CUMULUS CONVECTION AND LARGER-SCALE CIRCULATIONS

Part III

Broadscale and Meso Scale Considerations

by

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ABSTRACT

This manuscript (Paper III) discusses the magnitude and implication of the vertical circulation patterns of the summertime tropical atmosphere as derived from large scale consideration. It is to be compared with the vertical circulation patterns derived from cumulus scale considerations as discussed by López (1972a, 1972b) in Papers I and II. From the large scale it is shown that very significant extra up- and-down "local" vertical motion is occurring beyond what would be prescribed by the "mean" or synoptic scale flow patterns. This typical unresolved and mass compensating extra up- and-down vertical circulation pattern can be specified from satellite observed and calculated cloud cluster scale ($\sim 4^{\circ}$) mass, vapor, energy and rainfall-evaporation budgets. The method of determining this extra "local" vertical circulation and its magnitude are presented and discussed. Results are very closely comparable with those obtained by López (op. cit.) from an independent small scale approach through modeling of individual cumulus elements. These two independent approaches from different scales of consideration give very similar results. This "local" vertical circulation is shown to be fundamental for the mass, vapor and energy balances of the tropical atmosphere. Other discussions of the characteristics of the cumulus convective atmosphere are made.

I. INTRODUCTION

The manner by which cumulus clouds and the broader-scale flow patterns interact with each other is not well understood at this time. It is very important that we come to grips with the physics of this interaction problem. Many meteorologists feel this to be a fundamental requirement to improved understanding and prediction of large-scale atmospheric flow patterns.

The author feels that a significant expansion of our knowledge on this problem is possible from the meteorological information already on hand if we organize our various facets of information in a judicious way. This is the purpose of the following discussion.

This research discusses the required mass, water vapor, and energy budgets of the summer, oceanic Trade Wind-Equatorial Trough belt from about 5° to 25° latitude. These budgets are obtained from resolving into a mutually consistent pattern the available broad-scale (meso, synoptic, and zonal) observational knowledge on the mass, water vapor, and energy information in this region.

It will be shown that the accomplishment of these balances requires a very substantial "local" up- and-down vertical circulation with condensation and reevaporation rates much larger than the observed rainfall-evaporation. This paper specifies the magnitude of this vertical circulation and water vapor recycling, and then discusses the resulting energy requirements. It also compares these mass-vapor-energy budgets derived from broad-scale considerations for the cloud cluster with the same budgets obtained by López (Papers I and II) from incorporation of individual convection elements from his cumulus life cycle model. It will be shown that both approaches, one from the large scale going downward in scale (broad-scale approach), and the other using the individual convective elements and going upward in scale (cumulus scale approach) do indeed mesh with near identical mass-vapor-energy budget results. The meshing of these independent approaches from different scales of consideration lends confidence to the results to be shown.

II. DATA SOURCES

To deal with the tropical belt in a realistic way, it was indispensable that representative information be obtained on the typical tropical belt lapse rate conditions, vapor contents, divergences, shears, etc., associated with the satellite observed tropical cloud clusters, the other variable cloud areas, and with the clear regions. To accomplish this, the author performed extensive radiosonde data composite analysis of satellite observed cloud and clear areas in the Western Pacific and the West Indies. These regions have the only oceanic radiosonde networks from which associated wind-temperature-moisture information could be obtained.

Figs. 1 and 2 show the locations of the radiosonde networks which were used to composite the digatized satellite meso-scale cloud clusters, the other variable cloud regions, and the clear regions. This was accomplished for the three summer seasons of 1967-1969. Figs. 3-6 show the locations of the composited cloud clusters and clear areas within these networks. Clusters and clear areas were divided into two latitudinal groupings; those poleward and equatorwards of 18°. The methods of compositing, and a discussion of the data limitations, inaccuracies, etc. have been made in an earlier report by Williams (1970). The reader is referred to this report for more information. Fig. 7 shows typical views of these satellite observed tropical clusters and clear regions.

There were 557 clusters and 223 clear areas in the Western Pacific and 539 clusters and 212 clear areas in the West Indies networks which are included in the data composites.

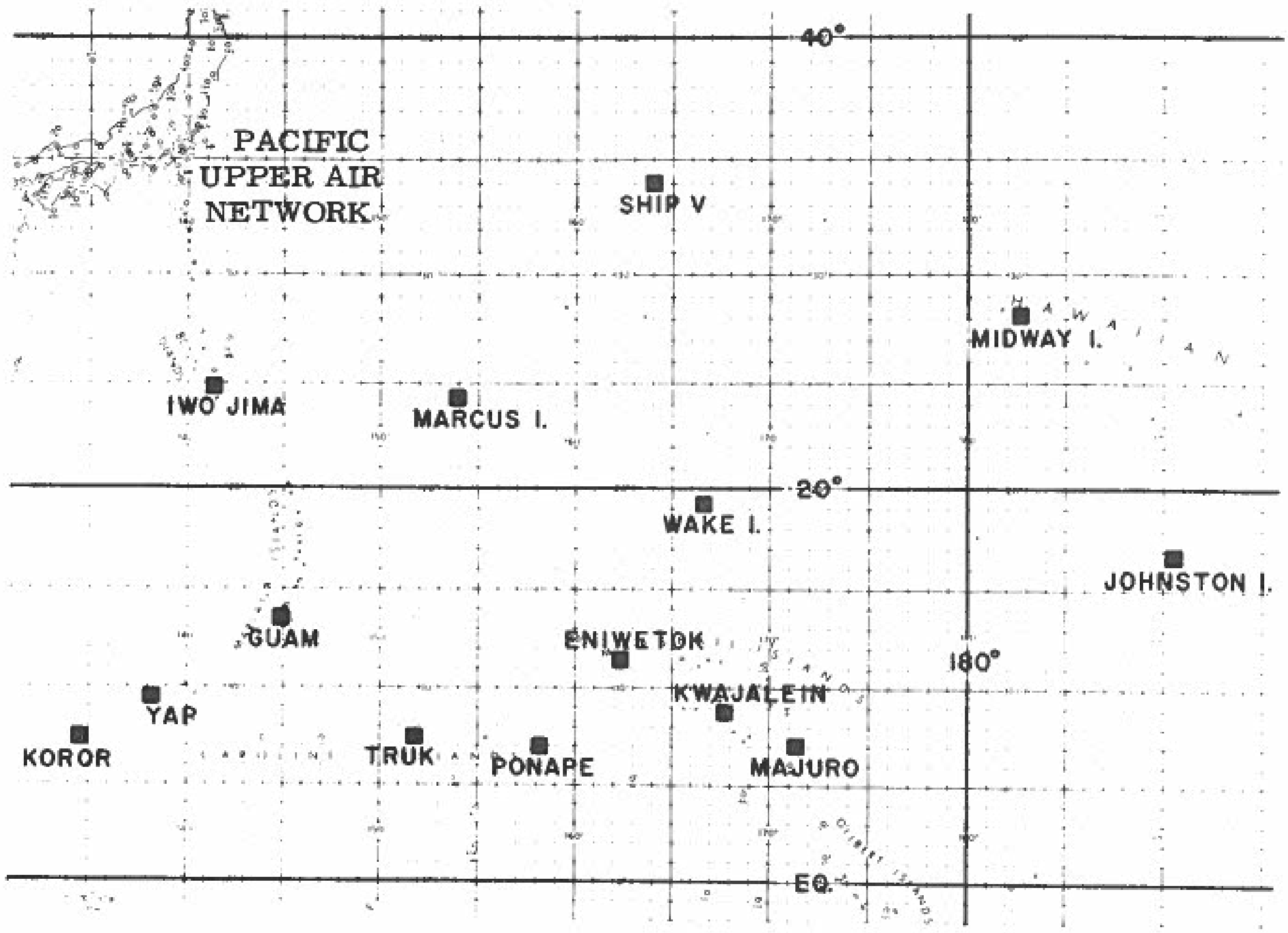

PACIFIC
UPPER AIR
NETWORK
40°
20°
EQ.
180°
SHIP V
MIDWAY I.
IWO JIMA
MARCUS I.
WAKE I.
JOHNSTON I.
GUAM
ENIWETOK
YAP
KWAJALEIN
KOROR
TRUK
PONAPE
MAJURO

Fig. 1.

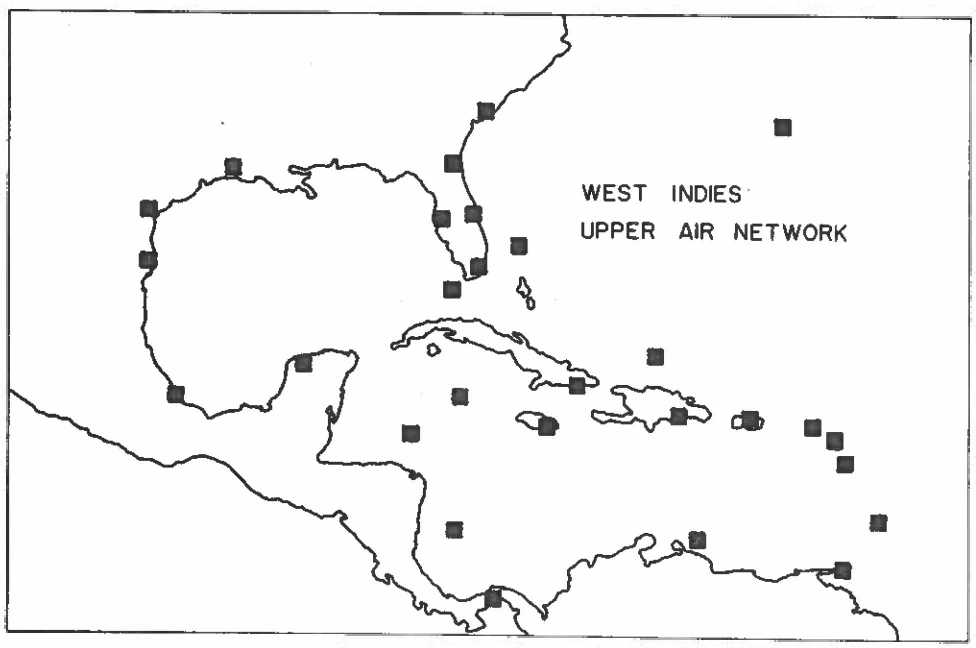

WEST INDIES
UPPER AIR NETWORK

Fig. 2.

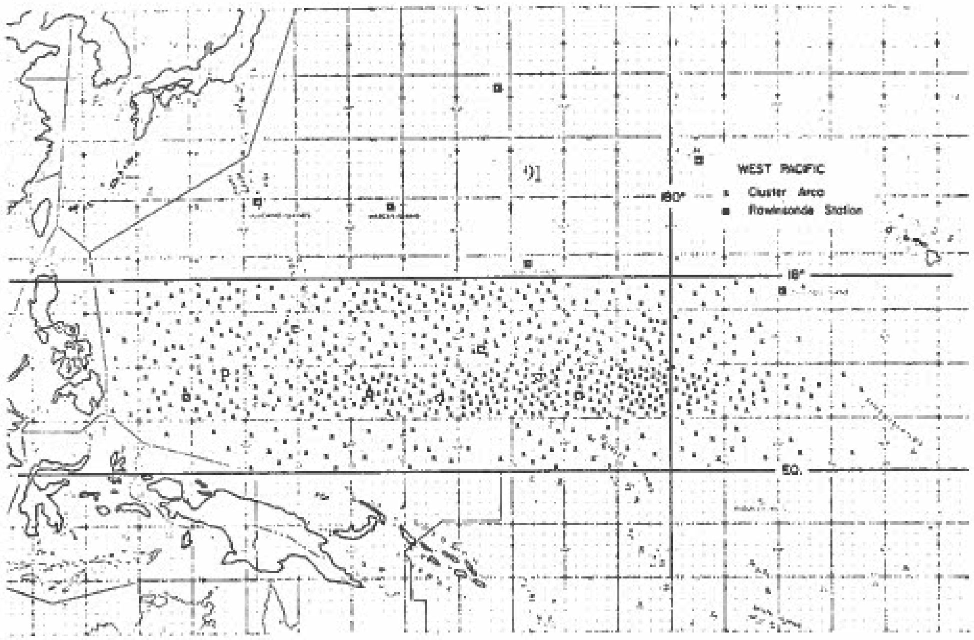


Fig. 3. Location of cluster regions in the West Pacific.

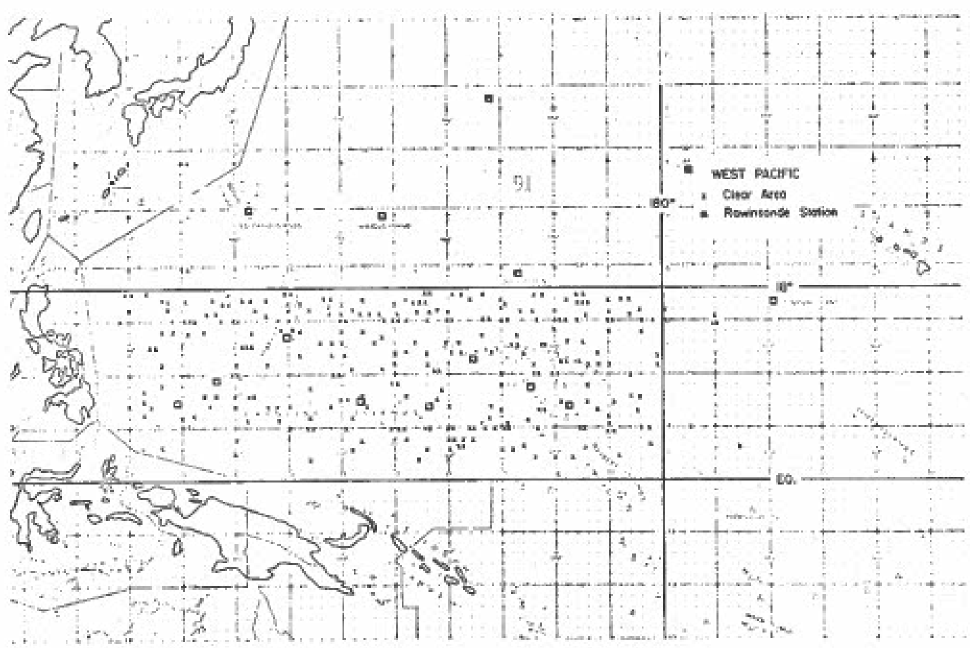


Fig. 4. Location of clear regions in the West Pacific.

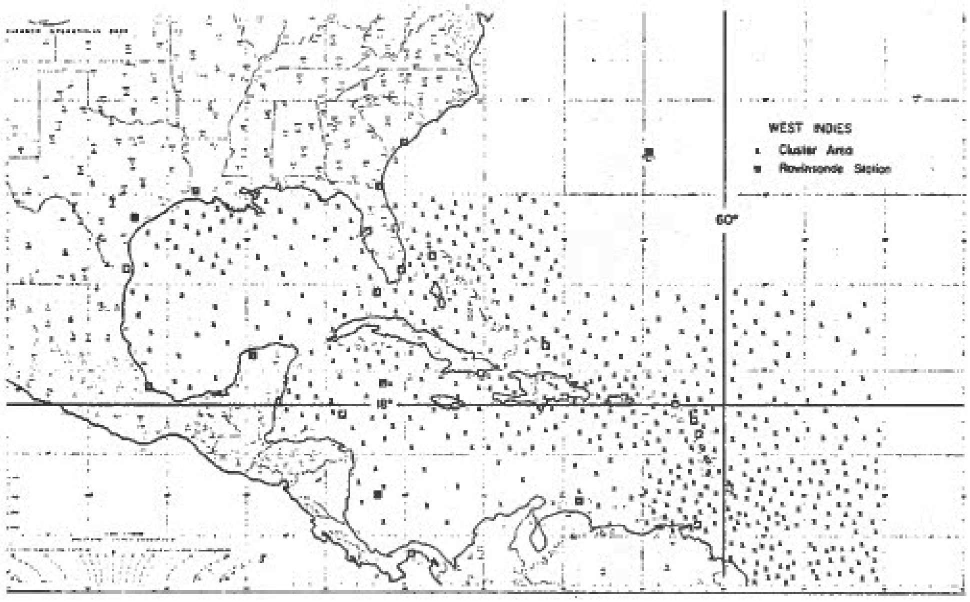


Fig. 5. Location of cluster regions in the West Indies.

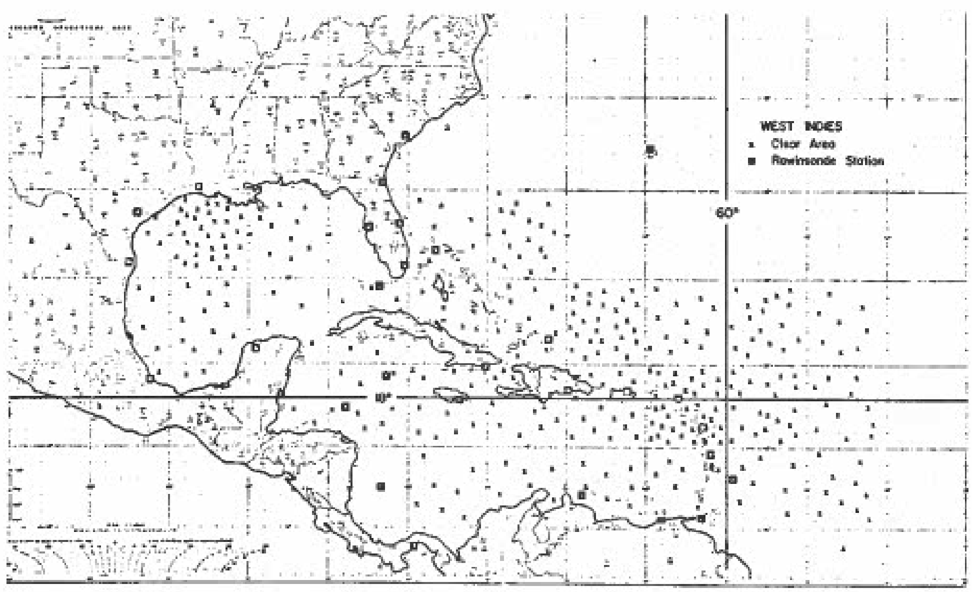
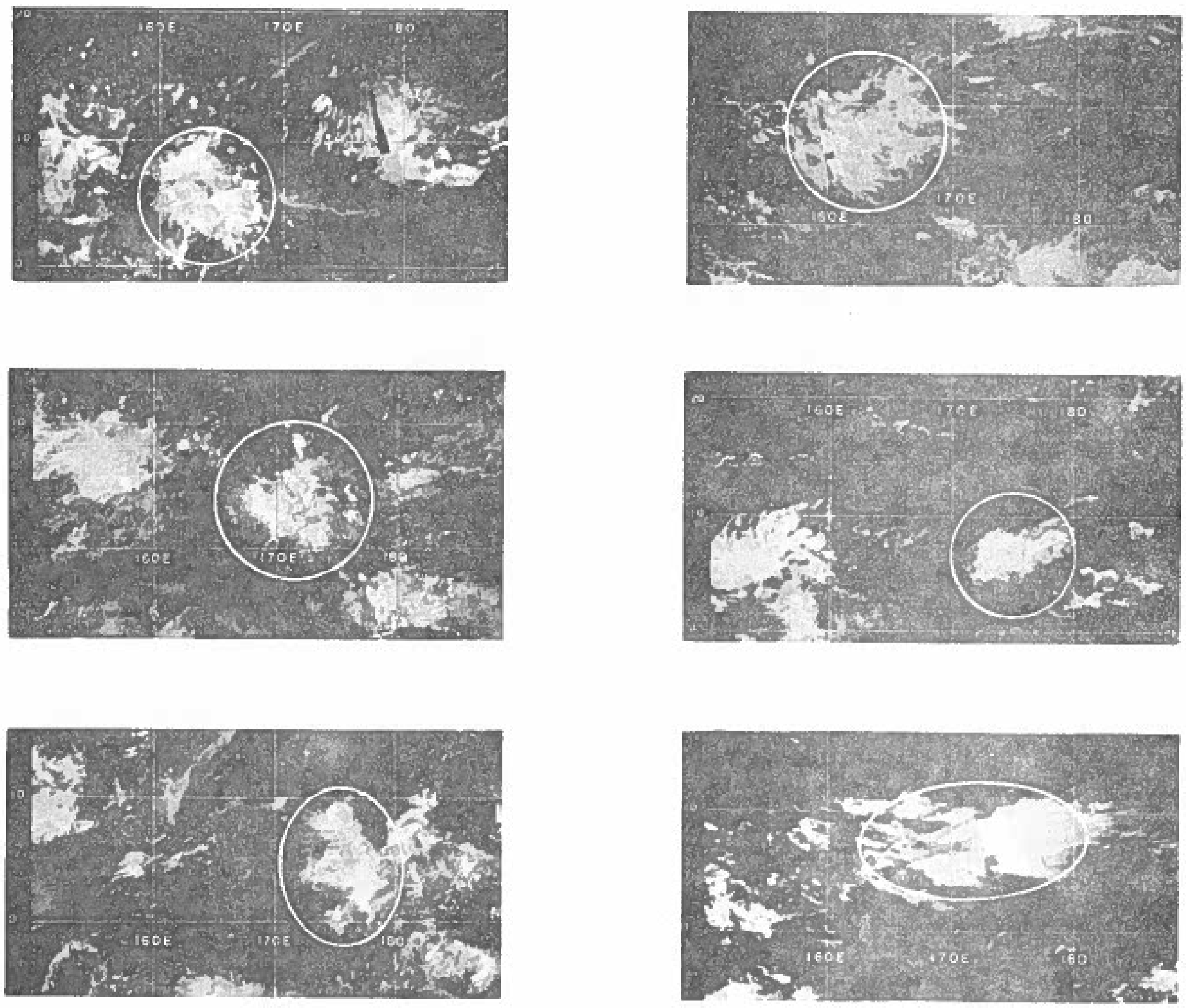


Fig. 6. Location of clear regions in the West Indies.

CLUSTER



CLEAR AREAS

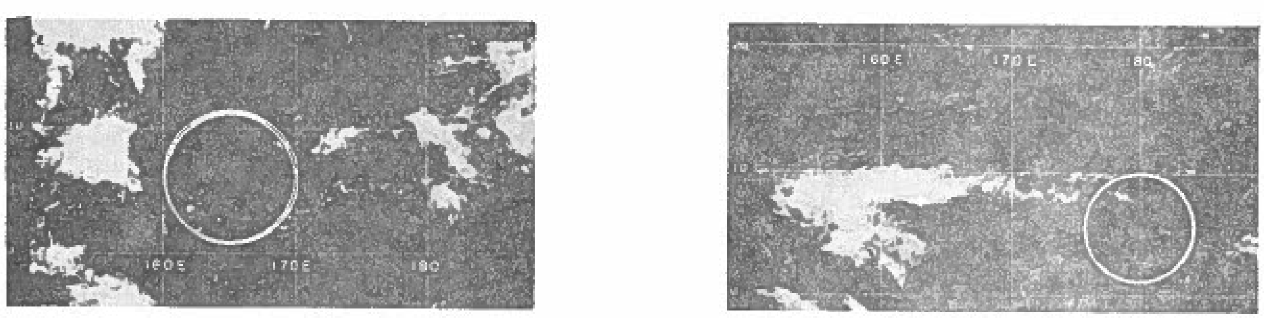


Fig. 7. Typical portrayal of cluster and clear regions.

