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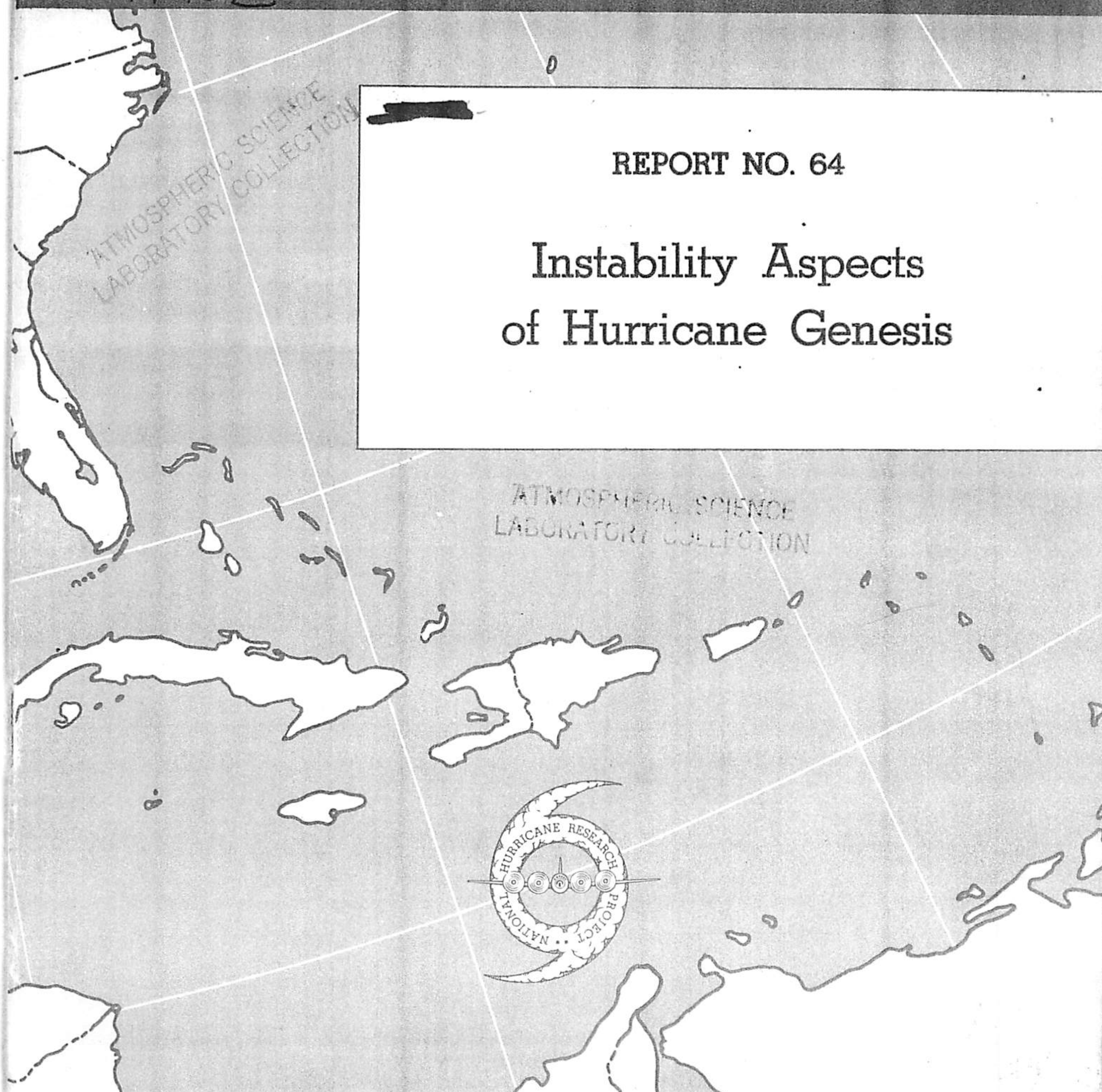

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REPORT NO. 64

Instability Aspects of Hurricane Genesis

CONTENTS

INSTABILITY ASPECTS OF HURRICANE GENESIS

M. A. Alaka
National Hurricane Research Project, Miami, Florida

ABSTRACT

The role played by various types of atmospheric instability in developing hurricanes is examined, and arguments are presented in support of the hypothesis that dynamic instability provides a trigger mechanism essential for development. This instability is shown to be released in the form of negative absolute anticyclonic winds when the upper-air anticyclone, under which hurricanes are known to form, reaches a critical intensity during the process of transformation of the initially cold-core disturbance into a warm-core system. From theoretical considerations, upper-air patterns favorable for hurricane formation and the optimum surface position of the disturbance with respect to these patterns are deduced and found to agree with observations and with previously published results.

A criterion for detecting areas of instability, based on a streamline analysis of the wind field in conjunction with the reported winds at particular stations, is developed and applied to the conditions prevailing during the formative stage of hurricane Carla (1961). It is found that intensification into a hurricane rapidly follows the occurrence of absolute anticyclonic winds over a sizable area in the upper troposphere, above and downstream from the surface position of the nascent storm.

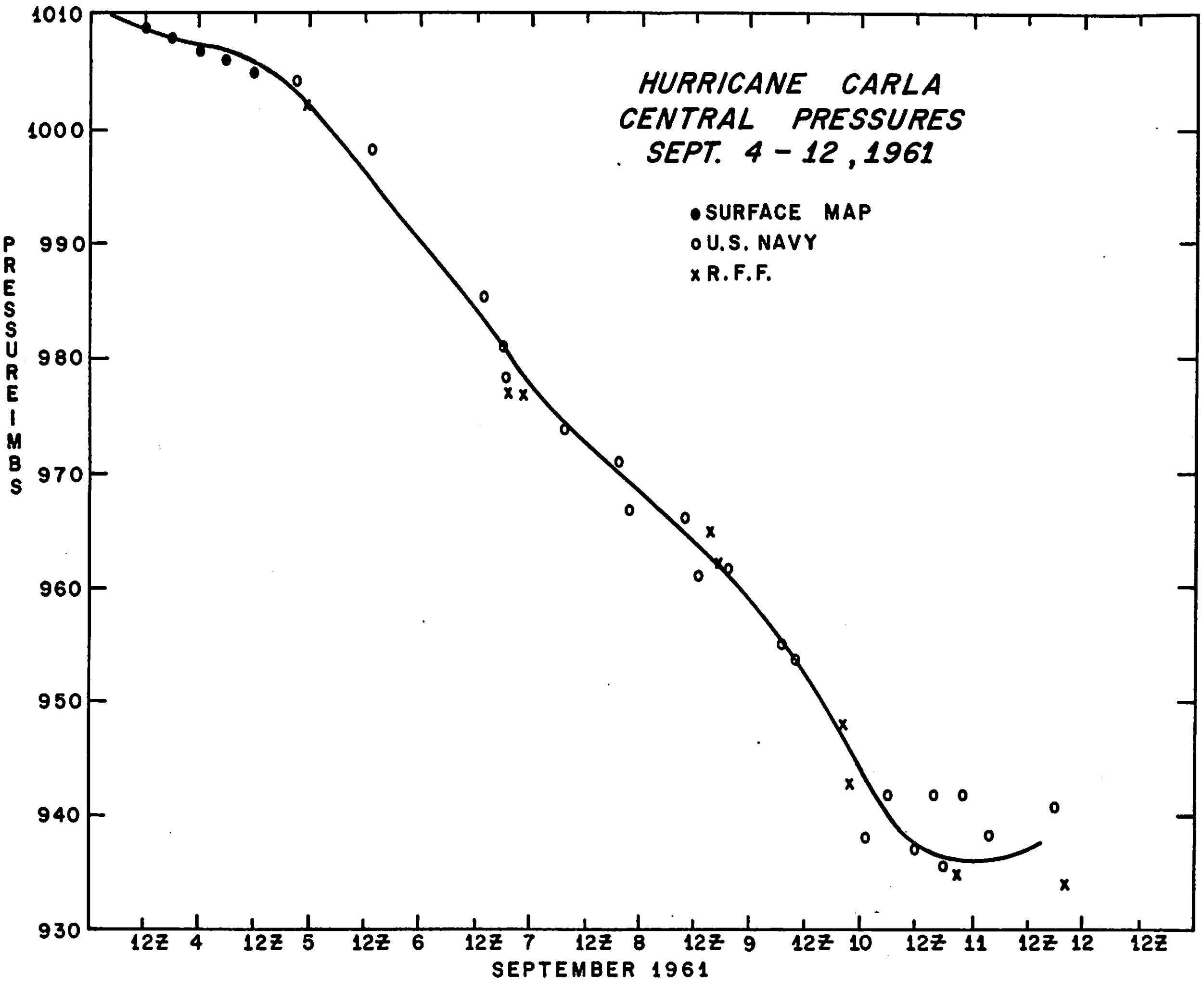


Figure 1. - Curve of central sea level pressure of hurricane Carla (1961) during its pre-formative and developing stages. The figure was adapted from one prepared by Dunn and Staff [7] with the addition of central pressures during the pre-formative period as estimated from surface synoptic maps.

1. INTRODUCTION

It has been observed that tropical disturbances which eventually develop into hurricanes often follow a typical course. A gradual intensification occurs until a short time before the eye is formed, then development accelerates considerably and deepening of the surface Low follows much more rapidly. This often sudden change of pace, clearly illustrated in figure 1, has suggested to meteorologists that the birth of a hurricane is marked by the release of an atmospheric "trigger," setting up processes which are self-maintaining and irreversible, as is evidenced by the apparent ability of hurricanes, once formed, to persist almost indefinitely, so long as they remain over tropical waters where they can draw into their circulation air with sufficient heat and moisture to replenish their basic energy source.

The nature of the triggering mechanism has been the subject of numerous studies. Our present purpose is to examine some of the pertinent ideas which have been presented in this respect and to offer further evidence that the development of dynamic instability is an integral part of the triggering mechanism.

2. INSTABILITY HYPOTHESES OF HURRICANE FORMATION

It was perhaps inevitable, in view of the observed change of pace of development at or about the time of hurricane genesis, that some form of destabilizing effect should be held to account for it. Indeed, gravitational (latent), dynamic (inertial, rotational) and baroclinic (Margulean) instabilities, singly or in combination, have been invoked by different authors in explanation of hurricane formation. A brief survey of the results of some pertinent studies and of hypotheses advanced in this connection follows.

