes but in different oceanic areas. Therefore, it is interesting to see how the large-scale mass, heat and moisture budgets differ between two typical ITCZ regions.

The budget results in GATE are similar to those in the western Pacific to some extent on the one hand, but there are several differences between these two regions on the other hand. The horizontal convergence and the cyclonic circulation take place in the lower troposphere, and the divergence and the anticyclonic circulation occur in the upper layer in both areas. Also the upward motion, large-scale apparent heat source and moisture sink are commonly observed in both areas. The low-level convergence layer is generally very shallow in GATE (below ~700 mb) as compared with the western-Pacific (below ~400 mb) and hence the maximum upward motion exists near 700 mb level in the former region but around 400 mb in the latter region. The layer structure of the divergence in the middle troposphere in GATE was not found in the western Pacific region. The relative vorticity possesses a maximum value in the middle troposphere around 650 mb in GATE but usually near the surface in the western Pacific. This difference of the relative vorticity would be mainly explained by the difference of the north-south horizontal shear of the zonal wind between the two different regions.

Because of these regional differences, we might expect that the influences of the cumulus ensemble of the large-scale heat and moisture fields in GATE would be also different to some extent from those in the western Pacific region. This will be more clearly shown in the following sections. It should be remarked that the comparison of large-scale budgets discussed above is not complete because of the difference in sample size and will be corrected by future analyses using data for all phases of GATE.

6. Cumulus updrafts and downdrafts in the western Pacific region

The budget results in the previous section clearly indicate significant effects due to cumulus ensembles on the large-scale mass, heat and moisture fields, but only show their net total effects. The detailed processes of the interactions which are taking place between the large-scale fields and the cumulus clouds cannot be cleared up until a diagnostic cloud model is applied.

Before investigating the detailed results in GATE, we shall examine some general effects due to downdrafts on the large-scale budgets by applying the diagnostic cloud model including cumulus downdrafts to the data in the western Pacific region and comparing with results obtained without downdrafts. The results in the western Pacific region will also be used for the comparison with those in the GATE area. Time-averaged large-scale budget results over the pentagonal area in the Marshall Islands region obtained by Yanai et al. (1973) are used for the cloud mass flux computations. This set of data was also used by Yanai et al. (1976) for the computation of the cloud mass flux using the spectral cloud model without downdrafts.

Fig. 9 shows the mass fluxes $M_B(p_p)$, at cloud base for the different cloud types, the total cloud mass flux $M_a$, and the mass flux $M$ in the environment obtained from the cloud model neglecting the downdraft (thick lines) which is essentially the same as that used by Nitta (1973) and Yanai et al. (1976). Nearly the same results are obtained as those of Yanai et al. (1976), although the conditions at the cloud top and base in this study differ slightly from those in Yanai et al. (1976) and they took time averages of the cloud mass flux after computations for each observation time. The bimodal structure of the cloud mass flux distributions with the dominance of very shallow and very deep clouds is noticeable. There exist very strong upward cloud mass flux and large compensating downward motion near the cloud base mainly due to the dominance of the shallow clouds.

Fig. 10 shows the results of mass fluxes given by including the downdraft. The downdraft air mass originating near 450 mb reaches to the cloud base where its magnitude is ~30% of that of the updraft mass flux. In comparison with the results without downdrafts, the magnitude of the mass flux for deep clouds is nearly the same as before, but the magnitude for shallow clouds is much reduced. As shown in Fig. 9 (thin lines), the inclusion of the downdraft decreases the net total cloud mass flux and thus decreases the subsidence in the environment especially near the cloud base. The amount of the net total cloud mass flux at the cloud base is reduced by ~40% if the downdraft is taken into account.

Fig. 11 shows the vertical distributions of moist static energy for different types of updraft $h_u$ and downdraft $h_d$, compared with that in the environment $h$. Since we do not specify any spectral cloud model for the downdraft, a single profile of the mean moist static energy in the downdraft is obtained from Eq. (36). The moist static energy for the updrafts decreases with height due to dilution through the entrainment through the environment. The downdraft originates in the middle troposphere around 450 mb from air in the environment and
Figs. 9 and 10. Cloud-base mass flux for updrafts as a function of detrainment level $M_B(p_0)$ (left) and net cloud mass flux $M$, large-scale mass flux $M = -\delta$, and mass flux in the environment $\bar{M}$ (right) obtained from a diagnostic cloud model without downdrafts in the Marshall Islands. Results for $M$ and $\bar{M}$ obtained by including downdrafts are also shown for comparison.