III. GENERAL TROPICAL BELT RAINFALL, WATER VAPOR, AND ENERGY BUDGET CONSIDERATIONS

Tropospheric conditions in the whole tropical belt of summer are considered. This includes the trade wind and equatorial trough regions from approximately 5° to 25° of latitude. Oort and Rasmusson (1971) have shown (from data derived from MIT Department of Meteorology rawinsonde tapes) that this tropical area is mass, vapor, and energy budget-wise a very self contained region. Their data indicates that the net tropospheric energy divergences in the summertime latitudes from 5-25° latitude are very small in comparison with the net tropospheric radiational cooling. They show,that for mean tropospheric conditions over this region,the vertical motion averages but a few mb/day, meridional induced water vapor changes average only about 0.1 gm/kg and that meridional induced advective cooling-warming rates (averaged through the troposphere) are less than $0.1^{\circ}C$/day. In comparison, local vertical motion to accomplish the required rainfall and balance the net radiation cooling must be of the order of 100 mb/day, and evaporation-rainfall must average about 0.5 cm/day (or $1.3^{\circ}C$ latent heat equivalent for the whole troposphere). Thus, in comparison with net tropospheric radiational losses, the summertime meridional fluxes in the broad tropical belt where cloud clusters exist can be largely neglected. Each latitudinal belt must very closely meet its own energy budget requirements. The longitudinal or Walker circulations allow for some individual longitude budget imbalances, but these largely cancel in the global latitudinal average. Fig. 8 shows that the net meridional divergence of energy is very small in comparison with the net radiation loss. It will be assumed that this broadscale estimate also applies to the summertime oceanic regions by themselves.

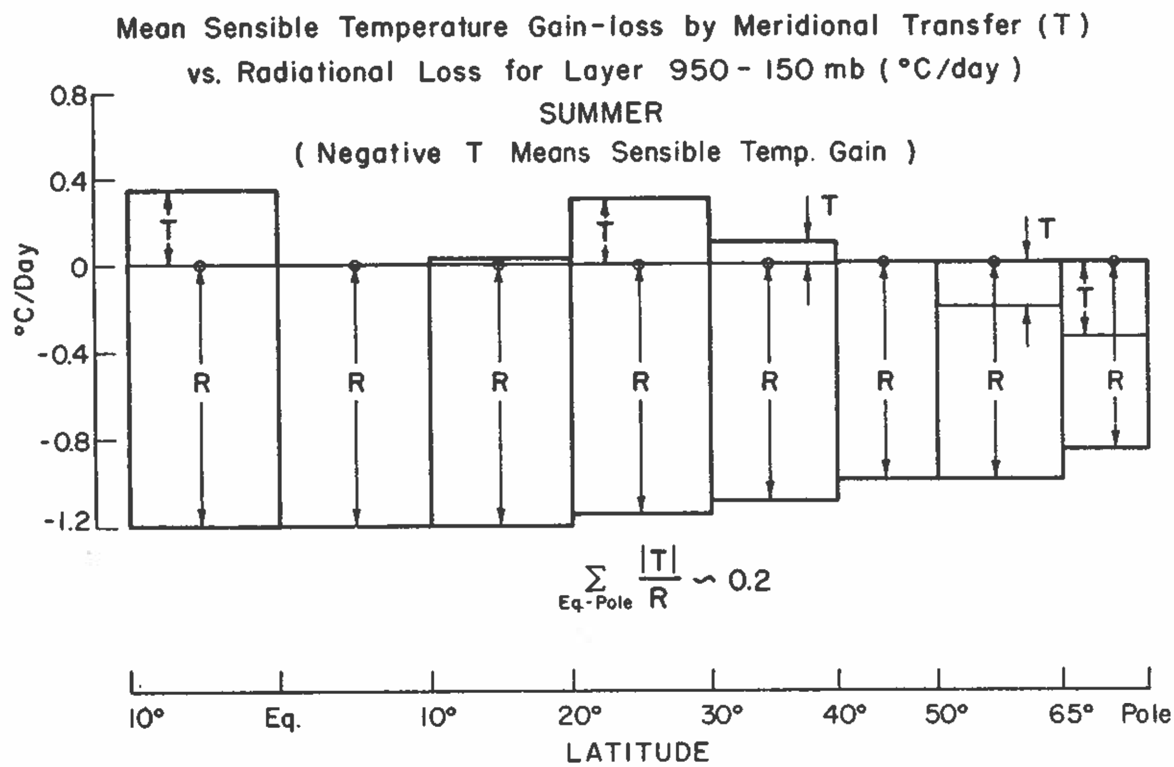


Fig. 8. Comparison of mean tropospheric temperature change due to all meridional energy convergence sources (T) by wind systems vs. net tropospheric radiation loss (R) in (°C/day) for summer. Meridional energy convergence obtained from Oort and Rasmusson (1971). Note lack of any appreciable energy convergence in Equator to 20° latitude belt.

Table 1

ESTIMATES OF EQUATOR TO 30°N NET RADIATION COOLING IN THE LAYER FROM 1000-150 mb. (°C/DAY)
(This tropical region has little seasonal variation)

		Eq-30°
Model Determinations	London (1957)	1.12
	Davis (1963)	1.11
	Rogers (1967)	1.18
	Dopplick (1970)	1.13
	Average	1.14
Measurements	Cox and Suomi (1969) Measurements (with Davis Short-Wave Values)	1.37
	Vonder Haar (1971)	1.26 (Eq-20°)

Over the tropical oceans the ratio of sensible to latent heat transport is felt to be very small (Sellers, 1965). Assuming negligible sensible heat transport, the self-contained nature of the tropical belt requires that the tropospheric radiation loss of about 1.3-1.4°C/day (see Table 1 for various net radiation loss estimates) be balanced by an average summertime rainfall-evaporation rate of about 0.5 cm/day. Previous estimates (Budyko, 1956) indicate that this is about what the summertime tropical rainfall-evaporation rates are. Evaporation of 0.5 cm/day will be assumed everywhere. For simplicity the net tropospheric radiation cooling and the energy of evaporation will be taken to be constant and everywhere balanced as in Fig. 9.

PRESSURE (10^2 mb)	CLOUD CLUSTER (20%) RADIATION (°C/day)	VARIABLE CLOUD REGION (40%) RADIATION (°C/day)	CLEAR REGION (40%) RADIATION (°C/day)
0.5			
1	0	0	0
2	-0.4	-0.4	-0.4
3	-0.9	-0.9	-0.9
4	-1.4	-1.4	-1.4
5	-1.5	-1.5	-1.5
6	-1.7	-1.7	-1.7
7	-1.9	-1.9	-1.9
8	-1.9	-1.9	-1.9
9	-1.6	-1.6	-1.6
10	-1.5	-1.5	-1.5
	EVAP. 0.5 gm/cm² day	EVAP. 0.5 gm/cm² day	EVAP. 0.5 gm/cm² day
	RAIN 2.5 gm/cm² day	NO RAIN	NO RAIN

Fig. 9. Assumed net radiation cooling, evaporation, and rain in the three tropical regions.

Although evaporation is rather uniformly distributed, rainfall is typically concentrated in meso-scale cloud clusters, taking up but 15-20% of the area of the tropical belt. Here the total tropical belt evaporation falls out as rain in average amounts of 2-3 cm/day (Williams, 1970). Other regions possess cloudiness but with negligible amounts of precipitation. The dynamics of the rain, cloud, and clear regions must be quite different. Each region has its own distinctive mass, vapor, and energy budget which should be individually treated.

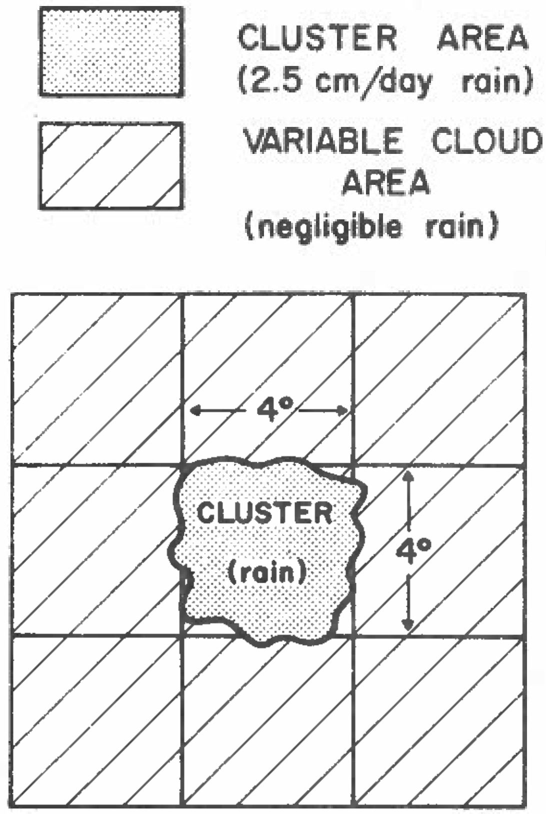


Fig. 10. Rawinsonde composite of variable cloud region was taken in the hatched region surrounding the cloud cluster (shaded area).

Simplified Division of Tropical Belt. To expedite understanding of the tropical belt it has been divided into three meteorological regions. These are:

1) Cloud cluster regions taking up about 20% of the area of the tropical belt and having average rainfall of 2.5 cm/day.

2) Variable cloud regions taking up about 40% of the area of the tropical belt which have scattered to broken cloud conditions but have no applicable rainfall. The average temperature and dew point conditions of this region were obtained from compositing the regions around the cloud cluster as seen in Fig. 10.

3) Clear regions taking up about 40% of the area of the tropical belt.

Fig. 11 and 12 show individual composites of the temperature and dew point for each region obtained from the rawinsonde compositing

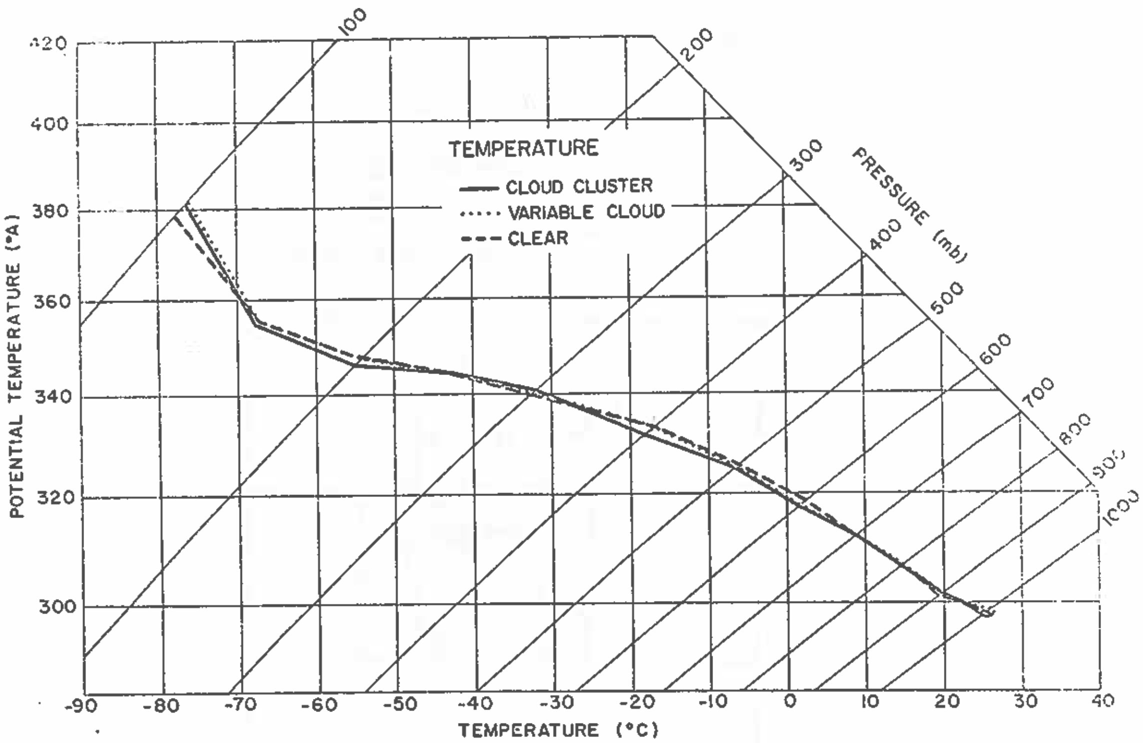


Fig. 11. Composite temperature soundings for three regions.

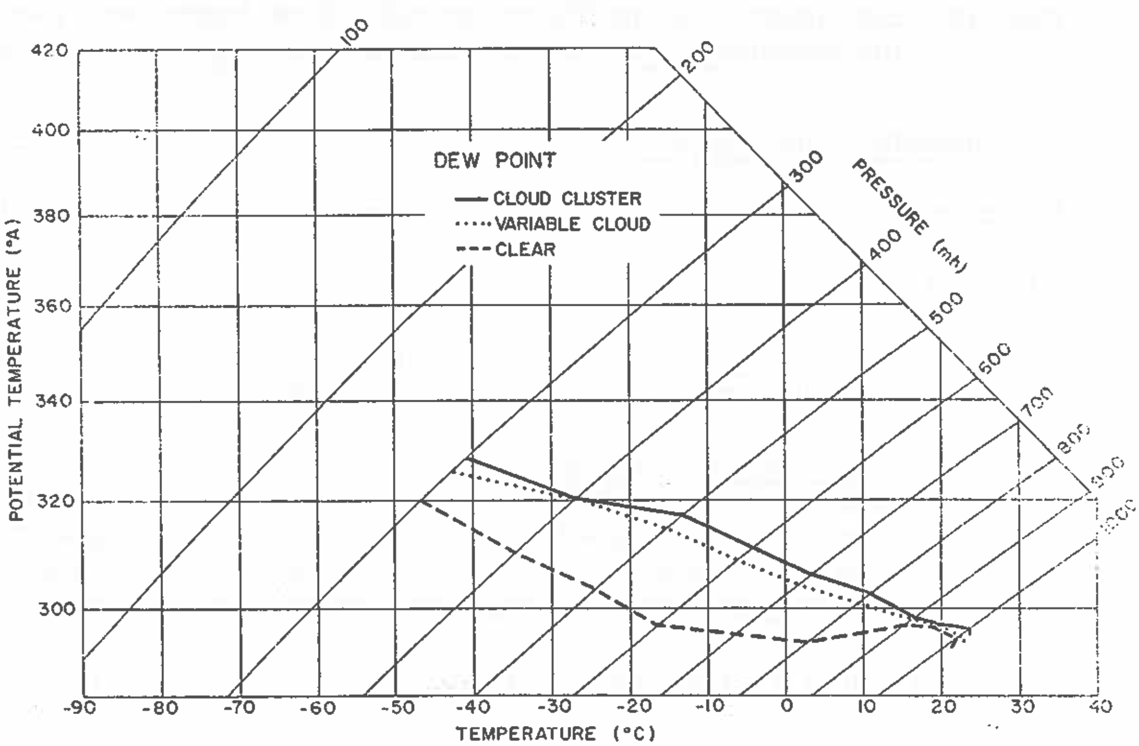


Fig. 12. Composite dew point soundings for three regions.

procedure. It is observed that no significant temperature differences exist between the three regions and that horizontal temperature advection must be near zero. The three regions are primarily contrasted by their middle and lower middle tropospheric moisture differences.

Fig. 13 portrays these three typical regions in idealized cross-section form. Observational research of Chang (1970), Frank (1970, 1971), Wallace (1970), Hayden (1970), and Martin and Suomi (1972) give general support to the above area percentage classification.

Each region will now be discussed in more detail. We will first establish the area divergence and "mean" vertical motion profiles.

Clear Area Region. This is the simplest region. The net radiation cooling as portrayed in Fig. 9 is assumed to be exactly balanced by a sinking compressional warming of about 25-30 mb/day. Composite of the winds around the clear areas independently verify this magnitude of sinking motion. The small decrease of water vapor content as a result of this sinking drying is made up at higher levels by vapor advection from the variable cloud regions.

Cloud Cluster Region. The composited cloud cluster data of the Western Pacific (Williams, 1970) and the West Indies (Gray, 1972 - report to be issued in the near future) show that the meso-scale cluster of about 4^{o} latitude width has a deep, nearly constant convergence up to about 400 mb and a strong concentrated divergent outflow layer at 200 mb. Fig. 14 shows the way in which the data was composited. Reed and Recker (1971) and Yanai (1971) have obtained or implied a similar mass composite profile for a number of easterly waves and/or

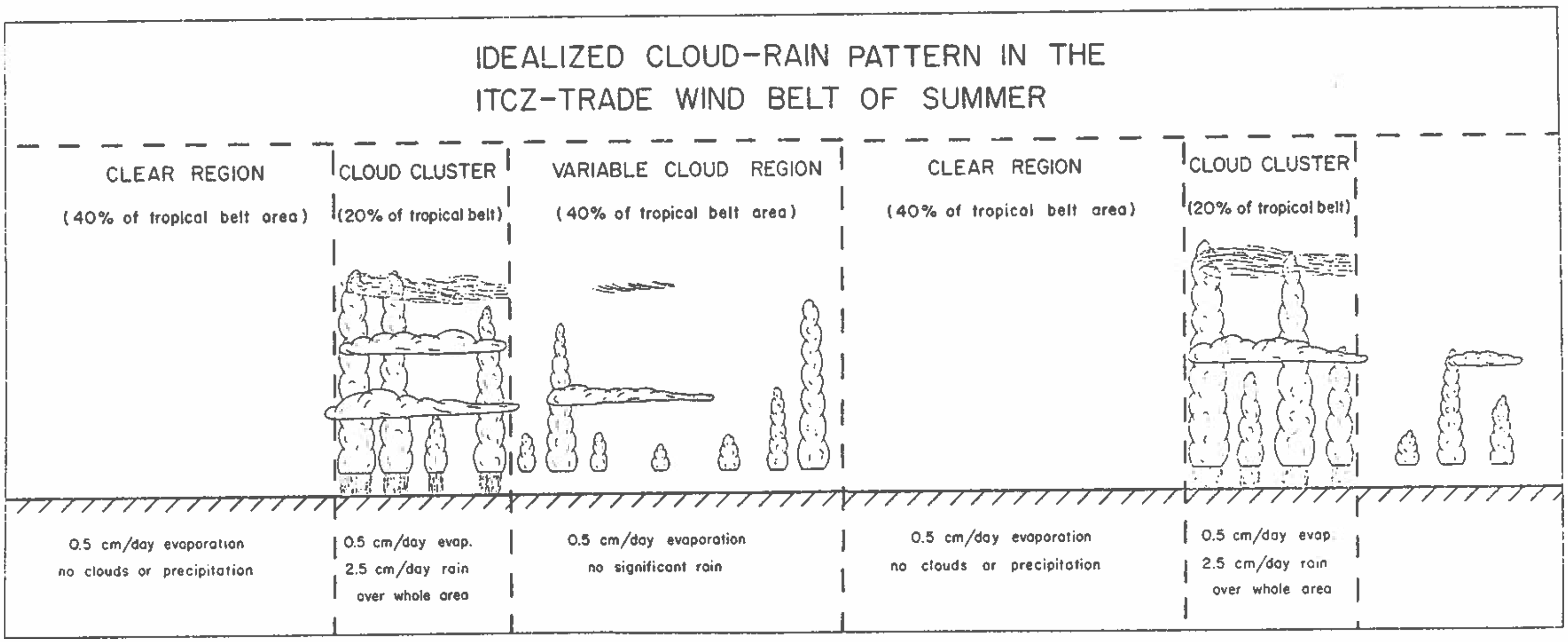


Fig. 13. Schematic picture of cloud and clear areas in the tropics.

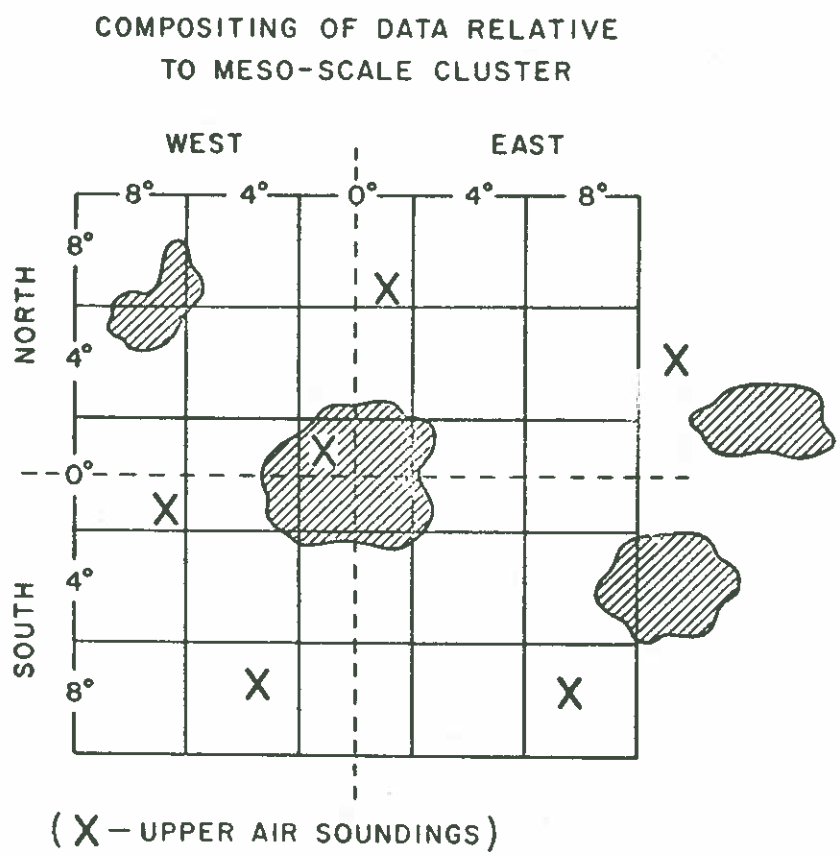


Fig. 14. Illustration of compositing scheme and rectangular grid centered on cloud cluster.

cluster systems in the central Pacific. Other information from tropical disturbances indicate a similar vertical convergence arrangement. This cluster convergence pattern up to 400 mb is responsible for an average 4^{o} cluster water vapor inflow of about 1.5-2.0 gm/cm^2 per day (see Williams, p. 46). This is necessary for a mean cluster rainfall of 2.0-2.5 cm/day if 0.5 cm/day evaporation is occurring underneath the cluster. Some eddy flux of water vapor into the cluster is also occurring. These cluster rainfall estimates have been independently verified from Pacific atoll and island rainfall data composited with respect to the clusters.