Latent instability. Since hurricanes form over tropical oceans, where the latent energy is large, and since moreover, it is an established fact that the basic energy source of hurricanes is the latent heat of condensation, it was logical to attribute their formation to latent instability in the atmosphere. Thus Haque [12] and Syōno [30] sought to understand the mechanism of hurricane formation by studying the behavior of a circular vortex embedded in a latently unstable atmosphere. The applicability of their results to hurricane formation was, however, disputed by Lilly [20] who, in extension of the above studies, found that an unstably stratified atmosphere favored a more rapid growth of cloud-scale motions than motion on the scale of a hurricane. In addition, Lilly pointed out that the smaller-scale processes tend to impede the development of the large-scale disturbances because of the dry downdraft regions associated with them. Essentially, the same inference may be drawn from the results of other authors, including Kuo [17, 18] and Kasahara [13, 14]. These results indicate that with latent instability as the main driving mechanism, disturbances on the scale of cumulus clouds tend to predominate unless exchange coefficients are excessively large.

On the other hand, the hurricane circulation is characterized by the simultaneous occurrence of vigorous convection and larger scale circulations without any indication that the former dominates or impedes the latter. Instead, both the smaller-scale and the larger-scale motion form integral parts of the hurricane structure. It has, therefore, been concluded that the formation of tropical storms cannot be directly attributed to growth of small-amplitude perturbations in a conditionally unstable atmosphere, even though it is known that the energy of these storms is derived from the release of the latent heat of water vapor. Some other mechanism is needed to channel the release of latent energy of unstable atmospheric stratification into a large-scale circulation.

While the evidence at hand largely supports the conclusion that hurricane formation cannot be solely attributed to latent instability, the picture presented above is not entirely accurate. Kuo [19] has recently shown that nonlinear diffusion effects cause the disturbances which are maintained by unstable atmospheric stratification to approach a steady state of equilibrium at a rate proportional to the square root of the instability. In other words, nonlinear frictional effects prevent cumulus-scale motions from growing indefinitely and thus dominating the larger-scale circulations. Moreover, the importance of latent instability in hurricane formation should not be underestimated, especially during the early stages of development. Hurricanes are warm-core circulations, whereas the initial tropical disturbance from which the hurricane develops is usually cold-core at the start. Kasahara [13,14] has shown that the release of latent instability is effective in forming a warm-core type circulation with cyclonic convergent circulation in the lower levels of the updraft areas and anticyclonic divergent circulation aloft. Indeed, Kasahara's results indicate that the efficient establishment of the warm-core circulation is dependent on the existence of unstable stratification throughout the troposphere.

But the warming which usually starts in the upper troposphere and gradually seeps downward tends to increase the vertical stability which has been observed to become nearly neutral at about the time the hurricane is formed (Yanai [31]). Thus, some other mechanism must be responsible for the ultimate step in the birth of the hurricane.

Baroclinic instability. To gain insight into the nature of this mechanism, Kasahara [14] studied the stability properties of symmetric disturbances in a conditionally unstable atmosphere subject to the destabilizing effects of a horizontal temperature gradient and the released latent heat. He found that the horizontal scale of motion produced by this type of instability is of the order of 100 km., and that the associated growth rates would increase the amplitude of the perturbation by a factor of 2 to 3 in a few hours - which is compatible with the observed growth of hurricanes. The required horizontal temperature gradient is of the order of 0.5° - 1°C. per 100 km. in the middle troposphere. Kasahara noted that this type of instability tends to produce a deep tropospheric cyclonic circulation and would account for the observed extension to the upper troposphere of the cyclonic circulation in the inner core of the hurricane.

Kasahara concluded that as soon as the buoyant instability is neutralized in the developing disturbance, baroclinic (Margulean) instability may provide the main driving mechanism through the developing stage. The destabilizing effect of a horizontal temperature gradient as a mechanism for the development of hurricanes has also been suggested by Kleinschmidt [15] and, more recently, by Yanai [31].

Yanai considers hurricane (typhoon) formation as a transformation process by which the dynamically forced convection of the cold-core stage changes into a free convection. The transformation occurs under the destabilizing influence of what he terms inertial instability along isentropes, which is released when the baroclinicity of the developing vortex exceeds a critical value. For vortices of 10^3 km. or less the critical baroclinicity depends mainly on the static stability (S)*, which for a value of 20×10^{-3} $m.^2 mb.^{-2}$, requires a minimum temperature gradient of 1.7°C./100 km. at 200 mb. Yanai's criterion is similar to the criterion developed by Kuo [16] for the onset of violent Hadley type convection as observed in dishpan experiments and described by Fultz and collaborators [9].

*See definition on page 6.

Dynamic instability. The concept of dynamic instability as a factor in hurricane development has been advanced by several meteorologists. Sawyer [28] postulated the occurrence of dynamic instability in the form of negative absolute vorticity, generated in a developing large cumulus cloud or a group of such clouds, to account for the rapid upper-air evacuation needed to reduce surface pressure in the incipient storm. Riehl [22] noted that disturbances which eventually develop into hurricanes intensify selectively when they become situated under an upper anticyclone. This led Alaka [1] to suggest that the mechanism which operates in triggering development is inherent in the dynamic properties of this upper anticyclone and that intensification into a hurricane occurs when, under suitable circumstances, dynamic instability is generated by anticyclonic winds exceeding $f\,R_t/2$. The occurrence of these winds and their properties were discussed by the author, and a measure of observational evidence attesting their occurrence in incipient and developing hurricanes was recently presented (Alaka [2,3]).

The importance of dynamic instability in hurricane genesis was also discussed by Kuo [18] who found that under the influence of dynamic instability, both disturbances with large horizontal dimensions and smaller disturbances tend to grow. But the growth of the smaller disturbances is not large and these disturbances are more likely to be affected by eddy diffusion. However, the amplification rate and the energy factors are small and Kuo, therefore, suggested that the importance of dynamic instability lies in the channelling of energy supplied by other sources into large-scale motion and not in the energy associated with the instability.

In line with these results, Kasahara [14] has shown that dynamic instability mainly increases the perturbation kinetic energy of radial motion and hence tends to create mass divergence, which would be instrumental in organizing upper-air outflow in the developing hurricane. But the direct source for this kinetic energy increase is the tangential component of the kinetic energy of the basic flow and, since the latter is limited, the effective operation of dynamic instability is dependent on some other mechanism to replenish the basic tangential circulation. Since latent instability tends to produce an anticyclonic circulation in the upper levels, Kasahara has suggested that the latter type of instability may provide a maintenance mechanism for dynamic instability.

It should be pointed out that in all the above studies, the criterion used for dynamic instability is an approximation based on the assumption that baroclinic effects are negligible. Formally, dynamic instability occurs in a symmetrical, stationary vortex, when

$$[\zeta_a \left(\frac{2V}{R_t} + f \right)]_\theta < 0 \qquad (1)$$

where,

V = wind speed
R_t = radius of curvature of the air trajectories

omit Pot.T.

The subscript θ indicates that the quantity in brackets must be computed on an isentropic surface. In actual fact, in view of the generally small angle between isentropes and the horizontal, the above criterion is held to be valid when applied to a level surface.

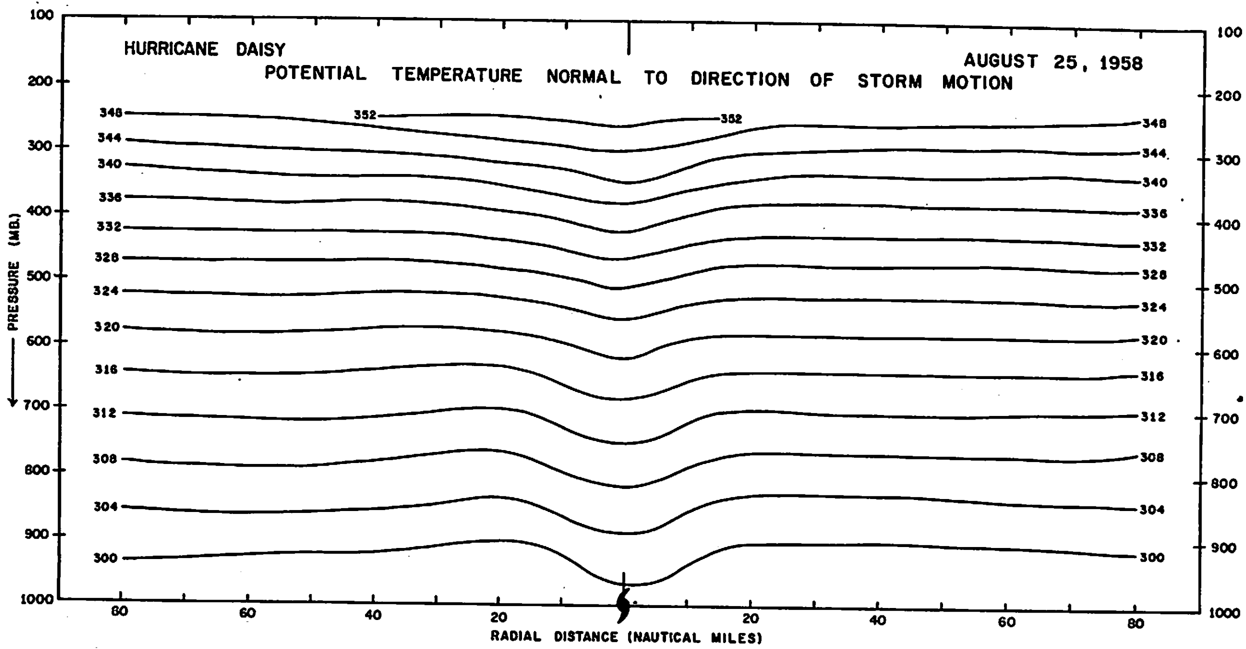


Figure 2. - Cross-section of the potential temperature (°K.) across hurricane Daisy on August 25, 1958, along a direction normal to the movement of the storm. The cross-section was constructed from observations made by Research Flight Facility aircraft flying quasi-simultaneously at 237, 570, and 830 mb., and two additional passes at lower levels (2800 and 1600 ft.).