Figures 9 and 10 illustrate the deviations of the temperature...
and water vapor mixing ratio in the downdraft from those in the environment; these are evaluated from Eqs. (30) and (31) by assuming the saturation condition (solid lines). The temperature in the downdraft is cooler than the environment through the whole layer with a maximum deviation at the cloud base of $\sim 4$ K. The downdraft air has mixing ratio excesses above 850 mb, but deficits of $\sim -1.3$ g kg$^{-1}$ near the cloud base. There were few direct measurements in the downdraft as compared with the updraft, but the results in this study generally agree with those measured by the Thunderstorm Project in 1946 and 1947. Byers and Braham (1949) observed $-3$ to about $-4$ K temperature deviations from the mean clear-air temperature near the cloud base and $-1$ to about $-2$ K deviations in middle and upper layers. Also Braham (1952) reported $-1.4$ g kg$^{-1}$ deviation from mixing ratio of the water vapor of the mean air mass at base of cloud. Similar cool downdrafts were observed near the sea surface over the tropical oceanic area by Zipser (1969) and Houze (1976), but their observed downdrafts had only $\sim 90\%$ relative humidity. To see the effects of the relative humidity on the computations of $T_d$ and $q_d$, $T_d$ and $q_d$ were recalculated under unsaturated conditions by setting the relative humidity $U$ at $90\%$ and $80\%$. Both results are also described in Fig. 12. As the relative humidity is decreased, the downdraft air becomes warmer and drier. However, even if the downdraft air has $U=90\%$, the main features of the temperature and the moisture in the downdraft do not change.

The downdraft air would most likely share the same temperature and moisture as those in the environment at the originating level and would become cooler and more humid due to the evaporation from falling precipitation during the descent. The values of the deviations of the temperature and the humidity near 450 mb in Fig. 12 evaluated from the assumption of constant relative humidity in the entire downdraft layer would be overestimated. The vertical distributions of the thermodynamic properties in the actual downdraft would be determined through intricate processes of the
cloud microphysics and the interaction between updrafts, downdrafts and the environment. The construction of a detailed downdraft model would be required, including the above processes, in order to get accurate profiles of the thermodynamic properties in the downdraft.

The computations of the individual cumulus terms (not shown) in the heat and moisture budget equations (13) and (14) indicate that the cloud downdraft affects the large-scale field as a heat sink and a moisture source mainly through the compensating uplifting. The downdraft detrainment effects are very small in the case of the western Pacific. These cooling and moistening effects due to the downdraft are especially large near cloud base. Therefore, the neglect of the downdraft in the previous diagnostic cloud model leads to the overestimation of the shallow cloud mass flux because only detrainment effects due to shallow clouds are heat sinks and moisture sources in the previous model.

All mass flux computations were performed with \( c_0 = 2.0 \times 10^{-3} \text{ m}^{-1} \) which was suggested by Schubert (1973) to obtain general agreement with the limited observations. In order to check the sensitivity of the cloud mass flux to \( c_0 \), we made similar computations by using \( c_0 = 1.0 \times 10^{-3} \text{ m}^{-1} \) and \( c_0 = 4.0 \times 10^{-3} \text{ m}^{-1} \), respectively (not shown). As \( c_0 \) is increased, the amplitude of the downdraft mass flux increases in general, but the maximum difference of \( M_d \) between \( c_0 = 1.0 \times 10^{-3} \text{ m}^{-1} \) and \( c_0 = 4.0 \times 10^{-3} \text{ m}^{-1} \) is about 0.5 mb h\(^{-1}\). The difference of \( M_d \) is generally smaller than that of \( M_d \). We might conclude that the results of the cloud mass flux by using \( c_0 = 2.0 \times 10^{-3} \text{ m}^{-1} \) would not lose generality; \( c_0 = 2.0 \times 10^{-3} \text{ m}^{-1} \) is used for the following computations in GATE.