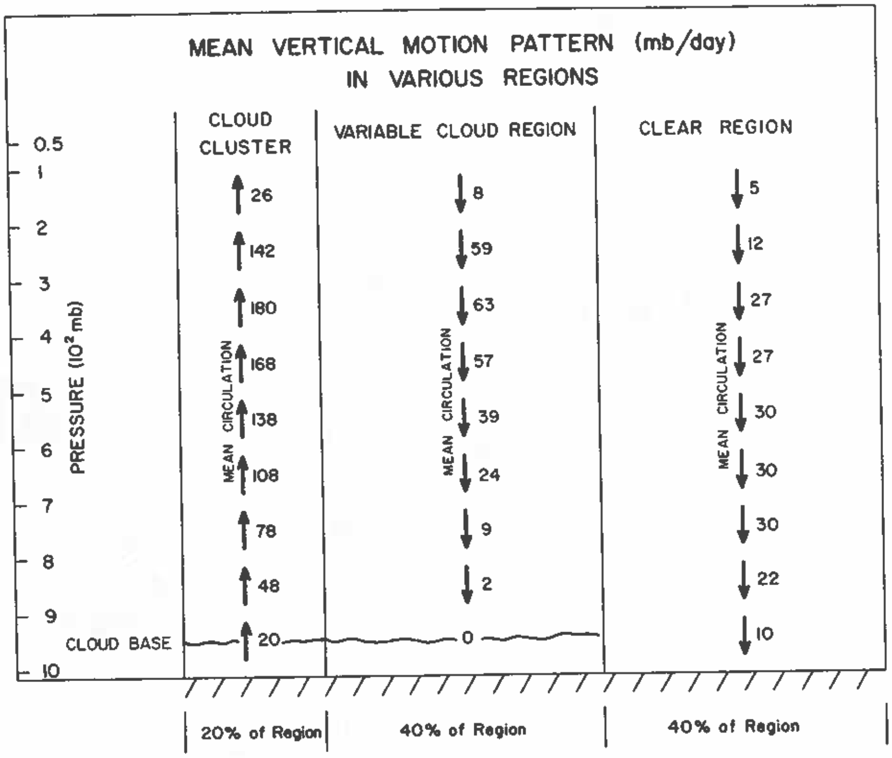


Fig. 15.

The cloud clusters and the clear areas are thus regions which import water vapor. The "mean" upward circulation at low levels as determined by the cluster convergence profile does not, however, carry sufficient water vapor upward to account for the observed upward water vapor advection at low levels. It is necessary to hypothesize a large extra "local" up-moist and down-dry circulation to accomplish all of the required upward vapor transport.

Variable Cloud Region. Given the average vertical motion pattern of the clear and cluster regions, the mean vertical motion of this region is solved for as a residual. This motion is downward everywhere except in the boundary layer. Fig. 15 portrays the mean vertical motion for the three regions in mb/day.

As the variable cloud region must export water vapor to both the cluster and the clear regions, it is also necessary (as with the cluster region) that a large "local" upward vapor transport against the downward vapor transport of the "mean" circulation takes place. The characteristics of these "local" circulations must now be discussed.

IV. DETERMINATION OF LOCAL CIRCULATION

The required local circulation in the cluster and variable cloud regions are determined from the need for upward vapor transport that cannot be accomplished by the "mean" or average upward circulation. The required upward water vapor transport (in gm/cm^2 per day) by the "local" circulation at any level ($\hat{q}_{z_i}$) is obtained by adding the surface evaporation (E_{sfc}) to the integrated net inward horizontal water vapor advection q_H and subtracting from these two quantities the vapor carried upward (in gm/cm^2 per day) by the mean circulation ($\bar{q}_z$) and the vapor which has been condensed to rain ($\dot{R}$). The required upward vapor transport by the local circulation at any level, ($\hat{q}_{z_i}$), is thus given by

$$\begin{pmatrix}\text{Upward}\\\text{Transport}\\\text{of Vapor by}\\\text{"local" cir-}\\\text{culation at}\\\text{level i}\end{pmatrix} = \begin{pmatrix}\text{Evapora-}\\\text{tion From}\\\text{sfc}\end{pmatrix} + \begin{pmatrix}\text{Horizontal}\\\text{Vapor Ad-}\\\text{vection}\\\text{From sfc}\\\text{to level i}\end{pmatrix} - \begin{pmatrix}\text{Vapor}\\\text{Carried}\\\text{by Mean}\\\text{Circula-}\\\text{tion at}\\\text{level i}\end{pmatrix} - \begin{pmatrix}\text{Vapor}\\\text{Conden-}\\\text{sation}\\\text{to Rain}\\\text{Below}\\\text{level i}\end{pmatrix}$$

$$\hat{q}_{z_i} = (E_{sfc}) + (q_H) - \bar{q}_{z_i} - \dot{R}, \quad \text{or}$$

$$\hat{q}_{z_i} = \text{Evap.}_{sfc} + \int_{sfc}^{p_i} q_{H_i} \frac{\delta p}{g} - \bar{q}_z - \int_{sfc}^{p_i} \left(\omega_m \frac{\partial q_s}{\partial p} + \omega_\ell \left(\frac{\partial q_s}{\partial p} - \frac{\partial q_e}{\partial p}\right)\right) \frac{\delta p}{g} \qquad (1)$$

where
- p is pressure
- g is gravity
- i highest level of consideration
- q_s saturated specific humidity

q_e	specific humidity of the environment
ω_m	mean upward moist circulation
ω_ℓ	is the local up-moist and compensating down-dry circulation
$\hat{}$	denote local circulation
$\bar{}$	denote mean circulation

Figs. 16 and 17 show comparisons of the required upward water vapor transport by the mean and by the local circulations for the cloud clusters and for the variable cloud regions. In the variable cloud regions all the upward vapor transport must be accomplished by the local circulation. In the cluster the local circulation is dominant at lower levels, the mean circulation at higher levels. The circulations have equal magnitude at 800 mb.

Once the required upward vapor transport by the local circulation ($\hat{q}_{z_i}$) has been specified at every level, the equal magnitude up-moist and down-dry local circulation (ω_{ℓ_i}) necessary to accomplish this upward vapor transport is determined at each level from the equation

$$\omega_{\ell_i} = \frac{(\hat{q}_{z_i})\,(g)}{(q_s - q_e)} \quad , \tag{2}$$

where $\hat{q}_{z_i}$ is in units of gm/cm^2 day and the other symbols have been defined with equation (1).

Fig. 18 shows the magnitude of this required local mass compensating up-moist and down-dry vertical circulation in mb/day (dashed lines)

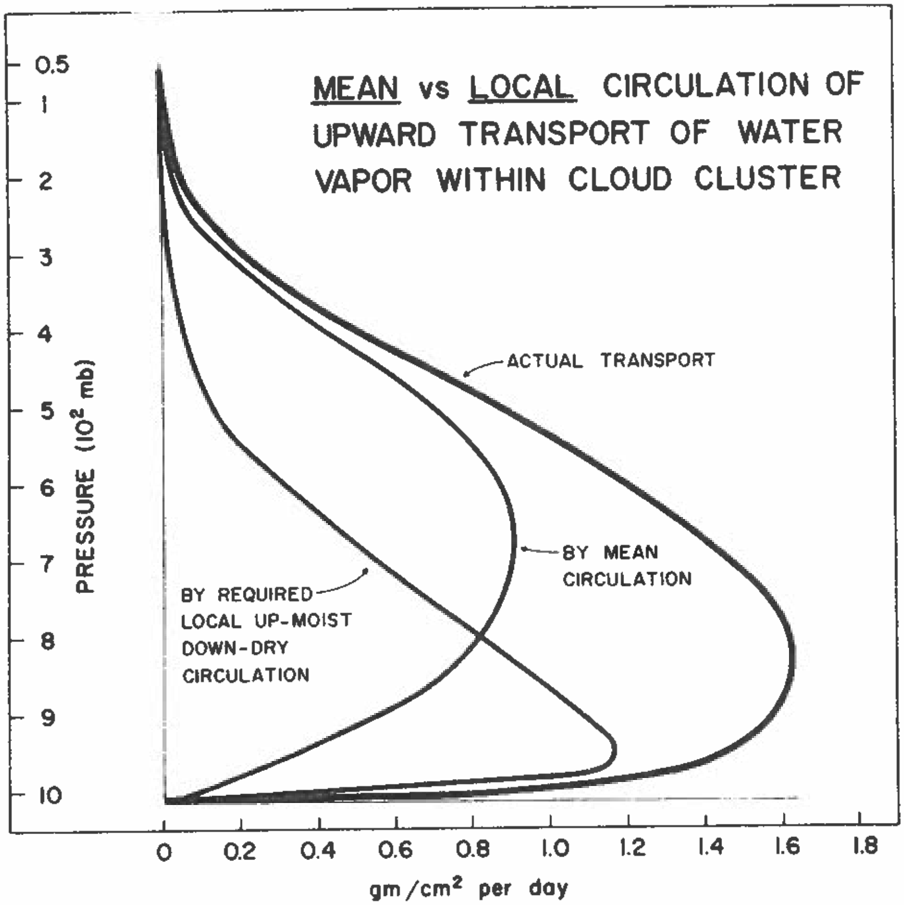

MEAN vs LOCAL CIRCULATION OF UPWARD TRANSPORT OF WATER VAPOR WITHIN CLOUD CLUSTER
ACTUAL TRANSPORT
BY MEAN CIRCULATION
BY REQUIRED LOCAL UP-MOIST DOWN-DRY CIRCULATION
PRESSURE (10² mb)
0.5
1
2
3
4
5
6
7
8
9
10
0
0.2
0.4
0.6
0.8
1.0
1.2
1.4
1.6
1.8
gm/cm² per day

Fig. 16.

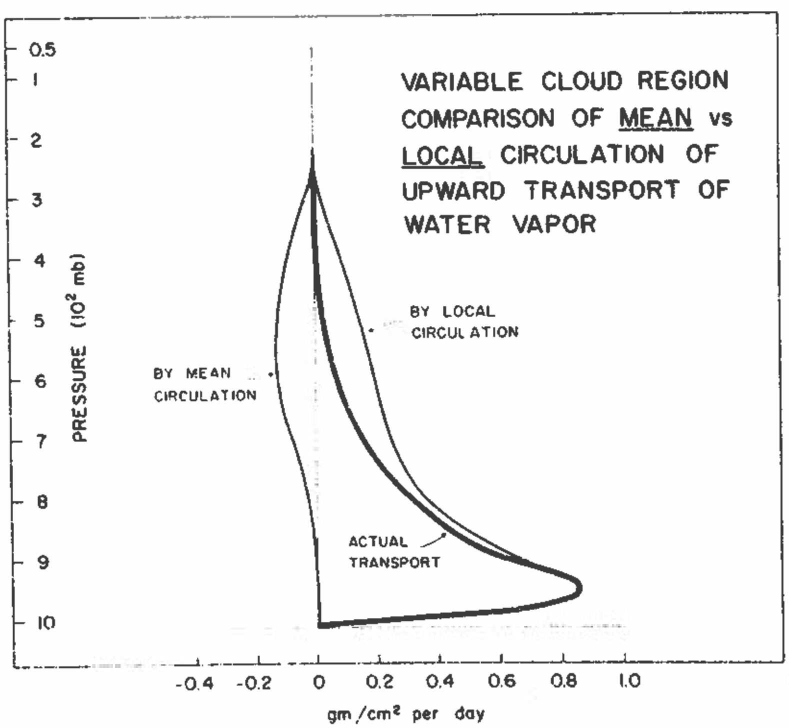

VARIABLE CLOUD REGION COMPARISON OF MEAN vs LOCAL CIRCULATION OF UPWARD TRANSPORT OF WATER VAPOR
BY LOCAL CIRCULATION
BY MEAN CIRCULATION
ACTUAL TRANSPORT
PRESSURE (10² mb)
0.5
1
2
3
4
5
6
7
8
9
10
-0.4
-0.2
0
0.2
0.4
0.6
0.8
1.0
gm/cm² per day

Fig. 17.

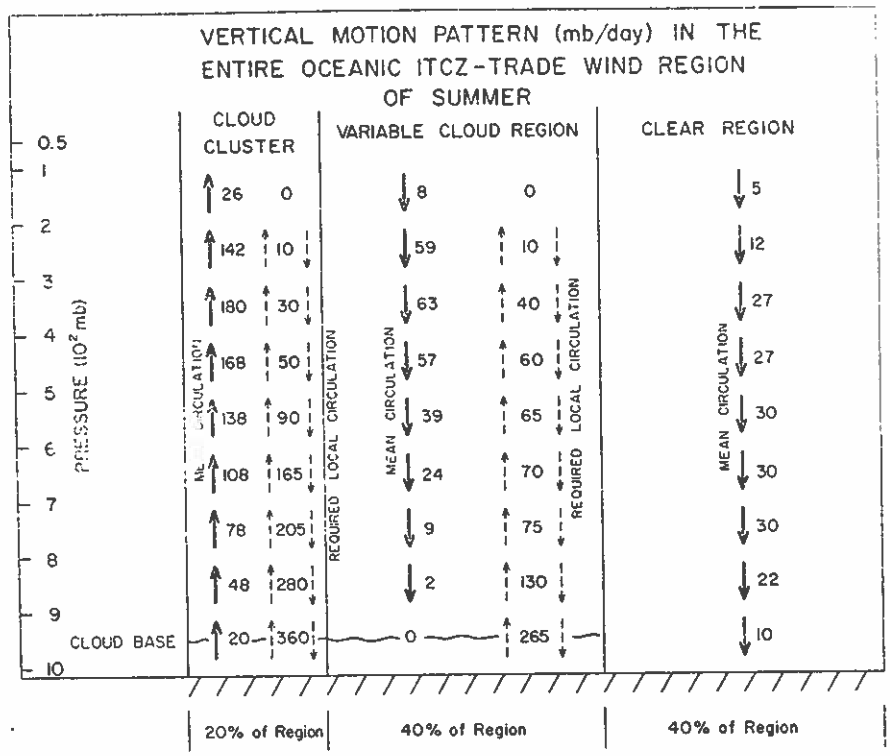


Fig. 18. Comparison of mean vs. local vertical circulation in the various tropical regions. The clear region has no local circulation.

for both the cluster and variable cloud regions and compares it with the mean circulation in these regions. At cloud base (~950 mb) this required local circulation is no less than 360 and 265 mb/day respectively. By contrast the mean circulation is only 20 and 0 mb/day respectively. Low level convergence gives little indication of the actual vertical circulation in operation.

The "local" circulation is thus seen to be a fundamental component to the upward vertical transport of water vapor in both the cluster and in the variable cloud regions. At lower levels it is responsible for most of the upward water vapor transfer. Despite its importance, this mass balancing up-moist and down-dry "local" circulation is typically unresolved by synoptic analysis. It is time that we come to a general understanding about its very large magnitude and probable important role in the vertical circulation processes.

V. COMPUTATIONAL PROCEDURES AND REQUIRED VERTICAL CIRCULATIONS

With the definition of the three distinctive tropical regions and the general discussion of the self contained nature (lack of significant outside meridional transports) of the tropical belt, we are now in a position to estimate the complete mass, vapor, and energy budgets for each region separately. A steady-state assumption for each region has been made.

The computational steps are made in sequence and are outlined by the flow diagram shown in Fig. 19. After the whole tropical belt budgets have been made, the individual region budgets are made for

1) divergence for mean vertical motion, then for
2) water vapor, then for
3) local circulation, then for
4) resulting energy requirements.

Individual Region Divergence and Mean Vertical Motion. The method of determining this has been previously discussed. Figs. 20 and 21 show the divergence and mean vertical motion profiles.

Determination of Individual Region Water Vapor Budgets. These were made in a way similar to the mass budgets. The cluster vapor convergence vertical profile was determined from the composite radiosonde information. The clear region vapor budget was specified by the rate of drying of each region due to the mean sinking motion. The variable cloud region vapor profile was obtained as a residual of the cluster and clear regions.

FLOW DIAGRAM OF COMPUTATIONAL METHOD

A	B	C	D	E
ASSUME TOTAL TROPICAL BELT ENERGY BUDGET REQUIREMENTS →	DETERMINE INDIVIDUAL REGION MASS BUDGETS AND MEAN VERTICAL CIRCULATIONS →	DETERMINE INDIVIDUAL REGION WATER VAPOR BUDGETS →	DETERMINE REQUIRED LOCAL UP-MOIST & DOWN-DRY VERTICAL CIRCULATIONS →	SOLVE INDIVIDUAL REGION ENERGY BUDGET FROM REQUIRED SINKING MOTION
Assume whole tropical oceanic belt a self contained energy system	Determine cluster divergence from obs. in W. Pacific and W. Indies	Obtain cluster rainfall and vapor convergence from obs. Determine vertical vapor transport requirements	Determine cluster "local" circulation from vertical vapor transport requirements that "mean" circulation does not accomplish	Assume negligible horizontal advection changes
Assume negligible sensible heat transfer from ocean	Assume clear region radiation cooling is balanced by subsidence sinking	Determine clear region vapor convergence from sinking drying	Determine variable cloud region "local" circulation from vertical vapor transport requirements that "mean" circulation does not accomplish	Determine compressional heating from local sinking rates
Balance net radiation with rainfall	Solve for variable cloud region mass profile as a residual of cluster and clear regions	Solve for variable cloud region moisture source requirements as a residual, then determine this region's vertical vapor transport needs	Clear region has no local circulation	Solve for evaporation cooling as a residual of compressional heating minus radiation
Balance rainfall with evaporation				

Fig. 19.

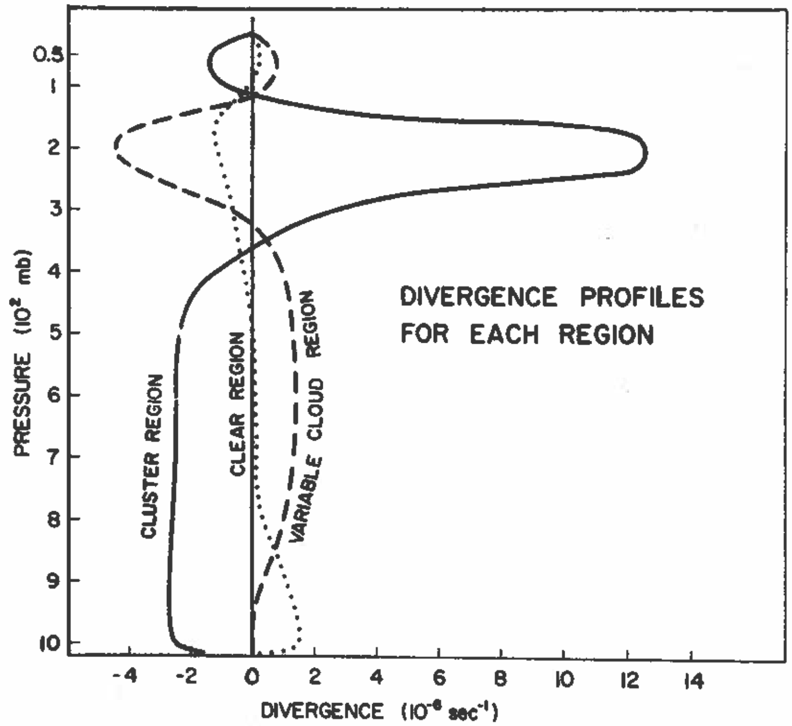

DIVERGENCE PROFILES
FOR EACH REGION
CLUSTER REGION
CLEAR REGION
VARIABLE CLOUD REGION
PRESSURE (10^2 mb)
0.5
1
2
3
4
5
6
7
8
9
10
-4
-2
0
2
4
6
8
10
12
14
DIVERGENCE (10^{-6} sec^{-1})

Fig. 20.

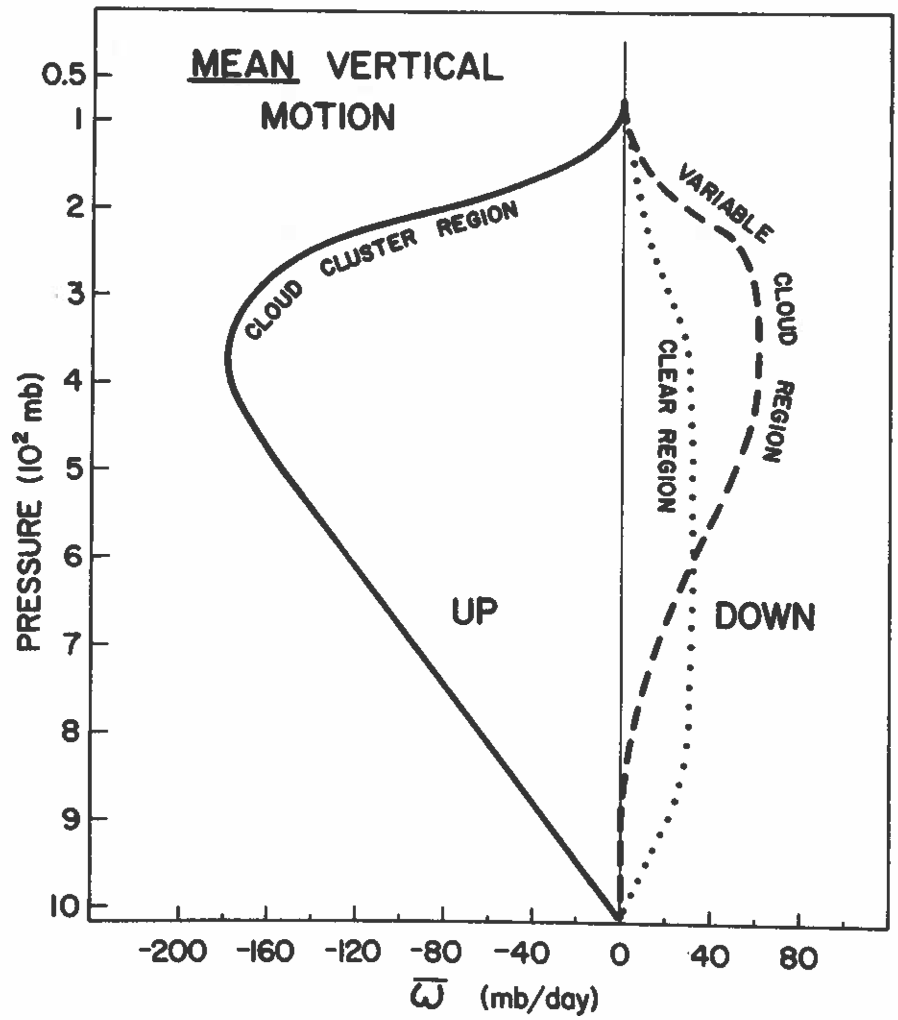

MEAN VERTICAL
MOTION
CLOUD CLUSTER REGION
VARIABLE CLOUD REGION
CLEAR REGION
UP
DOWN
PRESSURE (10^2 mb)
0.5
1
2
3
4
5
6
7
8
9
10
-200
-160
-120
-80
-40
0
40
80
$\overline{\omega}$ (mb/day)

Fig. 21.

As the rainfall of the cloud cluster is approximately five times the amount of the evaporation underneath it, it is obvious that a substantial vapor import must take place. This vapor import is accomplished by two processes:

1) Convergence of vapor from the surface to 400 mb due to the mean convergence profile as seen in Fig. 20. This accounts for most of the vapor convergence into the cluster.