If friction is neglected, Yanai's criterion for the growth of a symmetrical disturbance is given by

$$C = \left[\frac{B^2}{S} - \zeta_a \left(\frac{2\,V}{R_t} + f \right) - \left(\frac{1.2\ p_o}{r_o} \right)^2 S \right]^{1/2} \qquad (2)$$

where

C = a constant identified with the growth rate

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

$S = -\frac{RT}{p} \frac{\partial \ln \theta}{\partial p}$ (static stability)

p_o = the sea-level pressure

r_o = the outer radius of the closed vertical circulation

Thus the perturbation will grow if

$$\frac{B^2}{S} - \zeta_a \left(\frac{2V}{R_t} + f \right) > \left(\frac{1.2\ p_o}{r_o} \right)^2 S \qquad (3)$$

Yanai calls the expression to the left of the above inequality inertial instability along isentropes. The expression is, however, equivalent to that on the left side of inequality (1) which, as we mentioned, expresses the criterion for dynamic instability.

On the basis of his analysis of the development of typhoon Doris, 1958, Yanai[31] concludes that the baroclinic term (B^2/S) plays the determining role in satisfying inequality (3) and that the second term is negligible for vortices less than 10^3 km. in dimension. He finds that for a value of $S = 20 \times 10^{-3}$ m.^{2}mb.$^{-2}$sec.$^{-2}$, the minimum value of the baroclinicity corresponds to a temperature gradient of 1.7°C./100 km., a value which he considers attainable in a hurricane. For a barotropic vortex, the minimum value of $\zeta_a(\frac{2V}{R_t}+f)$ required to destabilize the vortex is 1.5×10^{-8} sec.$^{-2}$ if $S = 10^{-3}$m.^{2}mb.$^{-2}$sec.$^{-2}$ Yanai contends that such an excessive value is not encountered in the atmosphere.

Figure 2 represents a cross-section across hurricane Daisy on August 25, 1958, shortly after it developed an eye. It is seen that, aside from the innermost core of the storm where the isentropes show a slight dip, there is very little baroclinicity - indeed, not enough to satisfy Yanai's criterion. On the other hand, figure 3 shows the distribution of dynamic instability at 35,000 ft. (pressure altitude, U. S. Standard) around the core of the hurricane on the same day. The values shown are comparable with the minimum required by Yanai's criterion for destabilization.

Figure 3, however, is descriptive of conditions after the hurricane eye was formed and provides no certain clue as to whether dynamic instability occurred prior to the development of the hurricane, as such, and was thus instrumental in its formation, or whether it was a product of the hurricane itself. An unequivocal answer to this question can be made only by a study of the succession of events and processes which precede rather than follow the formation of the hurricane.

3. STAGES OF HURRICANE DEVELOPMENT

There are three main stages in the development of hurricanes:

The initial cold-core stage. Whether the initial disturbance is a wave in the easterlies, a shear line in the equatorial trough, a low-pressure trough previously associated with a front which had moved into the Tropics (Riehl [24]), or whether it occurs in any other form, it is usually a cold-core type circulation. Intensification into a hurricane is dependent on the transformation of the circulation to a warm core type, which would permit the conversion of the latent heat of condensation into the kinetic energy of the circulation (Palmén and Jordan [21]).

The warming stage. Even though precipitation in the Tropics tends to occur in and around cold-core troughs and cyclones, there is evidence that under suitable conditions, condensation heating may result in the development of an upper anticyclonic circulation which represents a warm-core circulation. Riehl and Burgner [26] and, more recently, Frank [8], have described two such cases. Frequently the warming process appears to be initiated by the drift of the low-level disturbance beneath a high-tropospheric ridge, as was shown by Riehl [25]. Yanai [31], in his more recent analysis of the development of typhoon Doris, 1958, similarly found that warming took place when a large anticyclone was situated above the low-level disturbance. As mentioned ear-

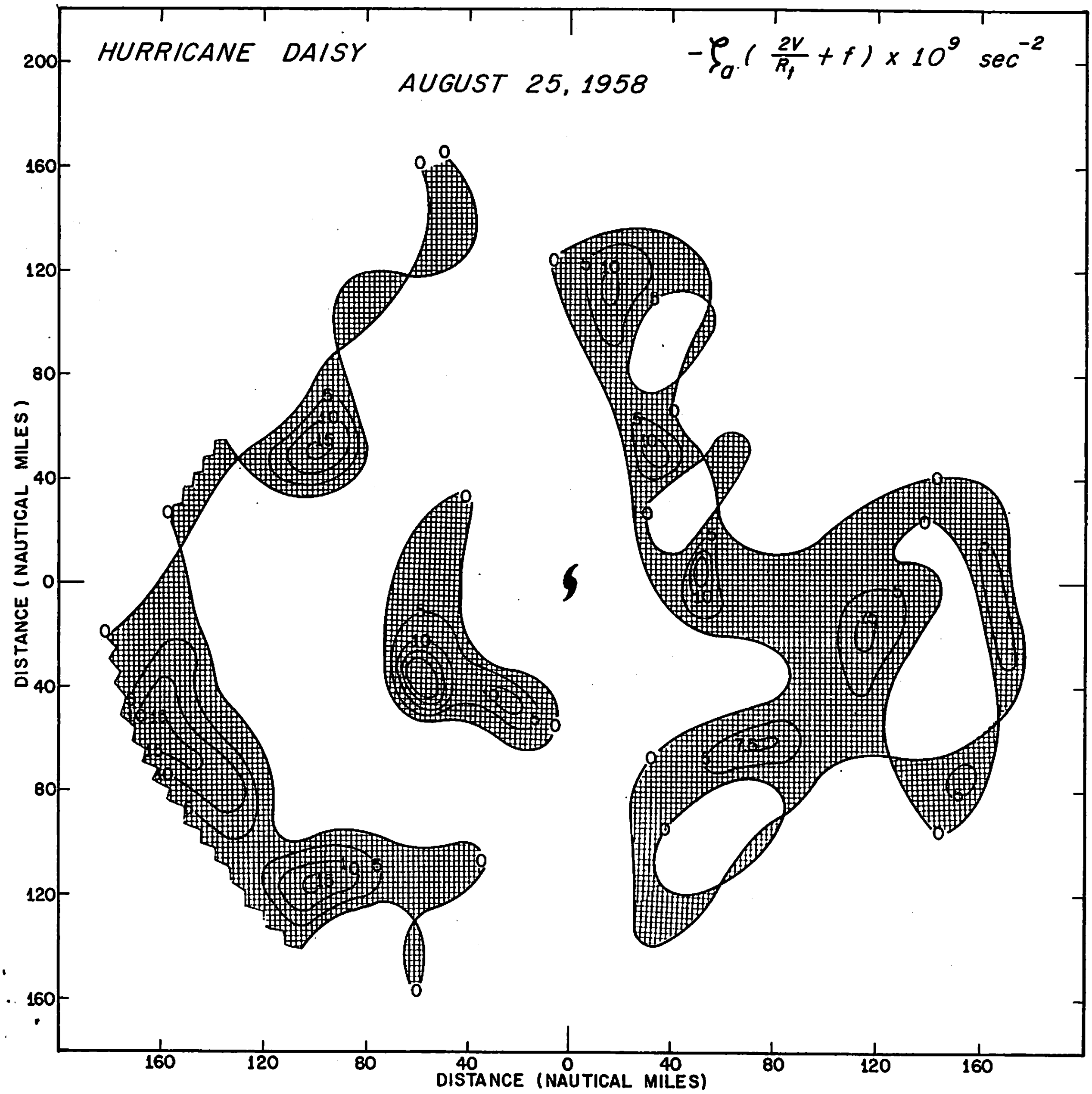


Figure 3. - Distribution of dynamic instability around hurricane Daisy at 237 mb. on August 25, 1958, computed from aircraft data.

lier, Yanai further observed that central warming proceeds from the upper to the lower troposphere.

<u>The "trigger."</u> Whether the warming occurs spontaneously through the release of latent heat or whether it is initiated in response to the drift of an upper anticyclone above the low-level disturbance, it is clear that the occurrence of an upper anticyclone above a low-level cyclone stimulates up-

Trigger effect

draft motion which, through the release of latent heat, warms the core of the circulation system. This, in turn, intensifies the upper anticyclone, so that there is a feed-back effect. It is observed that intensification into a hurricane, often very rapid, occurs at this stage, presumably when this process has continued until a critical condition is reached which releases the trigger mechanism. The trigger would thus appear to be activated by the effects of tropospheric warming beneath an upper anticyclone and, by the same token, must be inherent in the thermodynamic and dynamic properties of this anticyclone. To determine whether the trigger mechanism is in the nature of dynamic instability, it would be profitable to investigate the conditions attending the release of dynamic instability in intense anticyclonic vortices and then attempt to find out whether such conditions prevail at or immediately before the crucial time when the hurricane is formed.

4. INSTABILITY PROPERTIES OF ANTICYCLONIC VORTICES

According to inequality (1), dynamic instability is released if either, though not both, the absolute vorticity or the absolute angular momentum is negative. This criterion was derived for the special case of a stationary, circular, anticyclonic vortex in which the streamlines are identical with the particle trajectories. Hurricanes, on the other hand, develop under upper anticyclones which usually drift with the general circulation, and it is not a priori evident that this simplified criterion applies to such moving anticyclones. Indeed, it would be reasonable to expect that the criterion for the general case is more complicated, as is indicated by the results of Godson [11]. The latter, however, utilizes the geostrophic approximation in his derivation, thereby eliminating the all-important effect of the trajectory curvature. An exact and general criterion for the release of dynamic instability, taking into account the curvature effect, has so far defied derivation, but it would be reasonable to assume that the main factors of the stationary case, namely, negative absolute vorticity and negative absolute angular momentum should figure prominently albeit in a more complicated manner than in inequality (1). Such an assumption is justified by the results of Angell [4] who found that transosondes showed a distinct tendency for cross-contour flow where the absolute vorticity was negative ($\zeta_a < 0$) or where the absolute angular momentum was less than zero ($\frac{2V}{R_t} + f < 0$).

