The Effects of Release of Latent Heat on the Vorticity of a Tropical Storm over Land[1]

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ABSTRACT

The effect of the release of latent heat on the thermal vorticity of Hurricane Diane (1955) is examined when it passed over the east coast of the United States as a weakened tropical storm. A form of the thermodynamic energy equation is used as the diagnostic equation for this study. Because of uncertainty in the data, as well as incomplete physical understanding of the interaction between synoptic and convective scales, several alternative models are used which employ different assumptions on the influence of convection and the vertical distribution of humidity and temperature.

In agreement with the results of previous investigations, Diane was found to weaken after landfall as a result of the lifting of negatively buoyant air at the storm center. However, as the storm readjusted to its new environment the stability at the center again decreased. The latent heating, as represented by its horizontal Laplacian, then becomes effective in the maintenance and redevelopment of the storm.

1. Introduction

The release of latent heat is dynamically linked to a variety of scales of atmospheric motions. On the smallest scales, latent heating is coupled with the dynamics of convection, while on the synoptic scale latent heating is a critical factor in the thermal forcing of tropical storms and has an important effect upon extratropical storm development (Danard, 1964; Krishnamurti, 1968).

Although the importance of latent heating has been established, few quantitative studies have been made on the effects of latent heating in a hurricane after landfall, despite the fact that the precipitation may be intense in these storms (Dunn and Miller, 1960). Furthermore, although the horizontal distribution of precipitation with respect to the storm center has been described by several authors (e.g., Dunn and Miller, 1960), its implications have not been assessed. Bosart *et al.* (1972) have discussed qualitatively the influence of the precipitation maximum on surface vorticity development.

The dynamics and thermodynamics of the initial decay of the hurricane immediately after landfall have been the subject of several investigations. Hubert (1955) showed that the increased friction over land was not of sufficient magnitude to account for the increased rate of dissipation of hurricane energy which is observed over land. Miller (1963) determined that the kinetic energy production of Hurricane Donna (1960) was less over land than over the ocean. These results support the hypothesis of Byers (1959), Palmén (1956), Bergeron (1954), and Riehl (1954) that the absence of the sensible and latent heat flux from the ocean is responsible for the weakening of the hurricane over land. This hypothesis has been verified in a numerical experiment (Ooyama, 1969) in which the model hurricane decayed after the application of boundary conditions typical of a continental surface.

Although the dynamics of hurricanes have been investigated for the period immediately after landfall, there have been only a few studies of tropical storms after they have been over land for an extended period of time. Palmén (1958) computed the production of kinetic energy in Hurricane Hazel (1954) in a case where both baroclinic energy sources and latent heating made strong contributions. Bergeron (1954), in a study of a tropical storm over the southeastern United States, concluded that the warm core of the disturbance disappeared because the released latent heat could not overcome the adiabatic cooling induced by the vertical motion in a stable environment.

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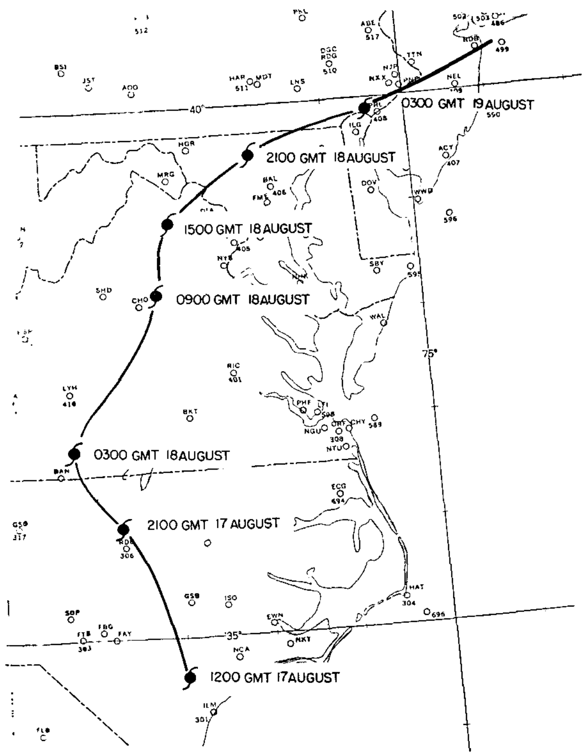

FIG. 1. The path of hurricane Diane, 17–19 August 1955.

The effects of latent heating on the temperature of the storm system are somewhat subtle, since the local temperature change caused by adiabatic cooling and positive vertical motion may almost compensate the temperature change induced by condensation. However, the relationship between the vertical motion and the latent heating is not a simple one, so that it is preferable to deal with these terms separately. The response of the circulation to the heating may be explained qualitatively by recognizing that the gradient of latent heating is proportional to the time-rate-of-change of the thermal wind in a balanced flow. Further,

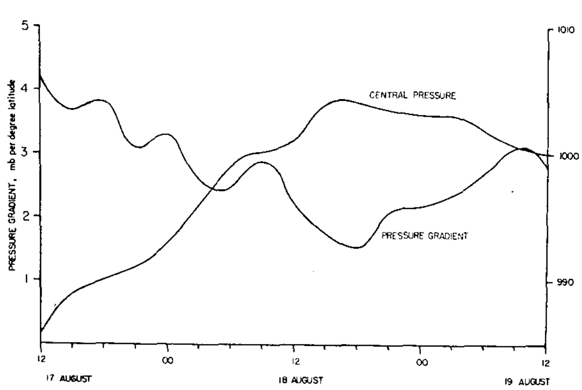

FIG. 2. The central pressure and average surface pressure gradient (as measured from the center to the last closed isobar of the storm) as a function of time (GMT). From Malkin (1959).