2) Eddy advection of vapor into the cluster beyond that specified by the divergence. This occurs primarily in the lowest 50-150 mb and is analogous to the extra vapor advection into squall lines when they overtake lower level air, as discussed by Newton, (1963). Typically, dryer downdraft air leaves the cluster on the upwind or east side while moister lower level air enters the cluster from the downwind or west side. This brings about a net moisture advection into the cluster. Zipser (1969, and personal communication) has discussed the large drying from downdrafts which occurs in clusters. Cluster surface winds are but half of the winds at the 800-900 level, as seen in Fig. 22.

An estimate of the vertical distribution of vapor advection into the cluster below any level p_i from these two processes can thus be obtained from the equation

$$\text{Vapor Advection into Cluster} = \underbrace{\int_{sfc}^{p_i} \nabla_2 \bullet q \, \frac{\delta p}{g}}_{\text{Vapor Transfer by Convergence}} + \underbrace{\int_{sfc}^{p_i} \Delta q_i \, \frac{|V_i - V_c|}{D} \, \frac{\delta p}{g}}_{\text{Extra Vapor into Cluster at Lower Levels by Relative Motion}} \qquad (3)$$

where p_i is the pressure at any level

Δq_i vapor difference between that entering and leaving cluster at any level - typically 1-5 gm/kg at lower levels.

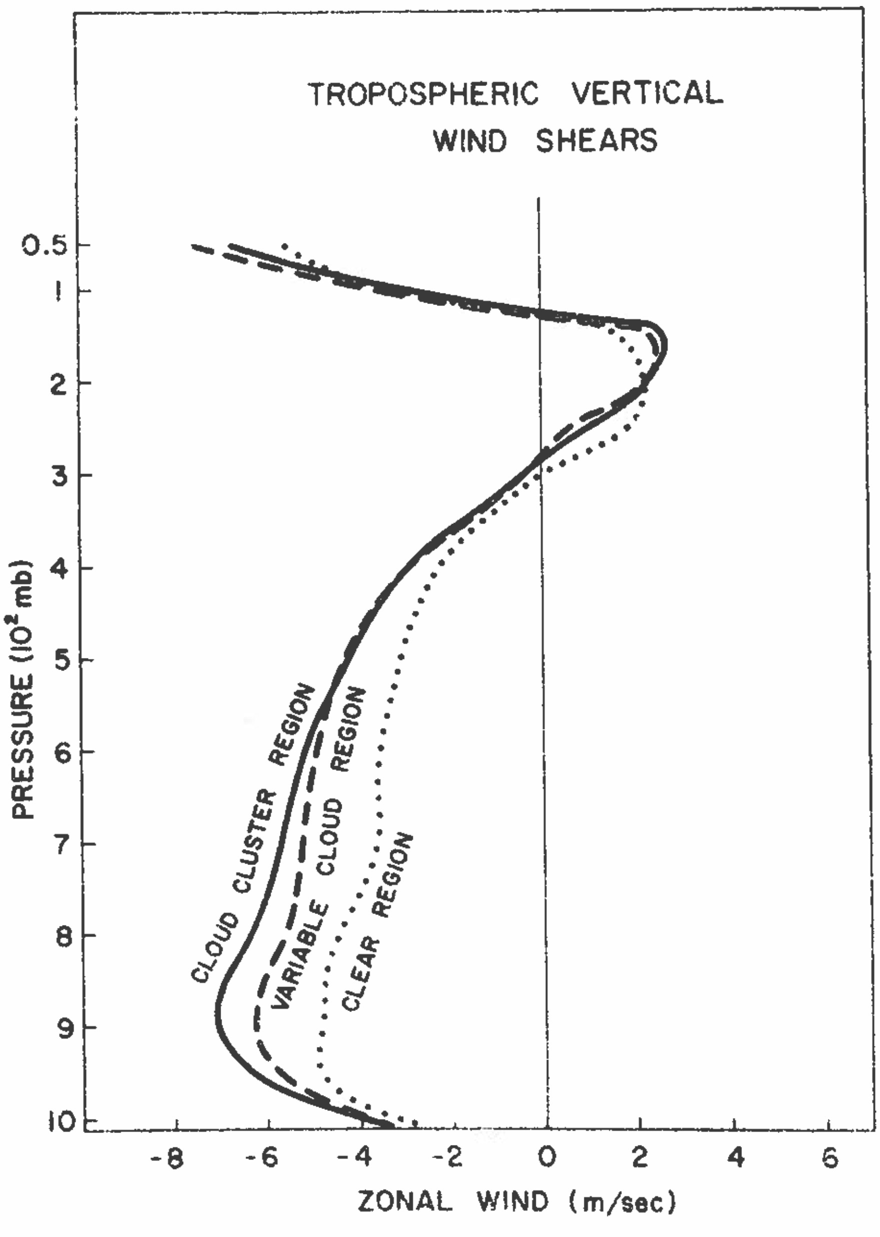


Fig. 22. Vertical profiles of average zonal wind in the three tropical regions.

V_i is the wind at any level in the cluster
V_c is the cluster velocity
D is the width of the cluster - taken as 4° latitude.

The main contribution of the second term on the right of (3) comes in the boundary layer. Here the speeds are substantially less than the cluster velocity. The vapor advection determined from (3) is shown for the cluster on the left portion of Fig. 23. Note that half of the vapor advection comes in the boundary layer (surface to 950 mb).

As previously shown in Fig. 16, the mean circulation in the cluster carries a small fraction of the required upward transport of vapor necessary to produce the 2.5 cm/day mean rainfall. The rest of the upward vapor transport must be carried by the "local" circulation.

Figs. 24-25 and Table 2 discuss in more detail the water vapor characteristics of the cloud cluster. Figs. 24-25 contrast level by level the upward vapor transports and the vapor released to rain by the mean and by the local circulation. It is to be noticed that about three-quarters of the cluster condensation going to rain comes from the mean circulation,while the local circulation is primarily responsible for the recycling of vapor as most of its condensation is reevaporated in its compensating sinking motion. Table 2 explicitly lists all of the parameters going into the cluster condensation, rainfall, and local circulation determination.

It should be realized that the 2.5 cm/day rainfall is not occurring everywhere within the cluster but is probably concentrated in local rain areas (some along squall lines) taking up but 10% or less of the 4° wide

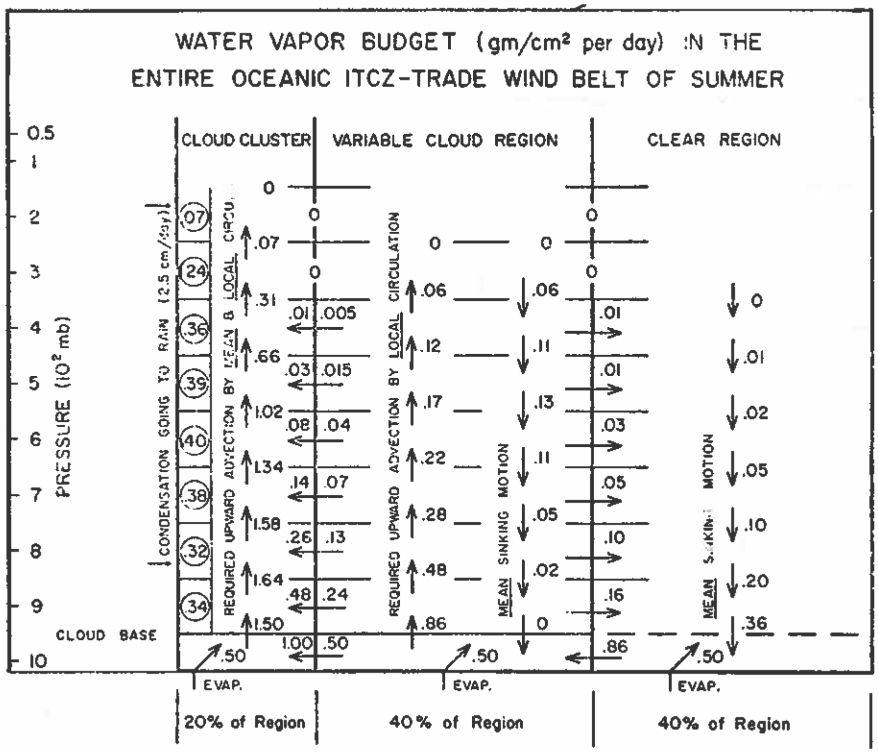


Fig. 23. Water vapor budget of the tropical belt expressed in gm/cm^2 per day. Doubling of vapor advection per unit area into cluster is due to variable cloud region being twice the area of cluster region.

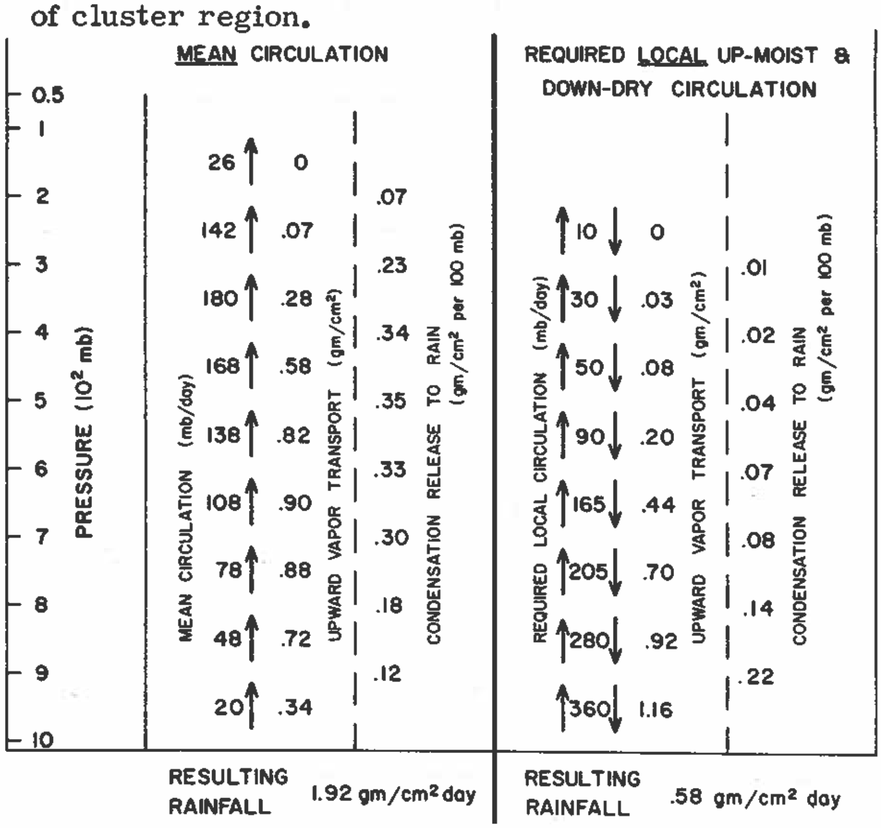


Fig. 24. Cloud cluster mean vs. local circulation values of upward vapor transport and of condensation going to rain.

Table 2

DETERMINATION OF CLUSTER REQUIRED LOCAL CIRCULATION

LEVEL (mb)	Mean Circulation (mb/day)	Condensation to Rain, Mean & Local Circulation (gm/cm^2 per 100 mb per day): Mean	Local	Tot.	Moisture Source: Evaporation and Horizontal Vapor Advection (gm/cm^2 per day)	Moisture Loss Through Condensation to Rain (gm/cm^2 per 100 mb per day)	Total Required Upward Vapor Advection after Subtraction of Condensation to Rain (gm/cm^2 per day)	Moisture Carried by Mean Circulation (gm/cm^2 per day)	Required Upward Vapor Advection at layer base by Local Circulation (gm/cm^2 per day)	Cluster Saturated minus Observed Specific Humidity (gm/kg)	Required Local Up-Moist and Down-Dry Circulation (mb/day)	LEVEL (mb)
sfc	0				EVAP. .50 sfc							
1000					1.00							
950	20				CU BASE		1.50	.34	1.16	3.2	360	950
900		.12	.22	.34	.48	.34						
850	48						1.64	.72	.92	3.3	280	850
800		.18	.14	.32	.26	.32						
750	78						1.58	.88	.70	3.4	205	750
700		.30	.08	.38	.14	.38						
650	108						1.34	.90	.44	2.7	165	650
600		.33	.07	.40	.08	.40						
550	138						1.02	.82	.20	2.0	90	550
500		.35	.04	.39	.04	.39						
450	168						.66	.58	.08	1.7	50	450
400		.34	.02	.36	.00	.36						
350	180						.31	.28	.0[illegible]	1.0	30	350
300		.23	.01	.24	.00	.24						
250	142						.07	.07	.002	0.2	10	250
200		.07	.00	.07	.00	.07						
150	26						.00	.00	—	0.0	—	150
100		.00	.00	.00		.00						
50	0											
		1.92	.58	2.50 VERTICAL TOTALS	TOTAL 2.50 cm/day	2.50 gm/cm^2						

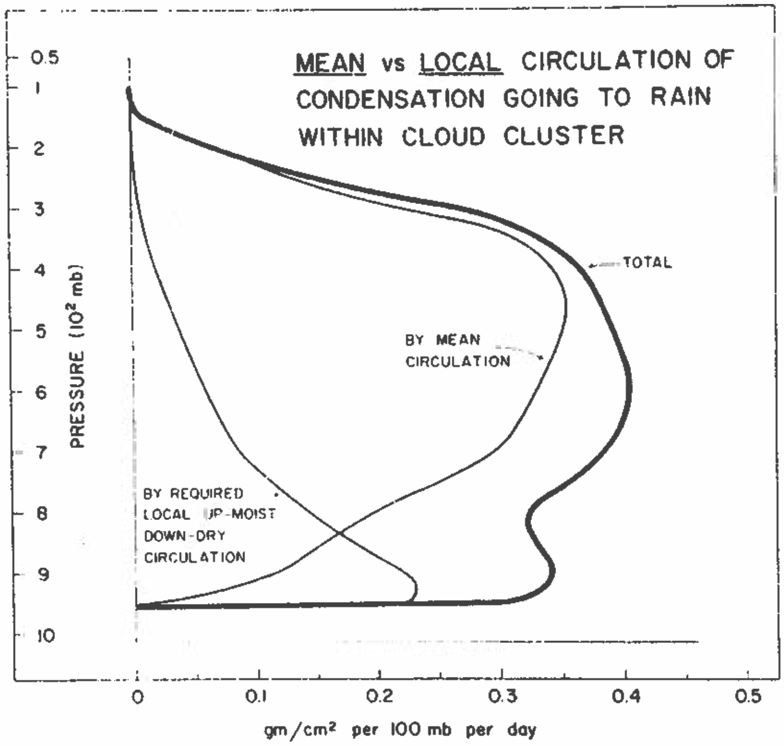


Fig. 25.

cluster. In addition it is likely that only about 10 to 20% (or 0.2 to 0.4% of the entire tropical belt) of the rain areas have active towering cumulus or cumulonimbus updrafts in operation. A similar scaling of the relative areas of the updraft, the rain regions, weather systems, etc., has been presented by Riehl and Malkus (1958). Even though the convective activity is highly concentrated within the cluster, the lack of appreciable temperature and water vapor gradients across the cluster allow for an area average treatment of the mass, vapor, and energy budgets.

The clear and variable cloud region vapor budgets are also shown in Fig. 23. The clear regions import vapor at upper levels to balance their sinking drying motion (.36 gm/cm^2 per day) and export this vapor into the boundary layer. Together with its evaporation of $0.5\,gm/cm^2$, the boundary layer accumulates vapor at a rate of $.86\,gm/cm^2$. The clear areas are thus source regions of water vapor. This is due primarily to their boundary layer divergence. At upper levels they import vapor.

The variable cloud region exports the vapor which it receives from its evaporation--that is $0.5\,gm/cm^2$ per day. Above the boundary layer, it exports $.86\,gm/cm^2$ per day which can only occur if the vapor is carried to upper levels by a substantial local circulation. Below cloud base the variable cloud region imports $.36\,gm/cm^2$ per day.

Fig. 23 illustrates the mode by which the clear and variable cloud regions supply their evaporated vapor to the clusters.

Individual Region Local Circulations. This has been adequately discussed in Section 3. The clear areas have no local circulation. The required up-moist and down-dry circulation of the other two regions is obtained from equation (2) with the required individual level upward transports specified in Fig. 23. As previously shown in Fig. 18, this local required vertical circulation is exceedingly large in the lower troposphere.

Determination of Individual Region Energy Budgets. Once the required regional net downward motion (ω_d) from both the mean and local circulations has been determined, the compressional warming, the water vapor recycling, and the net cloud cooling can be obtained for each region from the need to maintain steady-state conditions. As previously discussed, inter-regional temperature advection influences are negligible. Thus, at each level, the sensible temperature budget must be given as a balance between

$$\begin{pmatrix}\text{Sinking}\\\text{Warming}\end{pmatrix} + \begin{pmatrix}\text{Cloud Induced}\\\text{Temperature Changes}\end{pmatrix} + \begin{pmatrix}\text{Net}\\\text{Radiation}\end{pmatrix} = 0 \text{ , or}$$

$$\omega_d(\Gamma_d - \Gamma_e) + \begin{pmatrix}\text{Cloud Sensible}\\\text{Temperature Warming}\\\text{+ Evaporation Cooling}\end{pmatrix} + R = 0 \text{ ,} \qquad (4)$$

where Γ_d and Γ_e are the dry and environment lapse rates. López (1972a) has shown in his entire life cumulus model that direct sensible temperature diffusion from cumulus to their environment is very small, and in comparison with the evaporational cooling induced by the cumulus, can be neglected. Most of the cloud-induced sensible temperature changes result from liquid water evaporation on the sides of the cumulus during their growth and throughout the cumulus' interior as they die. This is pictorially portrayed in Fig. 26.

The sensible temperature budget of the summertime tropical belt is largely determined by the balance between sinking warming and by evaporational and radiational cooling. Thus,

$$\text{Sinking Warming} \approx \text{Evaporation} + \text{Radiation}$$

$$\omega_d(\Gamma_d - \Gamma_e) \approx L(\Delta q) + R \text{ ,} \qquad (5)$$

where Δq represents the amount of cumulus liquid water converted to vapor per unit mass and time. Throughout most of the summertime

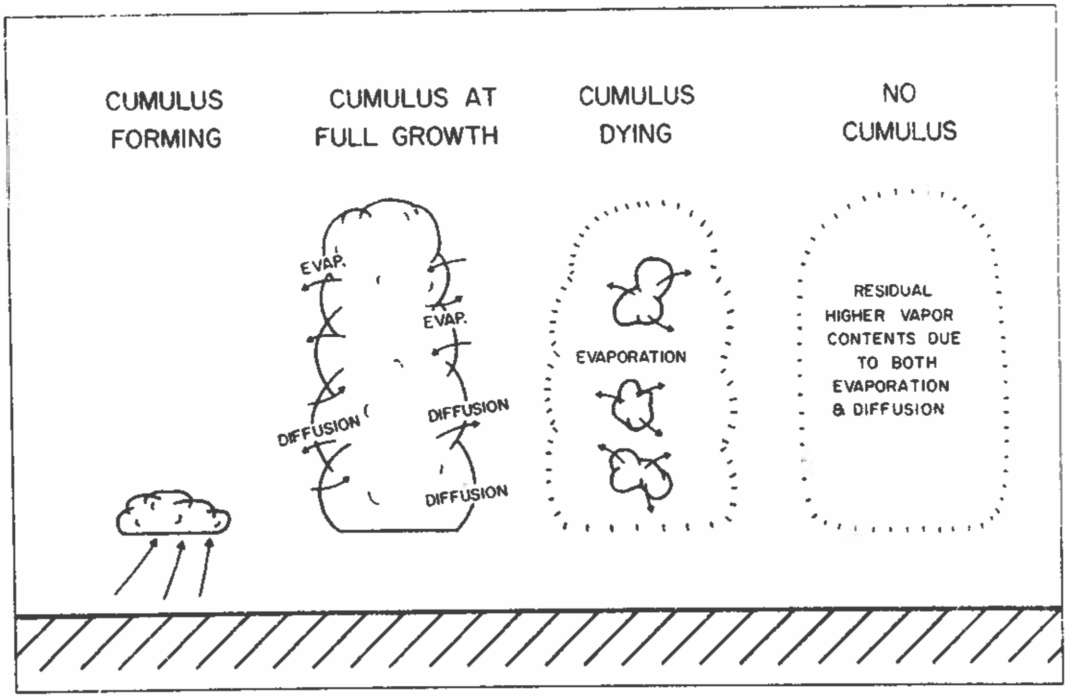


Fig. 26. Idealized picture of how the individual cumulus cools the atmosphere around it when it dies and how it raises the water vapor level.

tropical belt, the sinking warming and evaporation cooling are much larger than the radiation cooling.

Figs. 27-29 portray the vertical profile of the terms of equation (5) for each of the three tropical regions and for the entire tropical oceanic belt. In all but the clear regions, the primary balance is between the sinking warming and the evaporation cooling with the radiation cooling of much less local importance. It was not previously expected that the local cloud induced warming and cooling rates would be so large in comparison with the radiational cooling.

Fig. 30 shows the very large local recycling of water vapor for each of the tropical regions. Each region losses vapor through subsidence

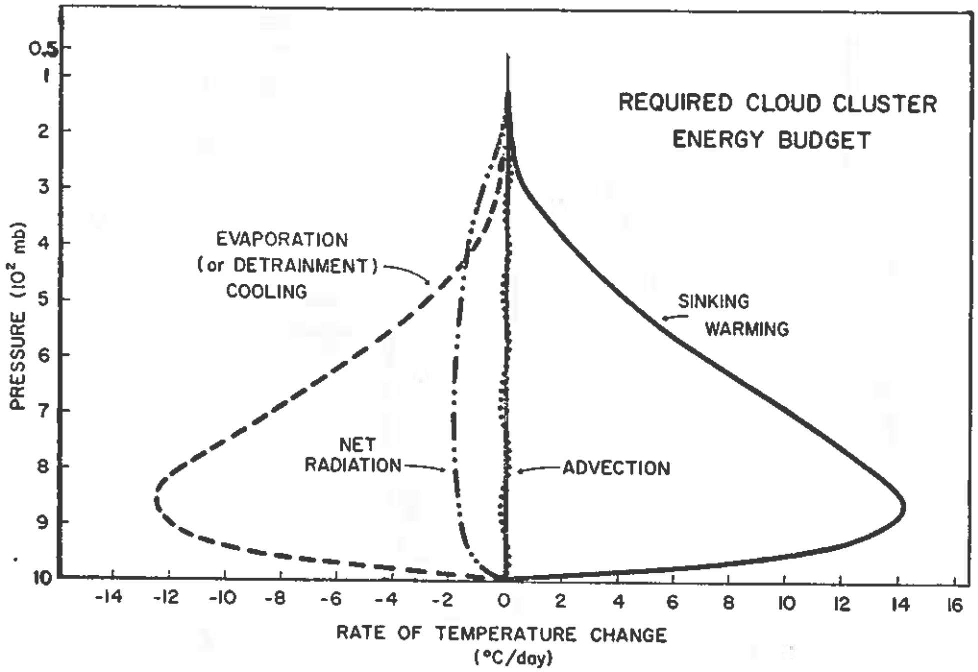


Fig. 27. Vertical distribution of required energy components in steady-state cloud cluster.

