The Mutual Variation of Wind, Shear and Baroclinicity in the Cumulus Convective Atmosphere of the Hurricane

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ABSTRACT

The enhanced cumulus convection in tropical disturbances acts in two opposing ways. In one sense the condensation heat from the cumulus acts to warm the inner portions of the disturbance and induce vertical shear of the horizontal wind through the thermal wind relationship. In the opposite sense the cumulus also act to suppress vertical wind shear by transfer of horizontal momentum within their up- and downdrafts. The operation of this dual or "paradox" role of the cumulus cloud is of fundamental importance for understanding the steady-state dynamics of the hurricane's inner core region and is also hypothesized to be of basic importance in the development of tropical storms and generation of easterly waves.

Observational information and calculations on the radial distribution of baroclinicity and vertical shear collected on 102 individual flight legs (on 19 flight levels) into hurricanes flown by the Research Flight Facility of the U. S. Weather Bureau are presented. The measured baroclinicity in the lower half of the troposphere at the inner 50-70 km radii is two to three times larger than cylindrical thermal wind balance would require. This baroclinicity may be 50 to 100 times larger in the inner areas of the hurricane than in the easterly wave, yet vertical wind shear in the hurricane may be but one to three times as large.

An extensive discussion is presented on the characteristics of the vertical motion within the hurricane. A steady-state model of mean flow conditions in the inner 80 km radius based on flight observations and best known surface stress conditions is assumed. Upon this mean vortex motion five sizes of cumulus up- and downdrafts are superimposed with their characteristic eddies and resulting stress. A discussion of the terms in the steady-state equations of motion reveals that sizable amounts of cumulus produced radial and tangential frictional acceleration is required to satisfy the broad-scale mean flow conditions. These cumulus produced horizontal accelerations account for the imbalance in the thermal wind relationship. Tropical storm development is not viewed as being possible unless the cumulus induced vertical momentum transfers act in a dominant way to oppose the thermal wind requirement and inhibit increase of vertical wind shear as baroclinicity increases. Vortex development thus requires a continual imbalance of pressure over wind acceleration.

1. INTRODUCTION

The crucial importance of realistically handling the tropical atmosphere and the effects of cumulus convection in treatment of the earth's global circulation has been discussed by the General Circulation Research Groups. Nevertheless, no established methodology for handling the cumulus convective atmosphere has yet been developed. The mutual interactive effects of the cumulus cloud with the synoptic-scale circulations are not well understood.

Recent numerical experiments on tropical storm genesis (Ooyama [36], Kasahara [23], Charney and Eliassen [7], Ogura [35], Kuo [24], Syono and Yamasaki [52]) have attempted to deal with the cumulus by parameterizating the condensation heating of the cumulus in terms of known variables of the larger scale.

These model experiments have not considered the possible interactive effects of momentum exchanges associated with the cumulus up- and downdraft elements. But, due to the probable high correlations of cumulus scale horizontal and vertical wind fluctuations in a vertical shearing flow, cumulus-scale momentum exchanges may play an equally important or dominant role in the realistic treatment of the cumulus atmosphere (Gray [14]).

With the establishment of the National Hurricane Research Project in 1955, a large quantity of observational information concerning the cumulus convective atmosphere is now available. Certain questions can be answered and other inferences drawn as to the physical processes operating on this hitherto incompletely observed convective scale of motion. The intensity of cumulus convection in the tropical storm (particularly near its center) is greater than in any other synoptic scale storm system. It may be easier to isolate the effects of the cumulus in this system. What we learn about cumulus convection in the inner region of the hurricane may have applicability to other less intense cumulus convective systems such as the easterly wave or squall line. The dynamics of the inner 100 km radius of the tropical storm where cyclonic circulation is present through the entire troposphere has yet to be completely studied.

Background: In comparison with the middle latitudes little horizontal temperature gradient is present in the tropical atmosphere. When large temperature gradients exist, they must be primarily produced by concentrated amounts of cumulus, even though necessary amounts of sensible and latent heat for cumulus buoyancy is received as a dynamic bi-product from the ocean. Advection of heat cannot produce the strong temperature gradients observed in the tropical storm which may exceed wintertime middle-latitude gradients by an order of magnitude. By contrast, middle-latitude gradients of temperature are primarily brought about through meridional advection of radiation-developed temperature differences.

The vertical wind shear near the eye wall in the lower half of the troposphere of the hurricane is only one to three times larger (Hawkins [17]) than the vertical wind shear in a similar layer in the typical easterly wave. Yet the baroclinicity present may be one hundred times as large (0.1°C/100 km vs. 10°C/100 km - Fig. 1). To understand the progressive development of the hurricane from the pre-existing disturbance, at least two perplexing questions must be answered:

1) How does it happen that the condensation-developed horizontal temperature gradients in the tropical storm are only slightly reflected by increase of vertical wind shear as occurs with middle latitude westerly circulations?

2) How is it possible for cumulus convection (which release condensation heat on a scale of 2-6 km) to generate a tropical storm which is of size two to three orders of magnitude larger than the individual cumulus elements?

This paper attempts to offer a physical explanation to the above questions by demonstrating the plausibility of cumulus up- and downdraft (2-6 km width) momentum transfers acting as a primary

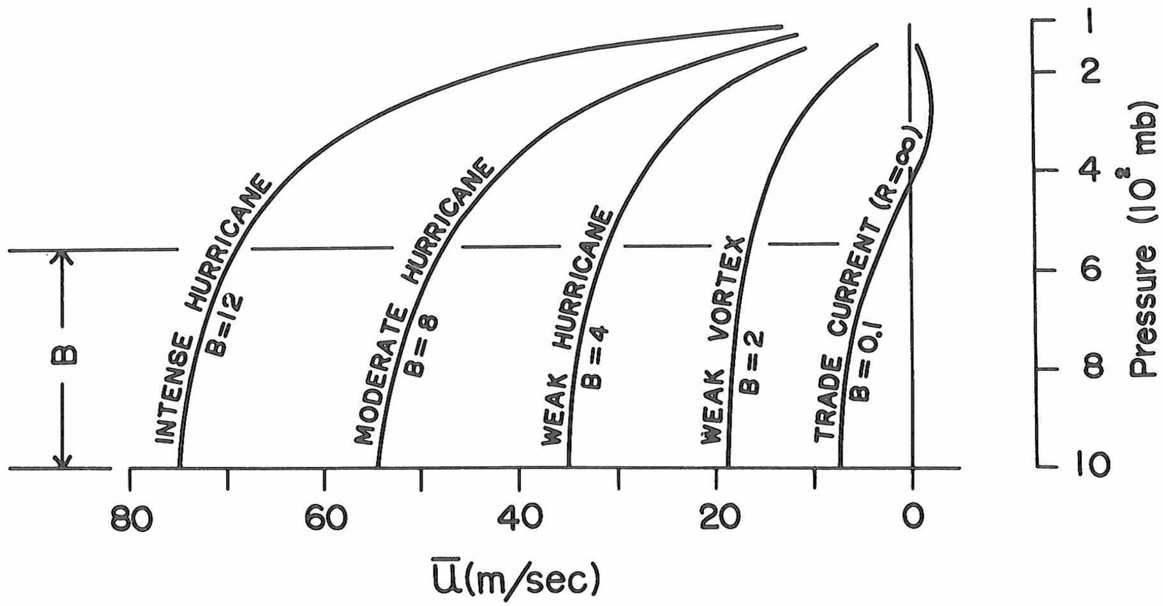

Fig. 1. Typical vertical shear of the horizontal wind existing at 40 km radius in tropical vortices of various intensities and in the usual trade current (radius = ∞). B is the average magnitude of baroclinicity present (°C/100 km along constant pressure surface) in the lower half of the troposphere (1000 - 550 mb) at 40 km radius under the assumption of cyclindrical thermal wind balance. f taken at 15°N

mechanism to inhibit increases of vertical wind shear as baroclinicity increases and couple the upper and lower troposphere and organize the circulation on a larger mesoscale. The cumulus up- and downdrafts also act to inhibit and regulate cross isobaric flow and thereby kinetic energy generation.

Qualitatively the intensification process is invisaged as follows: Differential surface convergence induces differential condensation heating and increase of horizontal temperature gradient between the cumulus and cumulus-free areas. Surface pressure is reduced in the areas of condensation (with assumed stratospheric level of no gradient). An imbalance between pressure and wind acceleration occurs. Although the atmosphere continually attempts to adjust itself to gradient wind balance, this adjustment is not instantaneous. In adjusting to the imbalance of wind and pressure, other accelerations are activated. For the usual increase of horizontal temperature gradient in middle-latitude circulations, an increase of wind shear takes place. But, the vertical momentum transports associated with cumulus-developed baroclinicity in the tropical wave or storm inhibit an increase of the vertical wind shear from acting as the primary mode of adjustment toward balance. The primary adjustment must then come from changes of the wind speed and/or curvature of flow. In this sense cumulus act to "force" a production of wind speed and/or increase of curvature. Cumulus induced velocity-curvature growth of this type might be termed "cumulus momentum exchange growth".

2. THERMAL WIND EQUATION WITH CURVATURE

Natural Coordinates: With the assumption of frictionless motion and with the hydrostatic approximation, the thermal wind equation in natural coordinates with p as the vertical coordinate may be written as

$$\left(f + \frac{2V}{R}\right) \frac{\partial V}{\partial p} = - \left.\frac{C}{p} \frac{\partial T}{\partial n}\right)_p \tag{1}$$

where f = Coriolis parameter

V = horizontal wind speed

R = radius of trajectory curvature which is assumed to be constant with height

p = pressure

C = gas constant

T = virtual temperature of air

n = distance along R , positive to the right of the direction of V in the Northern Hemisphere, and

$\Big|_p$ denotes differentiation along the constant pressure surface.

In this derivation it is assumed that there is no rotation of the coordinate system with height. The above equation may be more simply written as

$$W_R S_n = B_n \qquad (2)$$

where $W_R = \left(f + \frac{2V}{R}\right)$ = inertial parameter

$S_n = \frac{\partial V}{\partial p}$ = vertical wind shear parameter

$B_n = -\frac{C}{p} \frac{\partial T}{\partial n}\Big|_p$ = baroclinicity.

If differential condensation heating produces an increase with time of baroclinicity (denoted $\dot{B}_n$), changes in either wind speed ($\dot{V}$), curvature ($\dot{R}$), or wind shear ($\dot{S}_n$) must result (with constant f) as the atmosphere continually attempts to adjust itself to a new thermal wind balance.

If the cumulus acted to inhibit changes of S , substantial changes of V and/or R must result to balance even slightly horizontal temperature gradient increases. Thus for assumed tropical easterly wave flow with

1) radius of curvature of 10^3 km,

2) wind shear in the lower two-thirds of the troposphere of 6 m/sec,

3) mean wind of 5 m/sec (8 m/sec at the surface and 2 m/sec at 400 mb),

a 0.1°C/100 km temperature gradient increase with no change of vertical wind shear would require a layer mean wind increase from 5 m/sec to ~ 25 m/sec, or a decrease of the radius of curvature from 1000 to ~ 200 km for continued gradient wind balance (f taken at 20° latitude).

Cylindrical Coordinates: The complete cylindrical thermal wind equation with origin continually at the center of a vortex[1], can be written as

$$\left(f + \frac{2u}{r}\right) \frac{\partial u}{\partial p} = - \frac{C}{p} \left.\frac{\partial T}{\partial r}\right)_p - \frac{\partial F_r}{\partial p} + \frac{\partial}{\partial p}\left(\frac{dv}{dt}\right) \qquad (3)$$

where r = radius from coordinate origin

u = velocity in the tangential or θ direction relative to the moving storm center, positive counter clockwise

v = velocity in the radial direction relative to the moving storm center, positive outward

F_r = frictional acceleration in radial direction

$\frac{d}{dt}$ = substantial derivative

and the other symbols as before.

Disregarding the second and third terms on the right of (3) by assuming steady, frictionless motion, (3) can be written as

[1] Effects due to curvature of the earth may be neglected if consideration is given only to processes at the inner radii ($r < 150$ km).

$$W_r S = B \tag{4}$$

where $W_r = \left(f + \frac{2u}{r}\right)$

$S = \frac{\partial u}{\partial p}$

$B = -\frac{C}{p} \left.\frac{\partial T}{\partial r}\right)_p .$

If balanced conditions were to periodically prevail at incremental periods following increases of B with no increase of S , u would then be directly related to B .

3. OBSERVATIONAL INFORMATION

To what extent do the meteorological observations support the above hypothesis concerning the importance of vertical cloud-scale momentum transports? The aircraft flight tracks flown by the National Hurricane Research Project during the 1958-61 seasons (Fig. 2 is typical) permit direct determination of W_r and B along the individual flight legs. A calculation of the required vertical wind shear to accompany W_r and B may then be made.