Although the method for determining the downdraft properties in this study is quite different from that proposed by Johnson (1976), both results agree quite well. In his computation using the Reed and Recker (1971) western Pacific composite easterly wave data, Johnson found the magnitude of the downdraft mass flux at the cloud base to be about 30% of the updraft which is nearly the same as that obtained in this study. Both results revealed that the magnitude of the updraft mass flux by the shallow clouds is much reduced by inclusion of the downdraft. This general agreement between two methods indicates that the spectral cloud model for the downdraft adopted by Johnson (1976) could simulate the bulk characteristics of the downdraft in general.

7. Cumulus updrafts and downdrafts in GATE
   a. Cloud mass flux

The diagnostic cloud model including downdrafts is applied to the mean budget results for three different subperiods in GATE presented in Section 5. Figs. 13a–13c show the results of the cloud mass flux computations for the three subperiods. During the active period of the cloud cluster (Fig. 13a), there exists strong upward mass flux due to the updraft with the dominance of very deep clouds. The deepest cloud type which detrains between 250 and 200 mb accounts for nearly half of the total cloud mass flux at the cloud base and the deep clouds penetrating through 500 mb share more than 80% of the total cloud-base mass flux. The shallow clouds below 700 mb, on the other hand, share only 12%. The cloud downdraft originates from air near 400 mb and extends to the lower troposphere, but its amplitude is weakened in the layer between 600 and 800 mb. The amplitude of the downdraft mass flux is approximately half of the updraft.

During the dissipating period of the cloud cluster (Fig. 13b), shallow clouds predominate in the updraft mass flux and deep clouds are much suppressed. The magnitude of the downdraft mass flux exceeds that of the updraft mass flux in the middle troposphere and the net cloud mass flux is consequently downward. The downdraft mass flux does not reach the cloud base, but is terminated near 800 mb indicating that spreading from the downdraft to the surrounding air occurs between 550 and 800 mb. The occurrence of downdraft spreading above the cloud base is also found in Period 1.

During the relatively undisturbed period (Fig. 13c), bimodal distributions of the updraft mass flux are obtained which are similar to those in the western Pacific region, but the relative importance of the intermediate clouds which detrain in the middle troposphere increases as compared with that in the western Pacific region. The downdraft originates from air at 300 mb and increases its amplitude almost uniformly in descending to cloud base. There is no downdraft spreading in the middle troposphere as during the other subperiods. The amplitude of the downdraft mass flux is \( \sim 40\% \) of the updraft mass flux at the cloud base.

Fig. 14 shows the vertical profiles of the net total cloud mass flux and the mass flux in the environment for the three subperiods. The difference of the net cloud mass flux between the different periods is very clear at a glance. A large amount of mass is carried upward due to the dominance of the updraft for Periods 1 and 3. Since the deep clouds are much more enhanced during the former period than the latter period, larger mass flux extends to the upper troposphere in Period 1. Meanwhile, the downdraft dominates the updraft in the middle troposphere during the dissipating period and the net cloud mass flux is downward.

While the net cloud mass flux varies largely in proportion as the large-scale fields change, the mass flux in the environment appears to be insensitive to the large-scale situations and the cloud activities. Generally, weak subsidence of a few millibars per hour occurs in the environment. This apparent insensitivity of the environment mass circulation does not mean that the environment acts just as a passive region but strongly indicates the mutual interaction between the large scale and the
clouds. If the large-scale rising motion is forced, the cumulus clouds are amplified so that the cloud-induced subsidence approximately compensates the large-scale rising motion. This process has been properly formulated as a quasi-equilibrium condition between the large-scale forcing and the cumulus ensembles by Arakawa and Schubert (1974) in their parameterization of cumulus convection. If the cumulus ensembles become active, the large-scale circulation is organized so as to counteract the cloud-induced subsidence. Recently Yamasaki (1975) demonstrated in his numerical experiment that a mesoscale circulation system was generated through the interaction between cumulus convection and large-scale motion. As a result, large-scale motion and cumulus clouds are in a quasi-equilibrium state so that the weak subsidence in the environment simply nearly balances the radiation cooling.