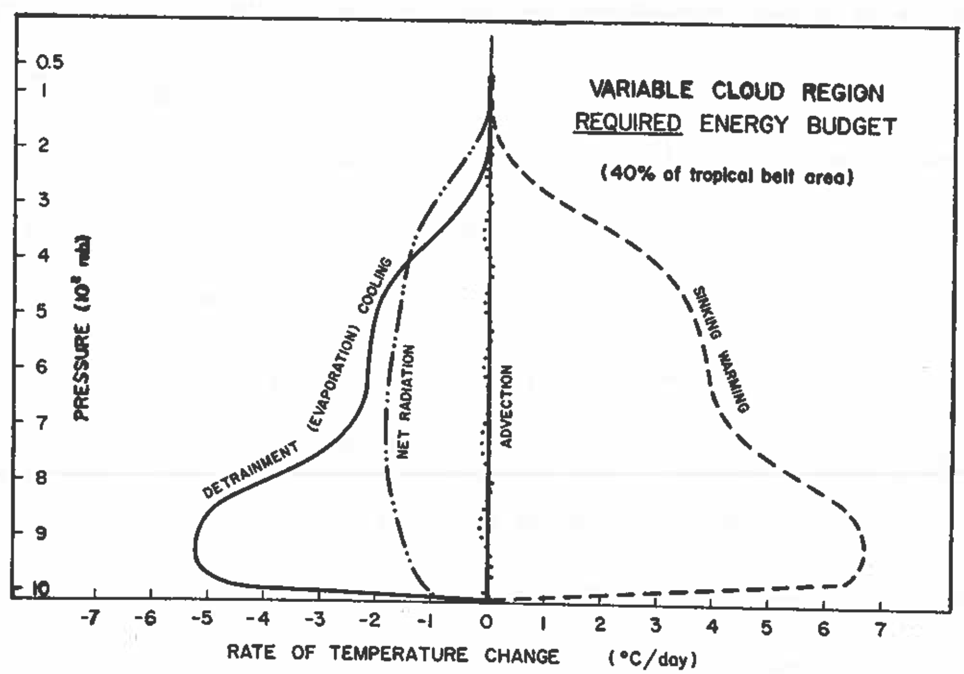


Fig. 28. Vertical distribution of required energy components in variable cloud region.

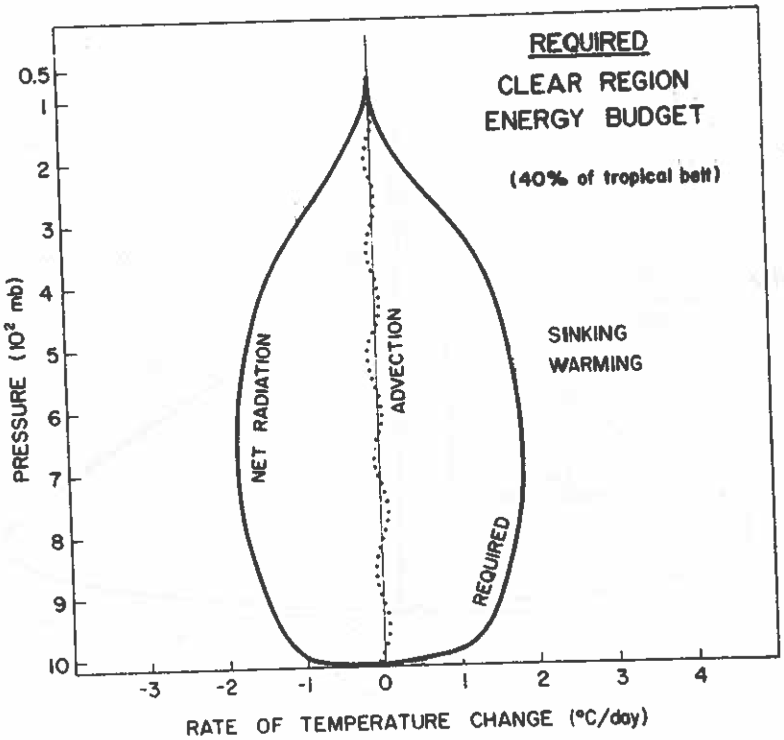


Fig. 29. Vertical distribution of required energy components in clear region.

drying. For steady conditions to prevail, the individual clouds must furnish an equal amount of vapor either by (1) direct vapor diffusion out from the clouds, or by (2) advection or mixing of liquid water (to be evaporated) out from the clouds. Both of these processes are accomplished from the sides of the clouds or throughout the region occupied by the clouds as they die, as implied in Fig. 26.

The relative magnitude of these two processes of vapor replacement to the sinking motion is specified in Figs. 31-33 for each of the tropical regions. It is seen that in the lower levels most of the vapor replacement comes from direct diffusion or advection of vapor out from the cloud. At middle and upper levels most of the vapor replacement comes

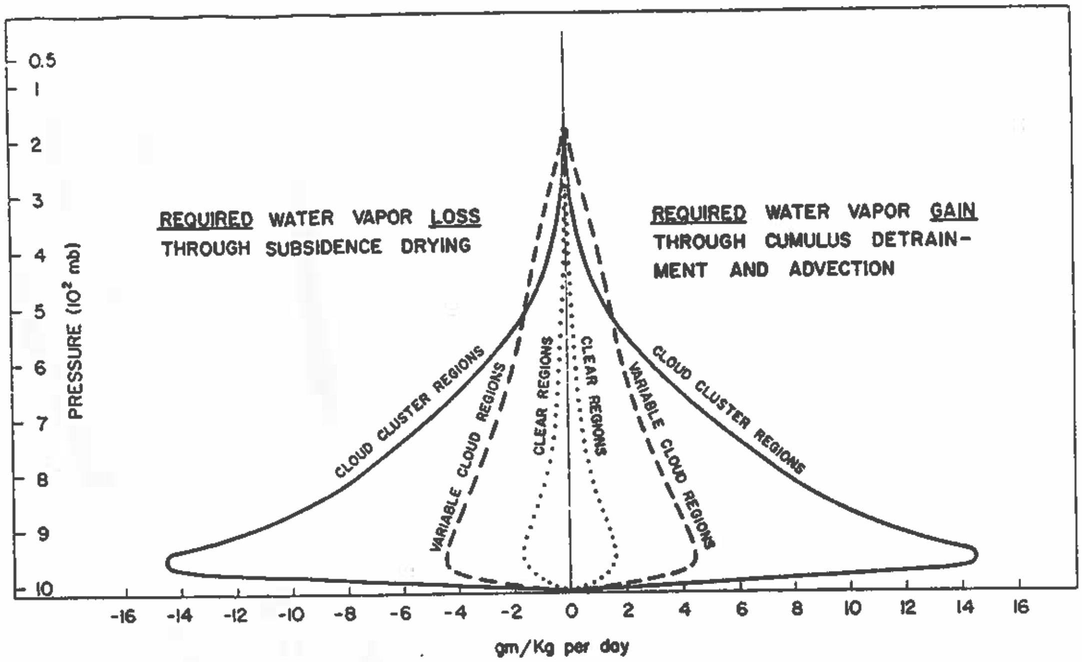


Fig. 30. Comparison of required water vapor gain and loss for each of three tropical regions.

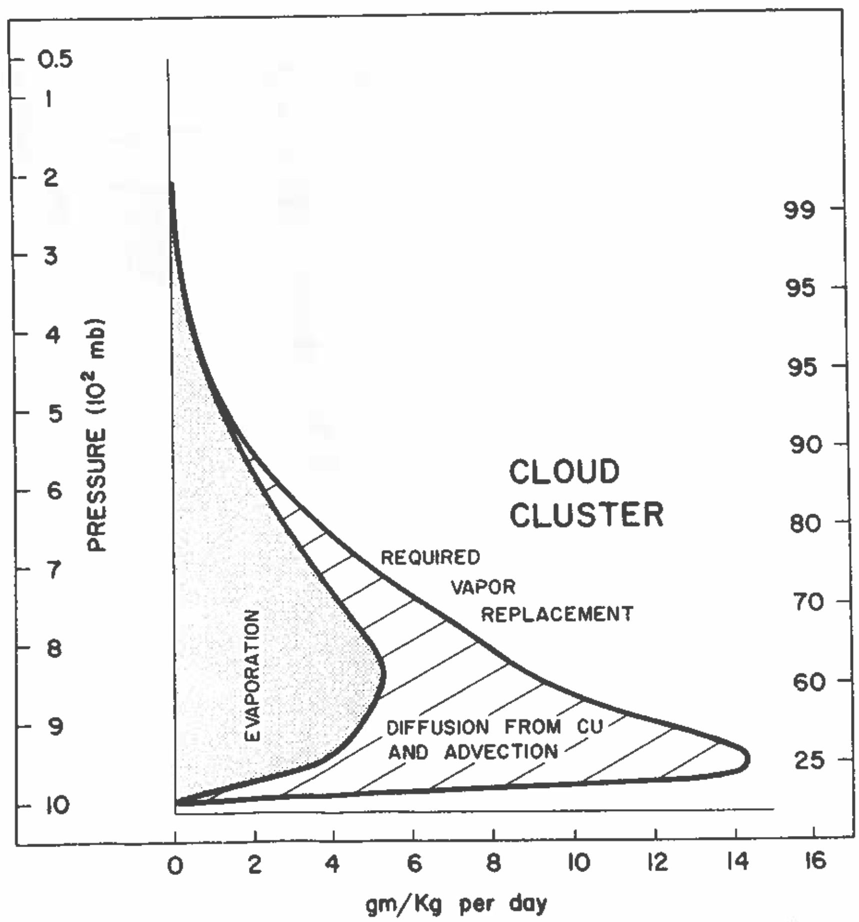


Fig. 31. Cluster required water vapor replacement to sinking motion to maintain steady state conditions and portion of vapor replacement coming from evaporation (on the right).

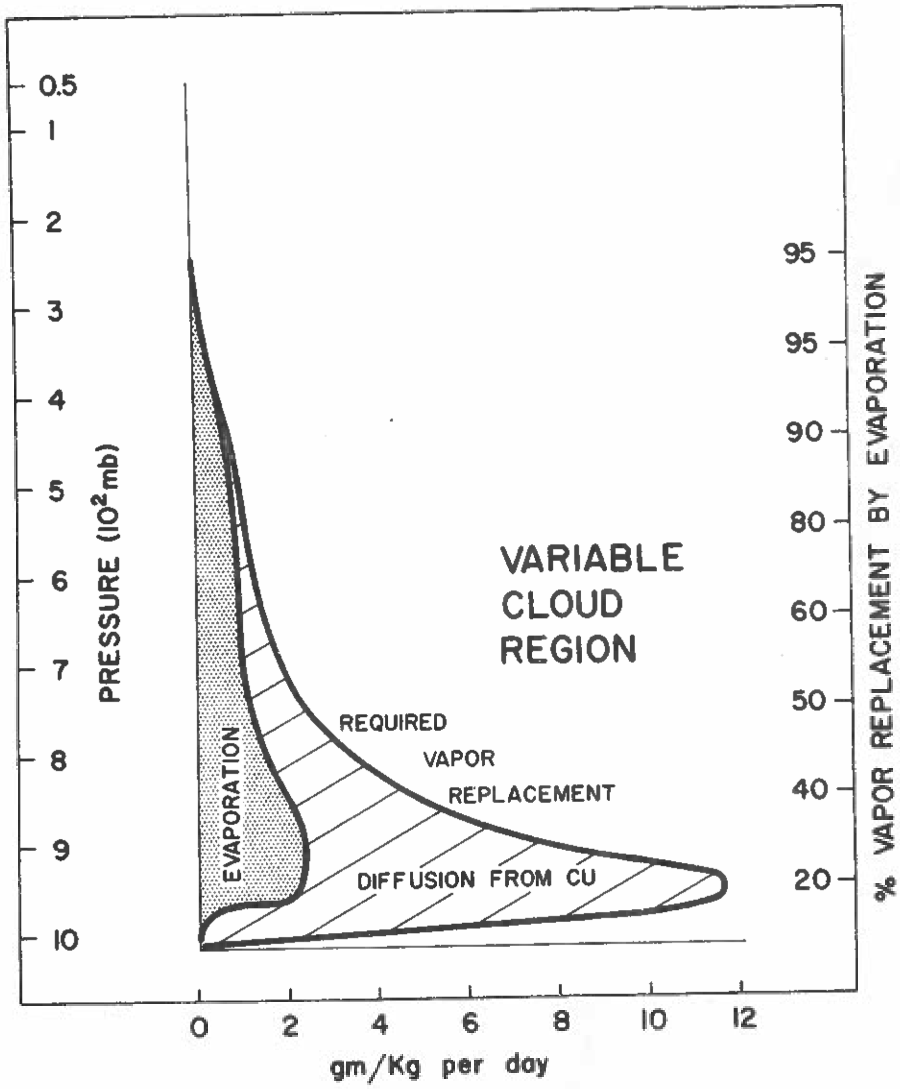


Fig. 32. Cluster required water vapor replacement to sinking motion to maintain variable cloud region conditions and the portion of vapor replacement coming from evaporation (on the right).

from evaporation. This is a result of higher liquid water detrainment from the cumulus at upper levels and the lower upper level water vapor contents. The percentage of water vapor replacement by evaporation to total required water vapor replacement is shown on the right side of these figures. It ranges from 20-25 percent at 900 mb to 80-90 percent at 500 mb.

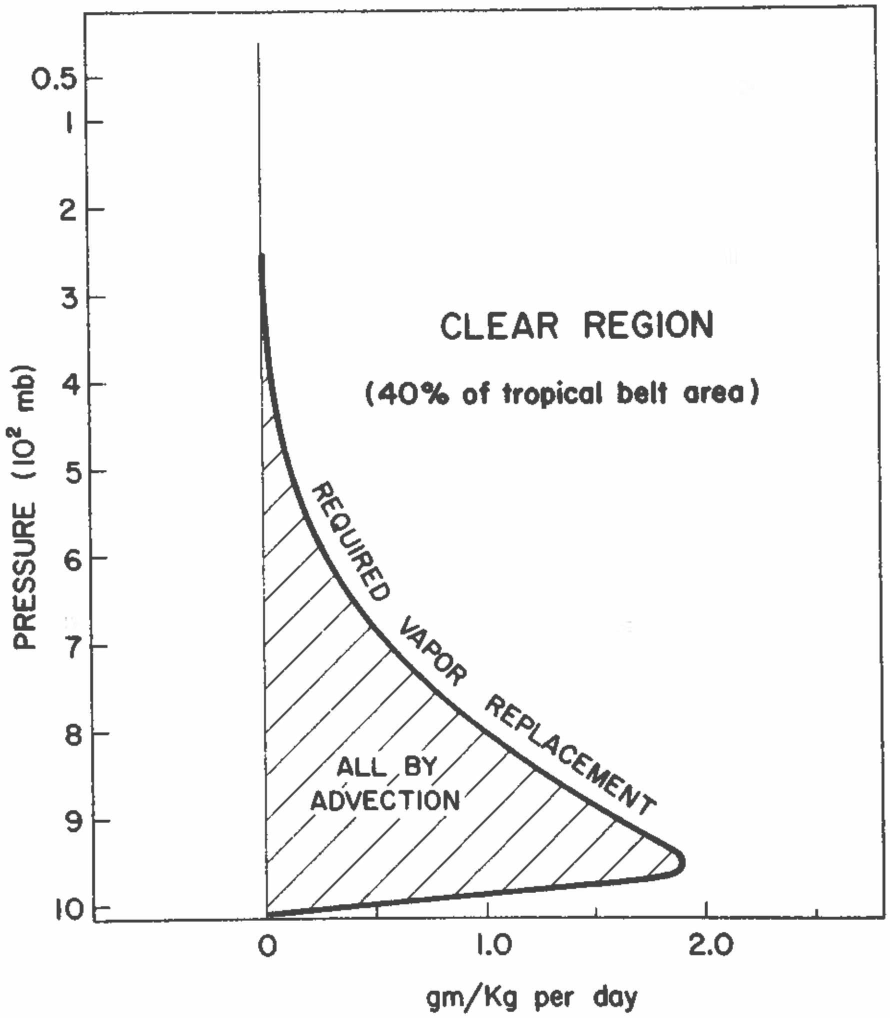


Fig. 33. Required advection of water vapor into the clear region to balance its sinking drying and maintain its steady state.

The large magnitude of this required vertical circulation and its consequent very large energy and vapor sources and sinks should be appreciated by those who wish to understand tropical circulations.

VI. ACTUAL OR PROBABLE VS. REQUIRED VERTICAL CIRCULATIONS

The previous sections have discussed the required "mean" and "local" vertical circulation necessary to satisfy simultaneously mass, vapor, and energy needs. These mean and local required circulations must continually be in operation. But this does not specify all the vertical motion going on. Besides these required circulations there is an extra or "probable" additional local circulation which does not influence the mass, vapor, or energy budgets and can only be roughly estimated. This extra mass balancing local circulation is the additional up-dry and compensating down-dry circulation which accompanies broadscale forced upward motion (often associated with layered clouds) and the up-moist and compensating down-moist downdraft circulations associated with heavy rainfall. Any additional forced up-dry circulation must be balanced by a compensating down-dry motion. Similarly, any additional moist downdraft motion must be balanced by a compensating moist upmotion. These additional mass balancing circulations, whose magnitude can only be roughly estimated, must be added to the already determined required up-moist and down-dry circulations to arrive at the actual or here defined probable real vertical circulation which is occurring. Fig. 34 portrays the author's estimate of the actual vertical circulation occurring in the cluster regions previously discussed. Seven classes of vertical circulations are shown, four

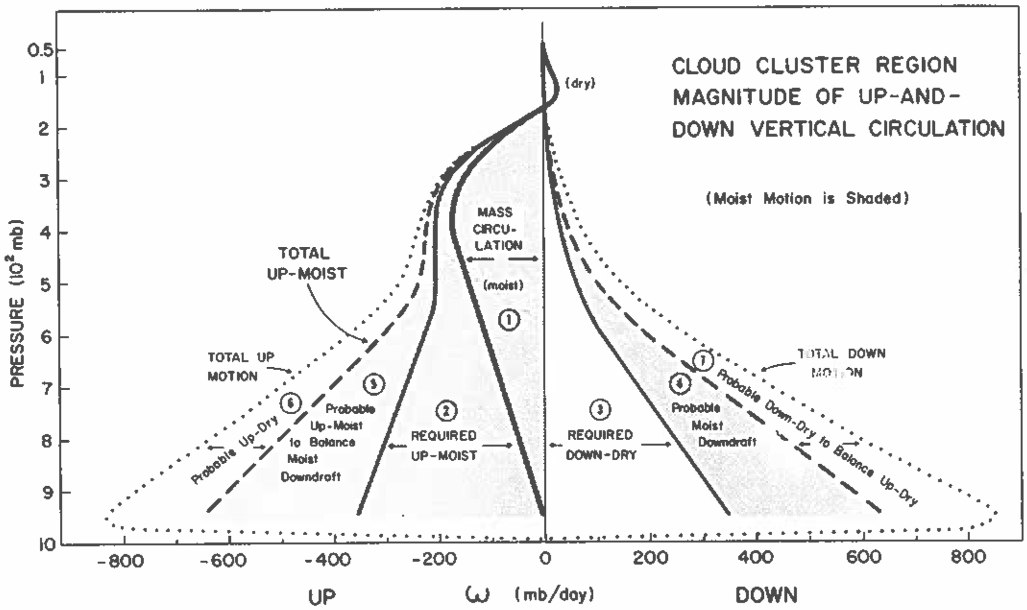


Fig. 34. Estimated complete vertical motion pattern of the cluster region. Numbered individual vertical motion components are discussed below.

moist and three dry. The moist or saturated motion has been shaded.

These seven vertical circulations are:

1) Mean upward motion as determined from the radiosonde cluster data (see Fig. 21). This has a maximum value of 180 mb/day at 350 mb. It has been assumed that the mean circulation goes up-moist.

2) Required local up-moist circulation with maximum value of about 350 mb/day at 950 mb.

3) Required local down-dry circulation with maximum value of about 350 mb/day at 950 mb. Circulations 2) and 3) must mutually balance each other.

4) Probable extra down-moist circulation as results with saturated downdrafts with rain. Maximum values assumed in lower troposphere of about 200 mb/day. This estimate is quite subjective.

5) Probable extra moist updraft to balance the moist downdraft. Motions 4) and 5) are mutually balancing.

6) Probable extra up-dry motion from meso or synoptic scale forced vertical motion as must be present to produce layered clouds.

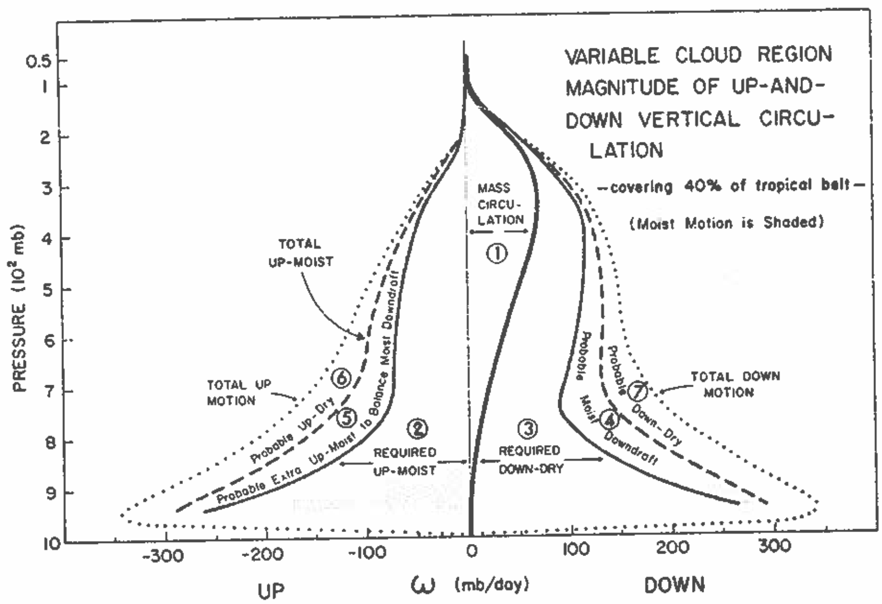


Fig. 35. Estimated complete vertical motion pattern of the variable cloud region. Numbered individual vertical motion components are as discussed for the cluster region.

7) Probable extra additional down-dry motion to balance the forced up-dry motion of 6). Motions 6) and 7) are mutually balancing.

If this estimate of the typical cluster vertical circulation is not too inaccurate, then we are forced to accept the reality of a very large recycling vertical circulation which, in the lower half of the troposphere, is from one to two orders of magnitude larger than the mean cluster upward motion. The implication of this very large local recycling circulation for vertical momentum and other dynamic influences may be substantial and requires careful consideration. The divergence pattern implied by the vertical motion of Fig. 34 exactly fits the mean divergence pattern of Fig. 20. Surface to 400 mb cluster convergence is due to the increase in downward motion being larger than the increase of upward motion.