In addition, if it is assumed that the individual leg observations on each flight level were taken over a quasi-constant time interval, an approximation to the tangential average parameters along the storm radius can be made by assigning weighting factors to each radial leg according to its percentage area of representation.

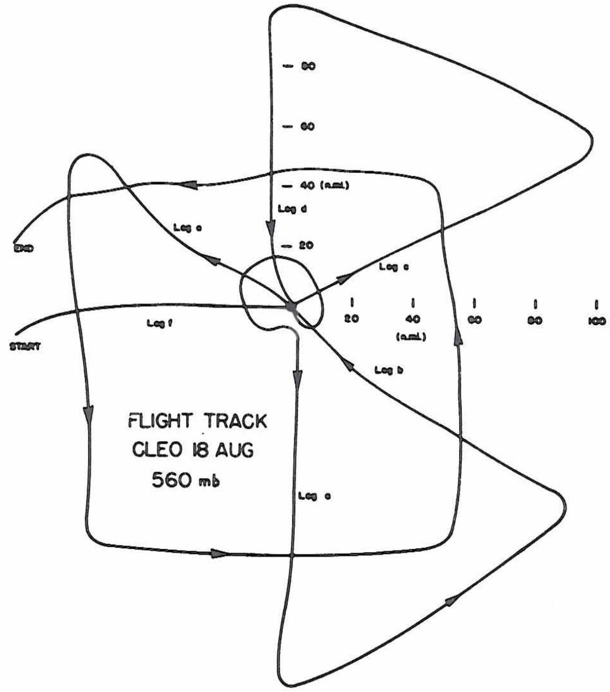

Fig. 2. Typical flight track at individual levels

Four to seven (usually 5-6) radial penetrations were flown over periods of 4-7 hours. Determination of the required vortex average vertical wind shear $\frac{\partial u}{\partial p}$ to accompany the vortex average measured values of W_r and B for steady, balanced, frictionless flow conditions may thus be made. A direct comparison of calculated with measured vertical wind shear values is possible in five cases where two aircraft operated together at different flight levels. Determination of the degree to which the cylindrical form of the balanced thermal wind equation (4) is applicable to the hurricane vortex may then be made.

Descriptions of the aircraft used, the meteorological instrumentation, the flight tracks flown, and the character of the data collected, etc., have been extensively discussed (Hawkins, et al. [16], Riehl and Malkus [42], Reber and Friedman [39], Colón [8], [9], Gray [12], [13], [14], Miller [31], La Seur and Hawkins [25], Gentry [10]). The reader is referred to these references for detailed information on the characteristics of the data.

Table 1 lists information concerning 19 flight levels on which individual radial leg and vortex average radial temperature and wind data are available. Table 2 lists observed flight level vortex average radial adjusted temperature decrease (or temperature gradient along the constant pressure surface) in 18 km segments from 18 to 90-126 km. In order that truly representative temperature gradients might be obtained, all spot temperatures used were obtained from averaging over 9 km on either side of the boundary value used. The temperature gradient between radius 36 and 54 km is thus obtained from the difference of the mean temperature in the radial interval from 45-63 km subtracted from that of the mean temperature in the interval between 27-45 km (Fig. 3).

Virtual temperature correction has not been made. Relative humidities are between 80 - 100 per cent at the inner radii in the middle and lower troposphere. This would account for a maximum error of 0.2°C in the temperature gradients. As temperatures

TABLE 1

FLIGHT LEVEL INFORMATION

Storm	Date	Flight Level mb	Radius of Data km	No. of Flight Leg Calculations	Approx. Max. Wind at Flight Level m/s	Approx. Central Pressure mb	Speed of Storm m/s
CARRIE	15 SEPT 57	605	18-108	6	42	965	6
CARRIE	17 SEPT 57	690	18-108	6	47	970	6
CLEO	18 AUG 58	810	18-126	5	45	970	7
CLEO	18 AUG 58	560	18-126	6	40	970	7
CLEO	18 AUG 58	255	18-180	4	20	970	7
DAISY	25 AUG 58	825	18-108	6	32	990	3
DAISY	25 AUG 58	555	18-108	6	30	990	3
DAISY	27 AUG 58	630	18-144	6	60	940	4
DAISY	28 AUG 58	620	18-126	7	52	945	9
HELENE	24 SEPT 58	635	18- 90	6	30	995	5
HELENE	25 SEPT 58	810	18- 90	6	40	980	5
HELENE	26 SEPT 58	580	18- 90	5	60	950	7
HELENE	26 SEPT 58	270	18-180	4	40	950	7
DONNA	7 SEPT 60	760	18-108	4	64	945	5
DONNA	7 SEPT 60	635	18- 72	4	60	945	5
CARLA	6 SEPT 61	905	18-108	6	32	980	6
CARLA	6 SEPT 61	585	18-108	6	25	980	6
CARLA	8 SEPT 61	860	18-144	4	42	955	5
CARLA	8 SEPT 61	720	18-144	4	39	955	5

TABLE 2
VORTEX AVERAGE RADIAL
TEMPERATURE DECREASES ALONG CONSTANT
PRESSURE SURFACES (°C)

Storm	Date	Flight Level mb	RADIUS (km)						
			18-36	36-54	54-72	72-90	90-108	108-126	18-72
CARRIE	15 SEPT 57	605	3.9	2.1	0.1	1.4	-0.4		6.1
CARRIE	17 SEPT 57	690	0.2	2.1	1.8	0.0	0.4		4.1
CLEO	18 AUG 58	810	2.2	1.2	-0.1	0.1	0.3	0.3	3.3
CLEO	18 AUG 58	560	3.1	1.9	-0.2	0.3	0.5	0.0	4.8
CLEO	18 AUG 58	255	0.1	0.7	1.3	1.0	1.0	0.2	2.1
DAISY	25 AUG 58	825	1.7	0.0	-0.3	0.0	0.0		1.4
DAISY	25 AUG 58	555	0.8	0.4	-0.1	-0.2	-0.1	-0.4	1.1
DAISY	27 AUG 58	630	2.3	0.4	-0.3	0.0	-0.1	0.1	2.4
DAISY	28 AUG 58	620	4.5	1.7	1.2	0.7	0.0	0.1	6.4
HELENE	24 SEPT 58	635	1.5	1.5	0.5	0.3			3.5
HELENE	25 SEPT 58	810	-0.1	1.9	1.0	0.3			2.8
HELENE	26 SEPT 58*	580	4.9	1.0	0.0	0.7	-0.6		5.9
HELENE	26 SEPT 58	270	1.3	2.7	1.5	0.9	0.6		5.5
DONNA	7 SEPT 60	760	3.4	1.1	1.1	0.7			5.6
DONNA	7 SEPT 60	635	5.5	2.1	0.2				7.8
CARLA	6 SEPT 61	905	1.0	0.5	0.2	0.1	0.0		1.7
CARLA	6 SEPT 61	585	0.9	0.5	0.3	0.2	-0.1		1.7
CARLA	8 SEPT 61	860	-0.2	1.0	0.6	0.4	0.0		1.4
CARLA	8 SEPT 61	720	1.1	2.5	1.3	0.5	0.2		4.9
		AVE.	2.0	1.3	0.5	0.4			3.8

*Southern Quadrant value taken from 710 mb flight leg

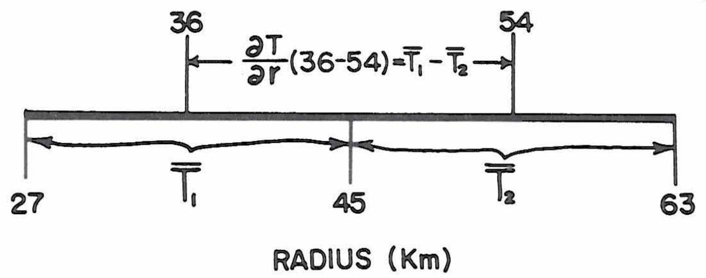

Fig. 3. Illustrating method of determining temperature gradient between 36 and 54 km radius.

are averaged over 18 km, radial gradients of any possible wetbulb effect to the vortex temperature reading are not thought significant.

The listed vortex average temperature gradients of Table 2 are demonstrated to be reasonable if a comparison is made of the observed layer average temperature gradients with the radial gradients of temperature computed from layer thickness differences.[2]

Table 3 shows a close comparison of observed layer average temperature gradients between radial intervals with the temperature gradients obtained from layer thickness gradients for seven available cases. Hydrostatic consistency was closely approximated. For these reasons the observed 18 km average temperature gradients are thought to be reliable in the inner area of the hurricane where sizable temperature gradients are present.

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[2] Layer mean temperature was obtained by averaging at the top and bottom.

TABLE 3

COMPARISON OF VORTEX AVERAGE OBSERVED AND COMPUTED TEMPERATURE GRADIENTS

Storm	Date	Between Pressure Levels mb	Radius km	Observed Temperature Gradient °C	Computed Temperature Gradient °C
CLEO	18 AUG 58	810-560	18-54	3.7	4.0
CLEO	18 AUG 58	560-255	18-90	3.6	4.0
DAISY	25 AUG 58	825-555	18-54	1.4	1.4
HELENE	26 SEPT 58	580-270	18-90	7.0	7.8
DONNA	7 SEPT 60	760-635	36-72	2.2	1.7
CARLA	6 SEPT 61	905-585	18-72	1.7	2.0
CARLA	8 SEPT 61	860-720	18-72	3.2	4.4
			AVE.	3.3	3.6

Computations of W_r and B terms were made along the radial legs on all the levels listed.[3] A determination was then made of the vertical wind shear which should accompany W_r and B if steady, balanced flow were present.[4] Table 4 lists typical measured values of temperature gradient and resultant S (in m/sec per 225 mb or one-quarter of the troposphere) for gradient balance along radial leg increments on each of the six flight legs of Hurricane Cleo at 560 mb on 18 August 58. Large variation of ΔT and S are present. The computed vertical shears ranged up to 23 m/sec per 225 mb. Computations on this level are closely representative of the results obtained on the other individual radial legs.

The area weighted vortex average computed shear in the inner two 18 km segments for Cleo at 560 mb was 15 and 8 m/sec per 225 mb. Yet the vortex average observed wind shear between 810 and 560 mb for the same radial segments were but 2.3 and 3.5 m/sec per 225 mb.

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[3] The relative wind with respect to the moving vortex center was used in the tangential average or symmetric vortex considerations. A small percentage correction to the listed relative tangential wind (i.e., u) was made to compensate the Doppler determined winds for ocean particle movement (see Appendix).

[4] Due to divergent radii lines the $(f + \frac{2u}{r})$ term was calculated as $\sum_i f + \frac{2\bar{u}}{\Delta r} \ln \frac{r_{2i}}{r_{1i}}$, where Δr is a 9 km interval and r_{2i}, r_{1i} are the outer and inner radii, respectively. $\bar{u}$ is the average value of u over 9 km radial interval.

TABLE 4

18 KM INCREMENTAL INDIVIDUAL RADIAL LEG TEMPERATURE GRADIENTS (°C) AND CALCULATED VERTICAL WIND SHEAR (m/sec per 225 mb - parentheses) FOR 560 MB FLIGHT LEVEL OF CLEO 18 AUG. '58.

Radius km	Storm Quadrant of Radial Flight Track: NE	SE	S	W	NW	N	Area Weighted Vortex Average
18- 36	2.5 (7.4)	3.6 (13.1)	2.9 (12.6)	2.8 (23.4)	3.0 (15.5)	2.9 (16.9)	3.1 (14.8)
36- 54	2.0 (6.8)	1.8 (5.9)	2.2 (6.8)	2.0 (10.4)	2.3 (7.4)	1.0 (9.9)	2.0 (7.8)
54- 72	-0.4 (-2.0)	0.1 (0.7)	-0.6 (-3.6)	0.2 (1.1)	-0.2 (-1.3)	-0.2 (-1.1)	-0.2 (-1.0)
72- 90	0.6 (4.1)	0.3 (2.2)	0.9 (6.5)	0.8 (5.6)	0.1 (0.8)	0.4 (3.2)	0.5 (3.7)
90-108	1.0 (8.1)	0.3 (3.1)	0.4 (4.1)	0.0 (0.0)	0.7 (6.5)	0.9 (8.3)	0.6 (5.0)
108-120	0.2 (1.8)	0.1 (1.2)	0.3 (-2.7)	-0.9 (-9.5)	0.1 (12.2)	0.5 (5.8)	0.1 (1.5)
120-138	0.1 (1.1)	-0.9 (-9.0)	0.2 (2.2)	-9.5 (-6.5)		0.1 (1.3)	

Table 5 summarizes results of vortex average calculated vertical wind shear for the highest 36 km radial segments on 17 middle and lower troposphere levels. Wind shears are given in percentage (or ratio) of amount of shear per 225 mb to layer average wind speed. Percentage vertical wind shear is defined as

$$\text{percentage shear} = \frac{u_b - u_t}{\bar{u}} \times 100 \qquad (5)$$

where u_t = tangential velocity at top of 225 mb layer

u_b = tangential velocity at bottom of 225 mb layer

$\bar{u} = 1/2\,(u_b + u_t)$.

