NOTES AND CORRESPONDENCE

The Dependence of Deep Cloud Mass Flux and Area Cover on Convective and Large-Scale Processes

G. S. BHAT

Centre for Atmospheric and Oceanic Sciences, Indian Institute of Science, Bangalore, India

19 November 1996 and 29 August 1997

ABSTRACT

A framework has been developed that brings together the important physical parameters and processes governing the vertical mass flux in deep convective clouds and their area cover. The main result is a simple relation for the cloud mass flux and area fraction in terms of the large-scale radiative cooling, environmental stratification, and the extent of lateral entrainment of the ambient air by the convective systems. It is shown that the contribution of the moist processes to the total vertical mass flux in deep clouds can become comparable to that of the large-scale radiative component, and thus the neglect of these subsynoptic-scale processes can severely underestimate the convective activity. Further, it is argued that the consideration of moist processes is not merely a question of the inclusion of a correction factor in the relationship, but the uncertainty that needs to be overcome before meaningful predictions of deep cloud area cover can be achieved.

1. Introduction

The vertical mass flux in convective clouds and the associated deep cloud cover are important atmospheric parameters because of their importance in vertical heat exchange, radiation, and possible role in climate change in the global warming scenario. Observations indicate that the fractional area cover (FAC) of deep convective clouds is small. Riehl (1979a) estimated the fractional area occupied by a hierarchy of cloud systems in a latitude belt of $\pm 10°$ about the equator; the active cumulonimbus (Cb) clouds occupy around 0.1%, mesoscale systems around 1%, and synoptic systems around 10%. Rossow (1994) considered both cloud optical thickness and cloud-top pressure in identifying deep convective clouds from satellite imageries and obtained a global mean deep cloud cover of 5.1% with a standard deviation of 0.9%. On a much smaller spatial scale, radar data can be used to obtain FAC of active clouds. For example, the average daily fractional area occupied by the precipitation echoes, inferred from the radar reflectivity data, is about 7% for the Darwin area (Williams and Renno 1993).

A simple physical explanation for the above observation has been provided by Riehl (1979b) in terms of the large disparity that exists in the atmosphere between the ascending and subsidence velocities or the corresponding timescales. The velocity of ascending air in cumulonimbus clouds is a few meters per second whereas the radiatively forced subsidence rate is about 30–40 mb day^{-1}, and the mass continuity demands that the area of ascent be small (Riehl 1979b). Theoretical studies also reveal that the horizontal scale of ascending air is indeed small compared to that of the subsiding air in a conditionally unstable atmosphere. One of the earliest studies is that by Bjerknes (1938), who considered the adiabatic ascent of saturated air through a dry-adiabatically descending environment. This work demonstrated that a convection arrangement in which appreciable ascending motion takes place in narrow cloud towers with slow downward motion in the wide cloud free spaces is favored in the atmosphere. Further, the solutions of the linearized equations governing the flow in the atmosphere indicate that unstable motions exist for horizontal scales of motion smaller than a limiting length of few hundred kilometers (Lilly 1960; Kuo 1961; Charney 1973). This length scale broadly corresponds to the horizontal size of the mesoscale systems observed in the atmosphere. Renno and Ingersoll (1996) derived an expression for the FAC of clouds assuming quasi-equilibrium conditions in which the mean entropy excess of the convective updrafts over the downdrafts is balanced by the radiative heat loss in the subsidence area. Substituting values representative of tropical conditions, Renno and Ingersoll (1996) obtained a fractional area cover of 0.05%, a reasonable value for the active Cb clouds. Recently, Craig (1996) employed dimensional

Corresponding author address: Dr. G. S. Bhat, Centre for Atmospheric and Oceanic Sciences, Indian Institute of Science, Bangalore 560012, India.
E-mail: bhat@caos.iisc.ernet.in

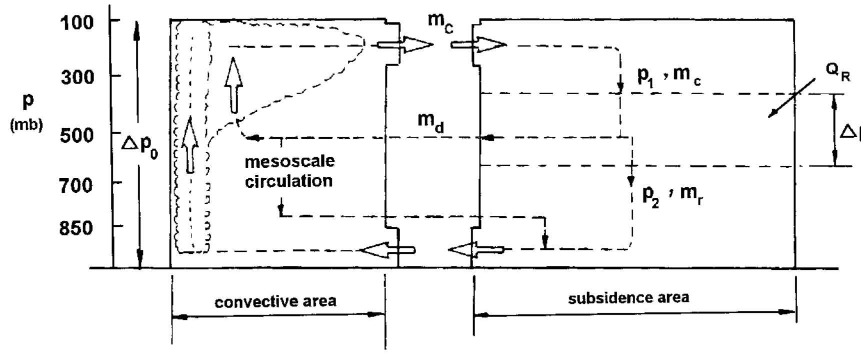

FIG. 1. A schematic of the one-dimensional model of the atmospheric circulation in the Tropics.

arguments under the assumption that an external forcing (sum of vertically integrated radiative cooling rate and adiabatic cooling rate due to large-scale convergence) and ambient stratification (expressed in terms of the enthalpy increase across the convecting layer) are the main factors governing deep convection and obtained results consistent with previous studies.