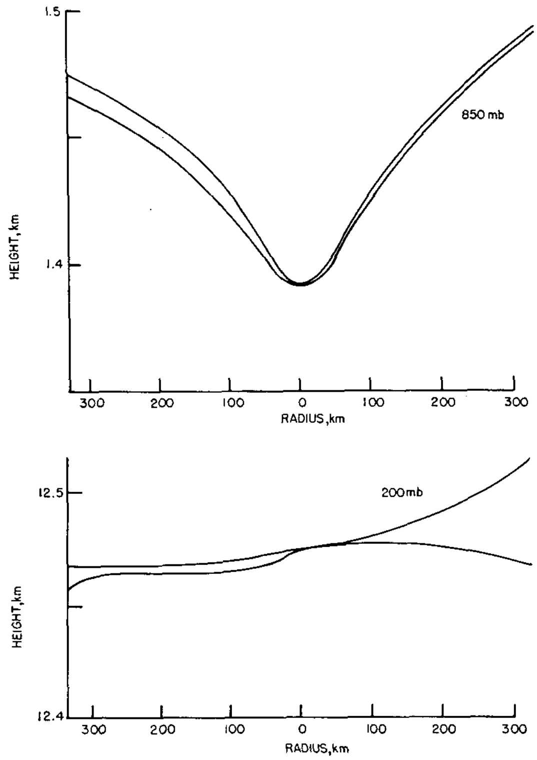

FIG. 3(a)

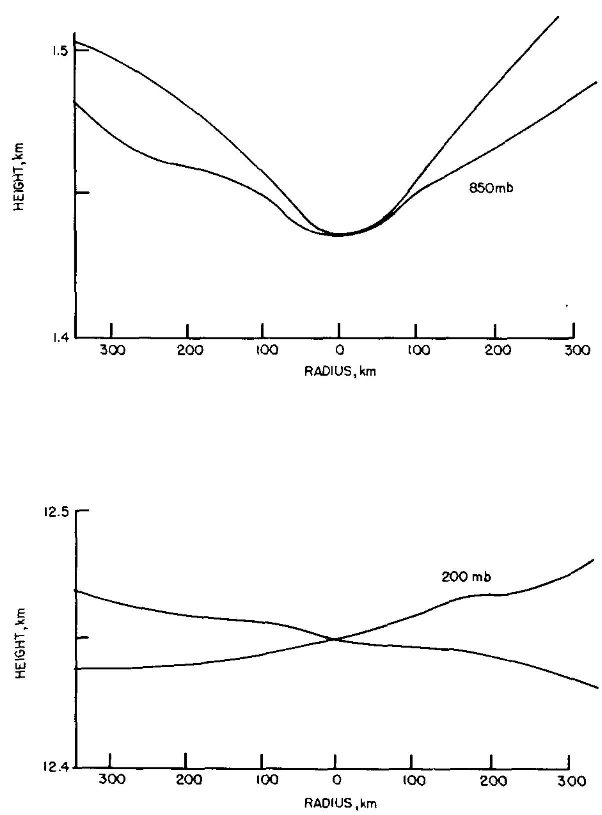

FIG. 3(b)

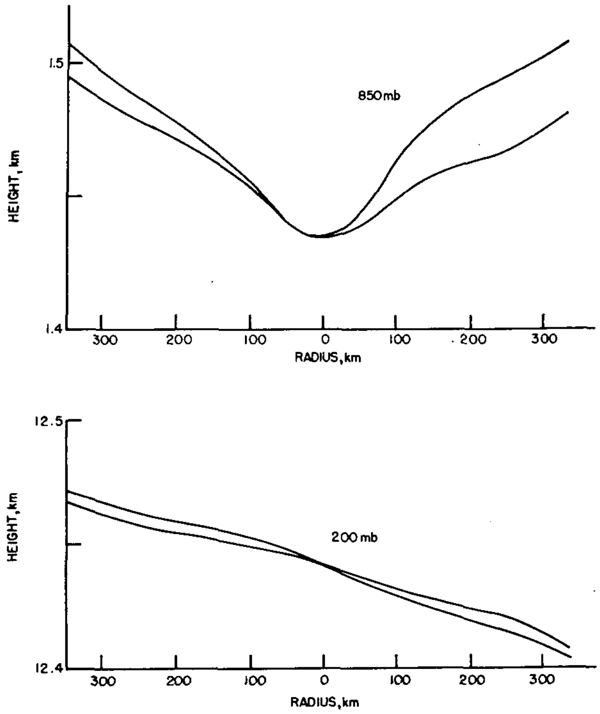

FIG. 3(c)

FIG. 3. Profiles of heights in orthogonal directions from the storm center at 850 mb and 200 mb: (a) for 0300 GMT 18 August; (b) for 1500 GMT 18 August; (c) for 0300 GMT 19 August.

the spatial changes of the gradient of latent heating (as expressed by the Laplacian of the heating function) must be balanced by temporal changes in thermal vorticity. The equation expressing this relationship was derived for a special case by Petterssen (1955) from the thermodynamic energy equation. In the present study this equation is applied to a case of a synoptic-scale storm. The temporal changes in thermal vorticity which are calculated are taken as a measure of changes in the storm intensity during its track over land.

After a study of the hurricanes which made landfall in the period 1950–1970, several were found to satisfy the criterion of weak interaction with the environment that is necessary for isolating the effect of heating upon the thermal vorticity. After initial weakening following landfall, these storms were observed (on a few occasions) to maintain their circulation or even reintensify without obvious interaction with baroclinic energy sources. Diane (1955) was chosen because it offered the opportunity of analyzing a tropical storm which had a lifetime over land that was considerably longer-than-average (Palmén and Jordan, 1955).

This diagnostic study has as its aim the determination of the effect of latent heating on the thermal vorticity of a hurricane after landfall as it readjusts to the continental environment. Of special interest is the influence of the horizontal distribution of gravitational stability, vertical motion, and latent heating, as measured by the Laplacian of these functions. An evaluation is made of the contribution of parameterized cumulus convection to the total heating and change in thermal vorticity.

2. Synoptic description

Hurricane Diane entered the United States mainland near Wilmington, N. Car., at about 1200 GMT 17 August 1955 (Fig. 1) and weakened rapidly as it passed northward through North Carolina into southern Virginia (Fig. 2). For a detailed description of the history of Diane see U. S. Weather Bureau Technical