b. Thermodynamic properties within downdrafts

The vertical distributions of the moist static energy in the downdraft and that in the environment during Period 1 are shown in Fig. 15. The downdraft originates
from air near 400 mb and descends to 600 mb while decreasing its moist static energy through entrainment. The moist static energy is nearly constant below 600 mb because of little entrainment as expected from the profile of $M_d$ shown in Fig. 13a. $h_d$ is larger than $\bar{h}$ above 800 mb and smaller below resulting in a downward total heat flux above and an upward heat flux below. Thus a large vertical flux convergence of moist static energy
due to the downdraft is expected in the layer below 650 mb.

Fig. 16 shows the deviations of the temperature and the water vapor mixing ratio from those in the environment for the three subperiods computed from the saturation condition. The air in the downdraft is colder than the surrounding air by a few kelvins and has mixing ratio excesses of a few grams per kilogram except near the cloud base where the downdraft air has slight mixing ratio deficits. The minimum temperature deviation and the maximum mixing ratio deviation occur around 650 mb where the moist static energy of the environment is minimum.

**c. Heat and moisture balances**

Figs. 17a and 17b illustrate the profiles of the individual cumulus terms during Period 1 which balance with the large-scale apparent heat source $Q_I - Q_E$ and the moisture source $-Q_E$. For the heat balance, the warming due to updraft compensating subsidence and the cooling due to downdraft compensating uplifting are major terms, but the former is about twice as large as the latter. The detrainment effects both from the updraft and the downdraft are very small. Since shallow clouds which lose their buoyancy below 500 mb have very small cloud base mass fluxes during Period 1 as shown in Fig. 13a, the cooling due to evaporation of detrained cloud droplets is quite small in the lower troposphere. For the moisture balance, the drying effect due to updraft compensating motion is the principal term, but nearly half of the drying effect is compensated by the moistening effects due to the downdraft and the detrainment from the updraft.

The results for the dissipating period are shown in Fig. 18. The warming and cooling effects due to compensating motions by updrafts and downdrafts are dominant but opposed to each other. The warming effect due to the updraft exceeds the cooling effect due to the downdraft below 700 mb, but the latter overcomes the former above 700 mb and a net large-scale apparent heat sink occurs in the middle troposphere. The situation is rather
complicated for the moisture balance. Two terms due to the updraft compete with each other in the lower layer near the cloud base, but the drying effect is larger than the moistening effect. In the middle troposphere the moistening effects due to the downdraft and the detrainment from the updraft exceed the drying effect and a large-scale apparent moisture source appears.

As shown in Figs. 17 and 18, the downdraft plays an important role in the large-scale heat and moisture budgets as the heat sink and the moisture source. The ne-
glect of the downdraft would lead to an overestimate of heating and drying in the environment.

d. Comparison with the cloud properties in the Marshall Islands area

We have discussed the regional differences of the large-scale budgets between GATE and the western Pacific region in Section 3 which suggest differences in the cumulus activity between these two different tropical ITCZ regions. However, the analyzed period in this study is too short to be used for complete comparison with results obtained from about 100 days averaged Marshall Islands data. Rather we shall point out general similarities and differences with regard to the cloud properties between the two regions in this subsection.

As for the updraft mass flux distributions, bimodal distributions similar to those in the Marshall Islands are obtained during Period 3 in which the GATE area is not affected by an organized cloud cluster. When a cloud cluster develops, the deep clouds are much enhanced, while during the dissipating stage of a cloud cluster, the deep clouds are suppressed and the shallow clouds dominate. Probably the bimodal distributions of the updraft mass flux in the Marshall Islands result from the long-time averaged data including both disturbed and undisturbed situations. General results of the response of the cumulus updrafts to the large-scale forcing in GATE are similar to those studied by Yanai et al. (1976).