Calculated shear as high as 37 per cent of the observed mean wind per 225 mb was obtained. Even though vortex averages were taken, unrealistically high computed percentage shear was obtained at the inner radii. Flight observations have shown, however, that measured wind shears in the lower half of the troposphere at the inner radii of the hurricane are considerably smaller. Hawkins [17] has constructed a nomogram of vertical wind shear obtained from available hurricane flight data of 1957-58. This diagram shows wind shear through the entire lower half of the troposphere (surface to 550 mb) within radii of 60-80 km to average but ten per cent of the mean wind speed (3-4 m/sec per 450 mb or five per cent per 225 mb). Flight observations since 1960 have further substantiated Hawkins' data.

TABLE 5

CALCULATED PERCENTAGE VERTICAL WIND SHEAR PER 225 MB FOR HIGHEST 36 KM RADIAL INTERVALS

Storm	Date	Level (mb)	Radius (n. mi.)	Ave. Percentage Vertical Wind Shear Per 225 mb
CARRIE	15 SEPT 57	(605)	18-54	31
CARRIE	17 SEPT 57	(690)	36-72	22
CLEO	18 AUG 57	(810)	18-54	12
CLEO	18 AUG 57	(560)	18-54	35
DAISY	25 AUG 58	(825)	18-54	12
DAISY	25 AUG 58	(555)	18-54	13
DAISY	27 AUG 58	(630)	18-54	6
DAISY	28 AUG 58	(620)	18-54	19
HELENE	24 SEPT 58	(635)	18-54	37
HELENE	25 SEPT 58	(810)	36-72	36
HELENE	26 SEPT 58	(580)	18-54	13
DONNA	7 SEPT 60	(760)	18-54	6
DONNA	7 SEPT 60	(635)	18-54	15
CARLA	6 SEPT 61	(905)	18-54	13
CARLA	6 SEPT 61	(585)	18-54	34
CARLA	8 SEPT 61	(860)	18-54	5
CARLA	8 SEPT 61	(720)	18-54	30
			AVE.	20 %

Table 6 lists ratios of percentage vortex average calculated shear to assumed vertical shear of five per cent per 225 mb for the 17 middle and lower tropospheric levels. In the radial segments between 18 and 54 km this ratio averages 4.4 and 3.3, respectively. This is 4.4 and 3.3 times larger than Hawkins' wind shear nomogram.

Five cases are available where the aircraft operated simultaneously at different levels in the lower troposphere so that direct measurement of actual wind shear could be made. Table 7 lists a comparison of computed to observed percentage vertical wind shears per 225 mb. For the inner two 18 km radial segments the ratio of computed to observed shear was nearly three and two, respectively. There seems to be no question that the observed vertical wind shears at the inner radii may be two to three times smaller than the computed shears obtained from the horizontal temperature gradients. Why should this happen?

A complete evaluation of eq. (4) could be accomplished in five cases where vertical shear was measured. Table 8 portrays computed values of $W_r S$ and B for these cases.

If B were larger or smaller than $W_r S$, then an excess (B_{ex}) or deficit ($-B_{ex}$) of baroclinicity above or below that required for balanced flow is hypothesized to be present. Thus

$$B_{ex} = B - W_r S \quad . \tag{6}$$

TABLE 6

RATIO OF CALCULATED TO AN ASSUMED VERTICAL WIND SHEAR OF 5 PER CENT PER 225 MB

Storm	Date	Level (mb)	Radius (km) 18-36	36-54	54-72
CARRIE	15 SEPT 57	(605)	9.4	3.2	0.4
CARRIE	17 SEPT 57	(690)	0.8	5.4	3.2
CLEO	18 AUG 58	(810)	3.6	1.4	-0.2
CLEO	18 AUG 58	(560)	9.4	4.6	-0.6
DAISY	25 AUG 58	(825)	10.0	-0.4	-2.6
DAISY	25 AUG 58	(555)	2.8	2.6	-1.0
DAISY	27 AUG 58	(630)	1.6	0.6	-1.4
DAISY	28 AUG 58	(620)	4.4	3.2	4.2
HELENE	24 SEPT 58	(635)	7.4	7.6	2.3
HELENE	25 SEPT 58	(810)	-0.8	8.8	5.3
HELENE	26 SEPT 58	(580)	3.8	1.4	0
DONNA	7 SEPT 60	(760)	1.6	0.8	1.6
DONNA	7 SEPT 60	(635)	3.6	2.2	0.4
CARLA	6 SEPT 61	(905)	3.0	2.1	0.8
CARLA	6 SEPT 61	(585)	7.6	6.0	4.4
CARLA	8 SEPT 61	(860)	-0.8	1.8	1.2
CARLA	8 SEPT 61	(720)	7.4	4.4	3.2
		AVE.	4.4	3.3	1.3

TABLE 7

RATIO OF CALCULATED TO OBSERVED PERCENTAGE WIND SHEAR PER 225 MB

Storm	Date	Between Pressure Level mb	Radius (km) 18-36	36-54	54-72
CLEO	18 AUG 58	810-560	25/10	22/11	9/ 3
DAISY	25 AUG 58	825-555	32/ 6	6/-7	-9/-10
DONNA	7 SEPT 60	760-635	13/-2	8/ 8	5/ 1
CARLA	6 SEPT 61	905-585	26/19	20/12	13/21
CARLA	8 SEPT 61	860-720	16/ 6	16/12	11/13
		AVE.	22/ 8	13/ 7	6/ 6

TABLE 8

LAYER AVERAGE CALCULATIONS OF

B , $W_r S$ AND B_{ex}

(units 10^{-6} cm^2/gm)

Storm	Date	Layer Thickness mb	Radius (km) 18-36 B			36-54 B			54-72 B		
			B	$W_r S$	B_{ex}	B	$W_r S$	B_{ex}	B	$W_r S$	B_{ex}
CLEO	18 AUG 58	810-560	6.5	1.9	4.6	3.6	2.8	0.8	-1.4	0.6	-2.0
DAISY	25 AUG 58	825-555	2.9	0.9	2.0	0.5	-0.1	0.6	-0.5	-0.3	-0.2
DONNA	7 SEPT 60	760-635	9.9	1.2	8.7	3.5	3.8	-0.3	1.3	0.4	0.9
CARLA	6 SEPT 61	915-585	2.3	1.1	1.2	1.1	0.6	0.5	0.5	0.4	0.1
CARLA	8 SEPT 61	860-720	0.9	0.3	0.6	3.4	2.0	1.4	2.0	1.4	0.6
		AVE.	4.5	1.1	3.4	2.4	1.8	0.6	0.4	0.5	-0.1

It is evident that excess baroclinicity of quite pronounced magnitudes are continuously present at the middle and lower tropospheric levels in the inner radii.[5] Values of B_{ex} were frequently as large or larger than the $W_r S$ and B terms themselves. An assumption of frictionless, gradient wind balance would be significantly in error in these areas where large B_{ex} exists.

It is surprising that the cylindrical thermal wind relationship should fail to describe the motion of the inner areas of the hurricane by margins as large as two to four hundred per cent. The large positive B_{ex} in the lower troposphere is consistent with an excess of pressure gradient over measured winds. In an earlier study (Gray [12]), the author noted a vortex average excess of pressure over wind acceleration in a similar data sample.

Lower Tropospheric Baroclinicity and Cloud Entrainment: Although the most intense baroclinicity in the developed storm is found at levels between 500 and 200 mb, Table 2 shows that large baroclinicity is also present in the lower half of the troposphere. This is consistent with the facts that

1) nearly all condensation heating occurs below 500 mb,

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[5] Numerous reconnaissance missions at the middle and lower tropospheric levels into the inner area of typhoons have also reported sharp increases of horizontal temperature right before and during entry to the eye-wall clouds.

2) a significant percentage of shallow cumulus is also present, and

3) observed magnitude of the deep vertical draft velocities and time scale of the deep cumulus cell is possible only if significant amounts of entrainment and turbulent mixing are taking place on the cloud boundaries.

Were "undilute" cumulus ascent hypothesized for the lower part of the troposphere, then vertical draft velocity and cell lifetime would have to be two to three times larger and smaller, respectively, than are observed (Byers and Braham [4], Gray [13], Simpson, et al. [53], Senn, et al. [49]).[6] By allowing shallow cumulus convection and some entrainment and turbulent mixing of the Cbs, substantial baroclinicity may be developed in the lower atmosphere. Ackerman [1] has measured liquid water content within cumulus along a number of the flight levels here discussed and found amounts below those required for undilute ascent.

Cumulus intensity is greatest in and near the eye wall area of the hurricane. It is in this region where the direct dynamic

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[6] If virtual temperature difference between cumulus cloud and environment averaged 2° through the layer from 950 to 500 mb (moist ascent from 950 to 500 mb in the West Indies summertime sounding requires 2.8° average difference of convective and environment virtual temperature), then the velocity of the vertical draft at 500 mb and the draft lifetime would be approximately 20 m/sec and 5 minutes, respectively. Higher draft velocities would occur at 400 and 300 mb.

influences of the cumulus are more clearly evident and can be more easily isolated from the other parameters. The above envisaged cumulus momentum exchanges are also acting within other areas and circulations, but their relative importance is smaller and present instrumental limitations do not allow for accurate isolation and assessment.

The following sections of this paper will attempt to explain the excess baroclinicity as a plausible consequence of the characteristic dynamical process associated with the hurricane's cumulus momentum exchanges.

4. CHARACTER OF VERTICAL MOTION AND STRESS IN THE HURRICANE

Qualitative inspection of hurricane radar echo information has shown the heavy liquid water (and undoubtedly the primary vertical motion) to be concentrated in the selective cumulus cloud areas which may occupy but 5-20 per cent of the hurricane's inner area (Jordan, et al. [20], Malkus, et al. [29], Riehl and Malkus [42], Malkus [30], Colón [8], [9], Ackerman [1], Gentry [10], and Simpson, et al. [53]). The author has made an investigation of the width and magnitude of hurricane cumulus drafts (Gray [13]). Radar observational studies of the motion of hurricane precipitation echoes have been made by Senn and Hiser [48], Senn, et al. [49], [50], [51], Watanabe [56], and others. The above and additional observational evidence allows an inductive determination of the character of the vertical motion at the inner radii of the hurricane.

Following Riehl and Malkus [41], [42], and Malkus [30], it is concluded that the vertical motion of primary significance is occurring within cumulus which often penetrate into the upper half of the troposphere. No dynamically significant vertical motion is believed to occur in the clear air or within the layered or stratus clouds between the cumulus. The lifetime of the deep cumulus cell averages 30 minutes. This cell life-cycle is pictured as having two periods, the first of 10-15 minutes duration when the vertical motion is predominantly upward and a similar latter period when the motion is predominantly downward. This is in agreement with the cell life-cycle invisaged by Byers and Braham [4]. The magnitude and width of the up- and downdrafts are of similar magnitude. The mass circulation through the hurricane results from a more intense and slightly greater number of updrafts than downdrafts. This mean vertical motion or mass-circulation through the storm system is not representative of the average absolute vertical motion occurring within the system if typical up- and downdrafts have a magnitude of 5-10 m/sec.[7]

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[7]The downdraft is established by frictional drag of the falling, strongly concentrated, heavy raindrops. This is assumed to be possible without invoking an evaporation-cooling mechanism.

Momentum Exchanges and Stress Associated with the Cumulus Drafts: In a previous observational paper on hurricane dynamics, the author (Gray [14]) has stressed the importance of vertical momentum exchanges associated with the correlation of cumulus-scale wind components. By performing space averaging, the wind could be separated into space-determined mean and cumulus-sized eddy components. The measured cloud-scale horizontal and vertical eddies (u', v', ω') were of magnitude 5-10 m/sec and the components were often highly correlated. Sizable horizontal accelerations could result from vertical gradients of the cumulus-induced stress.

The sign of the $u'\omega'$ correlation was typically negative in the middle levels indicating upward vertical momentum transports as is required by cumulus penetration through a layer where the mean wind decreases with height. The large horizontal momentum in the lower levels is carried by the updraft to higher levels. A resulting higher horizontal wind speed is present within and surrounding the updraft. The opposite process will be in operation with the downdraft (Fig. 4). A high correlation of $v'\omega'$ should also exist in the lower layers where the vertical gradient of radial wind is large (Fig. 5). Radar evidence is available (previously cited) which shows that the "hard core" echoes often have horizontal velocities considerably different than the surrounding cumulus-free air.