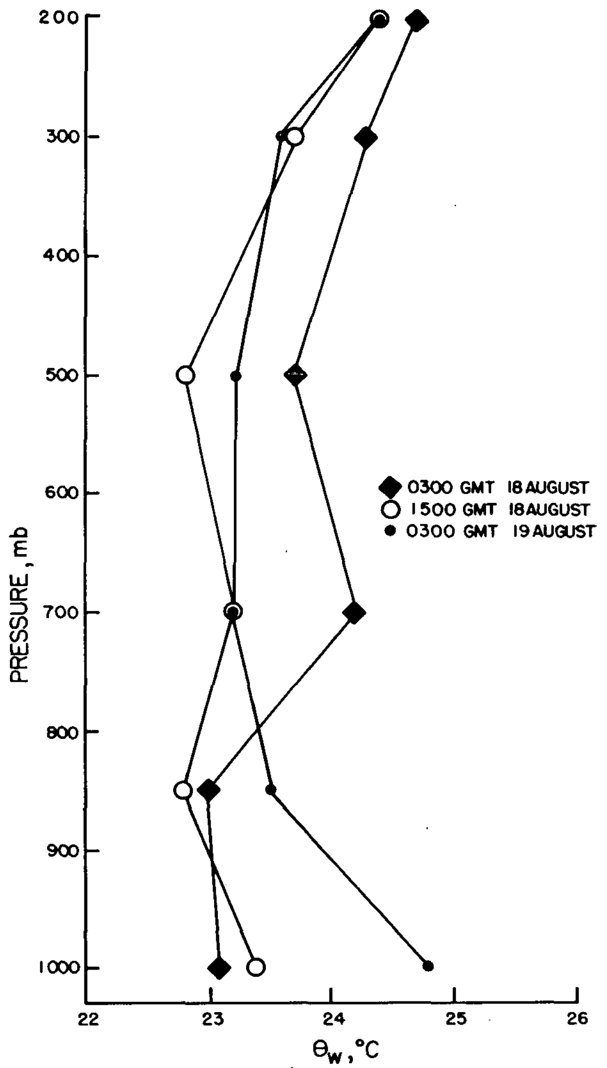

FIG. 4. Wet bulb potential temperature (θ_w) soundings at the storm center. These values were calculated for saturated conditions at a temperature obtained from horizontal temperature fields.

Paper No. 26 (1956) and Dunn and Miller (1960). At 0300 GMT on the 18th, Diane was centered about 120 km northeast of Greensboro, N. Car.

During the next 12 hours (Period I) the storm moved north-northeast at a speed of about 10 kt and was located about 100 km west-northwest of Washington on 1500 GMT 18 August. The central pressure increased and the height gradient weakened markedly at 850 mb (Fig. 3a, b), while at 200 mb the changes were small.

From 1500 GMT 18 August to 0300 GMT 19 August (Period II) the tendency for the storm to weaken was reversed (Fig. 2). A comparison of the 850 mb height profile for 0300 GMT 19 August (Fig. 3c) with earlier observations shows the circulation at 850 mb is no longer weakening.

An inspection of the interpolated sounding for 0300 GMT 18 August at the center of the storm (Fig. 4) shows the stability throughout a deep layer of the atmosphere is positive, although it should be noted that the 700–500 mb layer is conditionally unstable.

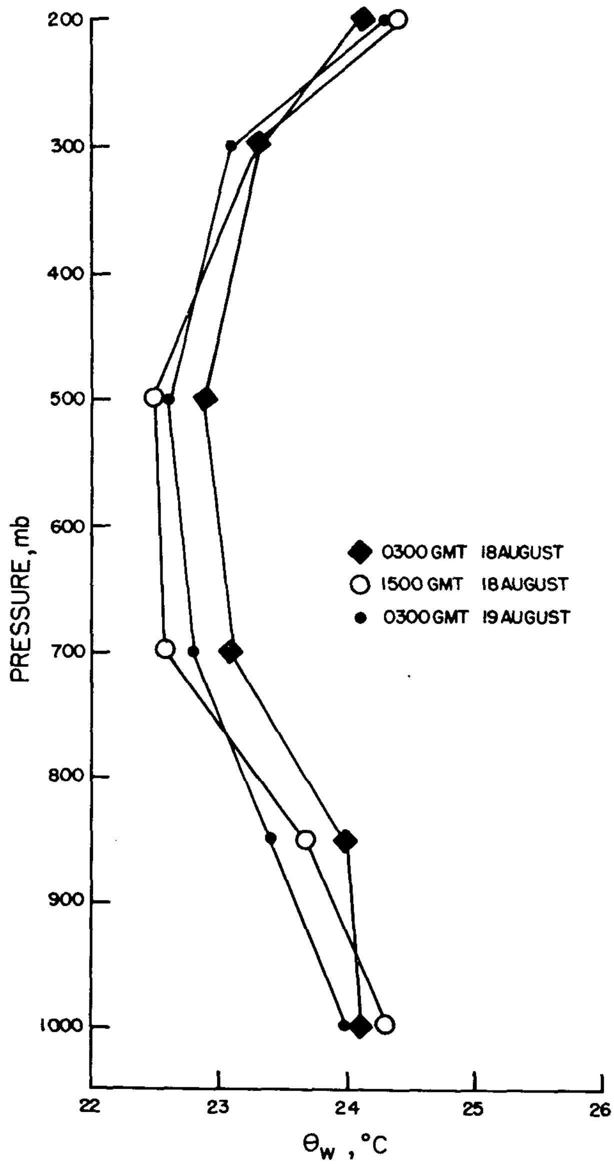

Fig. 5. Composite wet bulb potential temperature (θ_w) soundings at 240 km radius from storm center. These values were obtained in the same manner as in Fig. 4.

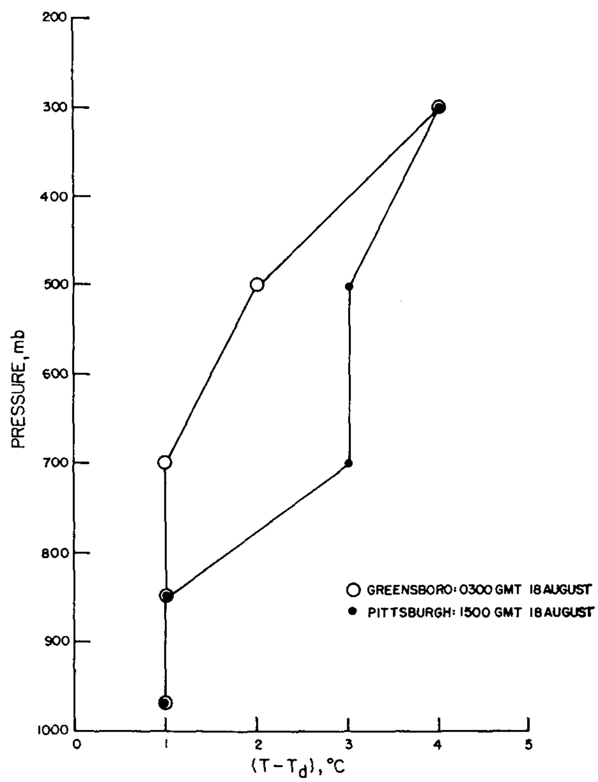

Fig. 6. Temperature-dewpoint depression ($T-T_d$) for selected stations.