As for the downdraft, the relative magnitude of the downdraft in GATE is greater than that in the Marshall Islands. The ratio of the mass flux of the downdraft to that of the updraft is \( \sim 40\% \) at the cloud base in GATE Periods 1 and 3, but 30% in the Marshall Islands. While the downdraft air smoothly descends below the cloud base in Period 3 as does the downdraft in the Marshall Islands, the downdraft air spreads out to the environment in the middle troposphere during Periods 1 and 2. The relation between the downdraft and the vertical wind shear will be discussed in the next section. The deviations of temperature and humidity within the downdraft from those in the environment are weaker in GATE than in the Marshall Islands. This is mainly caused by the wetter surrounding air in the former area than in the latter area.

8. Relation between the downdraft and the vertical wind shear

Takeda (1971) demonstrated a formation of a “long-lasting” cloud under a certain ambient wind shear in his two-dimensional numerical simulation. In his computations, the behavior of the downdrafts is very dependent upon the vertical wind shear in the environment. We have obtained different profiles of the downdraft mass flux for different large-scale situations in GATE and the Marshall Islands which may relate to the vertical profile of the ambient wind.

Fig. 19 illustrates the vertical profiles of the area-averaged zonal wind for the three subperiods in GATE and for the Marshall Islands. In the Marshall Islands,
a deep layer of easterlies extends to 400 mb and westerlies exist above, resulting in westerly wind shear in the whole troposphere except near the surface. On the other hand, the general wind shear in GATE is opposite with weak westerlies in the lower layer and easterlies above. However, a strong easterly jet appears near 600–700 mb for the cloud-cluster subperiods. This easterly jet is likely associated with easterly waves which pass through the GATE area with about a 3.5-day period as reported by Burpee and Dugdale (1975). According to their analysis, the convection is most intense just before or at the time of the trough passage at 700 mb. During the analyzed period, the trough axes at 700 mb passed the center of the GATE A/B-scale area at about 0000 GMT 6 September and 1800 GMT 13 September. Therefore, the cloud clusters analyzed in this study are those which developed ahead of the trough of the easterly waves. Northeasterly winds prevail ahead of the trough axes accelerating the easterly current during the period of the cloud cluster. During Period 3 from 0000 GMT 6 September to 0000 GMT 7 September, the GATE area is situated just on the center or to the east of the trough axes where southwesterly winds are induced by the waves. Therefore the easterly jet retreats during Period 3.

By comparing the vertical profiles of the downdraft mass flux with those of the zonal wind, the easterly jet appearing during the cloud-cluster subperiods seems to act as an obstacle for the downward penetration of the downdraft. The downdraft mass flux begins to decrease with decreasing height just above 650 mb where the amplitude of the easterly jet is maximum. For the cases of GATE Period 3 and the Marshall Islands in which there is no such strong jet in the middle troposphere, the downdraft smoothly penetrates down to the cloud base, even though the directions of the wind shear are opposite for these cases. The results shown in Fig. 19 suggest the possible link between the downdraft and the vertical wind shear, but the detailed mechanism of how these phenomena relate to each other is not clear yet. Not only the wind shear but also other large-scale parameters would vary according to the passage of the easterly wave and those variations may also cause the differing responses of cumulus ensembles including the downdraft. More data analyses for different large-scale conditions and more theoretical studies for the effects of the vertical wind shear on the clouds are needed in the future.

9. Conclusions and remarks

Mass, heat and moisture budgets over the GATE A/B-scale area are examined for two priority periods during which the cloud clusters develop and dissipate. The A/B-scale budget results vary in good accordance with the evolution of the cloud cluster revealed by the 3 h IR satellite pictures. During the developing and mature stage of the cloud cluster, large convergence exists in the lower layer below 700 mb and large divergence in the upper layer near 250 mb. There also exist weak divergence and convergence in the middle troposphere. The maximum upward motion occurs at 700 mb due to the low-level convergence and a secondary maxi-
The cumulus appears around 300 mb probably due to the enhanced deep clouds. Large amplitudes of a large-scale apparent heat source and a apparent moisture sink are observed in the whole troposphere and a great amount of the total heat energy is transported upward by clouds.