Fig. 35 shows the author's estimate of the actual or probable vertical circulations occurring in the variable cloud region. Again, it is

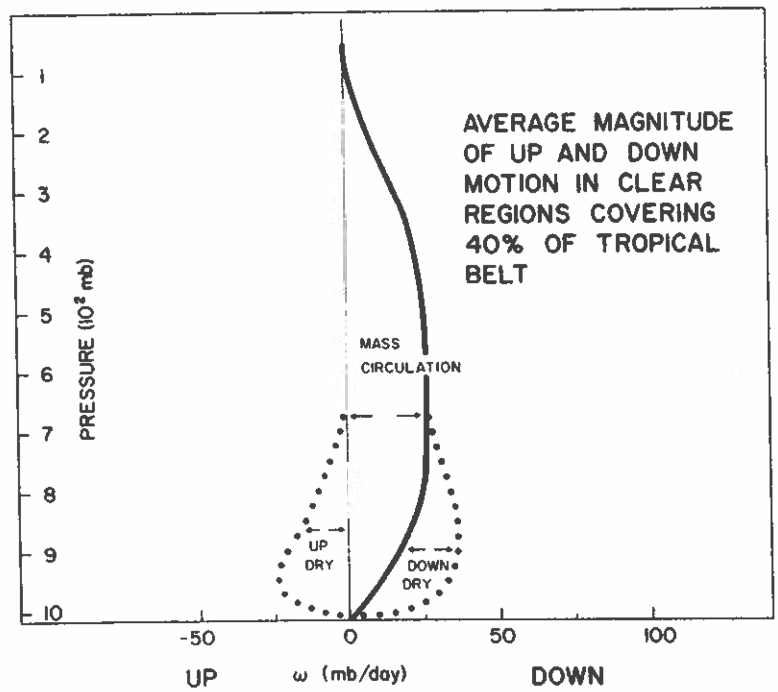


Fig. 36. Assumed total vertical motion in the clear regions.

to be noted how large the actual up-and-down local vertical circulations are in comparison with the mean vertical motion.

In the clear regions only a small extra or "probable" up-dry and down-dry circulation must be added to the mean circulation as shown in Fig. 36. This extra up- and-down dry circulation is thought to be due to boundary layer turbulent mixing.

Fig. 37 shows the author's estimate of the vertical profile of the average of the absolute magnitude of the mean, the required, and the probable vertical motion occurring for the entire oceanic tropical belt of summer. These local vertical circulations are very large even though we are averaging over the whole tropical region.

It was not expected that such a large required up-and-down compensating vertical circulation would be present at lower levels. Reed and

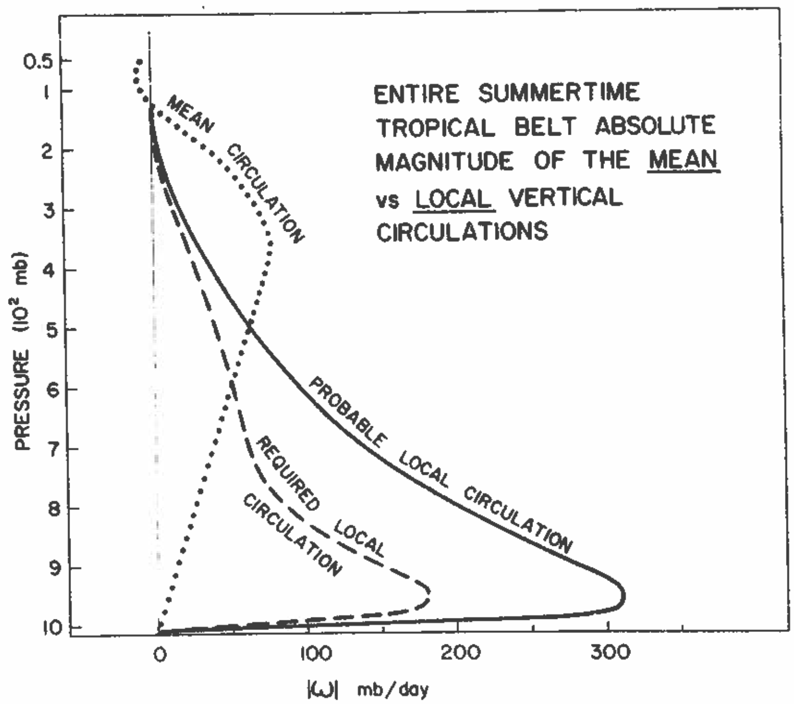


Fig. 37. Comparison of the magnitude of the mean and the local vertical circulations for the entire tropical belt of summer.

Recker (1971) have also deduced from broad-scale considerations that a large lower-tropospheric recycling circulation must be present, but their estimates are not as large. The implication of this large vertical circulation for the dynamics of the lower troposphere must be very great. The vertical momentum transfer by this extra circulation is probably very significant to inhibit establishment of vertical wind shears. It is observed that the vertical wind shear from cumulus cloud base to middle tropospheric levels in cumulus convective situations is typically very small. Table 3 shows the very high magnitudes of the "local" up and down circulation to the "mean" circulation at cloud base. An explanation of the lack of vertical vorticity balance and the need to hypothesize substantial sub-synoptic or cumulus vertical vorticity transfer, as has been discussed by Williams (1970), Reed and Recker (1971), Gray (1972b) and Holton and Colton (1972), is likely to be associated with this strong vertical circulation requirement.

This magnitude of cluster vertical circulation fits the estimates of López (1972b) which were derived independently from a cumulus modeling approach which will be discussed in Chapter VIII.

Table 3

The top of the boundary layer "mean" vs. "local" required and probable up and down vertical circulation. All values are in mb/day.

	Cluster Region	Variable Cloud Region	Clear Area	Whole Tropical Belt
A. Absolute value of mean circulation	20	0	10	10
B. Absolute value of required up or required down "local" circulation	360	265	0	200
C. Absolute value of probable up or probable down "local" circulation	820	340	25	300
Ratio of A to B	1/18	0	-	1/20
Ratio of A to C	1/40	0	2/5	1/30

VII. PROBABLE TROPICAL REGION VAPOR AND ENERGY BUDGETS

The extra vertical circulation beyond that required to meet mass-vapor-energy budgets results in additional warming-cooling and condensation-evaporation rates. These must be added to the required values. They are believed to be close to the actual (or probable) rates of change really occurring. Figs. 38 and 39 show the author's estimates of the probable energy and vapor balances occurring within the cluster. In comparison with the radiation cooling and rainfall rates, these mutually balancing cluster energy and vapor exchange rates are indeed very large.

Even for the entire tropical belt these <u>probable</u> energy and vapor exchanges are very large in comparison with the radiation cooling and rainfall rates. They are shown in Figs. 40 and 41.

Figs. 42-43 portray tropospheric averages of the mean, required, and probable vertical circulation and the same values for the sinking warming--evaporation cooling rates for each of the three regions and for the entire tropical belt. Fig. 44 shows the large magnitude of condensation-evaporation to rainfall. For the whole tropical belt probable condensation-evaporation is about 6 times larger than rainfall; <u>required</u> condensation-evaporation about 3-1/2 times rainfall.

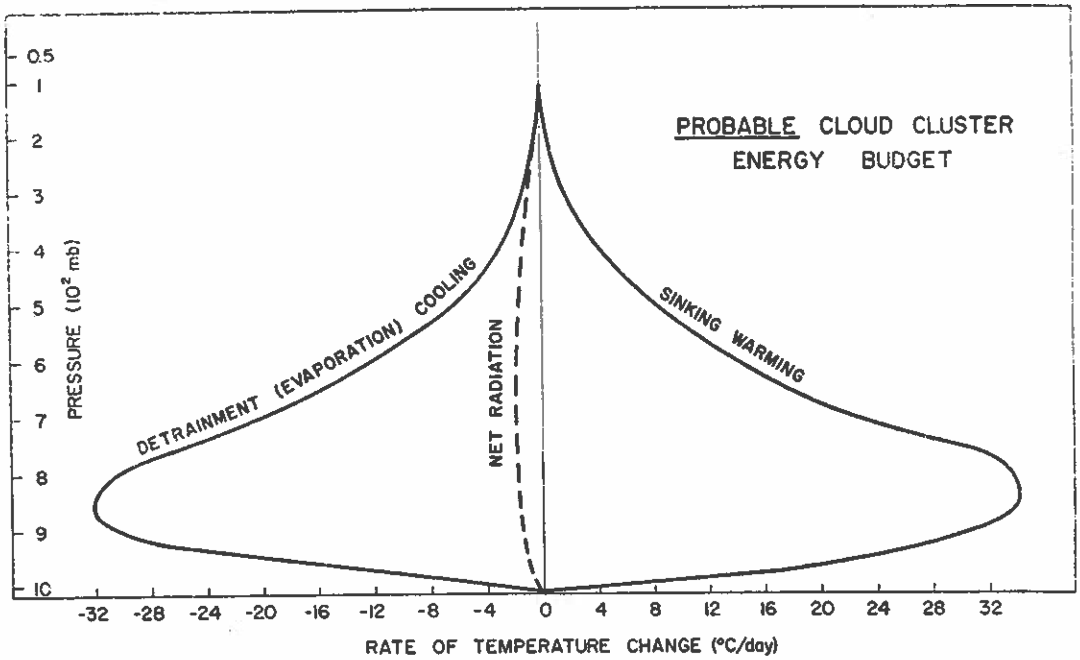

PROBABLE CLOUD CLUSTER
ENERGY BUDGET
PRESSURE (10^2 mb)
0.5
1
2
3
4
5
6
7
8
9
10
DETRAINMENT (EVAPORATION) COOLING
NET RADIATION
SINKING WARMING
-32 -28 -24 -20 -16 -12 -8 -4 0 4 8 12 16 20 24 28 32
RATE OF TEMPERATURE CHANGE (°C/day)

Fig. 38.

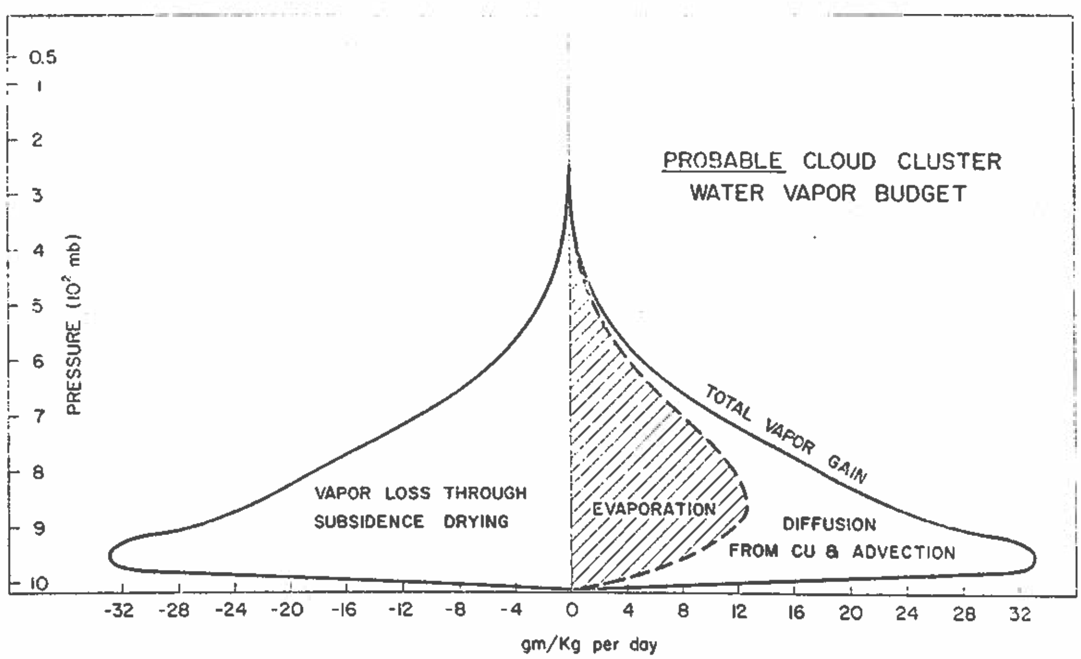

PROBABLE CLOUD CLUSTER
WATER VAPOR BUDGET
PRESSURE (10^2 mb)
0.5
1
2
3
4
5
6
7
8
9
10
VAPOR LOSS THROUGH
SUBSIDENCE DRYING
EVAPORATION
TOTAL VAPOR GAIN
DIFFUSION
FROM CU & ADVECTION
-32 -28 -24 -20 -16 -12 -8 -4 0 4 8 12 16 20 24 28 32
gm/Kg per day

Fig. 39.

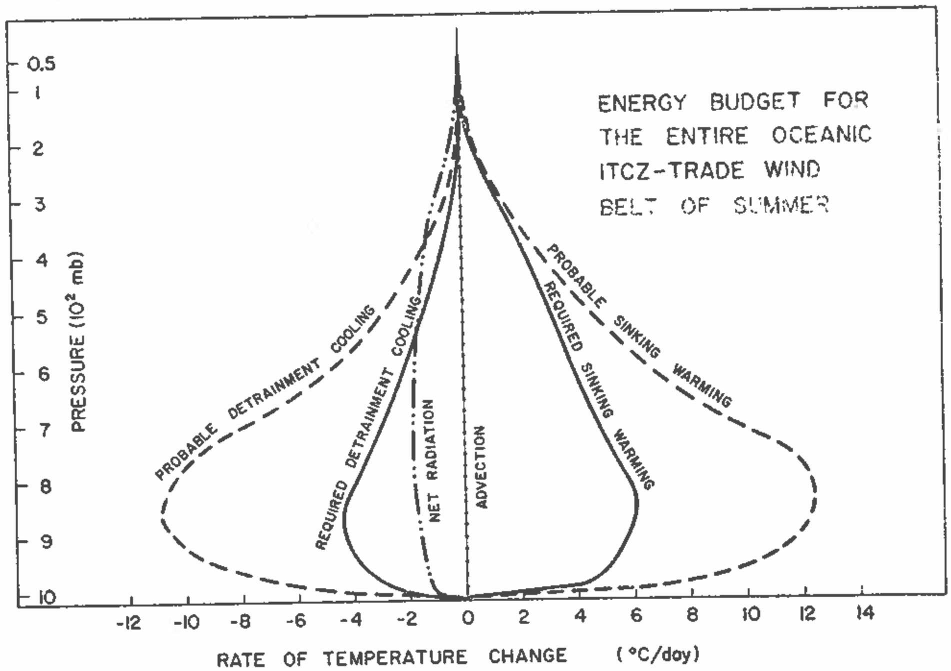


Fig. 40.

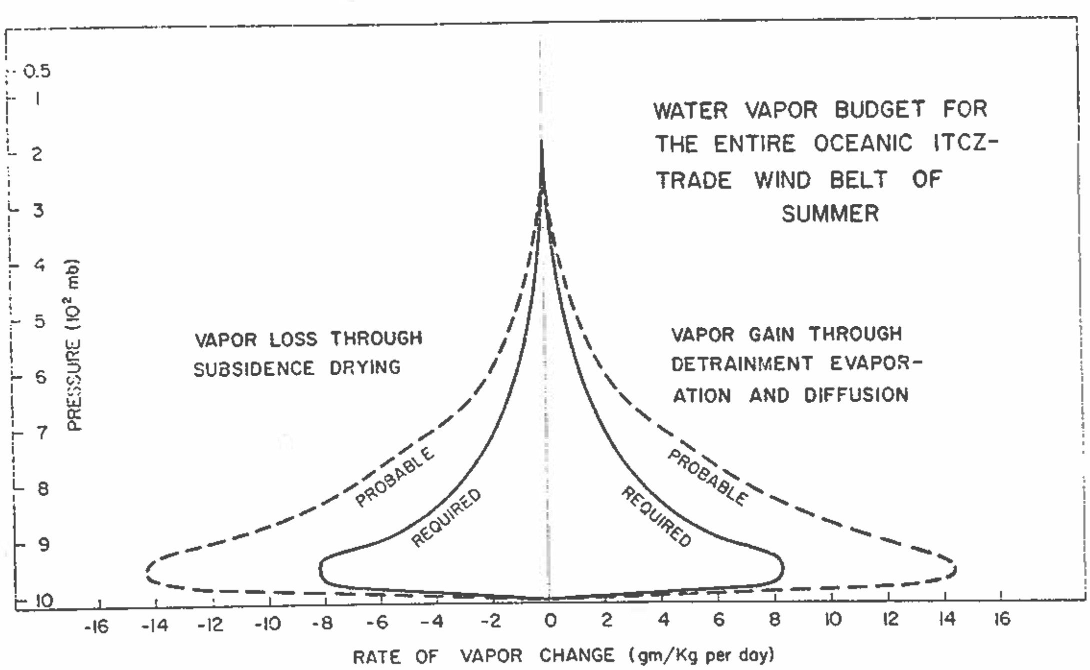


Fig. 41. Comparison of required and probable water vapor budgets for the entire tropical belt of summer.

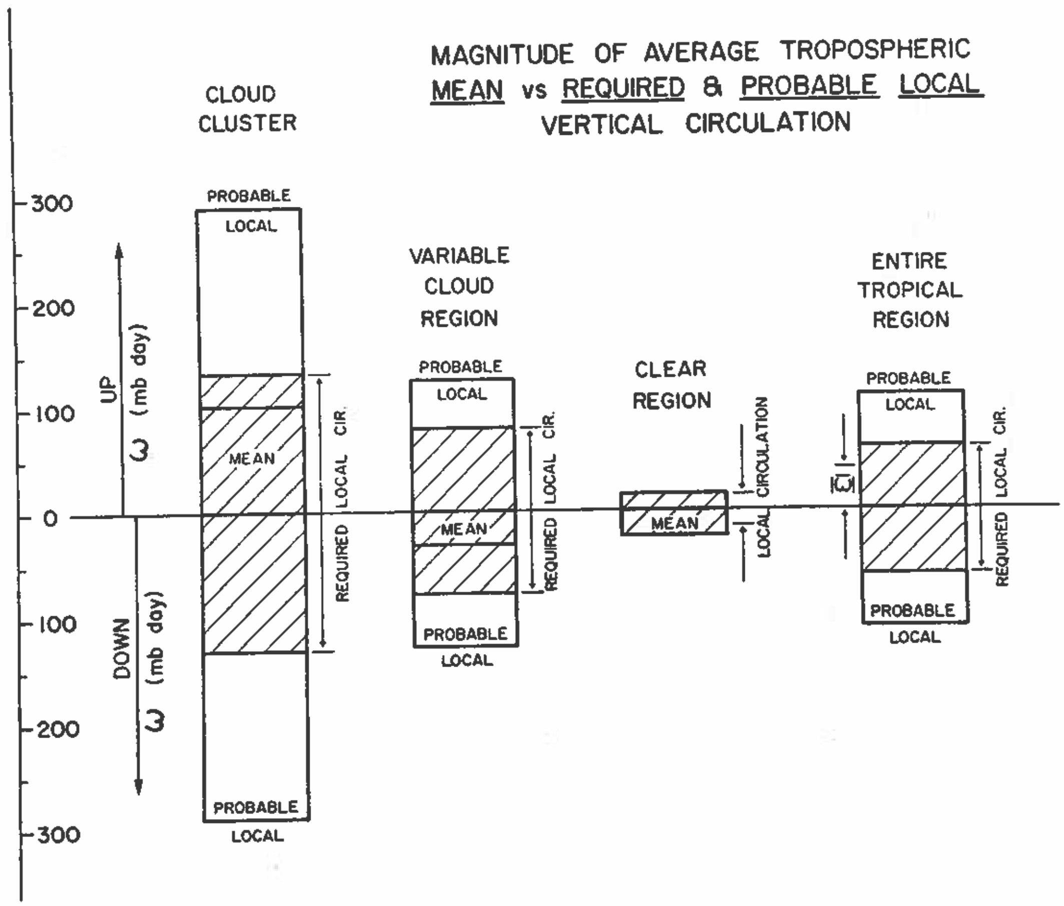


Fig. 42. Comparison of individual region mean vs. local circulations.

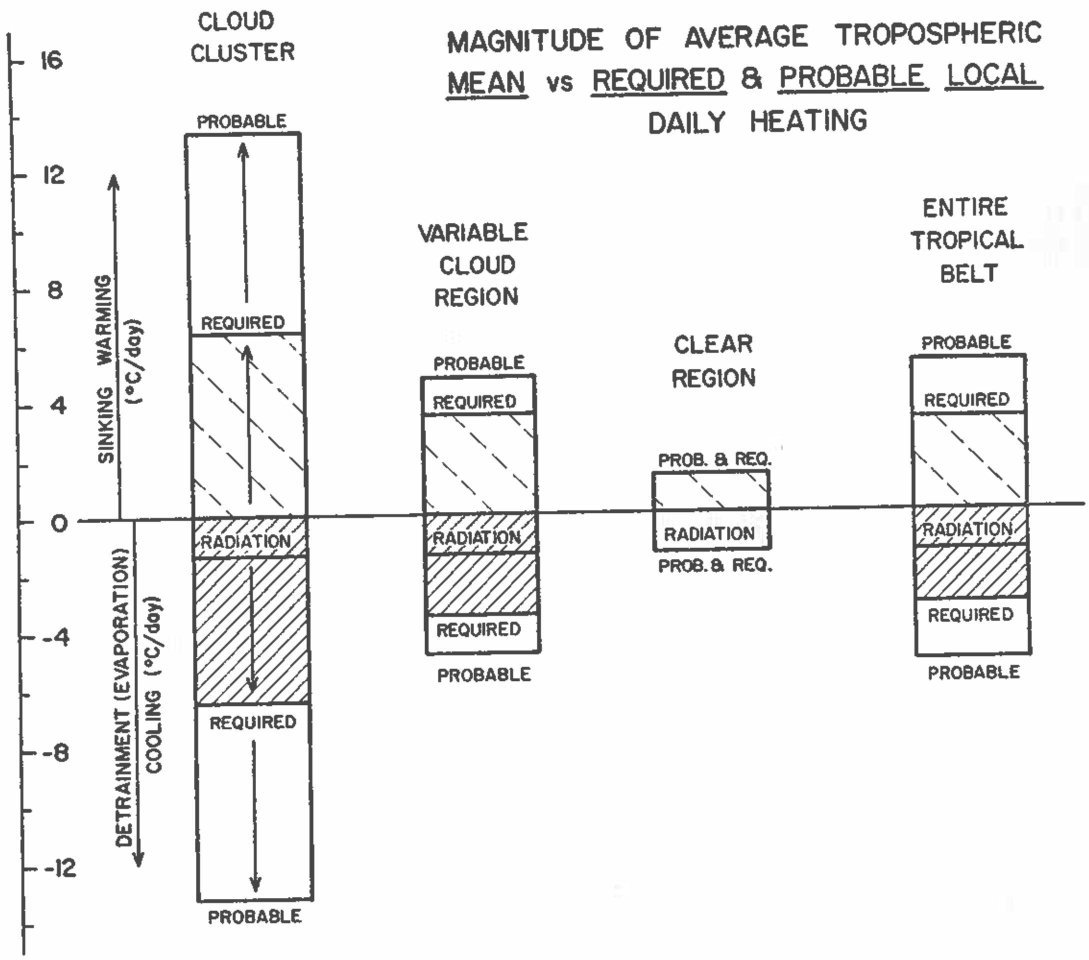


Fig. 43. Comparison of individual region sinking warming and evaporation cooling with the net radiation cooling.