To test the validity of the hypothesis that dynamic instability is a factor in hurricane development, our plan is to investigate the preferred location of negative absolute vorticity and negative absolute angular momentum in moving anticyclones and the conditions which favor their occurrence, and then compare the findings with actual upper-air conditions during hurricane genesis.

If friction is neglected, and n and s are coordinates, respectively, normal to and along the direction of flow and if N is a coordinate in the direction of the horizontal pressure gradient, α is the cross-isobar angle of the wind, and R_t the trajectory radius of curvature, the equation for the horizontal component of air motion may be written

$$\frac{dV}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial s} = -\frac{1}{\rho}\frac{\partial p}{\partial N}\sin\alpha = b_s \tag{4}$$

and

$$\frac{V^2}{R_t} + fV = -\frac{1}{\rho}\frac{\partial p}{\partial n} = -\frac{1}{\rho}\frac{\partial p}{\partial N}\cos\alpha = b_n \tag{5}$$

From (5), if anticyclonic flow is considered and if the absolute value of R_t is used,

$$V = \frac{f R_t}{2} \pm \left[\left(\frac{f R_t}{2}\right)^2 - R_t b_n\right]^{1/2} \tag{6}$$

Consider an anticyclone with circular streamlines, moving without change of shape. If R_s and Ψ, respectively, denote the radius of streamline curvature and the angle from the direction of motion of the streamline system to the wind direction, and if the change of wind direction with height is neglected,

$$\frac{1}{R_t} = \frac{1}{R_s}\left(1 - \frac{C\cos\Psi}{V}\right) \tag{7}$$

where C is speed of movement.

Substituting for R_t in (6), we have

$$V = \sigma \pm [\sigma^2 - R_s b_n]^{1/2} \tag{8}$$

where

$$\sigma = \frac{f R_s + C\cos\Psi}{2} \tag{9}$$

and if $V_g' = V_g \cos\alpha$, where V_g is the geostrophic wind,

$$V = \sigma \pm [\sigma^2 - f R_s V_g']^{1/2} \tag{10}$$

Figure 4 shows the wind distribution in an anticyclone with circular streamlines, for a given value of $f R_s$ and various values of $C \cos\Psi$. The wind curves have two branches corresponding to the two solutions of equation (10) which meet at the point where the pseudo-geostrophic wind (V_g') reaches its maximum. Curve "c" corresponds to the stationary case and it is seen that the value $f R_t/2 = 2 V_g'$ coincides with the maximum value of V_g' which, if gradient equilibrium exists, is equal to the geostrophic wind V_g.

It has been customary among meteorologists to reject the larger root of equation (6) which, on the basis of the stationary case, would preclude the occurrence of anticyclonic winds greater than $f R_t/2$; i.e., winds equivalent to a negative absolute angular momentum. Indeed, such winds have been termed abnormal or anomalous. The validity of the exclusion of the larger root has been discussed elsewhere (Alaka [2]). But as was shown by Godson [10], in moving anticyclones, winds greater than $f R_t/2$ are not restricted to the larger root of equation (10); nor are winds less than $f R_t/2$ restricted to the smaller root.

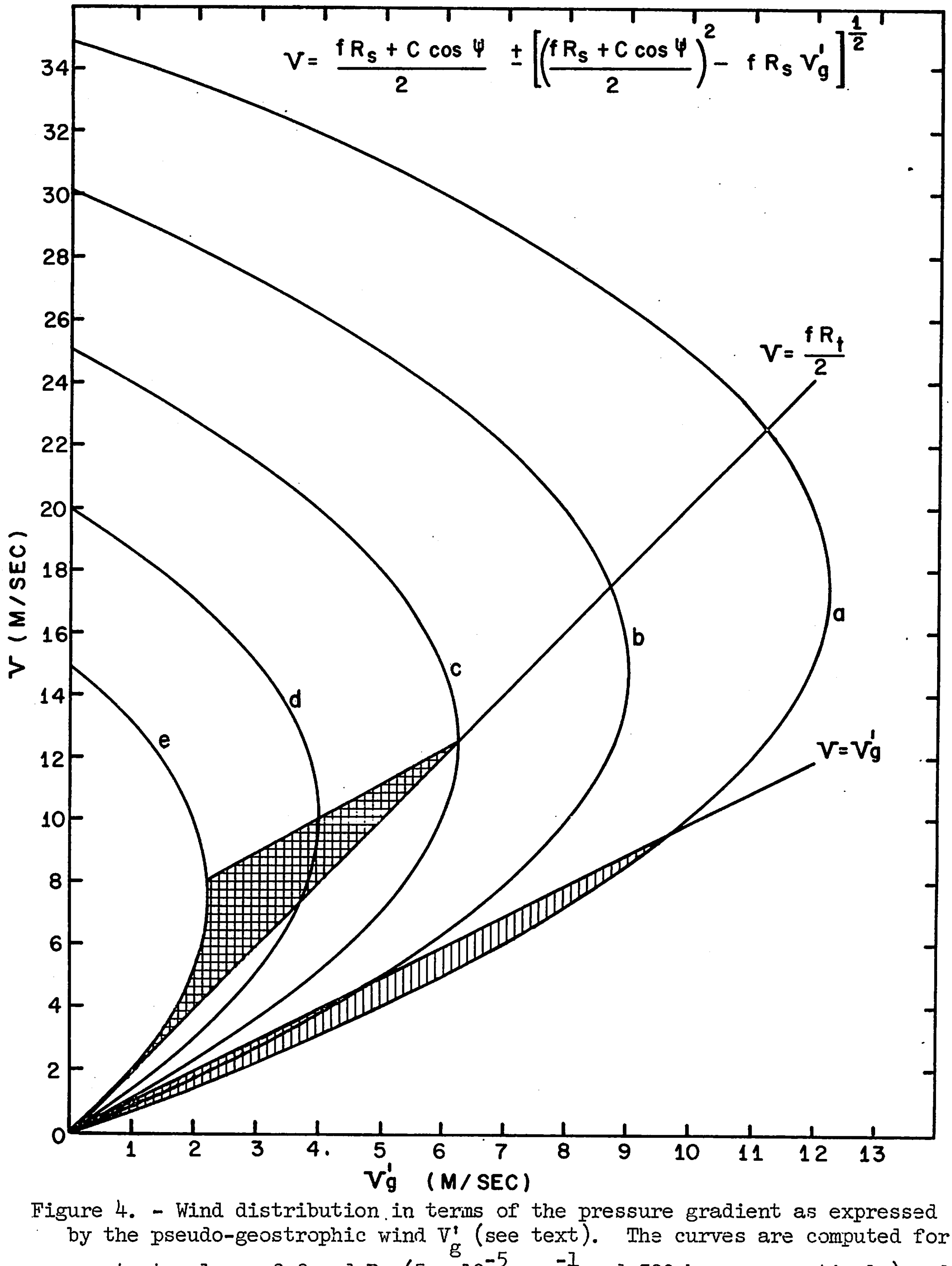


Figure 4. - Wind distribution in terms of the pressure gradient as expressed by the pseudo-geostrophic wind V_g' (see text). The curves are computed for constant values of f and R_s (5×10^{-5} sec.$^{-1}$ and 500 km., respectively) and for the following values of C cos Ψ : curve a: 10 m.p.s.; curve b: 5 m.p.s.; curve c: 0; curve d: -5 m.p.s.; and curve e: -10 m.p.s.

Curves "a" and "b" of figure 4 correspond to $C \cos \Psi > 0$, i.e., to the left of the moving anticyclone. Here even the larger root of equation (10) has winds numerically less than $f R_t/2$. The shaded portions of these curves correspond to wind speeds less than the pseudo-geostrophic wind speed and indicate a cyclonic curvature of the air trajectories at these points. On the other hand, curves "d" and "e" correspond to $C \cos \Psi < 0$. The shaded portions of these curves represent winds appropriate to the smaller root of equation (10); yet they are greater than $f R_t/2$. These winds occur near the maximum values of (V_g') which in view of the usually small cross-isobar angle of airflow, especially in the upper air, very nearly correspond to the maximum value of the geostrophic wind (V_g).

If we then visualize an anticyclone in which the pressure gradient at every point is a maximum, the wind at all points to the right of the path attains a value greater than $f R_t/2$. At the same time, the absolute vorticity is positive at these points, as can readily be deduced as follows:

From equation (10), if V_{max} denotes the wind corresponding to the maximum pseudo-geostrophic wind $V'_{g(max)}$

$$V_{max} = \sigma = \frac{f R_s + C \cos \Psi}{2} \tag{11}$$

$$-\zeta = \frac{V_{max}}{R_s} + \frac{\partial V_{max}}{\partial n} = f + \frac{C}{2}\frac{\cos \Psi}{R_s} \tag{12}$$

Since R_s and C are positive,

$$\zeta + f \lesseqgtr 0 \lesseqgtr \cos \Psi \tag{13}$$

We thus obtain the interesting result that in a moving maximum anticyclone, all points to the left of the path have negative absolute vorticity but positive absolute angular momentum (winds less than $f R_t/2$). On the other hand, at every point to the right of the path, the absolute vorticity is positive but the absolute angular momentum is negative (fig. 5a).