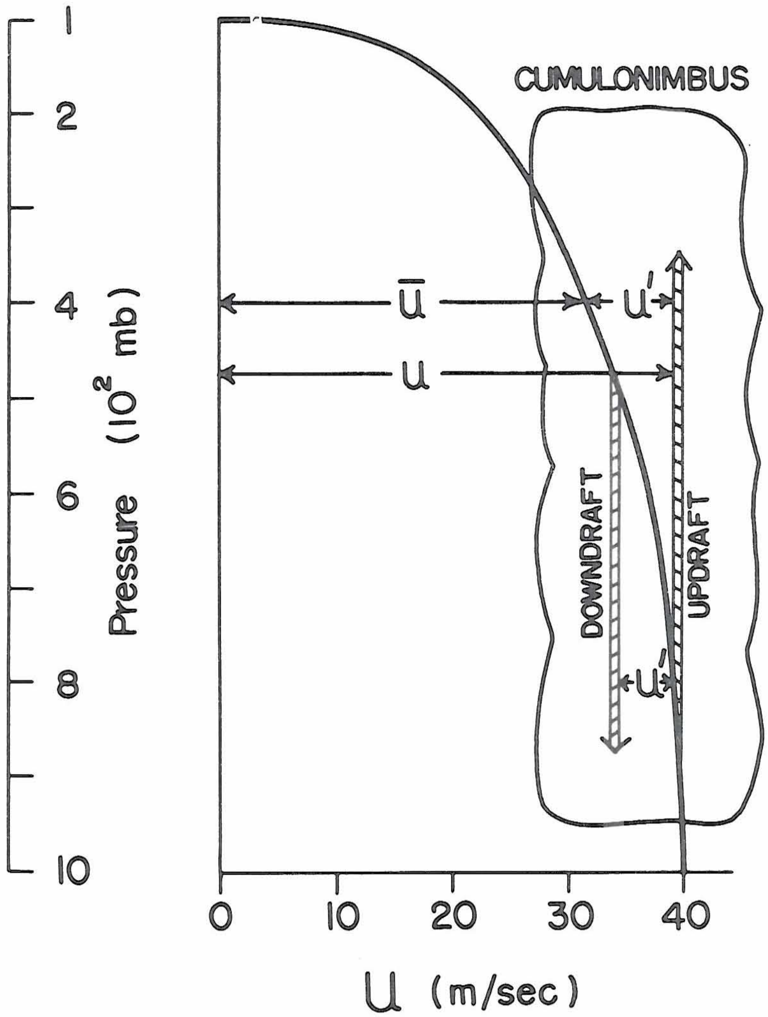

Fig. 4. Portrayal of how the horizontal velocity within the deep cumulus can be different than the velocity of the surrounding cumulus-free area if large vertical wind shear is present. The tangential velocity at any level within the draft (u) may be significantly different (u') than the horizontally averaged tangential wind ($\bar{u}$) surrounding the cumulus.

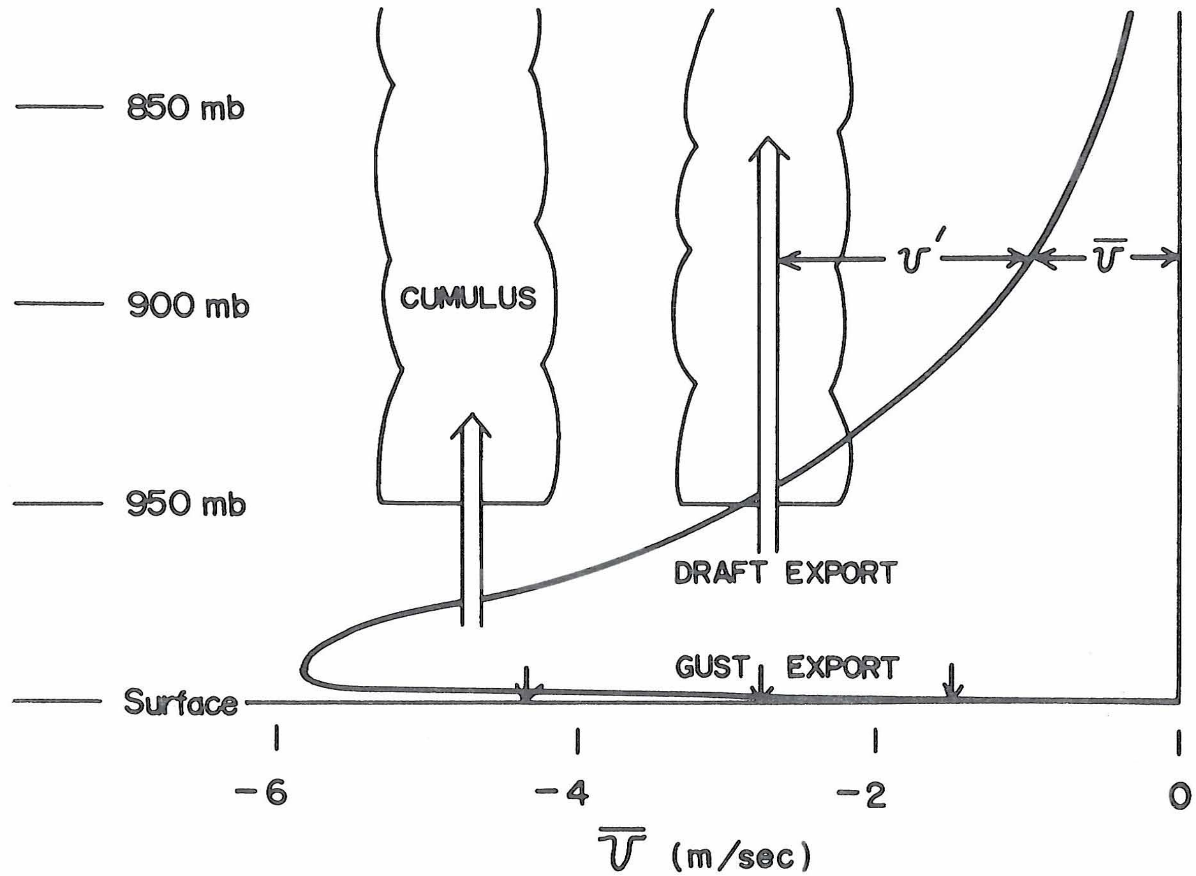

Fig. 5. Illustrating how cumulus updrafts from within the lower inflow layer can produce at higher levels significantly greater negative radial velocities (v') than the mean radial velocities ($\bar{v}$) surrounding the cumulus. Typical exports of momentum out of the inflow layer by the gust (below) and draft (above) processes are also shown.

In order to gain more insight into the internal dynamics of the inner area of the hurricane and to quantitatively explain the lack of thermal wind balance as a plausible consequence of the cumulus vertical momentum transports, a symmetric vortex flow pattern for this area based on recent flight observations will be assumed and discussed in the following three sections.

5. ASSUMED VORTEX MODEL

A symmetric steady-state vortex flow pattern is now arbitrarily specified. Upon this mean flow pattern a distribution of three sizes of cumulus convection will be superimposed. Plausible cloud-scale wind fluctuations associated with these cumulus will also be assumed. This complete specification of the mean flow and cumulus scale wind eddies will allow for calculations of the frictional and other terms in the equations of motion. The following parameters are specified to represent the conditions within the inner 80 km radii of the moderate, moving hurricane with cumulus convection concentrated in eye-wall clouds at 33 km.

a. <u>Tangential Wind in Horizontal.</u> Following Riehl [43], an

$$\bar{u}\, r^{x} = \text{constant} \tag{7}$$

relationship is taken as the functional form of the mean tangential wind ($\bar{u}$) with radius. x equals -1 for radii from 15 to 30 km, and 0.5 for radii beyond 30 km (Fig. 6). For the data sample here presented, Riehl (op. cit.) has shown that $x = 0.5$ closely fits a majority of the radial wind profiles beyond the radius of maximum winds. Maximum wind is assumed to be 40 m/sec at 30 km radius.

b. Tangential Wind in Vertical. The vertical distribution of the mean value of u is taken to be given as

$$\bar{u} = \bar{u}_o \cos^{1/3} \alpha \tag{8}$$

where $\bar{u}_o$ is the mean wind at the surface, and $\alpha = 0$ and $\pi/2$ at 1000 and 100 mb, respectively, as shown by the diagram on the left of Fig. 7. α and φ to follow are linear functions of pressure.

c. Radial Distribution of Radial Wind. Following Palmén and Riehl [38], Malkus and Riehl [28] and Miller [32], the surface stress ($\tau_{\theta p_o}$) will be assumed to vary according to the usual relationship

$$\tau_{\theta p_o} = \rho_o K \bar{u}_o^2 \tag{9}$$

where K is an arbitrarily specified exchange coefficient for gust-scale momentum transfer at the ocean surface. This constant is assigned the value 2.5×10^{-3}.

ρ_o is the surface density.

Assuming ninety per cent of the interface induced stress to dissipate in the lowest 100 mb, the average tangential frictional acceleration in the lowest 100 mb (F_θ) is assumed to be given by the relation

$$\widetilde{F}_\theta = (0.9)\, g \frac{\tau_{\theta z_o}}{\Delta p} \tag{10}$$

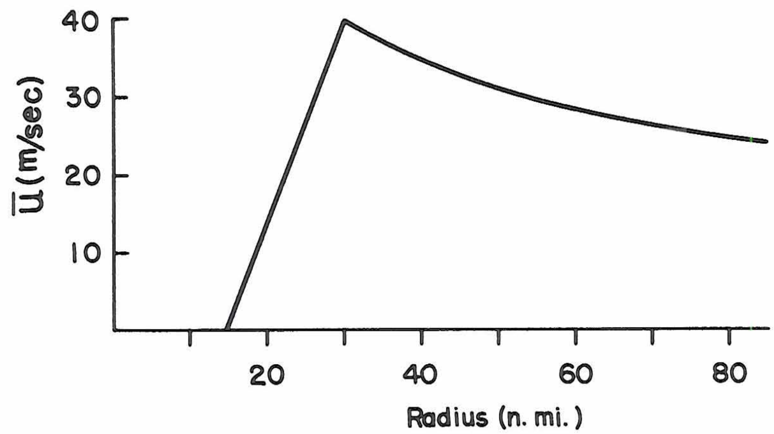

Fig. 6. Assumed radial profile of mean tangential wind ($\bar{u}$)

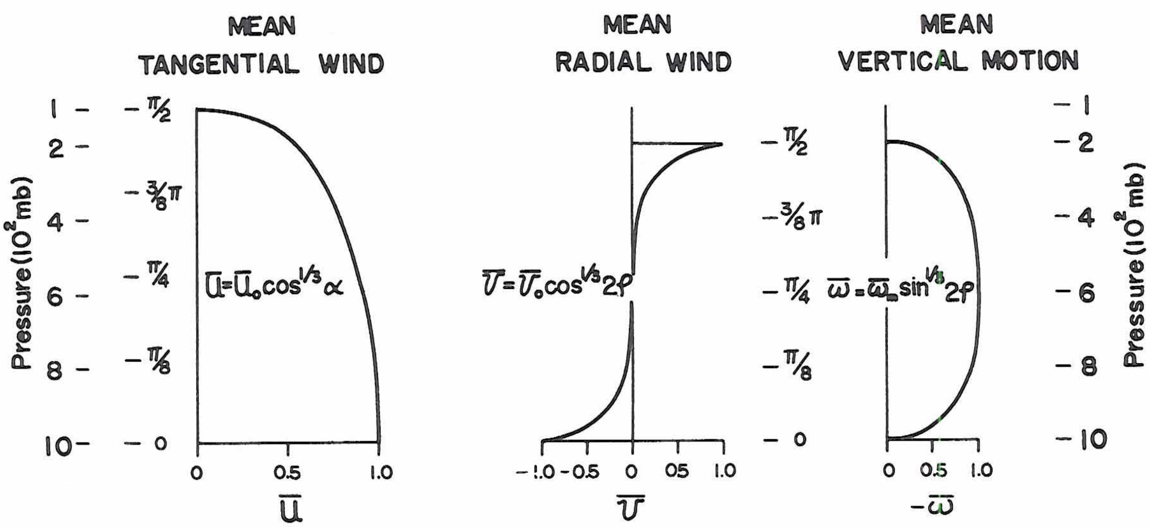

Fig. 7. Assumed vertical dependence throughout the troposphere of the mean tangential ($\bar{u}$), radial ($\bar{v}$), and vertical wind ($\bar{\omega}$).

where g is the gravitational acceleration,

Δp a 100 mb thickness, and

$\sim$ denotes vertical averaging.

Given $\tilde{F}_\theta$ the average radial velocity in the lowest 100 mb of the steady-state symmetrical hurricane is very closely determined from the cylindrical tangential equation of motion $(\underbrace{-\bar{v}\zeta_a}\underbrace{-\bar{\omega}\frac{\partial \bar{u}}{\partial p}} + \tilde{F}_\theta = 0)$ by the relationship

$$\tilde{v} \approx \left(\tilde{F}_\theta / \tilde{\zeta}_a\right) \qquad (11)$$

where $\tilde{\zeta}_a$ is the average absolute vorticity within this layer depth. The negligible vertical shear of $\bar{u}$ in the lowest 100 mb allows the $\bar{\omega}\frac{\partial \bar{u}}{\partial p}$ term to be neglected. The radial distribution of $\tilde{v}$ in the lowest 100 mb and resulting mean upward velocity ($-\bar{\omega}$) at 100 mb above the surface level is portrayed in Fig. 8.