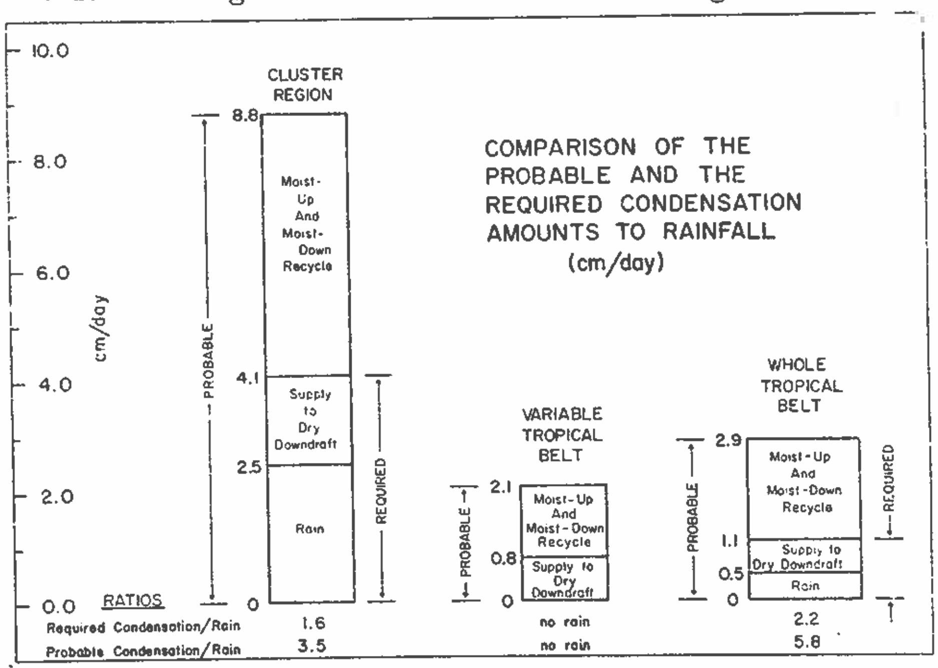


Fig. 44. Comparison of individual region condensation-evaporation vs. rainfall.

VIII. COMPARISON OF CLUSTER VERTICAL CIRCULATIONS OBTAINED FROM THE BROAD-SCALE MODEL OF THIS PAPER WITH THOSE OF CUMULUS SCALE MODEL OF LÓPEZ (Paper II)

López (1972b-Paper II) has discussed the cluster vertical circulation as determined from a life cycle cumulus model in conjunction with radar and composite radiosonde cluster data. This cluster model, derived independent of the approach of this paper from cumulus and cluster scale consideration alone, will now be compared with the cluster model of this paper, derived from tropical belt and cluster-scale consideration without any cumulus-scale knowledge. Fig. 45 briefly outlines in flow diagram form how the two models were developed and how they were meshed for comparison.

Fig. 46-48 show the comparison of the two models with regard to their local vertical circulations, their sinking warming and evaporation cooling rates, and their water vapor loss-gains. One can clearly see how close each of these models were in specifying the required mass, energy, and water vapor budgets. This lended consistency and more credibility to the results of each approach. To ultimately verify cumulus-broaderscale interaction models, this inward meshing of the dynamics from opposite scales of consideration should always be a desired goal.

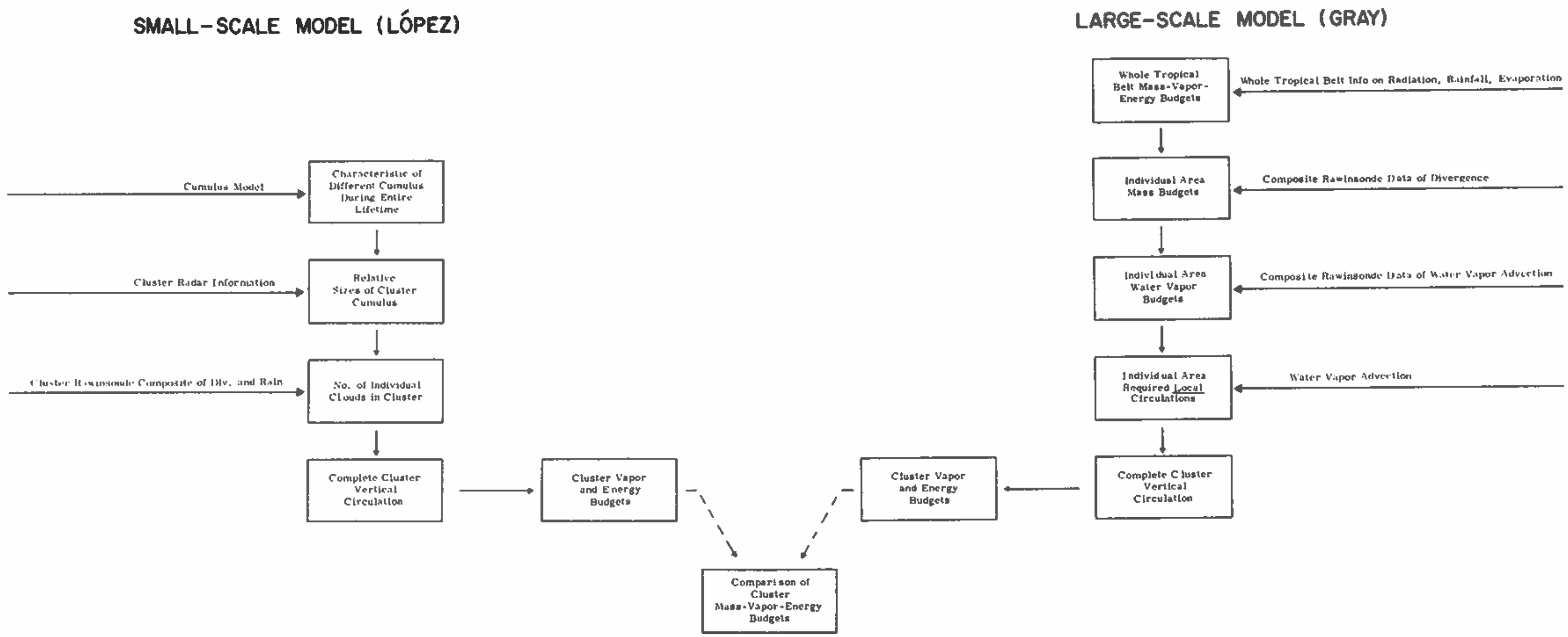


Fig. 45. Flow diagram of how the cloud cluster dynamics are independently determined and compared from the broadscale model of this manuscript (Paper III) and from the small scale model of López (1972b - Paper II).

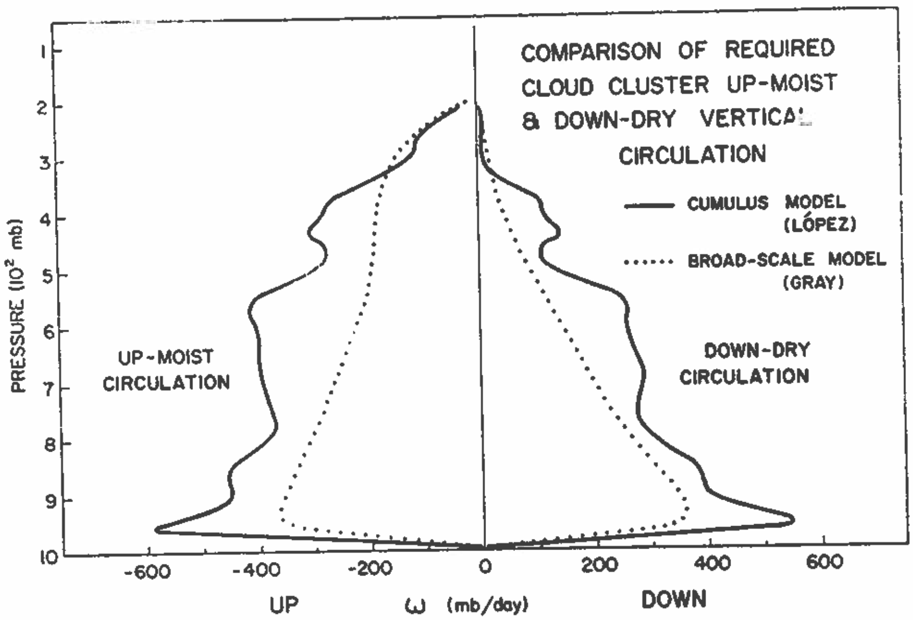


Fig. 46. Comparison of the cloud cluster vertical circulation as determined by the broad-scale model of this report and by the cumulus model of López (1972b).

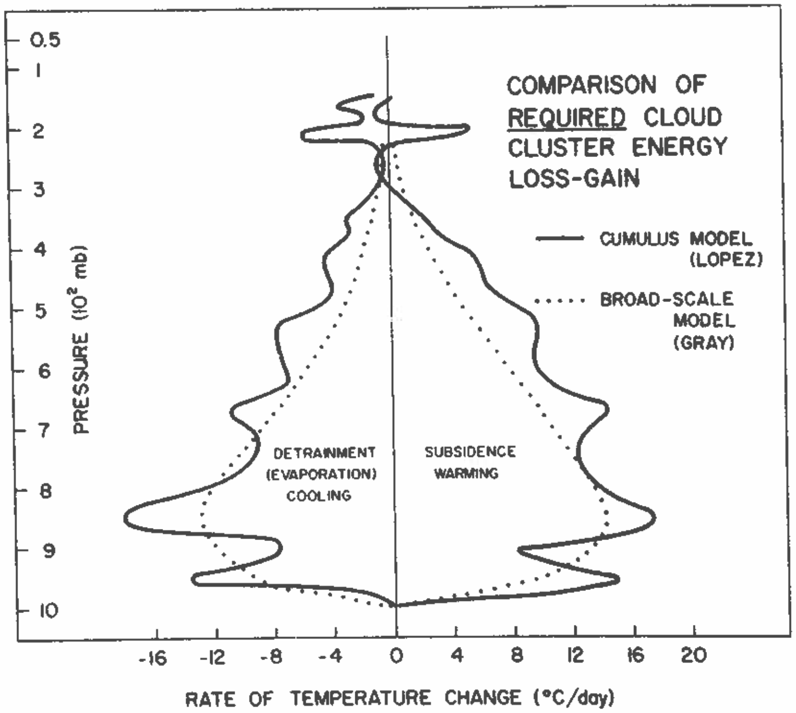


Fig. 47. Comparison of each model's subsidence warming and evaporation cooling.

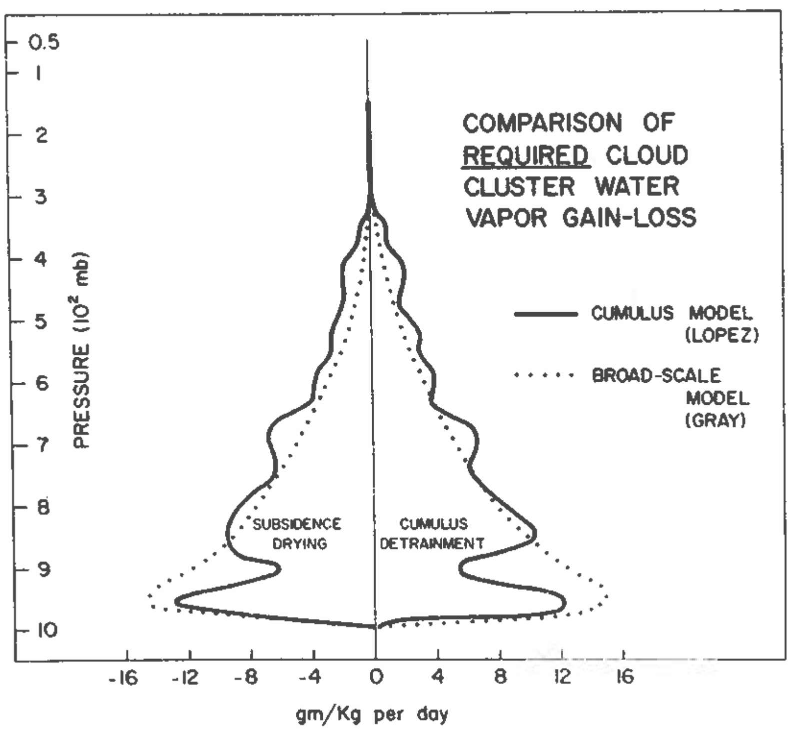


Fig. 48. Comparison of each model's water vapor budget.

IX. PARAMETERIZATION OF MOIST PROCESSES

This paper has demonstrated the very large vertical circulation of the tropical atmosphere. It has shown that the summertime tropical atmosphere meets its mass-vapor-energy budgets in a rather complex way. The authors (Gray and López) feel that it is very difficult (or impossible) at the present time for any numerical model of tropical motion to adequately come to grips with the real problems of incorporating the cumulus convection in terms of the broader-scale flow. This will probably not be fully possible until we have more empirical information and general scientific consensus and understanding of

1) the individual cumulus dynamics with their characteristic mass, momentum, vapor, and sensible temperature rearrangements. This requires a good whole-life individual cumulus model.

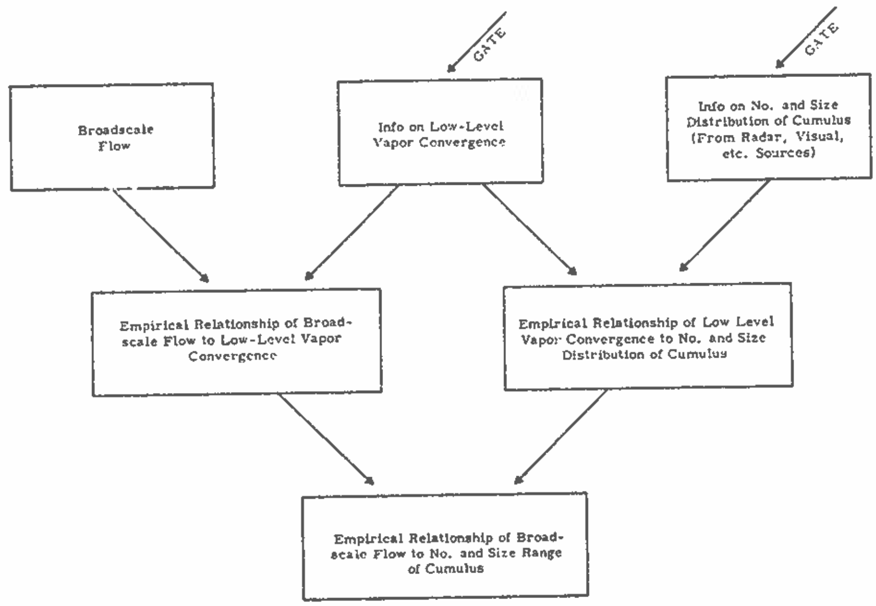
2) the large magnitude of the local vertical circulation here described

3) the association of cumulus number and relative size distribution with various broad-scale features such as

   a) low-level mass and moisture convergence
   b) other broader-scale flow features such as the low-level relative vorticity and the tropospheric vertical wind shears, etc.

This knowledge will only come from an increase of our observational sampling of the tropical atmosphere and the cloud cluster. A judicious sampling and study of our already collected data (radiosonde and satellite) in the Western Pacific and in the West Indies will help to answer many of these questions. The coming GATE experiment and other tropical land experiments will hopefully help to fill many of the empirical gaps in our observation knowledge.

REQUIRED BACKGROUND RESEARCH
RELATION OF BROADSCALE FLOW
TO NUMBER AND SIZE OF CUMULUS



MODEL B

Fig. 49. Outline of proposed scheme for relating the broadscale flow to the number and size distributed of the cumulus.

The authors feel that the cumulus parameterization scheme which will finally prove workable will be one which handles the cumulus processes separately from the main working model.

We must first establish empirical relationships giving the association of the broadscale flow to the number and size of the cumulus (as in Fig. 49--Model B) and also empirical information on the influence of the cumulus on the surrounding circulation (as in Fig. 50--Model C). Only when this type of empirical evidence is obtained can we proceed confidently to an actual parameterization scheme as is speculatively proposed in Fig. 51. Here, the cumulus induced changes as determined by

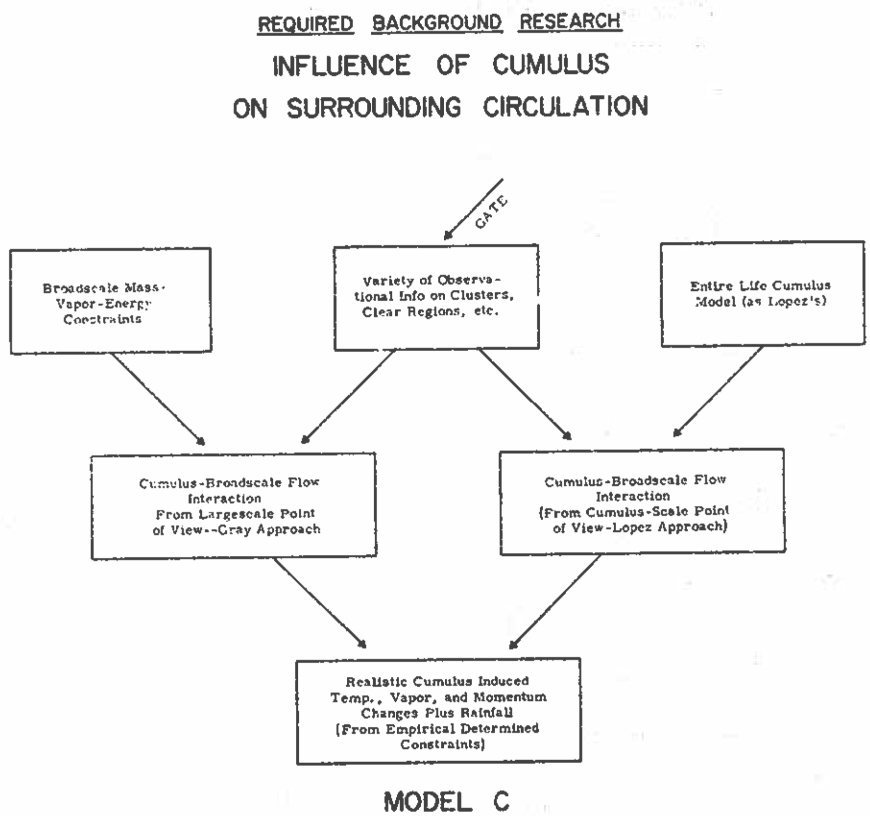


Fig. 50. Outline of proposed scheme for determining the influence of the cumulus on the surrounding circulation.

Models B and C are fed back to the basic working model (A) and new broadscale flow patterns specified at each time step.

In this figure the basic working model (A) uses already determined cumulus-broadscale relationships as given by Models B and C to specify its next timesteps. Empirical relationships of Models B and C are determined before Model A is run. Models B and C calibrate the cumulus with the broadscale flow. They must be based on empirical knowledge. They have yet to be realistically specified. Field programs and other research with observations should help to better specify the inputs of these auxillary models.

PROPOSED CUMULUS PARAMETERIZATION SCHEME

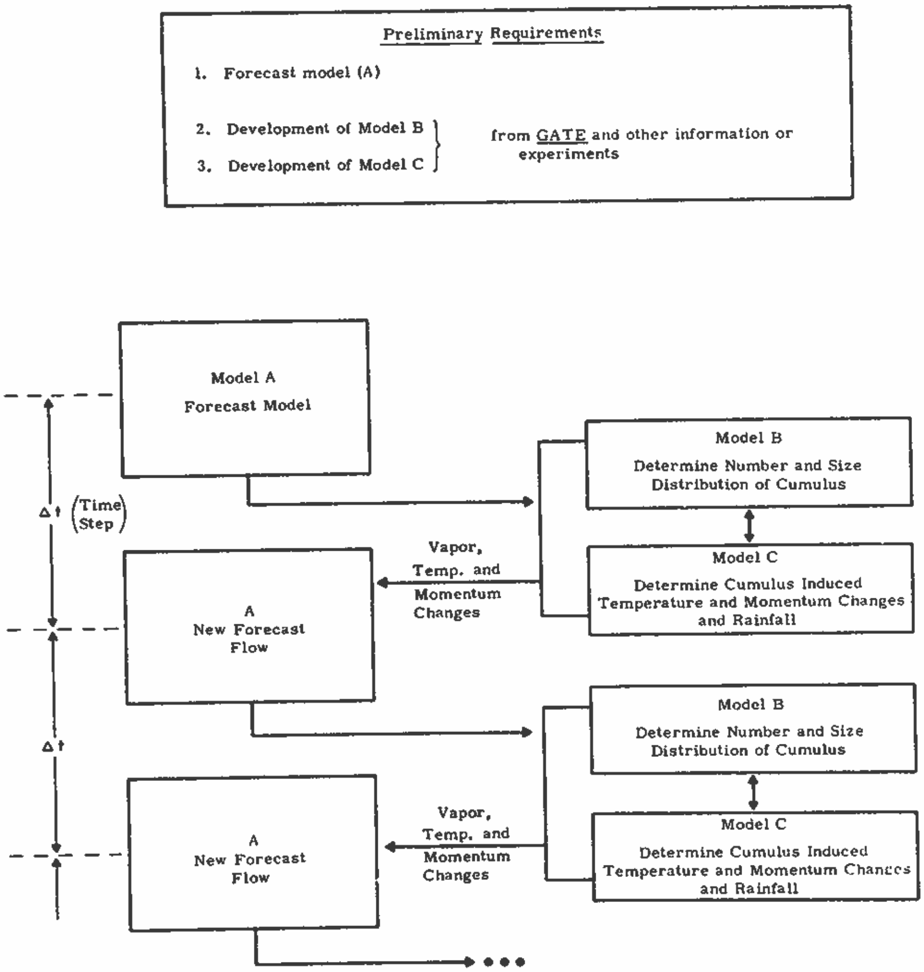


Fig. 51.

Separate Treatment of Cumulus Cloud. In that the cumulus cloud is such a distinctive physical unit, it would appear that it should be independently treated probably in a fashion similar to that discussed by López (Paper II). The physics of the cumulus-broadscale interaction may not necessarily be overcome by applying the primitive equations to even smaller grids and time steps without additional insight into the character of the individual convective elements. At the same time cumulus parameterization schemes based on abbreviated or oversimplified cloud model schemes may prove to be inadequate at really coming to grips with all of the real life cycle cumulus influences. The author feels that the minimum degree of cumulus model sophistication that will prove workable is probably that of whole-life one-dimensional model of López (1972a). In any rate, we will probably not be able to settle this question until we have a generally accepted sophisticated whole life cumulus model to which the simplified cumulus models can be compared.

X. DISCUSSION

Large Vertical Recycling. The large vertical recycling of air that has been discussed is a function of two primary tropospheric processes

1) the direct radiational cooling of the troposphere, and
2) the direct cooling nature of the cumulus clouds as they evaporate and die.