The main physical considerations in these studies are (i) the ascent is moist adiabatic, and (ii) the subsidence, either dry adiabatic or forced by radiative cooling, is in equilibrium with the large-scale environment. An important source, namely, the contribution of the moist convection itself to the mass flux in clouds, has not been taken into account. The moist processes bring down the mid- and upper-tropospheric air to low levels (moist downdrafts; e.g., Knupp and Cotton 1985; Houze 1989) and entrainment can substantially enhance upward mass flux in convective clouds (Arakawa and Schubert 1974). These processes operate on convective timescales (i.e., a few hours at most), and several diagnostic studies have shown that these subsynoptic-scale contributions are not negligible. For example, assuming a configuration in which the net radiative cooling of the atmosphere is balanced by the condensational heat release, Gray (1973) showed that the required upward mass flux in clouds is far greater than that supplied by the large-scale radiative subsidence. The balance, to be met by the subsynoptic-scale processes, is comparable to the large-scale component in the lower troposphere (Gray 1973). Several subsequent studies have suggested that the downward mass flux in convective systems can be as large as 25%–50% of the upward mass flux at cloud base (e.g., Nitta 1977; Leary and Houze 1980). However, this important aspect has not been taken into account by the simple theoretical models that addressed the problem of area cover of deep clouds so far.

The present study introduces a method to combine large-scale and convective-scale contributions. A simple relationship for the upward mass flux in clouds and their fractional area cover has been obtained in terms of large-scale radiative cooling, stratification in the environment, and convergence–divergence characteristics of deep convective systems. It is shown that the ambient stratification and radiative processes are basically responsible for the smallness of the convective areas, which is consistent with the previous studies. Moist processes, on the other hand, can enhance the area cover by a factor of 2 and thus be the main sources of fluctuations in convection.

2. Mass and energy balance in the subsidence area

The present study is based on the vertical mass circulation consideration. When air ascends in deep clouds, there has to be a compensating subsidence outside in order to satisfy the mass continuity. We can think of this ascending and subsiding motions as constituting a vertical circulation cell and expect the slower process, namely, the subsidence, to control the mass circulation rate (Riehl 1979b). As shown below, the mass and energy balance of the subsiding air leads to a simple formulation for the mass flux in deep clouds and cloud areal cover. The main advantage of this approach is that we get the required relationship without going into the complex details of moist convection. All that is needed is the mean horizontal mass divergence of convective systems, for which reasonable documentation exists in the literature now. To keep the matters simple, only tropical circulation is considered henceforth.

A schematic of the one-dimensional vertical circulation cell of the tropical atmosphere assumed here is depicted in Fig. 1. This is an extension of the single-cell model of the tropical circulation proposed by Sarachik (1978), and Betts and Ridgway (1988) to include the horizontal entrainment by the convective systems. The pressure scale at the left in Fig. 1 is suggestive of the normal vertical extent of deep convection in the Tropics and Δp_0 is the depth of the troposphere. The atmosphere is divided into two broad regions. All the deep convective areas are enclosed within the convective region, and the areas cloud free or with shallow clouds form the subsidence region. These two regions have distinctly different divergence characteristics. The convective region (includes the active Cb and the as-

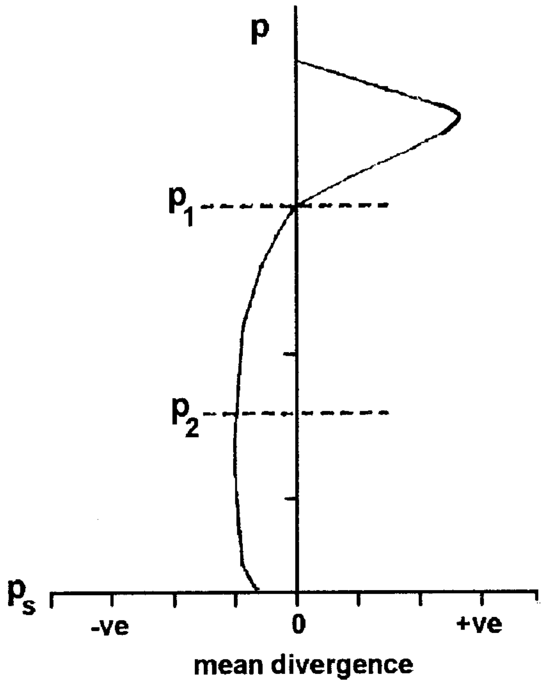

FIG. 2. A schematic of the mean horizontal divergence of tropical deep convective systems.