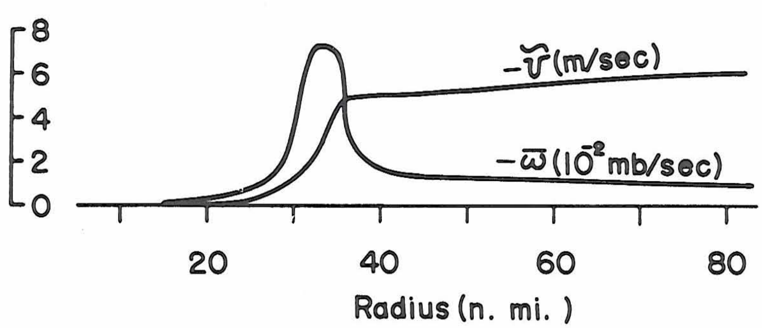

Fig. 8. Radial distribution of the average inward radial velocity ($\tilde{v}$) in the lowest 100 mb and upward vertical velocity at 900 mb ($-\bar{\omega}$) which results from tangential wind and stress assumptions.

d. Radial Wind in Vertical. The mean radial wind is assumed to vary with height at each radius as shown by the center diagram in Fig. 7. The functional form of this is given by

$$\bar{v} = \bar{v}_o \cos^{1/3} 2\varphi \qquad (12)$$

with $\varphi = 0$ and $\pi/2$ at 1000 and 200 mb.

$\bar{v}_o$ = the mean surface radial wind and was determined from frictional drag considerations in "c" above.

e. Vertical Wind. The mean vertical motion at each radii is assumed to vary as

$$\bar{\omega} = \bar{\omega}_m \sin^{1/3} 2\varphi \qquad (13)$$

as shown by the diagram on the right in Fig. 7. ω_m is the mean vertical velocity at 600 mb at any radius, $\varphi = 0$ and $\pi/2$ at 1000 and 200 mb, respectively.

f. Cloud Distributions and Up- and Downdrafts. A variety of cumulus cloud sizes exists in the tropical storm. It is difficult to deal with all of these. For simplicity it will be assumed that there are three typical cumulus sizes each existing in equal percentage area amounts. The effect of layered or stratus type clouds will be assumed to be negligible. The vertical motion within the cumulus will be given by the basic equation

$$\omega = \omega_o \sin^{1/3} 2\varphi \qquad (14)$$

The three types of assumed cumulus are

1) Small cumulus (Cu) which extend from 950 to 700 mb. Maximum upward velocity (at 825 mb) is taken to be $0.5\,\omega_o$, no downdrafts present. $\varphi = 0$ and $\pi/2$ at 1000 and 700 mb.

2) Towering cumulus (TWG Cu) which extend from 950 to 500 mb, maximum upward velocity is $1.0\,\omega_o$ and downdrafts $0.5\,\omega_o$. $\varphi = 0$, $\pi/2$ at 1000 and 500 mb, respectively, for the updraft and 0 and $\pi/2$ at 1000 and 700 mb for the towering cumulus downdrafts.

3) Cumulonimbus (Cb) which extend to the 200 mb level. Maximum updrafts and downdrafts are 2.0 and $1.0\,\omega_o$, respectively, where $\varphi = 0$ and $\pi/2$ at 1000 and 200 mb for the updraft and 0, $\pi/2$ at 1000 and 500 mb for the downdraft.

The magnitude of ω_o is specified such that the maximum velocities of the Cb updrafts, Cb downdrafts and TWG Cu updrafts, and Cu updrafts are always 8, 4 and 2 m/sec, respectively, or 4.8, 3.4 and 1.7 m/sec at 900 mb. The upward circulation through the vortex is carried by the Cb updrafts. The Cb downdrafts are balanced by the TWG Cu updrafts, the TWG Cu downdrafts by the small Cu updrafts.

Figure 9 portrays the relative magnitude of these five classes of cumulus up- and downdraft patterns. Figure 10 shows the percentage area of Cb updrafts and of each of the other types of cumulus up- and downdraft as a function of radius and also the total percentage area taken up by all five classes of updraft and downdraft. The magnitude of the draft velocities and mean vertical circulation determine these area percentages.

The above fits basic observational flight evidence of the characteristics of the typical moderate hurricane circulation which is moving with the cumulus convection concentrated in its center. Mass balance is present at each radii. The cumulonimbi extend to the upper troposphere and the mass inflow and outflow circulation is concentrated in the lowest 100 mb and 200 mb, respectively.

6. ASSUMED CUMULUS INDUCED WIND EDDIES

The above specification of cumulus up- and downdrafts which are superimposed upon the mean flow conditions (in which vertical shear is present) must of necessity require that the winds within the cumulus be different than the mean winds surrounding the clouds. As shown in Figs. 4 and 5, the horizontal momentum in the lowest layers can be advected upward within the cumulus updraft to produce a component wind at a higher level which is significantly

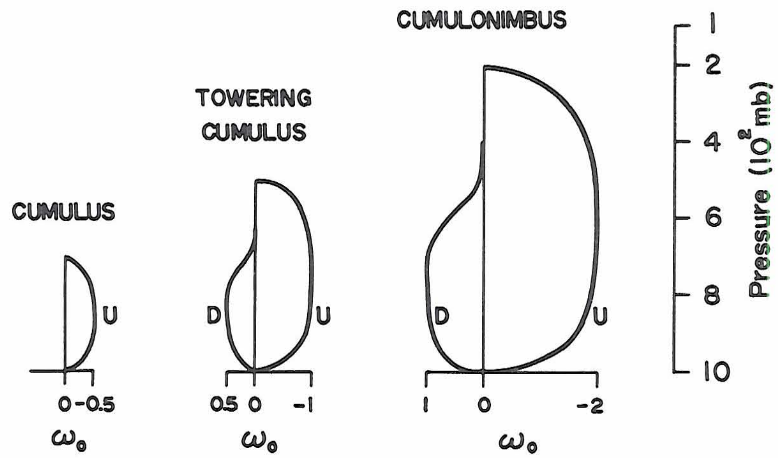

Fig. 9. Portrayal of the relative magnitude and depth of the five classes of assumed cumulus Updraft (U) and Downdrafts (D).

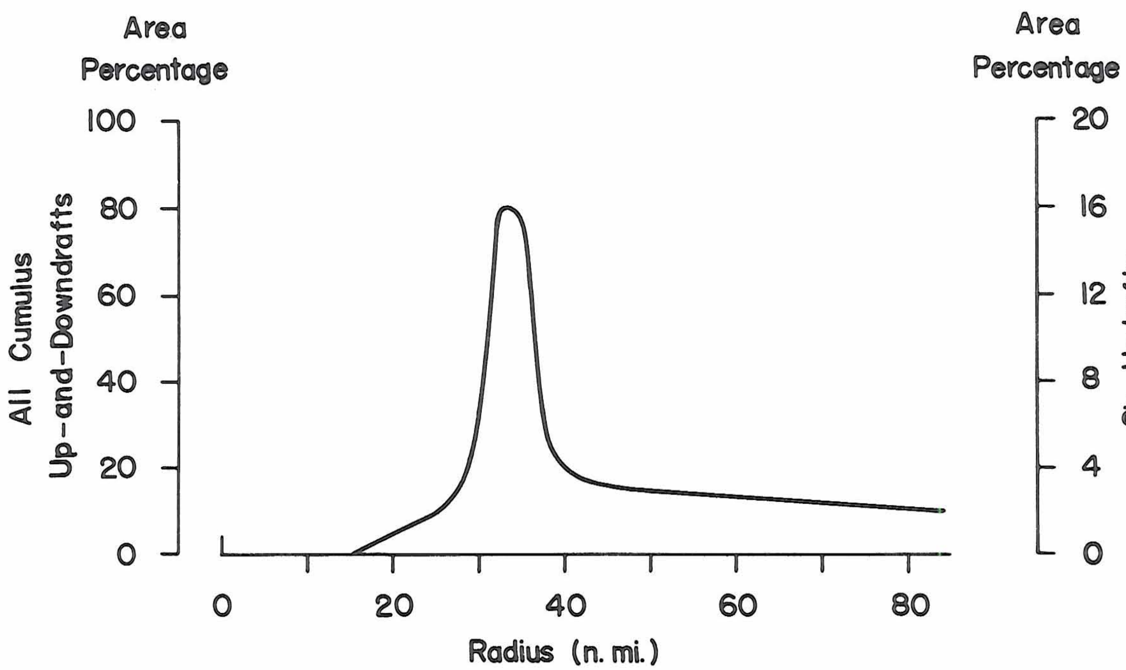

Fig. 10. Radial distribution of the percentage of area at each radius which is occupied by <u>each</u> class of cumulus up- and downdrafts (scale on right) and the total percentage of area which is occupied by all the five classes of cumulus up- and downdrafts (scale on left).

different than the mean wind component surrounding the cumulus. These cumulus-induced component wind variations will henceforth be called eddy winds.

Vertical Eddy Wind (ω'): With the previously specified functional form of ω with height, the magnitude of the typical vertical eddy wind associated with each cumulus up- and downdraft is portrayed in Fig. 11 and defined as

$$\omega' = \omega - \bar{\omega} \tag{15}$$

where $\bar{\omega}$ and ω have been previously specified in Sections "5 e" and "5 f".

The mean upward vertical velocity ($-\bar{\omega}$) is accounted for by the velocity of the Cb updrafts multiplied by the area made up by the Cbs (average approximately $6\frac{1}{2}$ per cent between 30-50 km radius). The mean vertical circulation through the system is thus but a small fraction of a m/sec and the vertical eddy wind is closely approximated by the velocity within the cumulus ($\omega' \sim \omega$).

Tangential Eddy Wind (u'): The tangential eddy wind is produced by the cumulus advecting larger tangential momentum from below in the updraft and smaller tangential momentum from above in the downdraft (Fig. 4). The Cb updrafts are most effective in this transfer because they penetrate through the greatest vertical shear. At the upper levels the difference between the tangential wind and the surrounding flow can be appreciable. At these levels

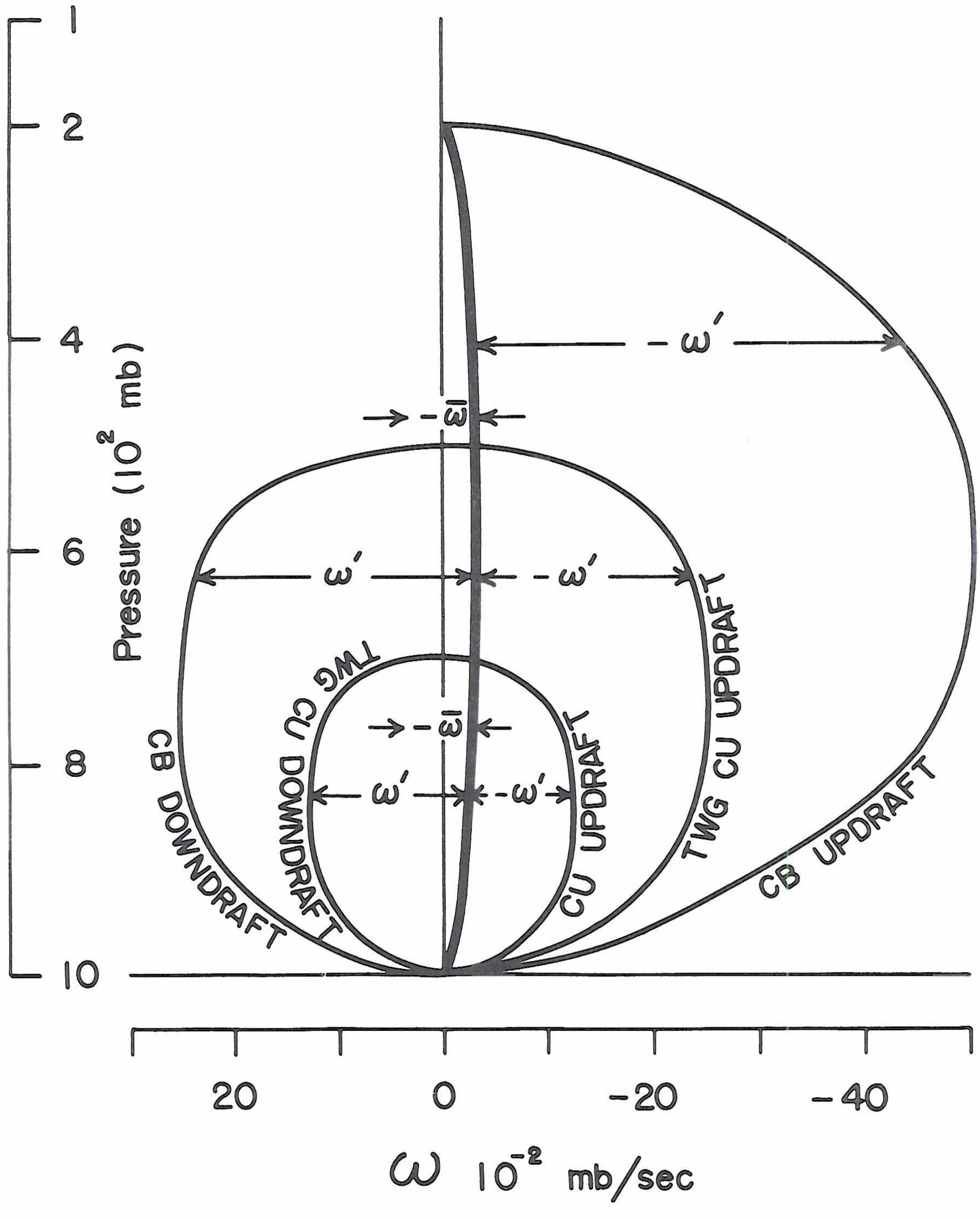

Fig. 11. Portrayal of the relative magnitude of the five assumed classes of cumulus up- and downdrafts for average conditions between 30 and 50 km radius and how each updraft and downdraft can be significantly different ($\pm\,\omega'$) than the mean vertical motion ($-\bar{\omega}$) through the system. [Cumulonimbus (Cb), Towering Cumulus (TWG Cu), and Cumulus (Cu).]