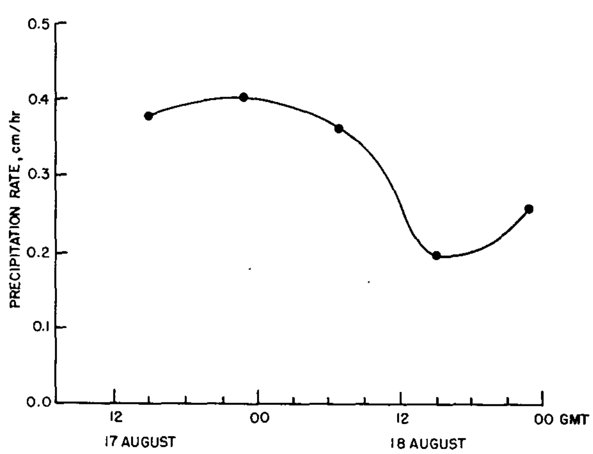

Fig. 7. Precipitation rate at center of Diane, 1500 GMT 17 August–2300 GMT 18 August. Data points represent 4 h averages.

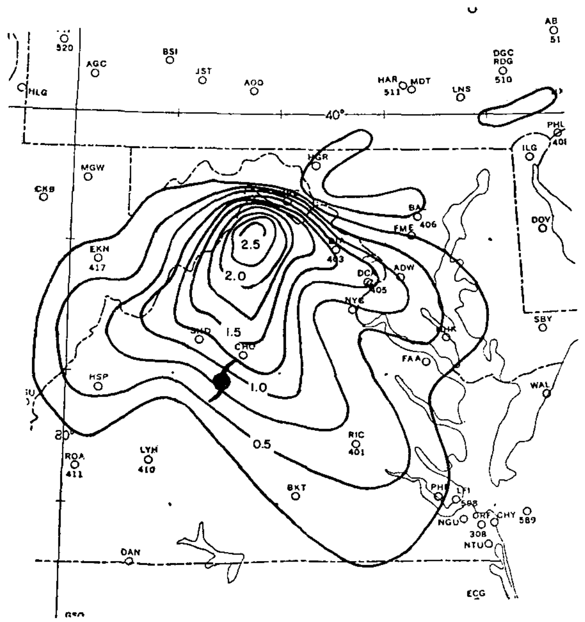

FIG. 8. Two-hourly rainfall distribution, 0700–0900 GMT 18 August. The position of Diane is indicated. Contour intervals are 0.25 cm.

Conditional instability is also present at some distance from the storm center (Fig. 5). For example, the Greensboro sounding taken on August 18, 0300 GMT, is conditionally unstable and shows nearly saturated conditions through a great depth of the atmosphere (Fig. 6).

In summary, Figs. 4 and 5 indicate that the stability decreased rapidly during Period I and was nearly neutral by 1500 GMT 18 August. This trend continued throughout Period II, so that by 0300 GMT 19 August the atmosphere was conditionally unstable in a deep layer.

During Period I the precipitation rate at the center decreased steadily (Fig. 7), while the maximum precipitation rate is located 80–160 km to the northeast of the storm center. Figure 8 shows a 2-hourly rainfall distribution typical of the first 12 hours. The spreading of the maximum precipitation away from the center after landfall has been well documented for other storms (Berg, 1954; Brooks, 1946; Miller, 1958). This feature has been observed even when the influence of the topography and interaction with preexisting extratropical storms are minimal. A partial explanation may be the increased stability which develops at the center after landfall due to the abrupt reduction of the sensible and latent heat fluxes at the lower boundary.

The maximum precipitation continued to occur well to the north and east of the storm center throughout Period II (Fig. 9). The precipitation at the storm center decreased through most of this time period but showed an increase towards the end (Fig. 7). This increase is probably a response to decreasing stability, since thunderstorms were observed later in the second period at stations near the center of the storm for the first time since landfall.

Throughout the entire period of analysis, interaction with the extratropical environment appears to be weak. For example, Fig. 10 shows that the only disturbance at 500 mb was located directly over the surface position of the storm, and thus was an integral part of the storm circulation.

3. The thermal vorticity equation

Using the hydrostatic relation and the ideal gas law the first law of thermodynamics may be stated

$$\frac{1}{c_p}\frac{dQ}{dt}=\frac{-g}{R}p\frac{\partial}{\partial t}\left(\frac{\partial Z}{\partial p}\right)+\mathbf{V}\cdot\nabla T+\omega\frac{\partial T}{\partial p}-\frac{1}{c_p\rho}\omega, \qquad (1)$$

where the notation is conventional.

Taking the integral of (1) over time and between 200 mb to 850 mb, and applying the operator ∇^2, we obtain

$$\zeta_{\mathrm{TH}}\Big]_{t_0}^{t_1}=\frac{R}{f}\int_{t_0}^{t_1}\int_{200}^{850}\nabla^2\Big[\underset{\text{(I)}}{\overline{\mathbf{V}}\cdot\nabla T}+\underset{\text{(II)}}{\omega(\Gamma-\Gamma_a)}-\underset{\text{(III)}}{\frac{1}{c_p}\frac{dQ}{dt}}\Big]\times d(\ln p)dt, \qquad (2)$$

where the thermal vorticity

$$\zeta_{\mathrm{TH}}=\zeta_{850}-\zeta_{200}=(g/f)\nabla^2(z_{850}-z_{200}),$$

and

$$\Gamma_a=RT/pc_p$$
$$\Gamma=\partial T/\partial p.$$

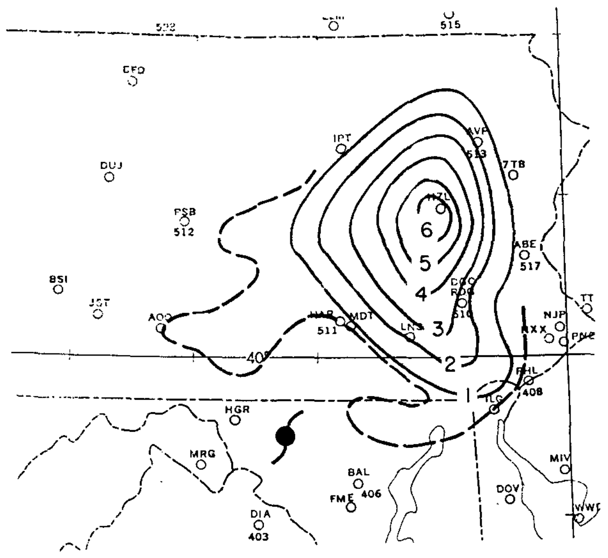

FIG. 9. Two-hourly rainfall distribution (cm), 1900–2100 GMT 18 August.

The effect of temperature advection is represented by the term (I) in (2). In this study, advection of the temperature field by the geostrophic component of the wind only has been taken into account. The ageostrophic advection should be small, since the storm which has been chosen for the analysis is embedded in a weak thermal field of the westerlies. Furthermore, Krishnamurti (1968) in a study of a strongly baroclinic extratropical storm with precipitation rates similar to those of Diane also found the ageostrophic temperature advection to be 1 to 2 orders of magnitude smaller than the contribution of latent heating.