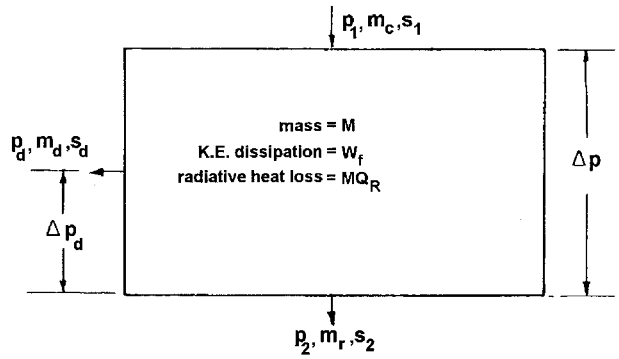

FIG. 3. The energy budget of the clear atmosphere. See the text for details.

sociated stratiform clouds) is characterized by a large-scale convergence from the surface to about 400–300-mb level and divergence above (e.g., Gray 1973; Frank 1977; McBride and Gray 1980a; Mapes and Houze 1995). A schematic of the mean divergence in the convective region is shown in Fig. 2; no restriction is imposed by the present method on the exact shape of the divergence except that there is one convergence and divergence layer each. The subsidence region is characterized by upper-level convergence, subsidence throughout the (upper) troposphere, and low-level divergence.

Let us enclose the mass M of the atmosphere in the subsidence region between pressure levels p_1 and p_2 by a closed surface and consider its mass and energy budgets (Figs. 1 and 3). It is convenient to choose p_1 as the level of nondivergence (Fig. 2). For convenience of reference, the mass M under consideration is referred to as the system and the enclosure as the control volume. The mass flux entering the control volume at the top surface is m_c having dry static energy s_1. (Dry static energy, denoted by s here, is given by $s = C_pT + gz$, where T is air temperature, C_p is specific heat at constant pressure, z is height above the sea level, and g is acceleration due to gravity.) Mass flux m_d leaves the control volume through the lateral sides due to entrainment by the deep convective systems, that is, detrained. The detrainment can take place at all levels and let s_d be the mass-weighted average dry static energy of the detraining mass flux (Fig. 3). The remaining mass flux, $m_r = m_c - m_d$, leaves the lower boundary with dry static energy s_2. It is assumed that p_2 is sufficiently above the surface so that there is negligible transport of energy flux into the control volume across the lower surface by shallow cumulus clouds. Then the energy balance for the system is (Fig. 3)

$$\partial(Ms)/\partial t = m_c s_1 - m_d s_d - m_r s_2 + MQ_R + W_f, \quad (1)$$

where s is the average dry static energy of the air in the control volume, Q_R is the average radiative heating rate (per unit mass), and W_f is the total dissipation of kinetic energy within the control volume. The term on the left-hand side (storage term) can be expressed as

$$\partial(Ms)/\partial t \sim M\partial s/\partial t \sim MC_p\partial T/\partial t, \quad (2)$$

where T is the average temperature of the mass in the control volume. The temperature of the tropical atmosphere is stable; even during the El Niño years, the increase in the temperature of the troposphere in the Tropics is about one degree Celsius (Philander 1990, 31). If this increase occurs gradually, say over a period of a month or a season, then the contribution of the storage term can be ignored compared to the radiative term, which is of the order of a degree Celsius per day (Dopplick 1979).

The kinetic energy of the air in the atmosphere is very small compared to the dry static energy (Neelin and Held 1987). Further, the dissipation of the kinetic energy takes place mainly in the atmospheric boundary layer and near cloud tops (e.g., Renno and Ingersoll 1996). The control volume shown in Fig. 3 is well above the surface layer and below the cloud top. Therefore, the contribution of the dissipation of kinetic energy can be neglected here. The neglect of W_f and the storage term leads to the radiative–convective balance, that is,

$$-m_c s_1 + m_d s_d + m_r s_2 = MQ_R. \quad (4)$$

Rearrangement of (4) gives

$$m_c[(s_2 - s_1) - \beta(s_2 - s_d)] = MC_p dT_R/dt, \quad (5)$$

where $\beta = m_d/m_c$, and Q_R is expressed in terms of the equivalent temperature change due to the mean radiative heating rate dT_R/dt. Let us assume that s varies linearly