It may be of interest to investigate the manner in which the areas covered by negative absolute vorticity and negative absolute angular momentum change with changing pressure gradients. Let us, for instance, consider an anticyclone in which the pseudo-geostrophic wind at all points is equal to 3/4 the maximum. Denoting the corresponding wind by V', we have

$$V' = \sigma \pm \left(\sigma^2 - \frac{3\sigma^2}{4}\right)^{1/2} \tag{14}$$

Taking the smaller of the two roots,

$$V' = \frac{\sigma}{2} = \frac{f R_s + C \cos \Psi}{4} \tag{15}$$

Equation (15) indicates that, other factors being equal, an anticyclone with 3/4 the maximum pressure gradient has winds only half as strong as the maximum anticyclone. From (10), if we put $V^* = f R_t/2 = 2V_g'$

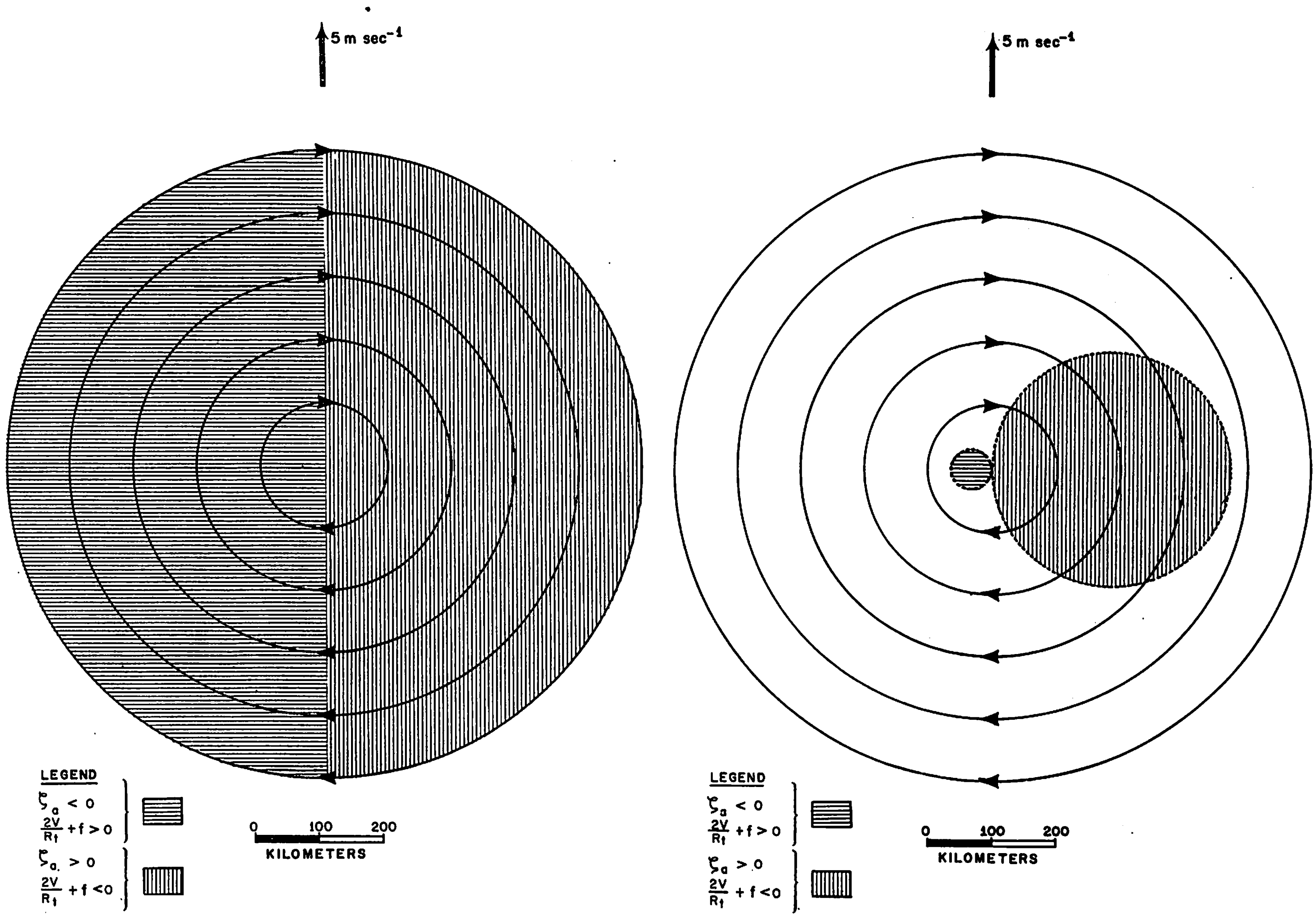


Figure 5. - (a) Distribution of negative absolute vorticity and negative absolute angular momentum in a moving maximum anticyclonic vortex with circular streamlines. (b) Distribution of negative absolute vorticity and negative absolute angular momentum in an anticyclone with circular streamlines, moving with a speed of 5 m.p.s. and having everywhere a pressure gradient equal to 3/4 of the maximum.

$$(2V_g' - \sigma)^2 = \sigma^2 - f\,R_s\,V_g'$$

and

$$V_g' = \frac{4\sigma - f\,R_s}{4}$$

Then

$$V^* = 2V_g' = \frac{f\,R_s + 2C\cos\Psi}{2} \qquad (16)$$

From (15) and (16), the condition that V' exceeds f $R_t/2$ is

$$\frac{f\,R_s + C\cos\Psi}{4} > \frac{f\,R_s + 2C\cos\Psi}{2} \qquad (17)$$

or

$$R_s < -\frac{3C\cos\Psi}{f} \qquad (18)$$

Figure 5b represents an anticyclone with circular streamlines and with 3/4 maximum intensity. If f is taken to be 5×10^{-5} sec.$^{-1}$ and C to be 5 m.p.s., the area in which the winds exceed f $R_t/2$ is the shaded area to the right of the direction of motion. It is seen that a sizable area is covered by such winds even when the prevailing pressure gradient is below the maximum.

By contrast, the area where the absolute vorticity is negative is rapidly reduced when the pressure gradients in the anticyclone are less than maximum. If ζ' denotes the relative vorticity associated with an anticyclone of 3/4 the maximum intensity

$$-\zeta' = \frac{V'}{R_s} + \frac{\partial V'}{\partial n} = \frac{f}{2} + \frac{C\cos\Psi}{4R_s}$$

For

$$\left.\begin{array}{r} +\zeta' + f < 0 \\ R_s < \frac{C}{2f}\cos\Psi \end{array}\right\} \qquad (19)$$

The area in which the absolute vorticity is negative is the shaded area to the left of the direction of motion in figure 5b. It amounts to a small fraction (1/36) of the area of negative absolute angular momentum to the right of the direction of motion. Yet as can be seen from figure 4, the pressure gradients required for an anticyclone of maximum intensity and, therefore, of 3/4 maximum intensity, are stronger to the left than to the right of the direction of motion, and the negative absolute angular momentum to the right of the direction of motion occurs not by virtue of the great speed of the anticyclonic wind but by virtue of the strong curvature of the trajectories of air particles.

5. HYPOTHESIS ON THE NATURE OF THE "TRIGGER" MECHANISM

It is clear from the above results that anticyclonic winds greater than $fR_t/2$ are not a rarity and that the term "abnormal" or "anomalous" applied to

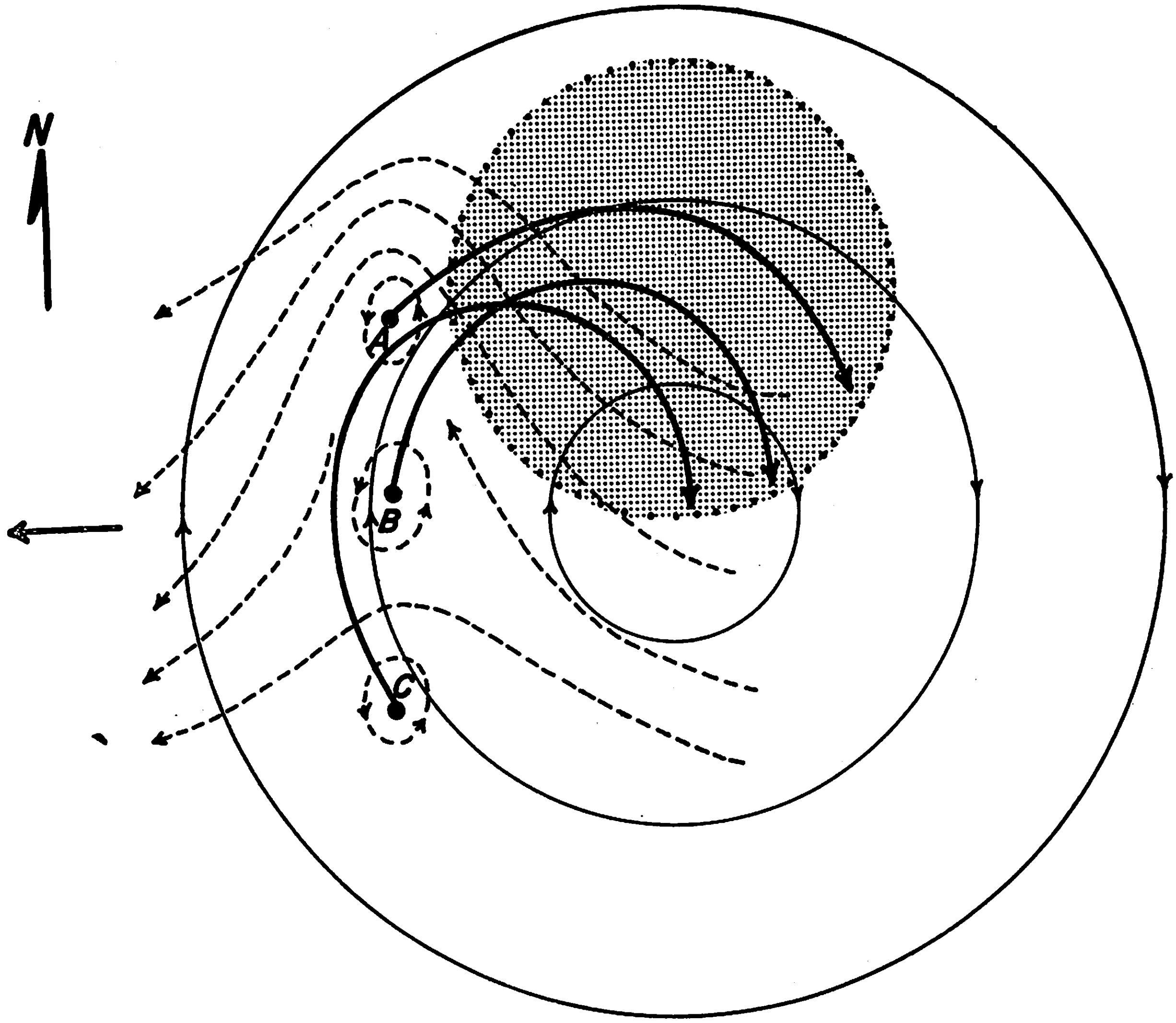


Figure 6. - An idealized picture of a westward-moving upper-tropospheric anticyclone showing the probable area of absolute anticyclonic winds (shaded). The dashed curves represent a low-level easterly wave with three closed circulation centers; the thick lines represent horizontal trajectories of air particles from the different centers.

these winds is a misnomer. A more appropriate designation would be "absolute anticyclonic winds" which we shall use henceforth in referring to these winds. It is equally clear that conditions for producing absolute anticyclonic winds in drifting upper Highs, such as those which are observed over the incipient hurricanes, are less stringent than those required for negative absolute vorticity. Therefore, if dynamic instability is a factor in hurricane development, it is more likely to be initiated by absolute anticylonic winds than by negative absolute vorticity.