large horizontal mixing must be occurring. The tangential eddy is arbitrarily specified as

$$u' = \bar{u}_o (1 - \cos^{1/3} \alpha)\, M_\theta \tag{16}$$

where $\bar{u}_o (1 - \cos^{1/3} \alpha)$ = the maximum eddy wind which would result from undilute ascent from the surface. $\alpha = 0$ and $\pi/2$ at 1000 and 100 mb.

M_θ = coefficient of mixing which is taken to be proportional to $\cos^{1/3} \wp$. The greatest mixing is thus assumed to occur at the upper levels of each cloud.

Then $u' = \bar{u}_o (1 - \cos^{1/3} \alpha) \cos^{1/3} \wp$, where $\wp = 0$, and $\pi/2$ at 1000 and 200 mb in the Cb updrafts; 1000 and 500 mb in the Cb downdrafts and TWG Cu updrafts, and 1000 and 700 mb in the TWG Cu downdrafts and Cu updrafts.

A vertical portrayal of the tangential eddy wind which results from a Cb updraft is shown in the upper left diagram of Fig. 13. Smaller horizontal eddies would occur for the four other classes of cumulus up- and downdrafts.

<u>Radial Eddy Wind:</u> The radial eddy is produced in a similar manner as the tangential eddy. However, the larger radial than tangential wind shear in the lower levels requires that the radial eddy (v') and the radial mixing be larger at these levels. The radial eddy wind is arbitrarily specified for levels above 950 mb as

$$v' = (v_i - \bar{v})\, M_r \tag{17}$$

where v_i = the average radial wind in the lowest 50 mb,

$\bar{v}$ = the mean radial wind at each level,

M_r = radial mixing which is assumed to be proportional to $\cos \wp$ instead of $\cos^{1/3} \wp$ as with the tangential eddy,

$\wp$ = similar values for the five classes of cumulus up- and downdrafts as were specified for the tangential eddy wind.

Figure 12 and the lower left-hand diagram of Fig. 13 portray the magnitude of radial eddy wind which would result from a cumulonimbus updraft through application of (17).

7. RESULTING CUMULUS-INDUCED STRESS AND HORIZONTAL ACCELERATION OF ASSUMED VORTEX

With the above specified eddy winds significantly large middle-level Reynolds stress values

$$\tau_{\theta p} = \frac{\widetilde{u'\omega'}}{g} \quad , \quad \tau_{rp} = \frac{\widetilde{v'\omega'}}{g} \tag{18}$$

where $\widetilde{\quad}$ denotes horizontal space averaging over distances considerably larger than the eddy

are present when the eddy wind components are appreciably correlated.[8] These stress magnitudes may be as large as the surface interface stress. Figure 13 portrays the specified vertical

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[8]See Gray [14] for a discussion of the treatment of cumulus-scale wind fluctuations as wind eddies from the Reynolds stress point of view.

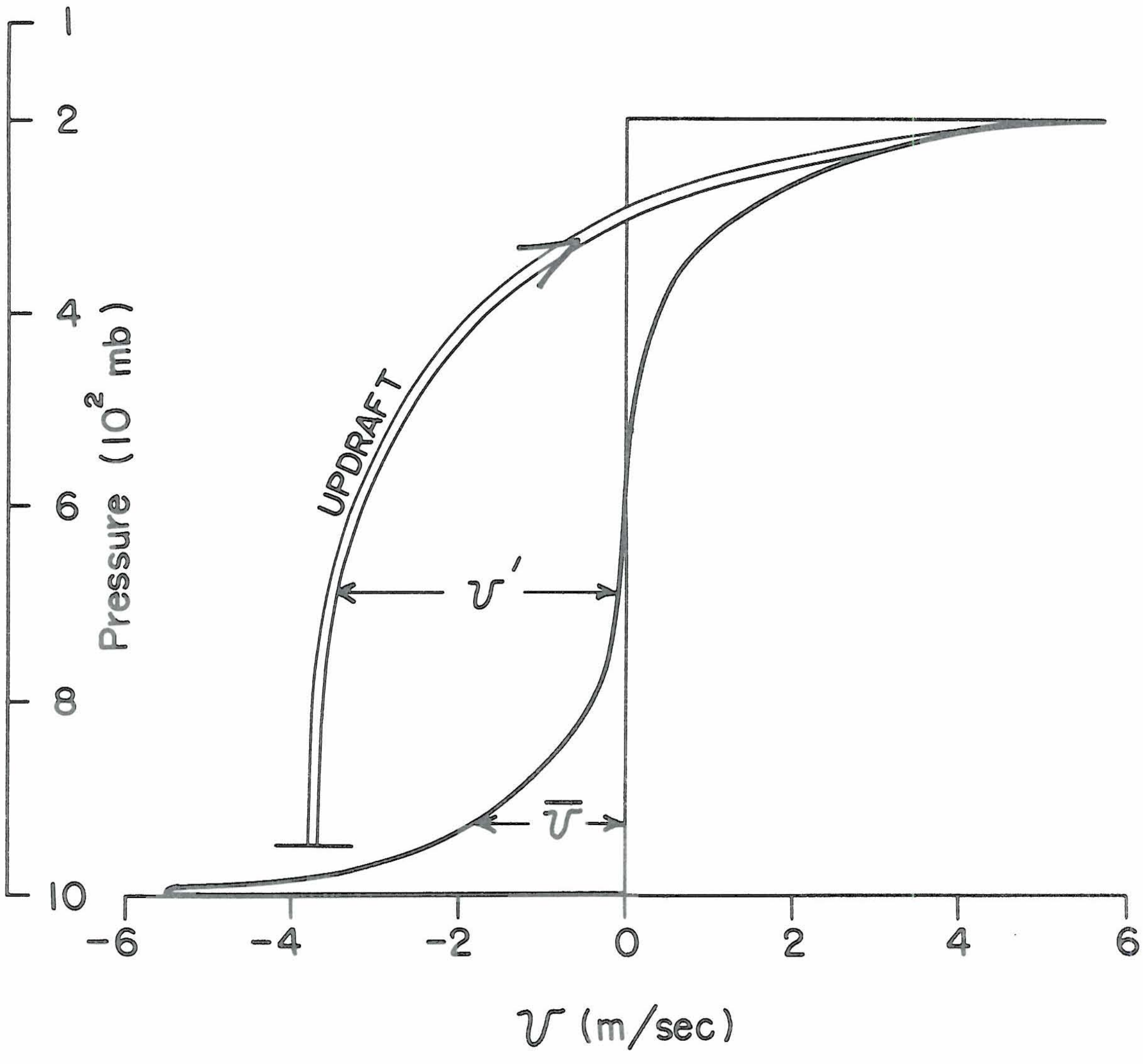

Fig. 12. Vertical distribution of the assumed radial eddy wind (v') which results from a cumulonimbux updraft which penetrates to 200 mb. Radial eddy wind $v' = (v_i - \bar{v}) \cos \varphi$.

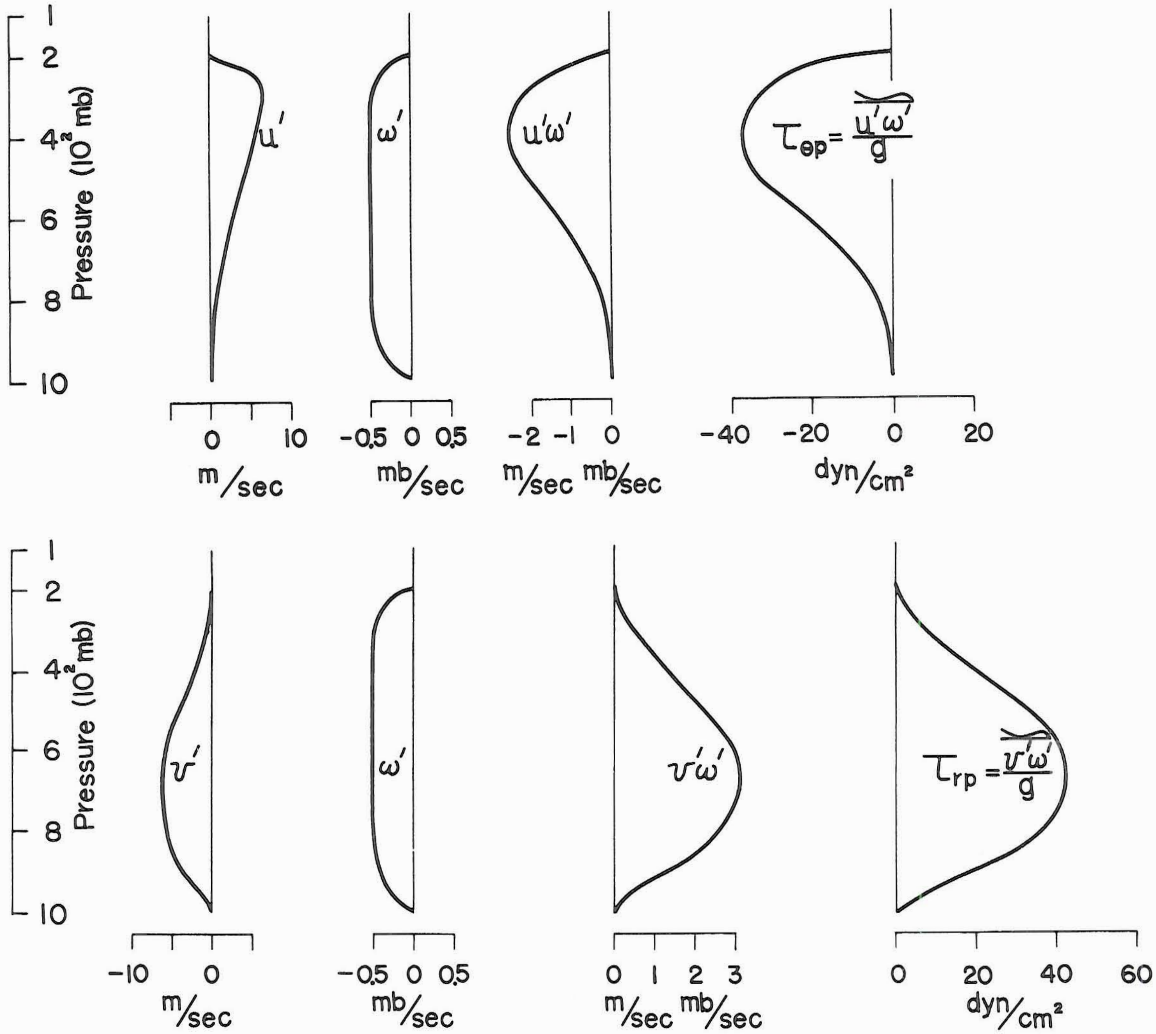

Fig. 13. Tropospheric distribution at 35 km radius of the magnitude of the resulting assumed tangential (u'), radial (v'), and vertical (ω') wind eddies, their product and consequent $\tau_{\theta p}$ and τ_{rp} stress which results from just the cumulonimbus updrafts if their components were correlated within the 15 per cent area occupied by these updrafts.

distribution u' , v' , and ω' wind eddies from the cumulonimbi updrafts at 35 km radius and the resulting Reynolds stress which would result from the correlation of the horizontal and vertical eddies within 15 per cent of the area which is occupied by Cb updrafts at the eye wall.