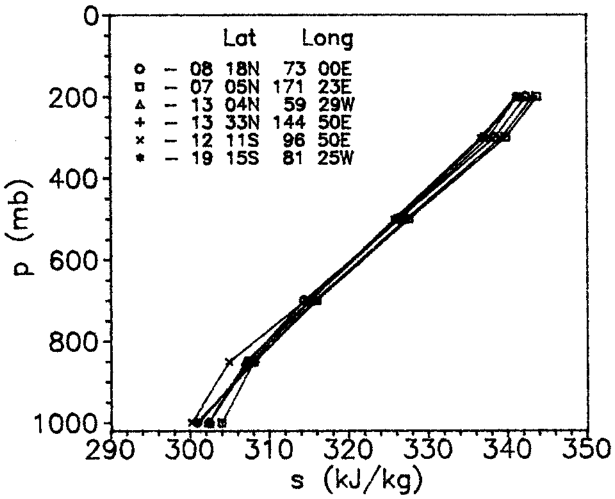

FIG. 4. Vertical distribution of the 10-yr average July, August, and September mean dry static energy at tropical island stations.

with pressure (observations show this to be a good approximation between 850 and 300 mb in the Tropics; see Fig. 4), that is,

$$s(p) = s_2 - (ds/dp)(p_2 - p), \qquad p_2 \geq p \geq p_1. \quad (6)$$

Hydrostatic equilibrium prevails in the subsidence region. Therefore, the mass M in the control volume and the total mass M_0 of the troposphere in the subsidence region are related by

$$M/\Delta p = M_0/\Delta p_0. \quad (7)$$

Combining (5), (6), and (7) we get

$$m_c = [M_0 C_p(dT_R/dt)]/[\Delta p_0(ds/dp)(1 - \alpha\beta)], \quad (8)$$

where $\alpha = (p_2 - p_d)/\Delta p$, p_d being the mean detrainment level defined by $s(p_d) = s_d$ (Fig. 3).

In the atmosphere, the mass contained in a column of unit area between the surface and the tropopause (i.e., in the layer of depth Δp_0) is constant within $\pm 3\%$ (hurricanes may be exceptions; however, such systems are rather rare). Therefore, the area fraction and mass fraction can be equated, that is,

$$M_c/M_0 = A_c/A_0, \quad (9)$$

where A_c is the area filled to the depth Δp_0 by the mass M_c that has converged into the convective region in time τ ($M_c = \tau m_c$) and A_0 is the subsidence area. Thus,

$$M_c/M_0 = A_c/A_0 = [C_p \tau(dT_R/dt)]/[\Delta p_0(ds/dp)(1 - \alpha\beta)]. \quad (10)$$

The product $\alpha\beta$ in Eq. (10) is a new parameter and its value for the tropical atmosphere needs to be calculated. A procedure to do this is given next under the assumption that the average horizontal mass divergence of convective systems is available either from observations or from model results. Then, for a given divergence profile, α and β can be calculated from the following expressions (see Fig. 3):

$$\alpha = (p_2 - p_d)/(p_2 - p_1),$$

$$p_d = \int_{p_2}^{p_1} (\nabla \cdot \mathbf{V})p\, dp \Bigg/ \int_{p_2}^{p_1} (\nabla \cdot \mathbf{V})\, dp, \quad (11)$$

$$\beta = m_d/m_c = \int_{p_2}^{p_1} (\nabla \cdot \mathbf{V})\, dp \Bigg/ \int_{p_s}^{p_1} (\nabla \cdot \mathbf{V})\, dp, \quad (12)$$

where $\mathbf{V}$ is the horizontal wind.

The linear dependence the of dry static energy on pressure has been used in obtaining the expression for p_d in Eq. (11). The physical interpretation of α, β and $\alpha\beta$ is as follows. The term $(1 - \alpha)$ is the fraction of the distance between p_1 and p_2 where the subsidence of the detraining mass flux m_d is driven by the radiative processes and α is the remaining fractional distance where the moist processes dominate. Here β represents the degree of lateral entrainment of the environmental air by the moist convective processes between p_1 and p_2 (sum of cumulus cloud entrainment and moist downdrafts). In the absence of moist downdrafts and lateral entrainment by Cb clouds, β is zero. In this limiting case, the entire mass diverging from the deep convective systems is brought down to p_2 by the radiative cooling, and $\alpha\beta = 0$. The other limiting case is $\beta = 1$ where all the subsiding air is entrained by the convective system between p_1 and p_2. For any reasonable divergence profile that can be thought of, α is always less than unity and so is the product $\alpha\beta$. In fact, αm_d ($=\alpha\beta m_c$), and not m_d, is the enhanced mass flux due to subsynoptic-scale moist processes and $(1 - \alpha\beta)m_c$ is the large-scale (i.e., radiative) contribution.

3. Results

a. Range of parameters in Eq. (10)

The deep clouds in the tropical convective systems extend up to about 100 mb. Hence $\Delta p_0 = 900$ mb is reasonable. The radiative cooling rate of the tropical atmosphere is in the range $-1°$ to $-2°$C day^{-1} (Dopplick 1979; Ackerman and Cox 1987). Here a value of $-1.5°$C day^{-1} is chosen as representative.