6. VERIFICATION OF THE HYPOTHESIS

<u>General considerations.</u> Since the main role of dynamic instability is to facilitate upper-air outflow, it would be expected that the most favorable lo-

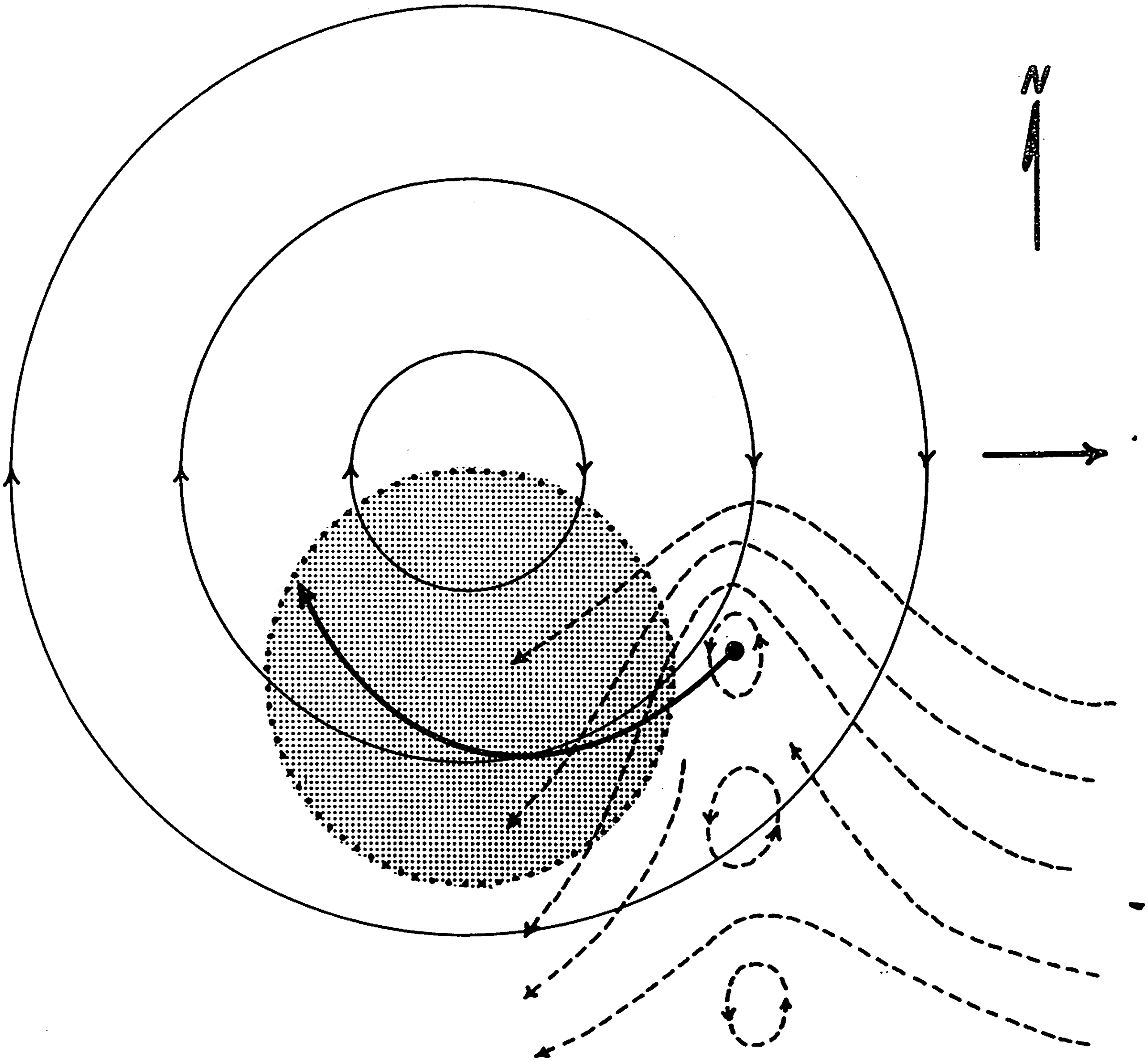


Figure 7. - Same as figure 6 for an eastward-moving anticyclone.

cation for incipient hurricanes should be such that air particles rising from the low-level disturbance will not be too far away from the area where dynamic instability prevails and will, moreover, have a long trajectory over this area. Thus the position A in figure 6 would be more favorable for development than positions B and C. This is compatible with results from a recent study by Colón and Nightingale [6] who found that a majority of disturbances, especially those at lower latitudes, which intensified under an upper anticyclone drifting in a westerly direction, reached hurricane intensity under a 200-mb. wind blowing from the southwest. The theoretically deduced cyclogenetic position A in figure 6 is also compatible with the findings of Riehl, Baer, and Veigas [27] who noted that hurricanes in the Gulf of Mexico formed only when the originating disturbance became located under the western fringe of a high-tropospheric anticyclone.

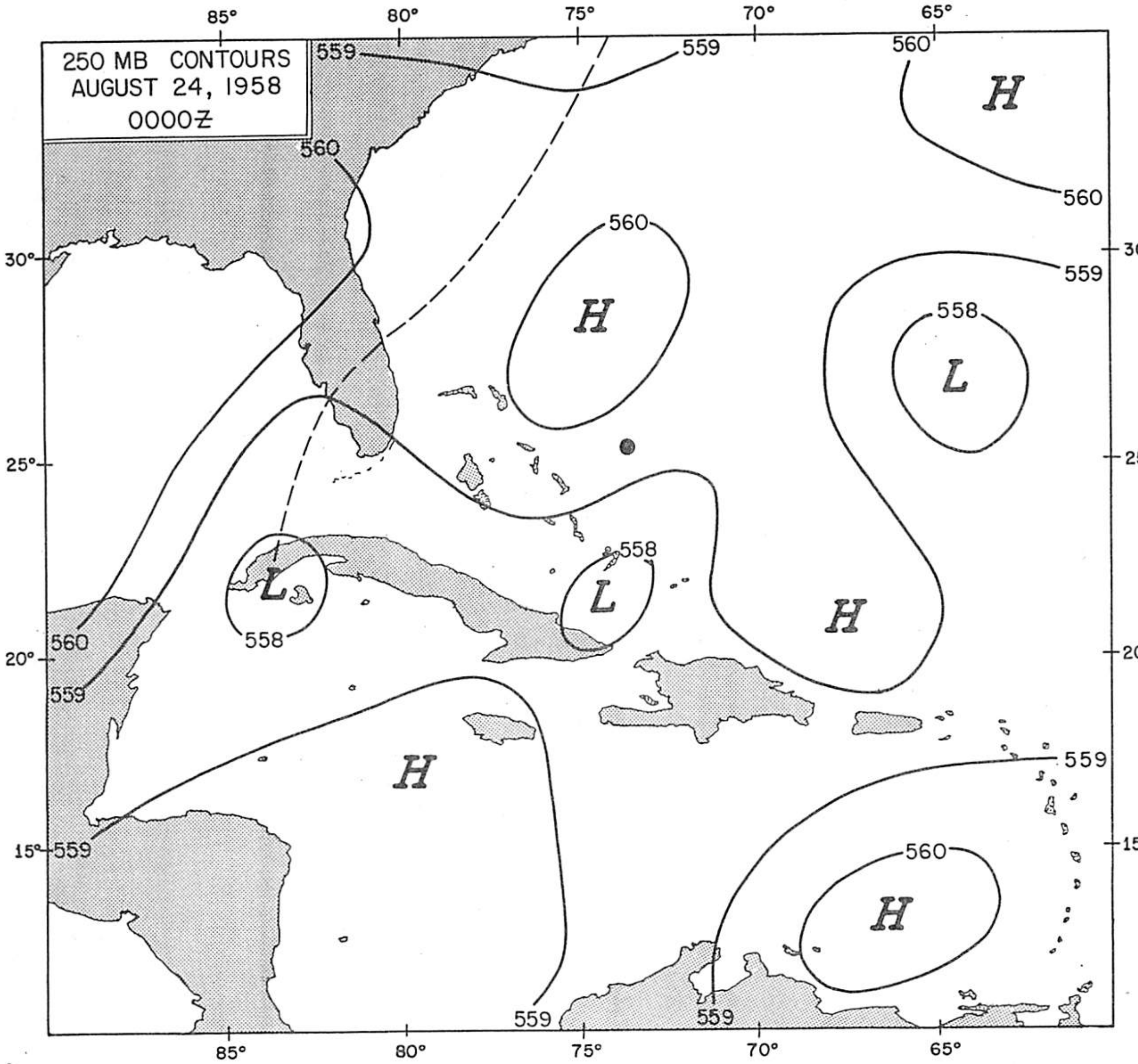


Figure 8. - 250-mb. contours on August 24, 1958 at 0000 GMT. Dot shows the surface position of hurricane Daisy in its incipient stage.

The dot in figure 7 shows a favorable location for intensification under an upper anticyclone drifting from west to east. This position agrees with that postulated by Riehl [23], and is probably mostly applicable to hurricanes which form at a relatively high latitude as in the case of hurricane Daisy, 1958 (fig. 8). This position also agrees with that found by Shinoda [29] in connection with typhoons forming in the vicinity of the Mariana Islands* away from the Intertropical Convergence Zone.