It is assumed in this model that in the first approximation to the turbulent stress that there is complete correlation of wind eddies within the restricted areas of the cumulus elements and no correlation of wind eddies within the much larger percentage areas with no cumulus. This is a different type of turbulence than the typical mechanically produced gust-turbulence of the planetary boundary layer. The size of the typical eddies in this kind of cumulus produced turbulent are much larger and are driven by the condensation developed buoyancy within the selective cumulus draft. The author has, in fact, measured this latter type of turbulence from the U. S Weather Bureau's specially instrumented aircraft observations in the hurricane. Draft-scale wind components often showed large correlation at selective regions within and surrounding the cumulus. Middle tropospheric stress values as high as 20-30 dynes/cm^2 were computed from the high correlation and magnitude of the draft-scale eddies within and surrounding the selective regions occupied by the cumulus. This scale of eddy was not observed in the cumulus-free areas. The reader is referred to the author's other paper [14] for a discussion of this latter type of turbulence.

$\underline{\tau_{\theta p}\text{ Stress:}}$ The three center diagrams of Fig. 14 portray the vertical distribution of the average $\tau_{\theta p}$ stress between 30-50 km which results from the up- and downdrafts within the three assumed classes of cumulus. The drawing on the left of Fig. 14 portrays the assumed vertical decrease of surface stress ($\tau_{\theta p}$). The drawing on the right of Fig. 14 shows the average stress between 30-50 km which results from the combined surface and all three sizes of cumulus. In the surface boundary layer tangential momentum is being transferred downward by gust-scale eddies and $\omega'u'$ is positive. Malkus and Riehl [28], Riehl and Malkus [42], and Miller [32] have estimated hurricane boundary layer stress values of 40-60 dyn cm^{-2} at inner radii. In the inflow layer where momentum is being transferred to the ocean, the eddy sizes are characteristically of gust length ($\sim$100 to 400 m). Negative correlation between the vertical and horizontal gust eddies is required for momentum transfer to the ocean surface. The few B-50 flights at 450 meters height that have been made in strong wind conditions have usually encountered a mechanical "washboard" type turbulence. U. S. Navy flights at sub-cloud levels in Typhoons have also reported this type of turbulence. Beginning at cloud base (400 to 600 m) and extending into the upper troposphere, a characteristically different type of "cumulus

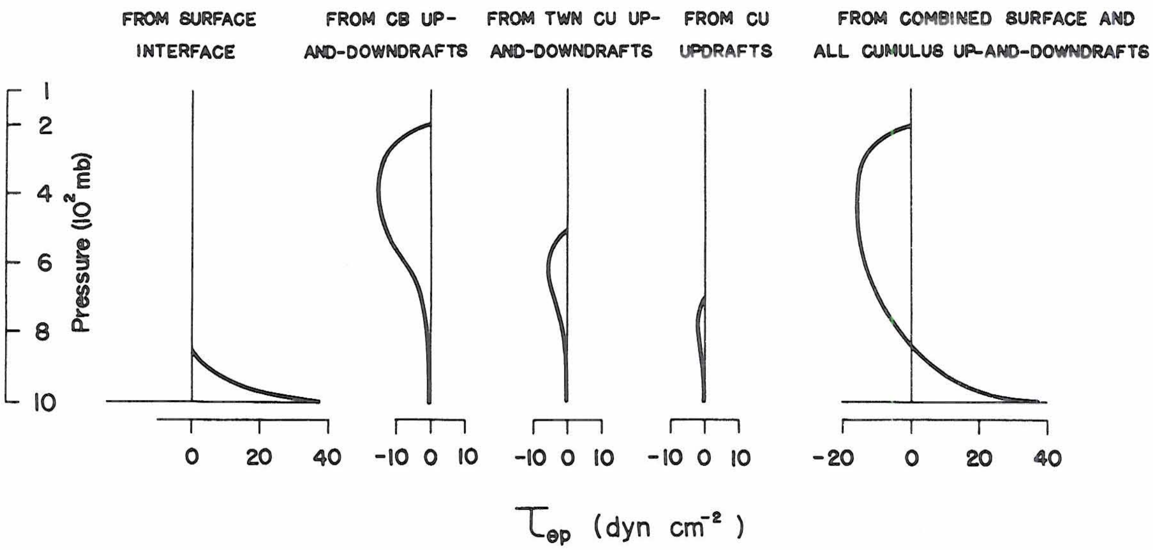

Fig. 14. Tropospheric distribution of average $\tau_{\theta p}$ stress between radii 30-50 km (right diagram) which results from the combined surface interface stress (left diagram) and the three sizes of cumulus (center diagrams).

scale" eddy wind turbulence is established. In this case momentum is being transferred upward by the positive correlation of cloud scale tangential and vertical eddies. These eddies are typically of much larger size than the eddies within the surface inflow layer.

Tangential Acceleration (F_θ): The left diagram of Fig. 15 shows the vertical distribution of the average combined $\tau_{\theta p}$ stress within the 30-50 km radius (a) and the $\tau_{\theta p}$ stress which occurs at the eye wall at 35 km (b). A larger magnitude of stress exists at the eye wall where the intensity of cumulus is greatest. The right diagram shows the tangential frictional acceleration (F_θ) which results between and at these same radii from the vertical gradient of this combined and magnitude to offer a plausible explanation for the large observed B_{ex}. It is thus concluded that the $\frac{\partial F_r}{\partial p}$ term of (3) is always of very large magnitude at the inner areas of the typical hurricane and can, in fact, account for most of the continuous steady-state imbalance of wind and pressure acceleration. The magnitude of the $\frac{\partial}{\partial p}\left(\frac{\partial \bar{u}}{\partial t}\right)$ term of (3) must be small in comparison of $\frac{\partial F_r}{\partial p}$ if unreasonably large growth rates are not to occur. The magnitude of the $\frac{\partial}{\partial p}\left(\bar{v}\frac{\partial \bar{v}}{\partial r}\right)$ and $\frac{\partial}{\partial p}\left(\bar{\omega}\frac{\partial \bar{v}}{\partial p}\right)$ terms of (3) are also of smaller magnitude above the surface inflow layer ($p < 850$ mb).

Given the lower stratospheric boundary condition of no pressure and wind gradient, a downward integration of $\frac{\partial F_r}{\partial p}$ through the troposphere produces the largest B_{ex} at the surface as shown in Fig. 18. This is produced primarily by the imbalance of the thermal-wind equation. Negative B_{ex} and stronger wind than pressure acceleration is hypothesized to be present above 450 mb. Below 450 mb, B_{ex} is positive and the pressure acceleration is stronger than wind acceleration.

A much larger pressure than wind acceleration is required in the lowest layer to maintain a continual inward acceleration of the surface wind in order to balance the continual drain of inward-directed radial momentum to the ocean surface below and by the cumulus above (Fig. 5). An excess of heating and pressure acceleration over wind acceleration must always be present within the steady-state storm to

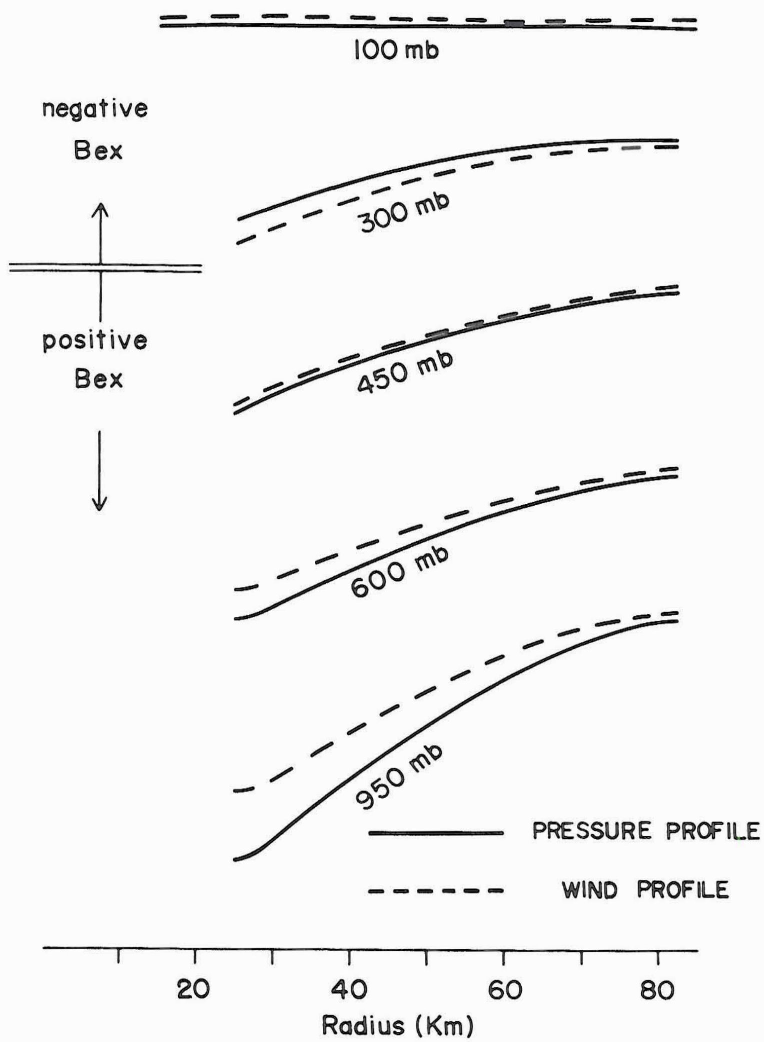

Fig. 18. Comparison of the unbalanced pressure and wind (converted into equivalent pressure gradient) profiles at various levels in the assumed steady-state vortex.

maintain this inflow. The observations of Graham and Hudson [11] and Schauss [46] have shown sizably larger cyclostrophic (or pressure gradient) than observed wind at the surface levels of the hurricane. In an earlier study the author found an excess of vortex averaged pressure over wind acceleration (Gray [12]). It is to be noted that the largest values of $\frac{\partial F_r}{\partial p}$ are found at the lowest level.

The cumulus produce the larger pressure gradients through condensation heating and also produce the mechanism whereby compensating radial accelerations are inhibited. In this way a steady-state imbalance of wind and pressure gradient can be maintained. The cumulus-induced radial accelerations act to inhibit compensating inward radial cross-isobaric motion which would generate large amounts of kinetic energy.

In the tangential direction the cumulus momentum exchanges also act to inhibit a larger generation of wind at the lowest levels. It is then possible for the inward directed cross-isobaric flow, both in the steady-state and intensifying vortex case, to be larger at the lowest levels, yet the vertical gradient of horizontal wind to remain small.

<u>Steady-State Tangential Equation of Motion:</u> As a check on the balance of accelerations and the significance of the cumulus induced accelerations, an examination was made of the terms in the symmetrical steady-state tangential equation of motion. This equation may be written as

$$0 = -\bar{v}\zeta_a - \bar{\omega}\frac{\partial \bar{u}}{\partial p} + F_\theta \tag{19}$$

Figure 19 shows the 30-50 km radial average of the vertical variation of these three terms for conditions of the assumed vortex with the superimposed cumulus convection. It is again seen that a sizable magnitude of F_θ must prevail through the middle and upper troposphere in order that balance prevail. It is impossible to prescribe a steady vortex with radial inflow near the surface and outflow concentrated at 200 mb without simultaneously prescribing a vertical distribution of frictional acceleration similar to Fig. 19 (with small $\bar{\omega}\frac{\partial \bar{u}}{\partial p}$). The magnitude and density of cumulus convection chosen was such as to just allow for balanced conditions. A balance of momentum is also presecribed by conditions of (19).

The magnitude of the cumulus downdrafts has been conservatively chosen to be but one-half of the updrafts and to exist only within the lower-half of the troposphere. With this specification the relative magnitude of $\left| F_\theta \right|$ to $\left| \bar{\omega}\frac{\partial \bar{u}}{\partial p} \right|$ through the troposphere is approximately three to one. A smaller updraft to downdraft ratio would make the above ratio considerably larger.

8. TANGENTIAL EQUATION FOR INTENSIFYING VORTEX

The symmetric cylindrical tangential equation for intensification is written as

$$\frac{\partial u}{\partial t} = -v\zeta_a - \omega\frac{\partial u}{\partial p} + F_\theta \tag{20}$$

Differentiating this partially with respect to pressure, we obtain

$$\frac{\partial}{\partial t}\left(\frac{\partial u}{\partial p}\right) = -\frac{\partial}{\partial p}\left(v\zeta_a\right) - \frac{\partial}{\partial p}\left(\omega\frac{\partial u}{\partial p}\right) + \frac{\partial F_\theta}{\partial p} \tag{21}$$

In the intensifying vortex there is little change of vertical shear in the lower half of the troposphere, and $\frac{\partial}{\partial t}\left(\frac{\partial u}{\partial p}\right) \sim 0$. The pressure gradient and inward radial velocity must be larger in the lower levels of the intensifying vortex. The vertical distribution of ζ_a and v requires that $-\frac{\partial}{\partial p}\left(v\zeta_a\right)$ be positive. $\frac{\partial}{\partial t}\left(\frac{\partial u}{\partial p}\right)$ can remain small only by $-\frac{\partial}{\partial p}\left(\omega\frac{\partial u}{\partial p}\right)$ and $\frac{\partial F_\theta}{\partial p}$ contributing in a negative sense to balance $-\frac{\partial}{\partial p}\left(v\zeta_a\right)$. In this way it is possible for the absolute vorticity and vertical gradient of radial wind to increase in the lower half of the troposphere, but the vertical gradient of tangential speed to remain small. It is to be noted (Fig. 19) that in the lower half of the troposphere the $\frac{\partial F_\theta}{\partial p}$ term resulting from the cumulus up- and downdrafts is of much larger magnitude than the $\omega\frac{\partial u}{\partial p}$ term.