For obtaining the mean gradient of the dry static energy, the radiosonde data taken at island stations and given in the Monthly Climatic Data of the World (National Climatic Data Center) have been used. In order to examine its dependence on relative distance from the equator, soundings from stations located between 30°S and 30°N and spread over the Indian, Pacific, and Atlantic Oceans have been selected. Table 1 gives the geographic locations of these stations. Figure 4 shows the 10-yr average (1977–86) July–August–September mean dry static energy profiles for six stations from this set. For all these stations, s varies more or less linearly with pressure for 850 mb $> p >$ 300 mb. Data for the remaining stations and periods revealed that

TABLE 1. Geographic locations of the stations selected for the calculation of the atmospheric mean vertical structure.

No.	Lat	Long	No.	Lat	Long
1	28°13′N	177°22′W	13	11°40′N	92°43′E
2	21°59′N	159°21′W	14	10°37′N	61°21′W
3	19°43′N	155°04′W	15	8°18′N	73°00′E
4	19°15′N	81°25′W	16	7°28′N	155°04′W
5	18°26′N	66°00′W	17	7°20′N	134°29′E
6	18°03′N	63°07′W	18	7°05′N	171°23′E
7	17°56′N	76°47′W	19	1°21′N	172°55′E
8	16°16′N	61°31′W	20	8°31′S	179°13′E
9	13°33′N	144°50′E	21	12°11′S	96°50′E
10	13°04′N	59°29′W	22	14°20′S	170°43′W
11	12°35′N	81°42′W	23	17°33′S	149°37′W
12	12°12′N	68°58′W	24	27°37′S	144°20′W

the near-linear variation of s between 850 and 300 mb observed in Fig. 4 is a rugged feature of the tropical atmosphere. The average slope, *ds/sp,* calculated for the same 10-yr period, is plotted against the distance from the equator in Fig. 5. It is seen that *ds/dp* depends on the latitude, especially in the Northern Hemisphere, and varies in the range -52 to -62 J kg^{-1} mb^{-1}. At each sounding station, the standard deviation of the temporal fluctuations is found to be not more than 8.5% of the mean, and for more than 60% of the stations, it is less than 5%. Since the temperature and geopotential height anomalies are small in the tropical atmosphere (Riehl 1979b, 73–76; Neelin and Held 1987), these limited number of soundings can be taken to be proxies for the entire Tropics. In the calculations below, *ds/dp* is set at -57 J kg^{-1} mb^{-1}.

In the following, $\alpha\beta$ is estimated for few tropical systems whose divergence profiles are known. From the observed wind field, horizontal divergence has been calculated for a variety of tropical convective systems (Reed and Recker 1971; Frank 1977; Thompson et al. 1979; McBride and Gray 1980a; Mapes and Houze 1995; etc.). For the present purpose, an average divergence for the entire life cycle of a convective system is desirable, which, to the best of the author's knowledge, is not available. In its absence, divergence calculated considering a large area is taken as a proxy. Divergence curves from Fig. 11 of Gray (1973), Fig. 18 of Frank (1977), Fig. 5 of McBride and Gray (1980b), and Fig. 13 of Mapes and Houze (1995) have been digitized and used for estimating $\alpha\beta$. While computing $\alpha\beta$, it is desirable to choose p_2 such that it corresponds to the upper limit of shallow cumulus convection in the atmosphere. Shallow convection extends up to about 700 mb (e.g., Ludlam 1980; Riehl 1979b), and $p_2 = 700$ mb is a reasonable value to take. However, in the present calculations, p_2 is varied between 650 and 850 mb so that we know the sensitivity of the results to the variation in the height of shallow clouds. Figure 6 shows the variation of $\alpha\beta$ with p_2 for the divergence profiles mentioned earlier. Note that $\alpha\beta$ increases with p_2 for all the cases considered and varies in the range 0.15–0.55 depending on the convective system and p_2. For p_2 between 750 and 700 mb, the normal upper extent of shallow

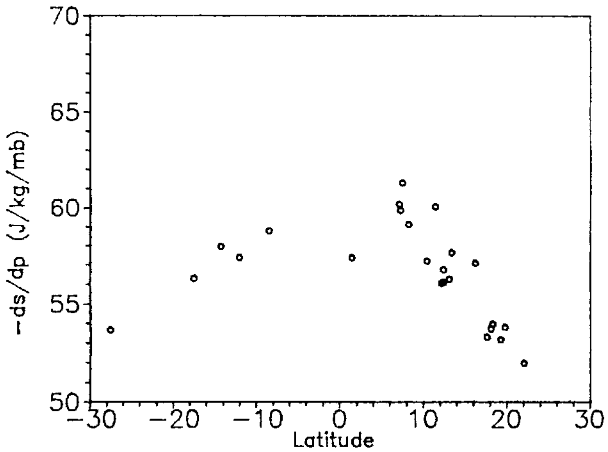

FIG. 5. Latitudinal variation of the average vertical gradient of the dry static energy.