The second term (II) in (2) represents the effects of vertical motion and stability, and term (III) represents the effects of diabatic heating on the thermal vorticity. For this study the only effects of latent heating have been considered in term (III), and the dense network of rainfall observations permits an estimate of this latent heat source to be made. By comparison, the turbulent heat flux at the surface should be small over land (Miller, 1963). The other sources that may be represented by the diabatic heating term, such as internal dissipation and solar and infrared radiation, are not the object of study in this paper, since their horizontal distribution would be extremely difficult to estimate.

4. Modeling of vertical motion and release of latent heat

It is often assumed that the release of latent heat is accomplished on two scales of atmospheric motion. In an absolutely stable environment, condensation is associated with the mean synoptic-scale vertical motion, while in a conditionally unstable atmosphere it also may be accomplished by small scale convective motions. Convection is not independent of the synoptic scale circulation, however, but is maintained by the associated moisture convergence. This parameterization of convection has been utilized in the hurricane models of Charney and Eliassen (1964) and Kuo (1965) and several others.

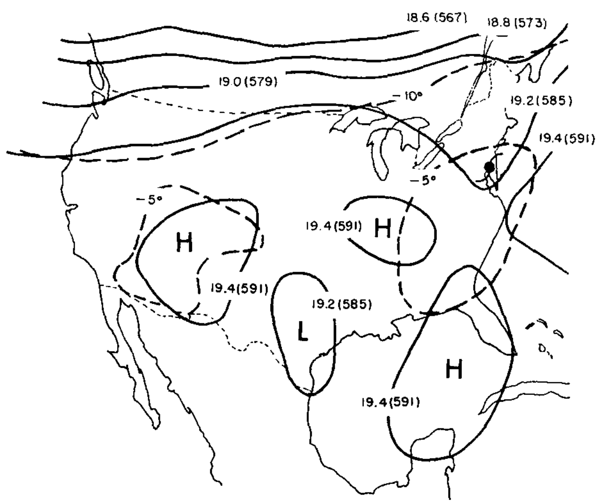

Fig. 10. U. S. Weather Bureau 500 mb analysis, 0300 GMT 19 August 1955. Height contours are labeled in thousands of feet (tens of meters) and isotherms in °C.

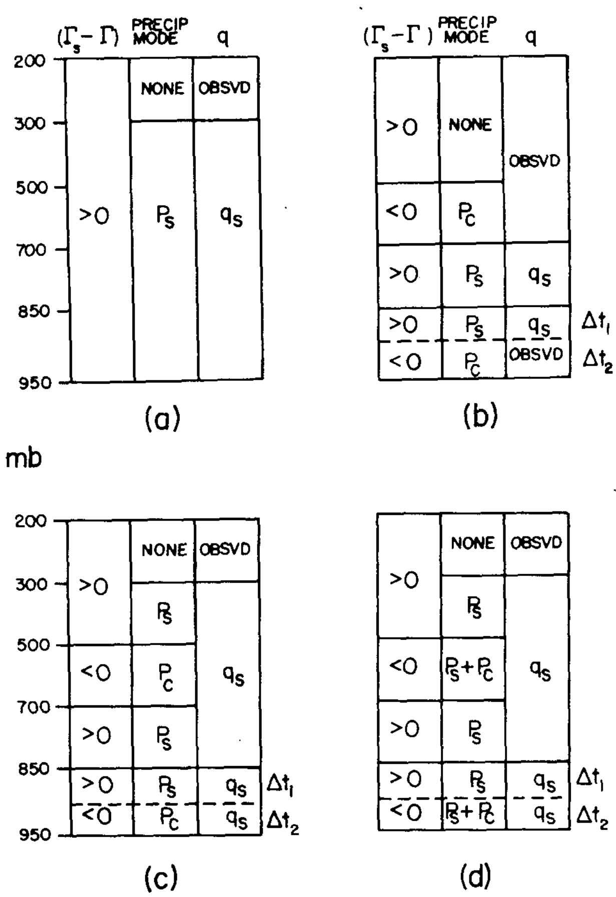

Fig. 11. Schematic diagram of precipitation models. Vertical profiles of stability, mixing ratio, and precipitation mode used for calculating the convective storage and heating: (a) Model I, (b) Model II, (c) Model III, and (d) Model IV.

When motion in a stable environment is considered, the diabatic heating

$$\frac{dQ}{dt} = -L\omega\frac{dq_s}{dp} \tag{3}$$

may be combined with the vertical motion term in the thermal vorticity equation to yield one term,

$$\frac{R}{f}\int_{t_0}^{t_1}\int_{200\text{ mb}}^{850}(\Gamma-\Gamma_s)\omega d(\ln p)dt, \tag{4}$$

where

$$\Gamma_s = \Gamma_a + \frac{L}{c_p}\frac{dq_s}{dp}.$$

This relationship is invalid in the case of buoyant convection, since the vertical velocity in (3) is the vertical velocity associated with the convective motion rather than the synoptic motion in (2). In that case the convective theory of Kuo (1965) may be used to estimate the latent heating independently of the vertical motion term.

Because of an incomplete physical understanding of the interaction between synoptic and convective scales and of the effect of large-scale synoptic vertical motion and the unsaturated environment upon the intensity of convection, several different models were used to calculate the components of steady and convective precipitation which separately affect the thermal vorticity through terms (II) and (III) in (2). A schematic diagram of these models is shown in Fig. 11.

Model I is constructed with the assumption that the latent heat was released by the large-scale synoptic vertical motion in a stable atmosphere. This assumption is reasonable in the first period of the analysis because the gravitational stability in a deep layer of the atmosphere is observed to be positive, although it should be noted that the air at the center of Diane was conditionally unstable from 700 to 500 mb (Fig. 4). In this model the atmosphere was assumed to be saturated from the lifting condensation level of 950 mb to 300 mb.

In order to calculate the vertical motion term in the vorticity equation, the value of omega must be determined from the steady component of the precipitation, P_S. Because of the lack of accurate wind data it was necessary to assume a vertical distribution of omega with a magnitude sufficient to produce the observed precipitation, P_S, where

$$P_S = \frac{1}{g}\int_{950\text{ mb}}^{300} \omega \frac{dq_s}{dp}\,dp. \tag{5}$$

For this study the level of non-divergence or maximum value of omega was chosen to be 700 mb (Hubert, 1959). The large-scale vertical motion may then be defined:

$$\omega(p) = \omega_0\left(\frac{1000-p}{300}\right) \quad 700\text{ mb} \leq p \leq 1000\text{ mb}$$

$$= \omega_0\left(\frac{p-200}{500}\right) \quad 200\text{ mb} \leq p \leq 700\text{ mb}$$

A value of ω_0 is then obtained which is consistent with the observed precipitation.