9. OTHER CONSIDERATIONS

With charcteristic cumulus induced stress as previously discussed, resulting values of vertical and horizontal kinematic eddy viscosity can range as high as 10^8 to 10^9 cm^2/sec. Figure 20 portrays the vertical distribution of kinematic viscosity (ν) coefficients (in units of $10^7 cm^2/sec$) which are defined as

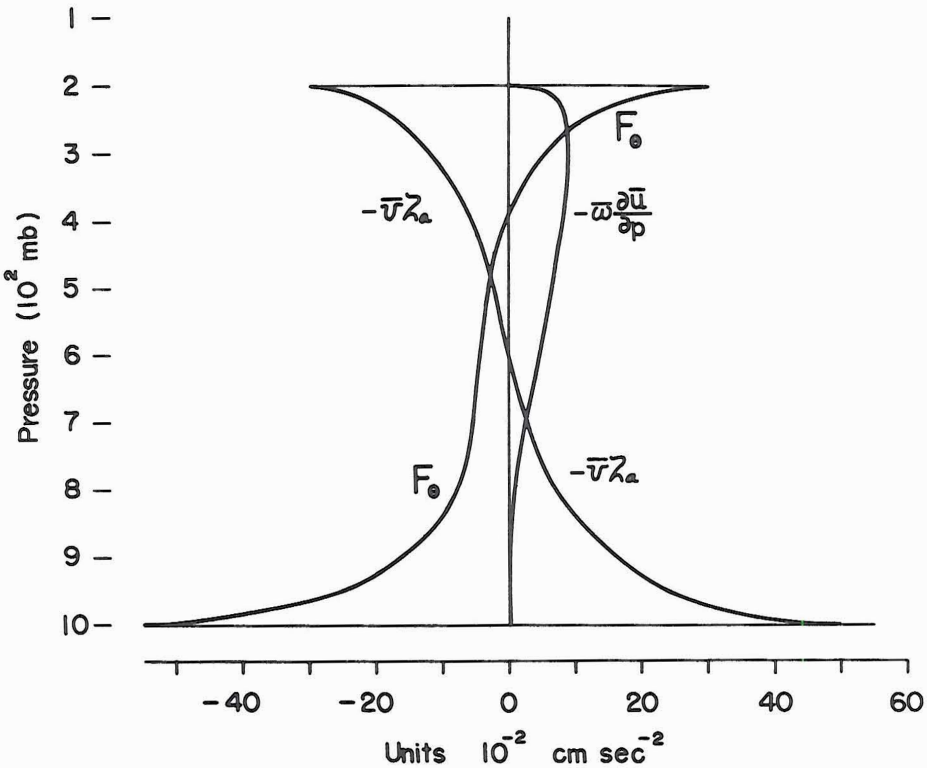

Fig. 19. Tropospheric distribution of the magnitude of the three terms in the steady-state tangential equation of motion which have been averaged between radii 30 and 50 km.

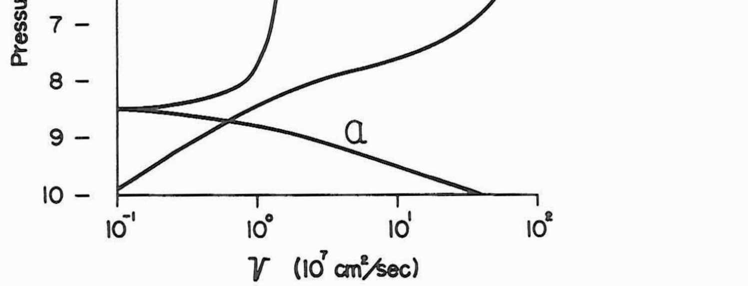

Fig. 20. Average vertical kinematic eddy viscosity coefficients (ν) between 30-50 km radius which would result from stress and mean wind shears of assumed vortex. Units 10^7cm^2/sec.

$$\nu\,(\text{vertical}) = \left|\frac{\tau_{\theta p}}{\rho^2 g \dfrac{\partial \bar{u}}{\partial p}}\right| , \left|\frac{\tau_{rp}}{\rho^2 g \dfrac{\partial \bar{v}}{\partial p}}\right| \qquad (22)$$

and are required with the various calculated values of stress and mean wind shear. A wide range and large magnitude of kinematic viscosity is required to satisfy conditions of the assumed vortex. This is particularly true at middle tropospheric levels in the cases where τ_{rp} is large and the shear of the mean radial wind small. The magnitude of kinematic viscosity can range from one to three orders of magnitude greater than the values which have been generally used in the previously cited numerical experiments. It would appear that the usually assumed mixing length theory which equates the magnitude of stress proportional to the mean shear (although relatively valid for the usual mechanical gust-scale mixing occurring in the atmosphere's surface layers) is not valid for fluid mixing accomplished by cumulus clouds where thermodynamic changes also occur. Attempts to account for the turbulent mixing by the cumulus in terms of mean shearing conditions appears not to be realistically valid.

Eddy momentum exchange coefficients characteristic of these higher magnitudes might shift the most preferred growth scale of cumulus-induced heating away from the cumulus to the more realistic meso-scale and offer a physical explanation of how the cumulus clouds on a scale of 1-10 km can generate a system which is two orders of magnitude larger. The buoyancy-produced draft scale eddy with a lifetime of but 10-15 minutes rapidly generates smaller turbulence eddies on its boundaries. These smaller gust scale eddies act to diffuse the heat and momentum of the draft eddy away from the cumulus to the larger system. This heat and momentum diffusion is viewed as occurring at a faster rate than the cumulus induced pressure gradients from the condensation heat are able to concentrate a separate cumulus-scale velocity system of their own. But the condensation heat is not lost. It remains within the larger system to maintain and intensify it. The momentum exchanges, on the other hand, do not accumulate and vanish or remain in direct relation to the continuance of the cumulus intensity and shearing flow.

10. SUMMARY DISCUSSION

The vertical momentum transports associated with cumulus up- and downdrafts offer a physical explanation for the observed two to three hundred per cent excess baroclinicity in the inner areas of the hurricane. These transports allow a steady-state vortex imbalance of wind and pressure acceleration. They allow maintenance of a vortex where the in- and outflow is predominantly concentrated in the lowest and upper troposphere. They offer a mechanism whereby the upper and lower tropospheric circulations can be coupled, and explain how vertical wind shear can remain small as baroclinicity increases. They further open up the possibility that cumulus momentum exchanges can act as a basic mechanism for the generation of tropical easterly waves and other phenomena.

The alteration of wind and pressure in response to cumulus condensation need not take place simultaneously. Condensation heat must initially go directly to lower the pressure surfaces (with assumed upper boundary) and increase the horizontal pressure and temperature gradients. Very little is known about the resulting time dependent adjustment of the wind changes in the real atmosphere. Adjustment solutions and discussions (Rossby [45], Cahn [6], Bolin [3], and Washington [55]) have been made only under highly idealized conditions. Numerical experiments should be undertaken to learn more of the details of the adjustment processes in the cumulus atmosphere. It is further suggested that future numerical experiments which incorporate cumulus convection also include the effects of cloud-scale momentum transports in addition to the effects of condensation heat.

It is important to appreciate the physical significance of cumulus-scale vertical momentum transports and of the dual or "paradox" role which the cumulus can play. It is hypothesized that if the cumulus induced baroclinicity in the lower half of the troposphere were to act in the usual sense to increase the wind shear, the development of the tropical wave and vortex could not proceed. Increased vertical wind shear would create an environment less conducive to further cumulus development (Asai [2], Hitschfeld [18], Malkus and Scorer [27]). The increased vertical shear would prevent concentration of temperature.

The above ideas are a further extention of the so-called "hot tower" hypothesis of Riehl and Malkus [30], [41], and [42], who emphasized the importance of selective vertical transport of heat within the cumulus up- and downdraft. In addition to the transport of heat, this paper has emphasized the fundamental importance of vertical transport of momentum by the cumulus.

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APPENDIX

Accuracy of Data. For information concerning the accuracy of the flight observations the reader is referred to the previously cited references on the Research Flight Facility instrumentation. Strong evidence points to the basic reliability of the instruments, especially to measured parameters which have been averaged over time and space such as the 18 km vortex averages of wind and temperature. Navigation corrections after a number of hours of flight with the measured winds were usually within a few km. When averaging was accomplished over the radial intervals at two levels, there was usually close hydrostatic consistancy between the measured temperature gradients and the gradients of layer thickness. Although some observational and computational shortcomings may be present, it is not thought possible that they could be of an extent as to alter the conclusions or implications which are advanced. Choice of a larger or smaller smoothing interval for determining the radial temperature gradients would likewise not significantly affect the results.

Correction of Winds Computed from Doppler Navigation System. The Doppler Navigation system allows computation of wind velocity from the vector difference of the True Airspeed and the aircraft motion relative to the ocean surface. It is reasonable that small percentage errors might occur in the wind determination if wind-driven ocean particle motion is present. Because of the difficulties of precise testing of this effect in hurricanes, no determination has yet been made of the small percentage correction which should be applied to the winds.

Quantitative information concerning this likely effect has, however, been gathered by a British Royal Air Force group (Grocott [15]) in wind conditions up to 30 m/sec over the North Atlantic.

Individual surface particles which reflect the electromagnetic energy for the Doppler determination do not move at the same speed as surface waves. They describe circles whose radii decrease with increasing depth. Theoretically determined curves are available which relate the particle speed to wave-length, wave-height, and wave-velocity. Observational relations of wave velocity with wind speed are also available. Combining these two relationships, Grocott, op. cit., presents an equation relating surface wind velocity and surface ocean particle velocity, thus

$$W = 0.92\,S^{1/3}$$

where W = water particle motion in m/sec, and

S = Doppler measured wind speed in m/sec.

Following Grocott this relationship is shown as the theoretical curve of Fig. 21 which has been extended up to velocities of 60 m/sec. Grocott also presents observationally determined water motion from a Doppler system which was obtained from flight navigation corrections.

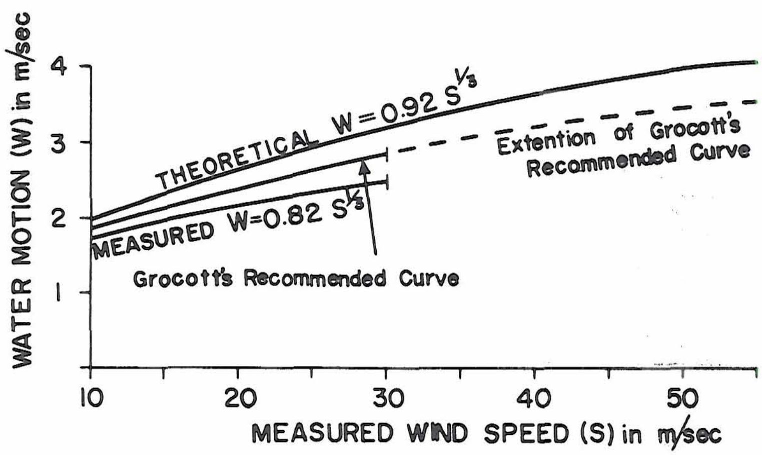

Fig. 21. Regression curve for Doppler determined wind speed vs. water motion for theoretical and for Grocott's [15] measured and recommended values.

For wind velocities up to 30 m/sec he found the relationship $W = 0.82\,S^{1/3}$. Considering all aspects, Grocott recommends use of a curve which is intermediate between measured and theoretically determined values ($W = 0.87\,S^{1/3}$). Grocott's recommended curve has been extended for this study to include velocities above 30 m/sec although it is uncertain that the functional dependence may not be altered at the higher velocities in tropical storms.

The Doppler Correction of $W = 0.87\,S^{1/3}$ requires an increase of the measured RFF winds in the data presented from seven to nine per cent. It seems reasonable that this effect would also occur with hurricane winds. To obtain the best possible winds it was deemed advisable to apply this correction. The results and implications of this paper are, however, in no significant way dependent on this correction being applied. The observed thermal wind imbalance of two to three hundred per cent in the inner area of the hurricane would not be substantially affected by 5-10 per cent alteration of the wind speed.

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