convection in the atmosphere, $\alpha\beta$ lies between 0.2 and 0.45. Among the systems considered, the cyclone (Frank 1977) has the lowest $\alpha\beta$ at any given p_2.

b. Fractional area cover

The predictions from Eq. (10) are shown in Fig. 7 as a function of $\alpha\beta$ for $\tau = 1$ day. The fraction of the mass of the atmosphere available for ascent in deep clouds in a day varies from 3% to 6.6% for $\alpha\beta$ in the range 0–0.55. Since mass and area fraction are comparable in the present study, the predictions shown in Fig. 7 are in good agreement with the results derived from satellite observations discussed in the introduction (e.g., Rossow 1994).

Next, let us consider the active cumulonimbus clouds. The typical timescale of cumulonimbus clouds is about half an hour to one hour (Ludlam 1980, 113), and substituting an average value of $\tau = 3/4$ hour in Eq. (10)

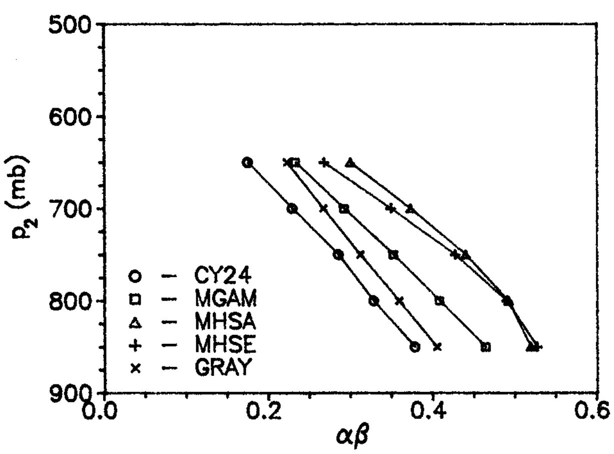

FIG. 6. Variation of $\alpha\beta$ with p_2 for some tropical convective systems. The divergence profiles are CY24, tropical cyclone 2°–4° radius (Frank 1977); MGAM, McBride and Gray (1980b) daytime; MHSA, Mapes and Houze (1995) active area; MHSE, Mapes and Houze (1995) extended area; and GRAY, Gray (1973).

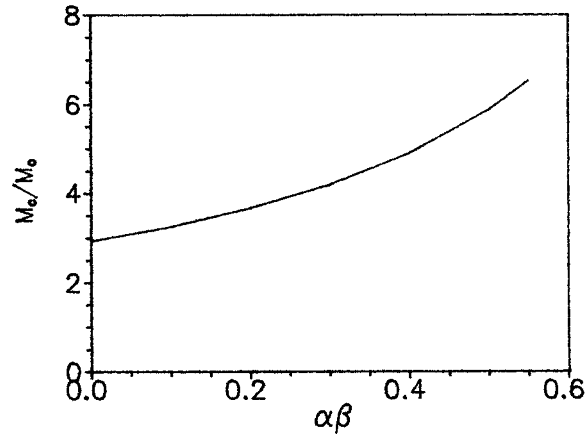

FIG. 7. Mass of air M_c available for ascent in deep clouds over a period of 1 day. Here M_0 is the mass of the tropospheric air outside the convective area, and M_c/M_0 is percent.

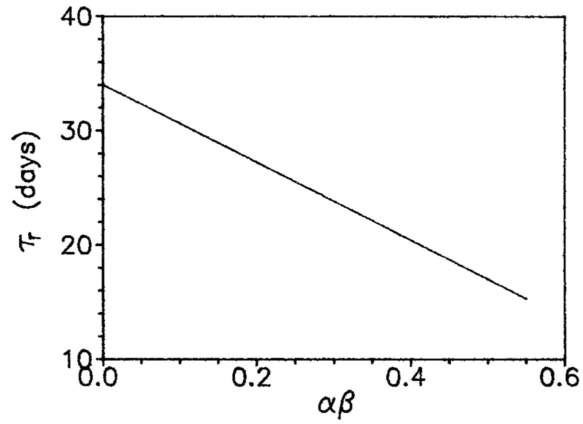

FIG. 8. The variation of the mass recycling period with $\alpha\beta$.

gives the fractional area covered by the active Cb clouds. The predicted mean value for a radiative cooling rate of $-1.5°$ C day^{-1} is 0.07%–0.15% for $\alpha\beta$ in the range 0–0.55. This is in good agreement with the value of 0.1% estimated by Riehl (1979a).

c. Recycling period

The time required for one complete recycling of the mass of the troposphere (denoted by τ_r here) can be taken as a measure of the degree of convective activity with larger values of τ_r being associated with lesser degrees of convection. In the present model, τ_r is obtained by setting $M_c/M_0 = 1$ in Eq. (10) and the result is shown in Fig. 8. In the absence of moist downdrafts, τ_r is about 34 days. In the event of downdrafts becoming very strong and the upper limit of shallow clouds not exceeding 850 mb, the convective recycling period is about 15 days. Thus, the moist processes can reduce τ_r by a factor of 1/2 and have the potential to contribute significantly to the variability in convection.