While the above considerations provide a qualitative support to the hypothesis that absolute anticyclonic winds are a factor in hurricane development,

*Information communicated orally by Kenji Shimada.

a definitive test of this hypothesis can be achieved only by applying the hypothesis to actual cases.

The case of hurricane Carla, 1961. Evidently a successful appraisal of the hypothesis requires an accurate upper-air analysis and this is not feasible for storms which develop far out at sea. Hurricane Carla, 1961, which formed in the northwestern Caribbean in a region where several upper-air stations exist, provides a case on which the ideas developed in the previous sections may be tested. Our aim is to find out whether, at the time of the formation of Carla, absolute anticyclonic upper winds did in fact prevail over a sizable area immediately downstream from the surface position of the nascent hurricane. At first sight, it would seem that the simplest way to delineate such an area is to find the locations where $V > 2V'_g \approx 2V_g$. The utilization of this criterion, however, depends on an accurate analysis of the upper contour field and this proved difficult, not so much because the observations were not numerous enough but mostly because they proved to be too inconsistent to permit a reliable contour analysis.

To circumvent this difficulty, we make use of equation (16) according to which absolute anticyclonic winds (V_a) satisfy the following criterion

$$V_a > \frac{f R_s + 2C \cos \Psi}{2} \tag{20}$$

The application of this criterion depends on an accurate streamline analysis which proved to be more feasible, especially with the help of isogons.

The analysis was performed on the mean wind for the layer 36,000-42,000 ft. and the results are shown in figures 9a-d. Mean layer winds rather than reported winds were chosen to eliminate short-period oscillations and features of shallow depth. Other advantages in tropical analysis of such winds over conventional reported winds were recently discussed by Colón and Zipser [5].

In figure 9, the areas where absolute anticyclonic winds prevail are stippled. The delineation of these areas was made on the basis of computations at locations and times given in table 1. The figures also show the position of the surface disturbance (black dot) plus the configuration of the innermost closed surface isobar (dashed lines).

Carla developed from a weak perturbation in the Intertropical Convergence Zone which manifested itself by a somewhat above-normal shower activity in the eastern Caribbean as early as September 1, 1961. The first indication of a closed circulation was noted on the 1200 GMT September 3 surface map and abnormal pressure and shower activity were mentioned in the tropical weather summary on that date (Dunn and Staff [7]). Rapid intensification, however, started only 24 hours later, as can be seen from figure 1.

Table 1. - Observed mean winds vs. computed criterion for absolute anticyclonic wind (m. $sec.^{-1}$) at various stations in the Caribbean.

Index Number	Station	September 4, 1961				September 5, 1961			
		0000 GMT		1200 GMT		0000 GMT		1200 GMT	
		V*	V	V*	V	V*	V	V*	V
501	Swan Island	36	35	17	16	22	14		
383	Grand Cayman	41	(33)	18	$\underline{29}$	30	20		
355	Camaguey	38	(46)			29	28	7	($\underline{25}$)
397	Kingston	43	24	31	$\underline{31}$	17	$\underline{19}$	0	$\underline{15}$
367	Guantanamo	28	(37)	18	$\underline{28}$	16	(10)	10	($\underline{28}$)
119	Turks Island	30	21	45	42	29	$\underline{37}$		
089	San Salvador			25	$\underline{49}$			60	(55)
201	Key West							17	$\underline{35}$
644	Merida							57	32

$$V^* = \frac{f\,R_s + 2C\cos\psi}{2}$$

V = Mean observed wind for the layer 37,000-42,000 ft. observed in the vicinity of an incipient hurricane.

Underlined observed mean winds exceed V* and are anticyclonic in space. Parentheses indicate winds at 200 mb.

Comparison of figures 9a and 9b is illuminating at this stage. We note that, on September 4 at 0000 GMT, the area of absolute anticyclonic wind is fairly small and, what is probably more important, far removed from the surface position of the disturbance. Twelve hours later this area had become much larger and its upstream edge much nearer the central portion of the disturbance. It is at this stage that conditions became critical as is shown by the marked steepening of the surface pressure curve of figure 1. Figure 9c shows that 12 hours later the upstream edge of the region of absolute anticyclonic winds was directly over the surface position of the disturbance and deepening continued at an even faster rate. At 0100 GMT (1 hour after map time), following aircraft reconnaissance, the first formal advisory was issued from the Miami Hurricane Center with a forecast for an increase to storm intensity, which was attained by 1000 GMT. Figure 9d shows the upper-air flow 2 hours later. By this time the surface cyclonic circulation had extended to the upper troposphere and, as noted by Dunn and Staff [7], the incipient hurricane had established an extensive outflow channel which permitted its continued development into a full-fledged storm in which peak gusts were estimated at 175 m.p.h. and the lowest reported pressure was 935 mb. It should be pointed out that the area of absolute and anticyclonic

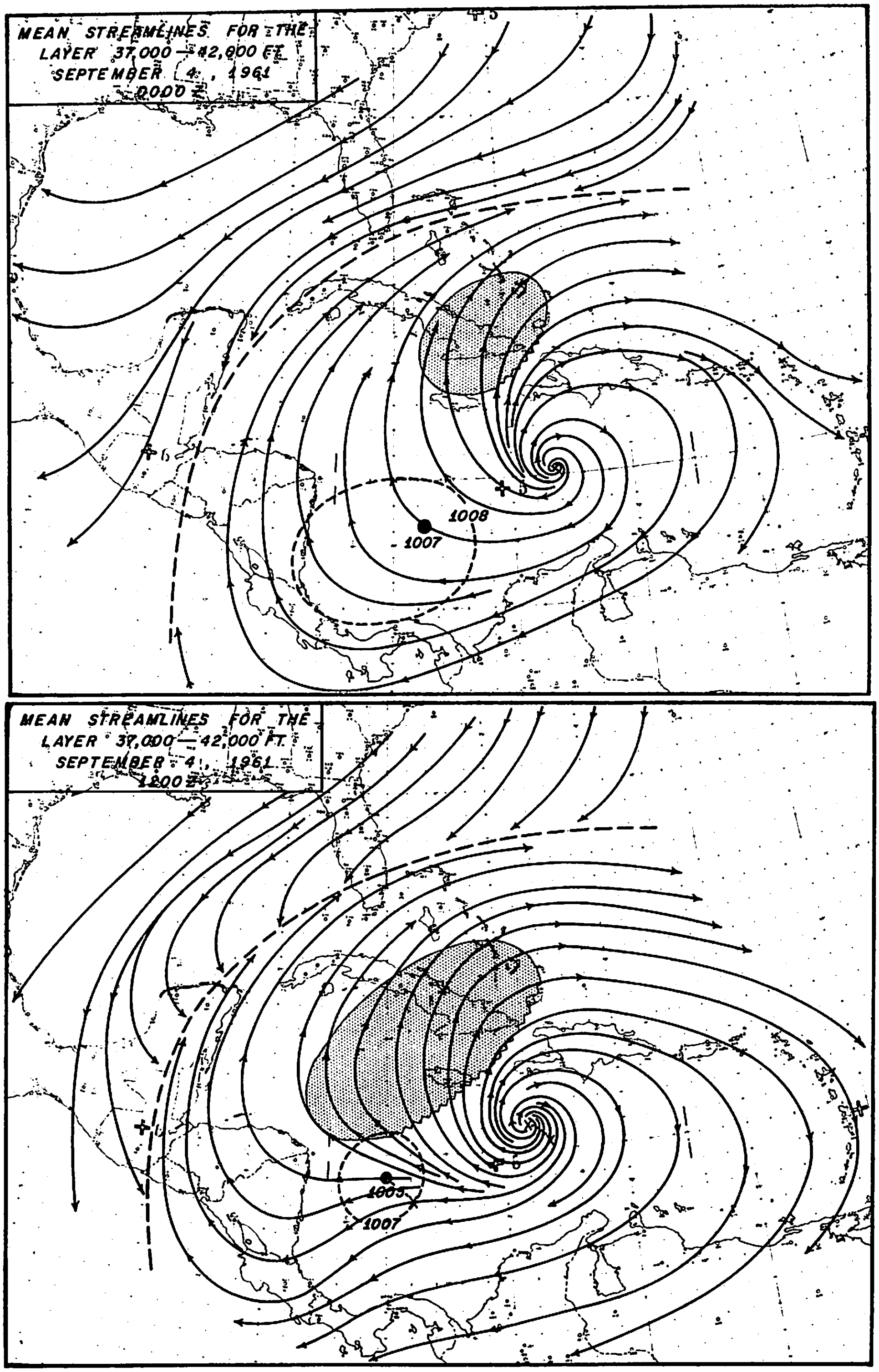
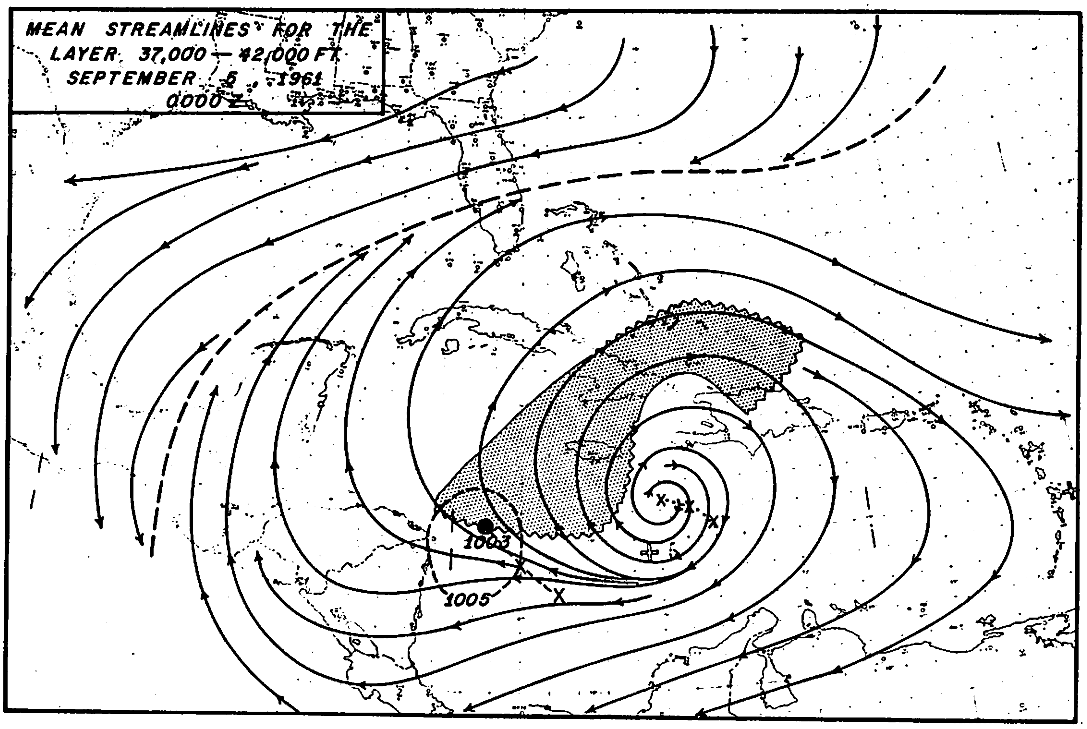
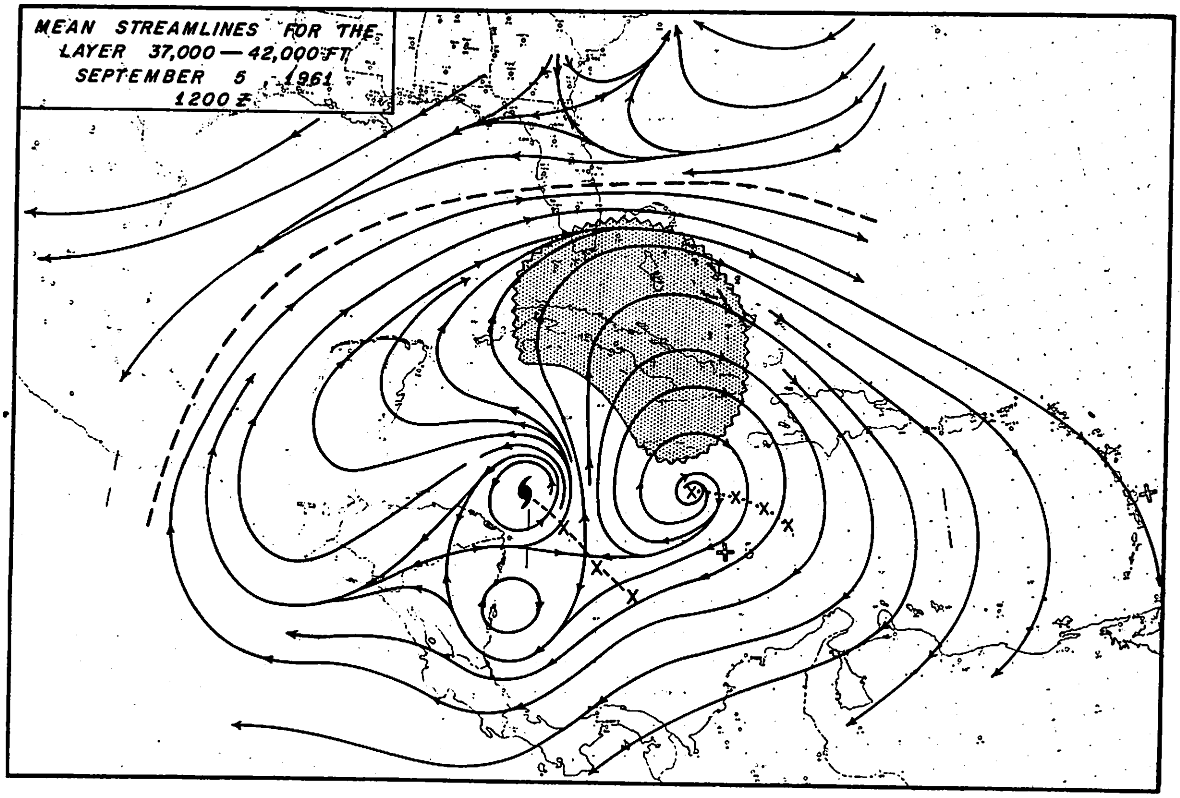