During the dissipating stage of the cloud cluster, the downward motion appears in the middle troposphere with strong convergence at 400 mb and divergence at 700 mb. Clouds act as a heat sink and a moisture source in the middle troposphere where the total heat flux due to clouds is downward primarily due to the dominance of the downdraft. During the period in which the GATE A/B-scale area is not directly influenced by an organized cloud cluster, general budget results are similar to those for the active cloud cluster period, but the amplitudes relating to the cumulus activity are much diminished and are confined to the middle and lower troposphere.

A diagnostic cloud model has been developed by including the cumulus downdraft. The same spectral cloud model is used for the updraft as in previous studies but a single bulk cloud model is introduced for the downdraft. The mass fluxes for different types of updrafts and the total downdraft mass flux are determined by solving coupled heat and moisture budget equations. This method is tested by using three months mean Marshall Islands data and applied to three cases in GATE which are classified according to the A/B-scale situations.

The major effects due to the downdraft on the heat and moisture fields in the environment are cooling and moistening, and the inclusion of the downdraft reduces the mass flux of the shallow clouds. The downdraft originates from environmental air somewhere in the middle troposphere and penetrates down below the cloud base. The amplitude of the downdraft mass flux is ~30% of the updraft mass flux at the cloud base and the net total cloud mass flux at the cloud base is reduced to ~60% of that computed without downdraft for the Marshall Islands data. The downdraft air is colder by a few kelvins than the environment and has mixing ratio excesses except near the cloud base. This cool and moist downdraft would be maintained by evaporation from the falling precipitation.

The response of the cumulus ensembles in GATE is very dependent on the large-scale situation. When the GATE A/B-scale area is not affected by an organized cloud cluster system, a bimodal distribution of updraft mass flux with the dominance of very shallow and very deep clouds is observed and the downdraft originating from the middle troposphere extends down to the cloud base with increasing amplitude. However, if a cloud cluster develops in the area, the very deep clouds are greatly enhanced and a strong downdraft whose amplitude is ~40–50% of the updraft is observed. During the dissipating period, the deep clouds are highly suppressed and shallow clouds predominate. The downdraft mass flux surpasses the updraft mass flux in the middle troposphere and the net cloud mass flux is downward.

The vertical distribution of the downdraft mass flux seems related to that of the ambient wind. When the ambient wind varies monotonically with height, the downdraft initiated from the midtropospheric air goes down smoothly below the cloud base regardless of the sign of the wind shear. However, during the cloud cluster period in GATE in which an easterly jet appears in the middle troposphere, the downdraft mass flux is weakened through the jet. The downdraft mass reaches cloud base in spite of the interference by the easterly jet during the active stage of the cloud cluster, but is terminated above the cloud base during the dissipating stage. This would occur probably because the supply of raindrops from the updraft is large enough to maintain the downdraft for the former stage, but not sufficient for the latter stage.

Although it is presupposed that the downdraft is maintained by the drag force and the evaporative cooling due to the falling precipitation generated in the updraft, the interaction between the updraft and the downdraft is not explicitly taken into account in this method. Malkus (1955) pointed out in her study of trade cumulus that the downdraft is mixed primarily with the air of the adjacent updraft. Recent studies of an evolving hailstorm by Browning et al. (1976) also suggested that some of the downdraft was generated within former updraft air. For realistic modeling of the downdraft the relation between the updraft and the downdraft must be considered explicitly. This may produce more realistic distributions of the thermodynamic parameters within downdrafts than those given by the saturation assumption used in this study.

A possible relation between the downdraft and the ambient wind profile is suggested in this study, but more data analyses and theoretical studies are required to understand the detailed interaction between these two phenomena. Not only the downdraft but also the updraft may interact with the ambient wind through momentum transports. Also as shown in Section 5, different distributions of the relative vorticity are associated with the different stages of the cloud cluster. Cumulus clouds may play an important role in the large-scale momentum and vorticity budgets as well as heat and moisture budgets. Large-scale vorticity budget analyses have been performed in connection with cumulus clouds by Riehl and Pearce (1968), Williams and Gray (1973), Reed and Johnson (1974), Ruprecht and Gray (1976a) and Chu (1976), but most of these studies were based on the interaction between the large scale and the cumulus updraft. Reexamination of the momentum and vorticity budgets including cumulus downdrafts is necessary in future.