4. Discussion

Without the $\alpha\beta$ term, Eq. (10) is a statement of the radiative–convective equilibrium (Emanuel 1994). In the earth's atmosphere, the radiative heat loss per day ($\sim$1.5 kJ kg^{-1}) is very small compared to the dry static energy stratification across the depth of the troposphere ($\sim$40 kJ kg^{-1}). This is the basic reason for the daily fractional area cover predicted by Eq. (10) being small. Here $\alpha\beta$, representing the contribution of the moist processes, is a correction factor and, as already seen (Figs. 7 and 8), its role is not negligible. The present results clearly demonstrate the need for incorporating proper downdraft physics in the cumulus parameterization schemes in numerical weather models. For example, if the cloud parameterization scheme in these models does not have the necessary physics to simulate realistic horizontal convergence profiles in the mid- and lower troposphere, it may lead to substantial underprediction of the amount of convection and its variability.

We can also think of $\alpha\beta$ as one of the the major uncertainties at present that limits the accurate prediction of the deep cloud activity for the following reasons. It is observed from Fig. 6 that $\alpha\beta$ depends, rather strongly, on the mean divergence profile for a fixed p_2. There is observational evidence that the large-scale wind field and humidity stratification are important factors in determining the organization of moist convection and the time response of the convection to large-scale forcing (Johnson 1978; Barness and Sieckman 1984; Frank and McBride 1989). We can expect the large-scale environmental conditions to influence the mean divergence profile also. However, a clear (numerical or theoretical) model capable of relating the mean divergence profile to large-scale conditions is yet to emerge. Therefore, $\alpha\beta$ is not merely a correction factor, but represents what needs to be understood better before accurate prediction of deep cloud activity becomes possible. This has implications for climate models, especially those aiming at the prediction of the consequences of global warming following the greenhouse effect. Suppose the average preference for convective organization changes, say from cloud clusters (which are relatively slow moving) to squall line type or cyclones, due to changes in the large-scale wind and moisture fields following global warming. Then the resulting differences in the downdrafts (downdrafts are observed to depend on the nature of convective systems; e.g., Betts 1976; Johnson 1978; Nitta 1977; Barness and Garstang 1982) can have a major consequence on the convective activity on the global scale.

5. Conclusions

The framework developed in the present study brings together the important physical parameters and processes affecting the vertical mass flux in deep clouds and their area cover. The main result is a simple relation for the cloud mass flux (and cloud area fraction) in terms of the large-scale radiative cooling, environmental stratification, and the extent of lateral entrainment of the ambient air by the convective systems. The steady-state model predicts the fractional area covered by deep convection in the right range for two important timescales. Therefore, Eqs. (8) and (10) can be accepted as very useful for describing the mass available for deep convection.

Acknowledgments. This work has benefited from the many discussions that I had with Prof. J. Srinivasan, and I thank him for several useful suggestions. I also thank three anonymous referees whose critical comments and suggestions have lead to improvements in the material presented.

REFERENCES

Ackerman, S. A., and S. K. Cox, 1987: Radiative energy budget estimates for the 1979 southwest summer monsoon. *J. Atmos. Sci.,* **44,** 3052–3078.

Arakawa, A., and W. H. Schubert, 1974: Interaction of a cumulus cloud ensemble with the large scale environment. Part I. *J. Atmos. Sci.,* **31,** 674–701.

Barness, G. M., and M. Garstang, 1982: Subcloud layer energetics of precipitating convection. *Mon. Wea. Rev.,* **110,** 102–117.

——, and K. Sieckman, 1984: The environment of fast- and slow-moving tropical mesoscale convective cloud lines. *Mon. Wea. Rev.,* **112,** 1782–1794.

Betts, A. K., 1976: The thermodynamic transformation of the tropical subcloud layer by precipitation and downdrafts. *J. Atmos. Sci.,* **33,** 1008–1020.

——, and W. Ridgway, 1988: Coupling of the radiative, convective and surface fluxes over the equatorial Pacific. *J. Atmos. Sci.,* **45,** 522–536.

Bjerknes, J., 1938: Saturated-adiabatic ascent of air through dry-adiabatically descending environment. *Quart. J. Roy. Meteor. Soc.,* **64,** 325–330.