These two processes combine to produce substantial cooling of the atmosphere, especially in the lower troposphere. The net radiational cooling of the troposphere of about $1^{\circ}C$/day is well known and accepted. The idea of the cumulus cloud as "a direct cooling agent — an indirect warming agent" has yet to be fully appreciated or accepted. To understand how the troposphere balances its radiation loss it is of fundamental importance to understand how condensation warms the atmosphere. Cumulus clouds, whether causing precipitation or not, do not directly act to warm the troposphere. In fact, cumulus clouds directly cool the environment in which they exist. By what mechanism then do the cumulus clouds act to balance the tropospheric radiation loss and at the same time directly cool?

As paradoxically as this may initially appear, this is indeed what goes on in the troposphere. The latent heat released from cumulus (both precipitating and nonprecipitating) goes primarily into potential energy gain and increasing the temperature of the rising parcel to that of the environmental temperature. The small extra (above environment) temperature increases of 1° to 2° of the rising parcel (which is required for buoyancy) does not warm the environment unless it directly mixes out from the cloud at a higher temperature. The rising parcel typically continues rising until it loses its buoyancy and temperature excess. It then mixes to the environment at a temperature little different (or even cooler) than that of the environment. This does not warm the environment. Any diffusional or advective heat transports out from the rising (and warmer) cloud parcel are more than overcome by the evaporation of the residual cloud liquid water particles around the cumulus or when they die. Individual small cumulus last only 5-10 minutes; Cb's only about 25-30 minutes. The residual liquid particles which remain after the vertical motion in the cumulus has stopped cool the environment in and around the dying cloud at a rate of 2.4°C for every gm/kg evaporated. Being an open system, the cumulus converts all its condensation heat to potential energy and exports this to the surrounding environment. Even though rainfall may have occurred, there is typically no warming, but a local cooling of the environment. This is not to say that the total effect of the condensation to the "closed system" of the globe or hemisphere has not been one of warming. It has. The dry adiabatic sinking at some other location (to satisfy mass balance) has more than compensated for the local cooling if there has been rainfall. In this way the cumulus act in a delayed action sense. They produce a local sensible cooling, but a global averaged warming can result.

There is supporting data for this physical explanation:

a) direct sensible temperature decreases around and after cumulus convection has occurred, as reported by Kininmonth (1971) from data of the VEMHEX project of 1969.

b) the lower tropospheric cool nature of the raining tropical cluster and easterly wave, as substantiated by Riehl (1945) and many others since. Why does this condensation not warm the environment?

c) the lack of any tropospheric warming (or surface pressure falls) in an observational study of one thousand Pacific trade-wind cloud clusters by Williams (1970). The average observed and calculated rainfall for these clusters was 2-1/2 cm per day for a 4^{o} latitude square area. This 1500 calories/cm^{2} per day condensation energy release led to no tropospheric warming even though vertical wind shears and horizontal sensible temperature advective influences were small. Where did the heating go? Again, primarily to potential energy to be export to the surrounding environment.

d) the direct cumulus cooling found by López (1972a) in his whole life cycle cloud model.

e) the maximum warming of the hurricane centers. Here the highest temperatures are found in the center subsidence region. The maximum upward motion in the eye wall convection area has lower temperatures. A diffusion of sensible temperature into the eye-wall cloud is necessary to maintain its temperature (Shea, 1972).

We must thus view the cumulus precipitation warming process as one in which the cumulus initially cools the local environment but at the same time acts to warm the outer environment by compensating dry subsidence. The subsidence warming is larger than the local evaporation cooling and leads to a net large scale tropospheric sensible temperature warming. The warming occurs at the place where the sinking motion exists. Local warming can, at times, occur if the compensating sinking motion takes place near the rising motion as is the case with tropical storm genesis in regions of small vertical wind shear (Gray, 1968).

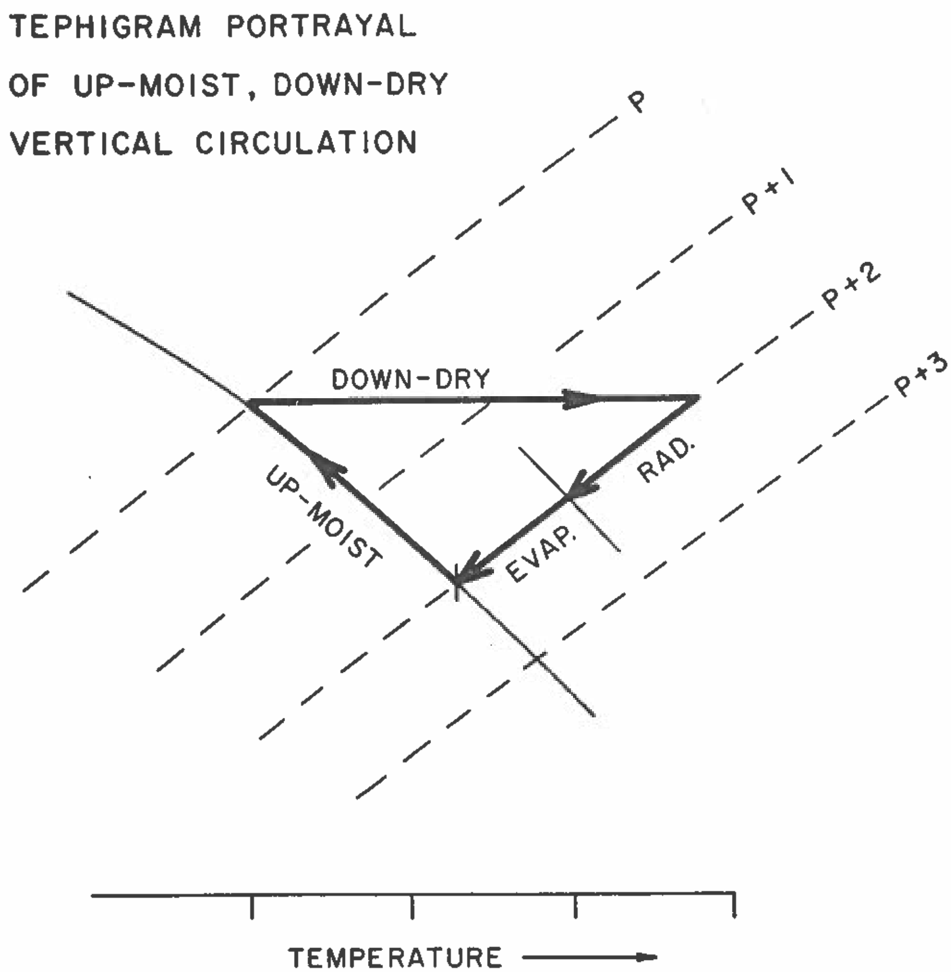


Fig. 52. Graphical portrayal of how an up-moist and down-dry vertical circulation can balance the radiation and evaporation energy losses.

The primary energy input to the troposphere can thus only come from a continuous up-moist and down-dry vertical circulation which acts to balance both the radiation and evaporation cooling as shown in Fig. 52.

This energy gain is paid for by a continuous water vapor loss. The clear area down-dry portion of this vertical circulation is continuously losing vapor which, for steady state, must be replaced by the cumulus. This is accomplished by vapor diffusion and by evaporation as shown in

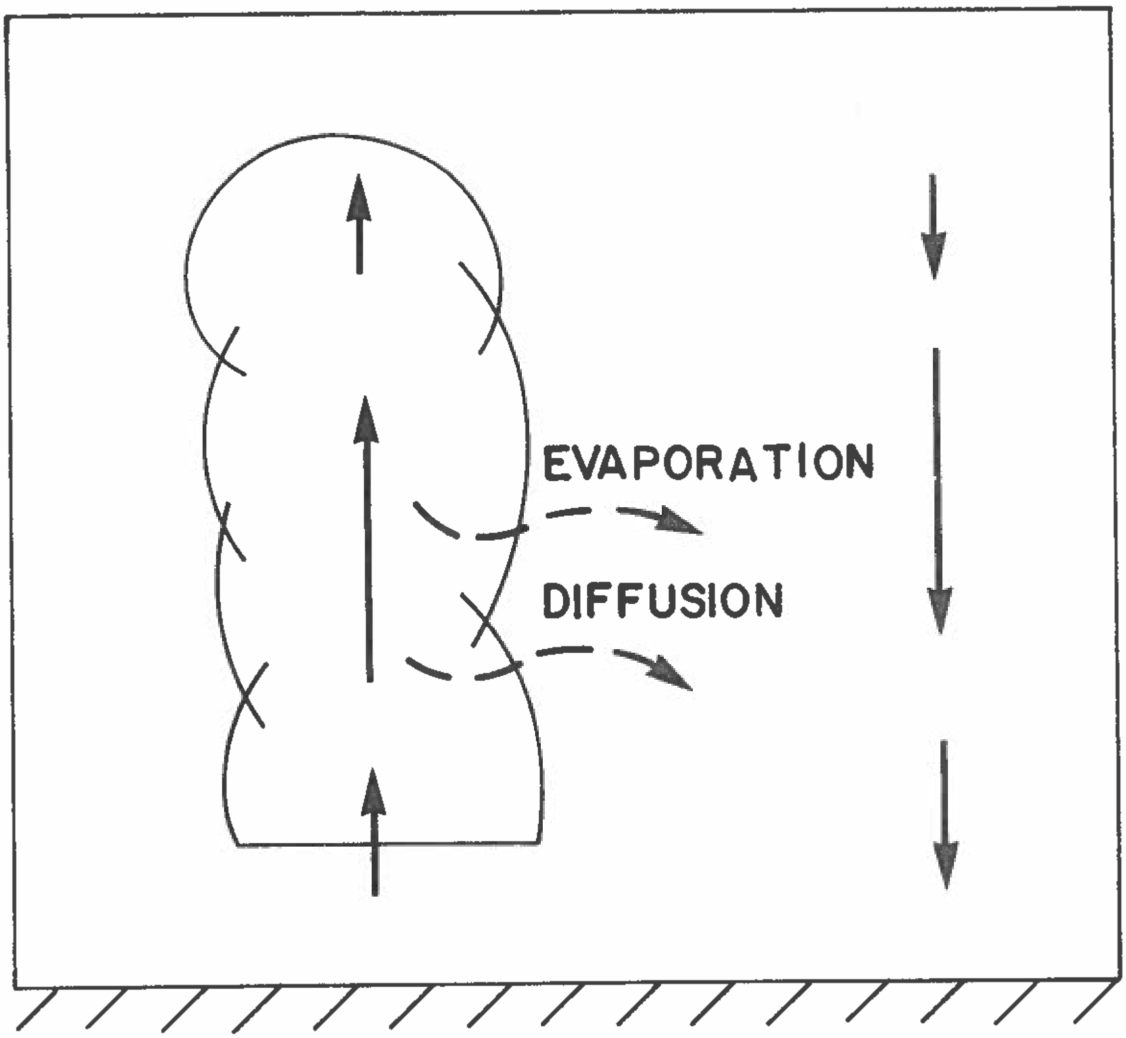


Fig. 53.

Fig. 53. This figure shows how the cumulus acts to increase the water vapor in the sinking air around it by both diffusion of vapor from the cloud and by liquid water evaporation on the edge of the cloud as the cloud dies.

There are now a number of meteorologists which are coming to think of the warming mechanism of the cumulus as resulting from compensating subsidence of the environment. The author has found general physical agreement on this idea in personal discussion and from the papers R. López (op. cit.), Pearce and Riehl (1969), K. Ooyama (1971, 1972), M. Yanai (1971, 1972), and J. Charney (1968). In addition, some of these researchers, like López (op. cit.), Kinnemonth (1971), and Betts (1971, 1972, and personal communication), are now also accepting the idea of the cumulus cloud acting as a direct cooling agent. The earlier findings of Riehl (1945, 1965) and Elsberry (1966) on the lower tropospheric cooling of easterly waves might also be interpreted as generally supporting this point of view. Caution in literal interpretation of the "hot tower" hypothesis of Riehl and Malkus (1958, 1961) needs to be exercised. Their demonstration of the dominant role of vertical transports of mass occurring in selective cumulus or cumulonimbus towers is indeed correct, but these Cb towers are (from the immediate environmental warming point of view) "cool" and not "hot". They act to directly cool and not warm the environment. A "cooling tower" hypothesis does not, however, imply that the individual cumulus towers play any less fundamental role than that envisaged by Riehl and Malkus. The main problem of the cumulus acting as a direct heating rather than as a cooling source is that the required vertical circulation is reduced to unrealistic low values.

At a recent NCAR workshop on cumulus parameterization at Boulder, Colorado*, a general concensus on the large required recycling of the cloud cluster atmosphere was arrived at by the author, López (1972a,

*July 10-14, 1972

1972b), Yanai (1972), Ooyama (1972), Rodenhaus (1972), and Zipser (1972).

Rate of Recycling in Generation of Individual Cumulus Elements. The general stability of the cloud cluster boundary layer and the need for a lifting mechanism to initiate parcel condensation until free convection is obtained is generally accepted. This requires that substantial mechanical forcing be applied in order that buoyant elements be initiated. As discussed by López (1972a), it is not possible to generate cumulus over the oceans without mechanically forced cloud base vertical velocities of 1-5 m/sec. Underneath his cumulus (cu), towering cumulus (TWG) and cumulonimbus (Cb), he must have average local convergences during their forced ascent stages of about $3\text{-}6 \times 10^{-3} \sec^{-1}$. These values are one to two thousand times larger than the synoptic convergence. Table 4 lists the required mean forcing parameters used in his simulation of the three sizes of cumulus. Even if the active cumulus take up but one percent of the cluster area, these convergences are 10-20 times larger than that specified by synoptic convergence. The only way the extra convergence under the cluster cumulus can occur is by a large mass recycling mechanism which is 10-20 times greater than the synoptic mass convergence. Betts (1972) has composited wind data around growing cumulonimbus clouds in Venezula and he finds convergence (for regions 2-5 times larger than López's Cb cloud base areas) in the subcloud layers of $\sim 2 \times 10^{-3} \sec^{-1}$, quite closely comparable with the required cumulonimbus boundary layer forcing of López's model. The recycling dry and moist downdraft air which penetrates into the boundary layer from upper levels is thus the primary source of mass forcing for buoyant parcel initiation. Were the cluster cumulus to be forced only by the synoptic convergence of $3\text{-}5 \times 10^{-6} sec^{-1}$, then the low level mass forcing requirements would permit only 1/1000 of the cluster area to be occupied by parcel ascent if all the mass sent into Cb's, or only 1/500 of the area if occupied by all towering cumulus or by all cumulus parcel ascent. López's (PaperII) radar data shows that the areas of the cluster occupied by active cumulus clouds are typically 10-20 times larger than these amounts. Given the typical cloud cluster synoptic convergences and stable lapse rate conditions, the fundamental requirement of large up- and-down mass recycling 10-20 times greater than the mean vertical mass flow at cloud base is clearly evident.

Table 4

Parameters used in Lopez's (1972a) simulation of cumulus

Cloud Type	Cloud Base Height (meters)	Mean Forced Convergence under Cloud during its initiation $-\nabla \cdot \mathbb{V}$ (sec^{-1})	Initial Radius R_o (km)	Duration of Forced up-draft through Cloud Base T_o (minutes)	Mean Value of Sinsoid-ually Forced Updraft $\bar{w}_o$ (m/sec)
Cb	520	$\sim 6 \times 10^{-3}$	2.0	20	3.3
TWG	520	$\sim 3 \times 10^{-3}$	1.0	15	1.7
Cu	520	$\sim 3 \times 10^{-3}$	0.5	15	1.7

Relationship of Cluster Mass Recycling to CISK Mechanism. It is obvious that synoptic-scale boundary layer induced frictional convergence from Ekman wind veering or the "so called Conditional Instability of the Second Kind (CISK)" mechanism as defined by Charney and Eliassen (1964) plays only a small direct role in the cluster upward mass transfers necessary to initiate the cumulus. This does not negate the fundamental need of the cluster for frictionally induced synoptic scale convergence, but it dictates that the primary direct function of the CISK process is to act as a mechanism for water vapor convergence. Unless the low levels of the cluster are continuously fed with water vapor, then the rain processes would act to decrease the cluster vapor and the resulting higher buoyant stability of the dryer environment would put unrealistically high requirements on the boundary layer mass forcing. Cumulus convection could not be sustained. This point has been emphasized by Rodenhaus (1971).

The very high correlation of boundary layer* relative vorticity with cumulus convection along frontal zones, squall lines, on the cyclonic shearing side of the trade wind, in tropical storms, etc., illustrates what a fundamental role the low level frictional forcing plays in allowing the cumulus recycling mechanism to get going. The general lack of cumulus convection in anticyclones and other regions of negative vorticity where weak lower level friction divergence is occurring is primarily due to the sinking induced drying of the environment.

*The terms "boundary layer" and "sub-cloud layer" are used synonymously. This refers to the layer from the surface to 950 mb.

To understand the high correlation of cumulus convection with boundary layer convergence and the CISK mechanism,one must fully appreciate the powerful influence of relative humidity on cumulus buoyancy. Figs. 11 and 12 showed that the major difference between the cluster and clear regions is in the relative humidity above 950 mb. Table 5 compares the temperature difference between a lifted parcel and the environment for the cluster and clear regions. In this table the parcel has been lifted from a condensation level of 950 mb with assumed entrainment rates of zero, 10 and 20 percent per 100 mb of parcel ascent. This table clearly shows the very strong dependence of parcel ascent on environment relative humidity and entrainment. A 10% per 100 mb entrainment rate would lead to a 450 mb parcel minus environment temperature of $-1.6^{\circ}C$ in the cluster as compared with $-6.0^{\circ}C$ in the clear areas. These are very large differences.

The cluster divergence profile of Fig. 20 is a result of the combined low-level frictional forcing mechanism and the cumulus feed-back of the up-and-downdrafts. Cluster convergence above the boundary layer is primarily a consequence of the cumulus. Middle level convergence is necessary to feed the accelerating downdrafts. Fig. 34 shows that the 400 to 950 mb increase of the total downward motion is greater than the 950 to 400 mb increase of the total upward motion. This middle level convergence leads to additional water vapor convergence.

After boundary layer convergence has initiated and sustained an updraft impulse to the level of free convection, an accelerating upward

TABLE 5

CLOUD CLUSTER VS. CLEAR REGION BUOYANCY PARAMETER FOR MOIST ASCENT FROM 950mb

Pressure Level	Relative Humidity (percent): Cluster	Relative Humidity (percent): Clear Areas	Cumulus Parcel Temperature Minus Environment Temperature (°C): No Entrainment: Clusters	No Entrainment: Clear Regions	For Environment Entrainment of 10% per 100mb: Clusters	For Environment Entrainment of 10% per 100mb: Clear Regions	For Environment Entrainment of 20% per 100mb: Clusters	For Environment Entrainment of 20% per 100mb: Clear Regions
950	80	66	0	0	0	0	0	0
850	77	55	+1.0	+0.7	+0.2	-0.8	-0.6	-2.6
750	70	40	+1.6	+1.3	0	-2.2	-2.8	-4.8
650	65	30	+2.0	+1.6	-0.2	-3.2	-3.7	-6.8
550	58	24	+2.7	+2.2	-0.8	-4.4	-5.2	-8.5
450	52	20	+3.0	+2.6	-1.6	-6.0	-7.0	-11.3

growth occurs which produces changes to the middle and upper tropospheric levels. If the updraft is sufficiently intense such that downdrafts and rain are produced, then further upsetting and influence to the middle and upper tropospheric levels occurs. The sharp increase of stability near the 200 mb level produces a rapid deceleration and mass divergence at this level. The accelerating downdrafts of the middle and lower troposphere produce cluster convergence at these levels. Even though middle and upper level broadscale responses to the individual cumulii occur, this does not negate the fundamental role of the boundary layer in initially dictating these responses.

Influence of Large-Scale Boundary Layer Convergence. The physics of the individual cumulus (as discussed by López), shows that the cloud cluster's mass convergence does not produce enough mass forcing underneath the individual cumulus such that it can directly release very much of the vapor convergence to rain. The cluster's convergence is too weak. Vapor accumulation is prevented and a steady state is obtained only by the establishment of an additional vigorous vertical mass recycling into the boundary layer. This extra up-moist and down-dry recycling would carry vapor upward and act to reduce the boundary layer vapor. The intensity of the recycling would increase until a balance was obtained between the overall cluster vapor convergence-evaporation and the cluster rainfall. At this point the downward mass flux into the boundary layer would be strong enough to produce enough mass forcing under enough ascending parcels such that cluster rainfall of 2.5 cm/day can occur.

Degree of Recycling. This degree of recycling may be defined as the ratio of the required absolute upward vertical motion at the top of the boundary layer, $|\omega_{up}|$, to the mean vertical motion, $-\overline{\omega}$, at this level, or $|\omega_{up}| / -\overline{\omega}$. The ratio is dependent upon the magnitude of the cluster water vapor accumulation from evaporation and convergence to the rainfall which the boundary layer mass convergence would produce from parcel forcing without recycling. Using López's cumulus model and the cluster observed mass convergence of $3 \times 10^{-6} \sec^{-1}$, the ratio of vapor accumulation to rainfall induced from the broadscale boundary layer forcing by itself was found to be about 20 or 30 to 1. Boundary layer convergence of $3 \times 10^{-6} sec^{-1}$ produce only enough mass forcing to the ascending parcel for a rainfall of about 0.1 cm/day. For cluster rainfall of 2.5 cm/day, an additional mass source about 25 times greater than that of the mean boundary layer convergence is required. The downward mass penetration from upper levels associated with the recycling mechanism is this additional mass source.

The CISK mechanism must thus be viewed not as a direct cumulus producing process, for in this it is much too weak, but instead as a required "trigger" for recycling. As far as cumulus convection is concerned, the most important aspect of the CISK mechanism is not the magnitude of the direct boundary layer mass convergence but rather the associated water vapor convergence.

Possible Importance of Recycling for Vertical Momentum Transfer. In that the cluster sinking warming and the evaporation cooling largely balance each other and in that the up- and-down recycling mass largely balance, one might argue that it is not important to deal with the recycling mechanism directly but only with its net influence. From the cluster mass and energy budgets this may be a reasonable conclusion; for the water vapor and probably for the momentum budgets it is not. The correlation of the recycling up and down motion with specific humidity has been shown to be very large. It seems likely that there would also be a significant correlation of the recycling motion with horizontal momentum. If this proves to be the case, then the recycling process has feed-back momentum influences which would also have to be parameterized along with the energy ones.

Definition of Planetary Boundary Layer. In this paper the planetary boundary layer is viewed as that layer where surface mechanical-driven gust-scale eddies (~50-500 m size) exist and decrease in density and intensity with height. Over the oceans this layer is typically one-half to one km thick (Gray, 1972c) and is primarily associated with the subcloud layer. To define the boundary layer with respect to a mixed layer which is dependent on the depth of the cumulus convection, as some researchers propose, is the obscure the important difference between gust-scale mechanically driven processes and cumulus thermally driven processes. The large vertical mass recycling portrayed in this paper show that a definition with respect to the top of the cumulus always puts the boundary layer near the tropopause in cluster situations and near 950 mb in clear situations. The top of this mixed layer is too variable. For simplicity the author proposes that the term "boundary layer" be used to apply only to the usual Ekman layer where surface generated mechanical gust-scale eddies are dominant. As long as surface winds are present, the mechanically driven gust-scale exchanges of this lowest km layer are always present regardless of the existence or degree of cumulus activity above.

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