In Model II (Fig. 11b) the mechanism for latent heating is through parameterized convection in those layers where the sounding is unstable, and through large-scale vertical motion where the atmosphere is observed to be neutral or stable. For calculations involving the mean motion, saturation is assumed to exist up to 700 mb, which is consistent with the observation of high relative humidities in the lower layers (Fig. 6). Above 700 mb, where the observed relative humidity is decreasing, unsaturated conditions are assumed for large-scale motion.

The convective theory used in Models II–IV is similar to Kuo's (1965); i.e., the moisture convergence must balance the sum of precipitation and storage due to convection,

$$I(\omega) = P_c + S_c. \tag{6}$$

Here the three-dimensional moisture convergence in a convective column is

$$I(\omega) = -\frac{1}{g}\int_c \nabla\cdot(q\mathbf{V})dp - \frac{1}{g}\int_c \frac{\partial}{\partial p}(q\omega)dp,$$

where the integration is performed over increasing pressure, over the layer, c, in which the convection originates. Since horizontal variations of q are small near the storm center we have

$$I(\omega) \approx -\frac{1}{g}\int_c \omega\frac{\partial q}{\partial p}dp.$$

An expression for S_c, the rate of moisture storage associated with convection, may be derived in the following manner. From Kuo (1965) the rate of convective rainfall is

$$P_c = I(\omega)\frac{m}{m+1}, \tag{7a}$$

where

$$m = \delta q_2/\delta q_1,$$

and δq_1 and δq_2 are, respectively, the moisture required to saturate and warm an atmospheric column to the saturated conditions of the hypothetical cloud. The properties of the cloud (T_s,q_s*) are determined by the temperature and mixing ratio of a saturated parcel lifted moist-adiabatically from the unstable layer, so that

$$\delta q_1 = \frac{1}{g}\int_K (q_s^* - q)dp$$

$$\delta q_2 = \frac{1}{g}\int_K \frac{c_p}{L}(T_s - T)dp,$$

where the integrals above are taken over the layers, K, in which $T_s \geq T$. Upon substituting into (6) for P_c we have

$$S_c = \left(\frac{1}{m+1}\right)I. \tag{7b}$$

Kuo's theory demands that convection takes place if

TABLE 1. The calculated values of the terms in Eq. (2) for different precipitation models. The residual term and the rate of cloud production ($a/\Delta t$) are indicated in the last two rows.

Model number	I	II	III	IV	I	II	III	IV
	0300–1500 GMT 18 August				1500 GMT 18 Aug–0300 GMT 10 Aug			
$\Delta\zeta$ observed $\times 10^{-3}$ s^{-1} (12 h)$^{-1}$	−1.2	−1.2	−1.2	−1.2	+0.3	+0.3	+0.3	+0.3
(I) Laplacian of thermal advection $\times 10^{-3}$ s^{-1} (12 h)$^{-1}$	+0.2	+0.2	+0.2	+0.2	+0.3	+0.3	+0.3	+0.3
% Convection for compatability	—	21%	48%	5%	—	28%	56%	8%
(III) Laplacian of latent heating $\times 10^{-3}$ s^{-1} (12 h)$^{-1}$	—	+3.4	+7.7	+0.9	—	+1.5	−2.9	+0.4
(II) Laplacian of vertical motion $\times 10^{-3}$ s^{-1} (12 h)$^{-1}$	−1.4	−37.9	−9.7	−2.9	−0.3	−12.3	−2.4	−0.6
$\Delta\zeta$ Residual $\times 10^{-3}$ s^{-1} (12 h)$^{-1}$	0.0	−33.1	−0.6	−0.6	−0.3	−10.8	+0.5	−0.2
$a/\Delta t$ (10^{-4}) s^{-1}	—	2.2	5.0	0.5	—	1.2	2.5	0.4

and only if the atmosphere is conditionally unstable and the net moisture convergence in the atmospheric column is positive.

The convective model used in this study differs from Kuo's in that the rate of convection and moisture storage will equal the three-dimensional moisture convergence in the unstable layers only. This is necessary to accommodate both steady and convective forms of precipitation.

The total precipitation, P_T, can be decomposed into the steady (P_S) and convective (P_c) precipitation components with the use of (6) and (7). Since P_T is equal to the sum of the steady and convective precipitation, an assumed value of $P_c = P_c^*$ yields a value of P_S which is then used to determine $I(\omega)$. The quantities $P_c(I)$ and $S_c(I)$ are then obtained from (7a) and (7b). In general, P_c will not be equal to $P_c(I)$. However, a unique and compatible value of $P_c^* = P_c(I)$ will exist somewhere in the range $0 \leq P_c^* \leq P_T$, since the lower limit implies that the calculated values $P_c(I) > 0$, while the upper limit implies that I and $P_c(I)$ are zero. These calculations are shown in Tables 2, 3, and 4.

The steady and convective components determined above were then used to calculate the heating and vertical motion terms in the vorticity equation. The heating term can be computed directly from P_c. The vertical motion term can be calculated from the P_s component by the procedures described previously using (5).

Another measure of the intensity of convection is the ratio ($a/\Delta t$), first defined by Kuo (1965) as the percentage, a, of the synoptic scale area covered by newly formed convective elements in time, Δt. This fraction is shown by Kuo to be equal to the quantity $I_K/\delta\bar{q}$, where I_K is the moisture convergence being used for convective storage and convective precipitation in the entire column, but does not include the moisture convergence used for steady precipitation. Kuo defines $\delta\bar{q}$ as the sum of the two integrals δq_1 and δq_2 discussed previously.

The tendency of hygrometers to yield humidity measurements which are too low results in an overestimate of the storage term S_C. Model III has been developed in order to examine the effect of this error (Fig. 11c). This model is similar to Model II, except that saturated conditions are assumed for the convective as well as the nonconvective layers, and S_C is zero. In addition, the stable 500–300 mb layer is assumed to be saturated with respect to the mean vertical motion, where previously the observed humidity sounding had ruled it out. In this layer the rate of cooling accompanying the large scale upward motion will be at the moist-adiabatic lapse rate, instead of at the dry-adiabatic as in Model II.

In order to examine the more realistic case of simultaneous steady and convective precipitation, a fourth model has been analyzed. In Model IV, as in Model III, both stable and unstable layers up to 300 mb are assumed to be saturated. However, the mean motion also releases some latent heat in the unstable layers. It will be assumed that the latent heating associated with this mean upward motion will be at a rate necessary to maintain the observed lapse rate. Furthermore, the moisture convergence within unstable layers is now used for both steady and convective precipitation, and $I(\omega)$ must be replaced by $I_K = I(\omega) - P_S$ in (6), (7a), and (7b).