Charney, J. G., 1973: Planetary fluid dynamics. *Dynamic Meteorology,* P. Morel, Ed., D. Reidel, 337–343.

Craig, G. C., 1996: Dimensional analysis of a convective atmosphere in equilibrium with external forcing. *Quart. J. Roy. Meteor. Soc.,* **122,** 1963–1967.

Dopplick, T. G., 1979: Radiative heating of the global atmosphere: Corrigendum. *J. Atmos. Sci.,* **36,** 1812–1817.

Emanuel, K. A., 1994: *Atmospheric Convection.* Oxford University Press, 580 pp.

Frank, W. M., 1977: The structure and energetics of the tropical cyclone I. Storm structure. *Mon. Wea. Rev.,* **105,** 1119–1135.

——, and J. L. McBride, 1989: The vertical distribution of heating in AMEX and GATE cloud clusters. *J. Atmos. Sci.,* **46,** 3464–3478.

Gray, W. M., 1973: Cumulus convection and larger scale circulations I. Broad scale and mesoscale considerations. *Mon. Wea. Rev.,* **101,** 839–855.

Houze, R. A., Jr., 1989: Observed structure of mesoscale convective systems and implications for large-scale heating. *Quart. J. Roy. Meteor. Soc.,* **115,** 425–461.

Johnson, R. H., 1978: Cumulus transports in a tropical wave composite for phase III of GATE. *J. Atmos. Sci.,* **35,** 484–494.

Knupp, K. R., and W. R. Cotton, 1985: Convective cloud downdraft structure: An interpretive survey. *Rev. Geophys.,* **23,** 183–215.

Kuo, H. L., 1961: Convection in conditionally unstable atmosphere. *Tellus,* **13,** 441–459.

Leary, C. A., and R. A. Houze Jr., 1980: The contribution of mesoscale motions to the mass and heat fluxes of an intense tropical convective system. *J. Atmos. Sci.,* **37,** 784–796.

Lilly, D. K., 1960: On the theory of disturbances in a conditionally unstable atmosphere. *Mon. Wea. Rev.,* **88,** 1–17.

Ludlam, F. H., 1980: *Clouds and Storms.* Pennsylvania University Press, 405 pp.

Mapes, B. E., and R. A. Houze Jr., 1995: Diabatic divergence profiles in western Pacific mesoscale convective systems. *J. Atmos. Sci.,* **52,** 1807–1828.

McBride, J. L., and W. M. Gray, 1980a: Mass divergence in tropical weather systems paper I: Diurnal variation. *Quart. J. Roy. Meteor. Soc.,* **106,** 501–516.

——, and ——, 1980b: Mass divergence in tropical weather systems paper II: Large-scale controls on convection. *Quart. J. Roy. Meteor. Soc.,* **106,** 517–538.

Neelin, D. J., and I. M. Held, 1987: Modeling tropical convergence based on the moist static energy budget. *Mon. Wea. Rev.,* **115,** 3–12.

Nitta, T., 1977: Response of cumulus updraft and downdraft to GATE A/B-scale motion systems. *J. Atmos. Sci.,* **34,** 1163–1186.

Philander, S. G., 1990: *El Nino, La Nina, and the Southern Oscillation.* Academic Press, 289 pp.

Reed, R. J., and E. E. Recker, 1971: Structure and properties of synoptic-scale wave disturbances in the equatorial western Pacific. *J. Atmos. Sci.,* **28,** 1117–1133.

Renno, N. O., and A. Ingersoll, 1996: Natural convection as a heat engine. *J. Atmos. Sci.,* **53,** 572–585.

Riehl, H., 1979a: Recent tropical experiments in the general context of tropical meteorology. *Meteorology over the Tropical Oceans,* D. B. Shaw, Ed., Royal Meteorological Society, 9–30.

——, 1979b: *Climate and Weather in Tropics.* Academic Press, 611 pp.

Rossow, W. B., 1994: Cloud variability depends on how you look at it. Preprints, *Eighth Conf. on Atmospheric Radiation and Sixth Conf. on Climate Variations,* Nashville, TN, Amer. Meteor. Soc., J2–J4.

Sarachik, E. S., 1978: Tropical sea surface temperature: An interactive one-dimensional atmosphere ocean model. *Dyn. Atmos. Oceans,* **2,** 455–469.

Thompson, R. M., S. W. Payne, E. E. Recker, and R. J. Reed, 1979: Structure and properties of synoptic-scale wave disturbances in the intertropical convergence zone of the eastern Atlantic. *J. Atmos. Sci.,* **36,** 53–72.

Williams, E., and N. O. Renno, 1993: An analysis of the conditional instability of the tropical atmosphere. *Mon. Wea. Rev.,* **121,** 21–36.