If the downdraft brings potentially cool heat energy down to the cloud base as demonstrated in this study, the large-scale field in the subcloud layer may be af-
fected significantly by the downdraft as studied by Betts (1976) and Johnson (1976). The modification of the subcloud layer due to cumulus ensembles including downdrafts is now being investigated at UCLA.

Because of limitations of the data in space and time used in this study, the A/B-scale fields are classified primarily by the activity of the cloud clusters. However, recent studies by Burpee and Dugdale (1975) and Reed et al. (1977) revealed that activities of cloud clusters in GATE were strongly related to the passage of easterly waves. The cloud cluster behaves as a part of the easterly wave interacting with individual cumulus clouds but has its own life period of about one day which is shorter than that of the waves. More studies are needed to clarify how wave disturbances, cloud clusters and cumulus clouds are coupled with each other.

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APPENDIX

Discrete Version for Computation of Cloud Mass Flux

For the computation of cloud mass fluxes, we discretized the heat budget equation (13) and the total heat budget equation (15) with a vertical interval of 50 mb. The large-scale parameters and the updraft parameters are defined at $p_i = p_0 - 50(i-1) [\text{mb}]$ or $p_i = 0.5 (p_{i-1} + p_{i+1}) [\text{mb}]$, $i=1,2,\ldots,n$, as the same way as Nitta (1975) and Yanai et al. (1976), where $p_0 = 950 \text{ mb}$ and $n$ is the total number of the updraft cloud types. $M_B(\lambda_i)$ is defined as the cloud mass flux at the cloud base for the type of the updraft which detains in the layer between $p_i$ and $p_{i+1}$ having the entrainment rate $\lambda_i$ (positive upward), and $M_d(p_{i+1})$ is defined as the total downdraft mass flux at $p_{i+1}$ (positive downward).

The discrete representations of Eqs. (13) and (15) lead to linear algebraic equations for $M_B(\lambda_i)$ and $M_d(p_{i+1})$, $i=1,2,\ldots,n$, of the form

$$A X = b,$$

where $A$ and $b$ are a coefficient matrix and a coefficient vector given by known large-scale parameters, respectively, and $X$ a vector composed of $M_B(\lambda_i)$ and $M_d(p_{i+1})$ to be solved. At the beginning of the iteration, $M_B(\lambda_i)$ and $M_d(p_{i+1})$ are calculated from (A1) by neglecting the detrainment from the downdraft. Then the coefficient matrix is recomputed by including the detrainment terms at $p_{i+1}$ only if

$$M_d(p_{i+1}) - M_d(p_{i+1}) < 0 \quad \text{for} \quad i > 1$$

$$M_d(p_{i+1}) - M_d(p_{i+1}) < 0 \quad \text{for} \quad i = 1$$

and new $M_B(\lambda_i)$ and $M_d(p_{i+1})$ are obtained. The same procedures are repeated until the solutions converge.

For the actual computation, the following two alternatives were used for solving (A1):

1) The solutions of (A1) in the upper troposphere above $\sim 300 \text{ mb}$ are unstable and cause troubles in iteration, because the contributions due to the moisture to the total heat budget equation (15) are very small in the upper troposphere and Eqs. (13) and (15) loose their independence easily due to observational errors. To eliminate this problem, only Eq. (15) is used for the computation above 300 mb by neglecting the downdraft. This alteration would be justified by the observations suggesting insignificant downdrafts in the upper troposphere.

2) From the physical point of view, $M_B(\lambda_i) > 0$ (upward) and $M_d(p_{i+1}) > 0$ (downward) are required, but there is no guarantee that solutions of (A1) satisfy these conditions because of the observational errors and the simplicity of the model. To satisfy the above conditions, we solve (A1) as to minimize

$$E = ||A X - b||,$$

subject to

$$x_i \geq 0 \quad \text{for} \quad i = 1,2,\ldots,n,$$

where $E$ is an Euclidean length of a vector $A X - b$ and $x_i$ are the elements of the solution vector $X$.

REFERENCES