5. Results

The results of the calculations for each term in (2), shown in Table 1, indicate that the thermal advection was not responsible for the decrease in vorticity during the first time period, but contributed to the small increase in thermal vorticity observed during the second time period. In Krishnamurti's (1968) study of a mid-latitude storm using winds from the balance equation and Palmén's (1958) study of Hazel using observed winds, the thermal advection made the largest contribution to the development and maintenance of the circulation. This contrast in results is a consequence of the very weak interaction of Diane with the baroclinic westerlies.

The following sub-sections describe the contributions of the heating and vertical motion terms in (2) to the

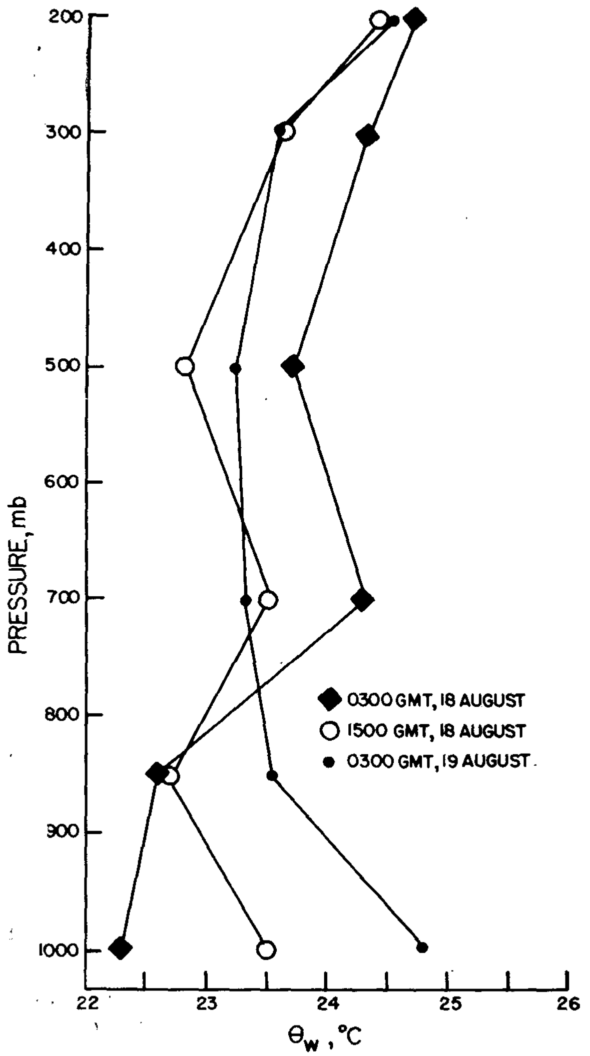

FIG. 12. Composite wet bulb potential temperature (θ_w) soundings at 48 km radius from storm center. These values were obtained in the same manner as in Fig. 4.

production of thermal vorticity in Diane for the four models presented in Section 4.

a. Model I

The calculations of Model I without convection (Table 1) show that the vertical motion term is responsible for the large decrease in the thermal vorticity during the first time period. These results also show that the major reason for the sharp reversal of the rapid rate of decrease of thermal vorticity during the second period of analysis was the large decrease in the magnitude of the vertical motion term. An inspection of the 0300 GMT and 1500 GMT soundings (Fig. 4 and Fig. 12) shows greater positive stability within a deep layer at the center than at the 48 km radius. In addition, the precipitation amounts and the vertical motion computed from it are greater at the center. Therefore $|(\Gamma-\Gamma_s)\omega|$ is more positive at the center of the storm, implying a decrease in thermal vorticity due to the Laplacian of the vertical motion term. During the second time period, the vertical motion term decreased considerably in magnitude. This result is a response to the relative decrease in stability at the center of the storm (Figs. 4 and 12), as well as the relative decrease in precipitation and, hence, the calculated vertical motion at the center. Consequently, the Laplacian of $|(\Gamma-\Gamma_s)\omega|$ is much smaller than during the first time period.

A non-convective model is appropriate for the stable atmosphere observed during the first time period. However, it cannot represent the contribution of convective heating in the production of vorticity within a conditionally unstable environment during the second time period.

b. Model II

Table 2 gives a listing of convective precipitation amounts, moisture convergence, and convective storage for Model II. The results shown in this table indicate that the main source of three-dimensional moisture convergence for convection was the 700–500 mb layer. This result also holds for Models III and IV. (If two-dimensional horizontal moisture convergence had been used there could have been neither steady nor convective precipitation originating from the unstable moisture divergent 700–500 mb layer. However, the mean vertical motion would have still contributed to a considerable moisture increase in this layer, and it was not observed during the period of study.)

The large vorticity imbalance (Table 1) occurs because the bias in the humidity sounding results in an excessive amount of the convergent moisture being

TABLE 2. Values of assumed (P_c^*) and computed [$P_c(I)$] convective precipitation, moisture convergence, and convective storage for selected fractions of convection in Model II. L_1 and L_2 refer to the 1000–850 mb layer and 700–500 mb layer, respectively. Double asterisk indicates the internally consistent solution.

Assumed fraction of convective precipitation for Δt_1	Components of the water budget (cm/12 h)					
	P_c^*	$I(\omega)$		$S_c(I)$		$P_c(I)$
			L_2		L_2	
0	0.00		5.44		4.60	0.94
**0.21	0.75		4.82		3.63	0.75
0.25	0.90		4.60		3.45	0.71
0.50	1.80		2.72		2.30	0.47
0.75	2.70		1.36		1.15	0.24
for Δt_2		L_1	L_2	L_1	L_2	
0.0	0.00	1.16	6.25	0.86	5.31	1.25
0.25	0.78	0.88	4.68	0.66	3.99	0.91
**0.28	0.89	0.86	4.50	0.65	3.84	0.89
0.50	1.56	0.58	3.12	0.43	2.66	0.62
0.75	2.34	0.29	1.56	0.22	1.33	0.31

used for saturating, rather than heating the upper layers.

c. Model III

Model III, with its assumption of a saturated atmosphere and no convective storage, S_c, gives somewhat higher values (approximately 50%) of percentage convection needed to satisfy the compatibility condition (Table 3). According to Table 1, this percentage of convection also leads to a balance in the thermal vorticity equation. However, the lack of any local observations of thunderstorms makes it unlikely that such large percentages of the observed precipitation were convective, although it is possible that there was widespread, weak convection.

d. Model IV

Model IV (Table 4) is the most satisfactory because only a small fraction of convection is required for internal consistency. This occurs because all layers make some contribution to the steady precipitation, which reduces the amplitude of the vertical motion and the computed moisture convergence for convection. Furthermore, the approximate balance in the vorticity equation achieved in Model III is preserved, since the strong adiabatic cooling in the 700–500 mb layer can now be offset by heating from steady rather than convective precipitation as in Model III. Finally, there is reasonable agreement with the value of $a/\Delta t = 0.5\times10^{-4}$ s^{-1} obtained by Krishnamurti (1969).

The results of Models III and IV for the second time period show that when the stability at the center of the storm becomes neutral or unstable and decreases radially, as in Diane, the latent heating may balance or overcome the vertical motion term and maintain or intensify the warm core (positive thermal vorticity).

TABLE 3. Values of assumed (P_c*) and computed [$P_c(I)$] convective precipitation and moisture convergence for selected fractions of convection in Model III. L_1 and L_2 refer to the 1000–850 mb layer and 700–500 mb layer, respectively. Double asterisk indicates the internally consistent solution.

Assumed fraction of convective precipitation	Components of the water budget (cm/12 h)			
for Δt_1	P_c*	$I(\omega)$		$P_c(I)$
		L_1		
0.00	0.00	3.40		3.40
0.25	0.90	2.54		2.54
**0.48	1.78	1.78		1.78
0.50	1.80	1.70		1.70
0.75	2.70	0.84		0.84
for Δt_2		L_1	L_2	
0.00	0.00	3.40	0.66	4.06
0.25	0.79	2.54	0.51	3.05
0.50	1.58	1.70	0.33	2.03
**0.56	1.78	1.50	0.28	1.78
0.75	2.36	0.84	0.18	1.02

TABLE 4. Values of assumed (P_c*) and computed [$P_c(I)$] convective precipitation, steady precipitation, and moisture convergence for selected fractions of convection in Model IV. L_1 and L_2 refer to the 1000–850 mb layer and 700–500 mb layer, respectively. Double asterisk indicates internally consistent solution.

Assumed fraction of convection precipitation	Components of the water budget (cm/12 h)					
for Δt_1	P_c*	P_s		$I(\omega)$		$I_K(\omega)=P_c(I)$
		L_2				
0.00	0.00	1.68		1.88		0.20
**0.05	0.17	1.60		1.78		0.17
0.25	0.90	1.26		1.41		0.15
0.50	1.80	0.84		0.94		0.10
0.75	2.70	0.42		0.47		0.05
for Δt_2		L_1	L_2	L_1	L_2	
0.00	0.00	0.23	1.42	0.30	1.63	0.28
**0.08	0.25	0.21	1.37	0.28	1.58	0.25
0.25	0.79	0.18	1.07	0.22	1.23	0.21
0.50	1.58	0.12	0.71	0.15	0.81	0.14
0.75	2.36	0.06	0.36	0.07	0.41	0.07

6. Discussions and conclusions

In agreement with previous studies of hurricanes over land by Miller, Bergeron, and others, Diane was found to weaken initially because of the lifting of a stably stratified column of air near the center of the storm. However, weakening of the storm ceased when neutral conditions were reached at the storm center during the second time period. Further destabilization signified the renewal of active convection closer to the center of the storm, which contributed to the reintensification of Diane.

The accuracy of the calculated values of the terms in the vorticity equation is certainly affected by the assumptions regarding the vertical velocity profile and the existence and distribution of the diabatic heat source. Nevertheless, the relative importance of the vertical motion and diabatic heating may be estimated.

An important contribution to the change in vorticity was made by the Laplacian of vertical motion and latent heating. In the early stages of readjustment of the storm, in spite of intense precipitation, the distribution of the latent heating was unable to compensate for the vorticity decrease caused by rising motion in a stable environment. Later, during the period of intensification, when the air in the center of the storm had become conditionally unstable, the vertical motion term became less important and the parameterized convective heating occurring near the storm center contributed to the intensification. Similar results were obtained by Tracton (1973) in a study of extratropical storms. Advection of cold air at the periphery of the storm, although weak, may also have played a role in the reintensification of the storm.

The initiation of thunderstorm activity near the storm center was related to a marked increase in the magnitude of the Laplacian of the heating function at the end of the second time period and thus contributed to the reintensification of Diane.

Throughout these calculations, changes in the Laplacian of such parameters as vertical motion and latent heating are more important than changes in the values of the parameters themselves. For example, the maximum rainfall actually increased with time, while rainfall near the center decreased throughout most of the period (Fig. 7). Therefore, the Laplacian of the precipitation makes a smaller contribution to thermal vorticity during the second time period. In fact, although the precipitation at the center of the storm decreased by about 15% from the first 12 hour period to the second, the Laplacian of the precipitation and hence ω decreased by almost 70%. Similarly, the Laplacian of the stability decreased to 90% of its value during the first time period, while the stability itself only decreased to 50% of its previous value.

Several different convective models employing Kuo's parameterization were studied. Models II and III, which included only convective precipitation in the unstable layers, require excessive convection or fail to lead to a realistic balance in the vorticity equation. Model IV, which provides for both convection and steady precipitation in unstable layers, appears to be the most effective in describing the dynamics and thermodynamics of Diane. Consequently, a more general parameterization than Kuo's seems to be necessary when dealing with a well-developed circulation which contains both conditionally unstable and stable, unsaturated layers.

The results of this study provide some insight into the general behavior of a particular class of tropical storms over land and, hence, may improve our ability to forecast their occasional intensification. This study has dealt with a storm which did not interact with a strong, upper-level trough and baroclinic field in the middle-latitude westerlies. When strong interaction occurs, a more complex combination of baroclinic and convective effects contribute to the reintensification.

The analysis has shown that as the storm readjusts to its new boundary conditions over land, continued weakening can be expected with a diminution of the rainfall at the storm center where static stability has increased rapidly. Consequently, the maximum in the rainfall pattern is displaced from the storm center.

Farther along its trajectory the sounding near the center may again approach neutral or unstable conditions due to sustained moisture inflow and mixing with the environment. Cold advection at upper levels will hasten this transformation. Under these conditions the precipitation may become effective in the intensification of the storm if the *distribution* of sensible temperature change maintains a warm core relative to the surroundings at the storm center. The effectiveness of this process depends upon the relative location of the precipitation maximum and the extent to which latent heating is compensated by adiabatic cooling. Therefore, cumulus convection (which is not simply connected to the synoptic scale vertical motion) can be very critical in controlling the sign of the Laplacian. This was the circumstance that influenced the redevelopment of Diane.

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