Figure 9. - Mean streamlines for the layer 37,000-42,000 ft. Shading covers area where absolute anticyclonic winds prevail; the pinked edge indicates uncertainty in delineating this area because of lack of observations. Dot represents the surface position of the disturbance and the surrounding dashed curve represents the innermost closed surface isobar. X's in-





dicate successive 12-hr. previous position of incipient hurricane and of upper anticyclone. (a) Mean streamlines for September 4, 1961, 0000 GMT. (b) Mean streamlines for September 4, 1961, 1200 GMT. (c) Mean streamlines for September 5, 1961, 0000 GMT. (d) Mean streamlines for September 5, 1961, 1200 GMT.

winds on September 5, 1200 GMT was probably more extensive than that depicted on figure 9d. Judging by the strong curvature of the streamlines, absolute anticyclonic winds may have prevailed to the west and southwest of the shaded area. This possibility which cannot, however, be ascertained owing to lack of data, is indicated on the figure by the pinked edges of the shaded area.

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REFERENCES

1. M. A. Alaka, "Dynamics of Upper-Air Outflow in Incipient Hurricanes," Geophysica, vol. 6, 1958, pp. 133-146.
2. M. A. Alaka, "The Occurrence of Anomalous Winds and Their Significance," Monthly Weather Review, vol. 89, No. 11, Nov. 1961, pp. 482-494.
3. M. A. Alaka, " On the Occurrence of Dynamic Instability in Incipient and Developing Hurricanes," Monthly Weather Review, vol. 90, No. 2, Feb. 1962, pp. 49-58.
4. J. K. Angell, "The Influence of Inertial Instability Upon Transosonde Trajectories and Some Forecast Implications," Monthly Weather Review, vol. 90, No. 6, June 1962, pp. 245-251.
5. J. A. Colón and E. J. Zipser, "Mean-Layer Wind Charts in Tropical Analysis," Monthly Weather Review, vol. 90, No. 11, Nov. 1962, pp. 465-470.
6. J. A. Colón and W. R. Nightingale, "Development of Tropical Cyclones in Relation to Circulation Patterns at the 200-Millibar Level," Monthly Weather Review, vol. 91, No. 7, July 1963, pp. 329-336
7. G. E. Dunn and Staff, "The Hurricane Season of 1961," Monthly Weather Review, vol. 90, No. 3, Mar. 1962, pp. 107-119.
8. N. Frank, "Synoptic Case Study of Tropical Cyclogenesis Utilizing TIROS Data," Monthly Weather Review, vol. 91, 1963 (to be published).
9. D. Fultz, R. R. Long, G. V. Owens, W. Bohan, R. Kaylor, and J. Weil, "Studies of Thermal Convection in a Rotating Cylinder With Some Implications for Large-Scale Atmospheric Motion," Meteorological Monographs, vol. 4, No. 21, American Meteorological Society, Boston, 1959.
10. W. L. Godson, "Some Remarks Concerning the Gradient Wind Equation," Bulletin of the American Meteorological Society, vol. 30, No. 12, Dec. 1949, pp. 342-346.
11. W. L. Godson, "Generalized Criteria for Dynamic Instability," Journal of Meteorology, vol. 7, No. 4, Aug. 1950, pp. 268-278.
12. S. M. Haque, "The Initiation of Cyclonic Circulation in a Vertically Unstable Stagnant Air Mass," Quarterly Journal of the Royal Meteorological Society, vol. 78, July 1952, pp. 394-406.
13. A. Kasahara, "A Numerical Experiment on the Development of a Tropical Cyclone," Journal of Meteorology, vol. 18, No. 3, June 1961, pp. 259-282.
14. A Kasahara, "A Study of the Stability of Thermally Driven and Frictionally Controlled Symmetrical Motions with Application to the Mechanism for Development of Tropical Cyclones," Dept. of Meteorology, University of Chicago, Technical Report No. 16, under Contract Cwb-9941, 1961, 63 pp.

15. E. Kleinschmidt, "Grundlagen einer Theorie der Tropischen Zyklonen," Archiv für Meteorologie, Geophysik und Bioklimatologie, Serie A, vol. 4, 1951, pp. 53-72.
16. H. L. Kuo, "Further Studies of Thermally Driven Motions in a Rotating Fluid," Journal of Meteorology, vol. 14, No. 6, Dec. 1957, pp 553-558.
17. H. L. Kuo, "Convection in Conditionally Unstable Atmosphere," Tellus, vol. 13, No. 4, Nov. 1961, pp. 441-459.
18. H. L. Kuo, "Mechanism Leading to Hurricane Formation," Proceedings of the Second Technical Conference on Hurricanes, June 27-30, 1961, Miami Beach, Fla., National Hurricane Research Project, Report No. 50, Part II, U. S. Weather Bureau, 1961, pp. 277-283.
19. H. L. Kuo, "Controlling Influences of Eddy Diffusion on Thermal Convection," Journal of Atmospheric Sciences, vol. 1, No. 3, May 1962, pp. 236-243.
20. D. K. Lilly, "On the Theory of Disturbances in a Conditionally Unstable Atmosphere," Monthly Weather Review, vol. 88, No. 1, Jan. 1960, pp. 1-17.
21. E. Palmén, and C. L. Jordan, "Note on the Release of Kinetic Energy in Tropical Cyclones," Tellus, vol. 7, No. 2, May 1955, pp. 186-188.
22. H. Riehl, "On the Formation of Typhoons," Journal of Meteorology, vol. 5, No. 6, Dec. 1948, pp. 247-264.
23. H. Riehl, "A Model of Hurricane Formation," Journal of Applied Physics, vol. 21, No. 9, Sept. 1950, pp. 917-925.
24. H. Riehl, Tropical Meteorology, McGraw-Hill Book Co., Inc., New York, 1954, 392 pp.
25. H. Riehl, "On the Production of Kinetic Energy from Condensation Heating," The Rossby Memorial Volume, The Rockefeller Institute Press and Oxford University Press, New York, 1959, pp. 381-399.
26. H. Riehl and N. M. Burgner, "Further Studies of the Movement and Formation of Hurricanes and Their Forecasting," Bulletin of the American Meteorological Society, vol. 31, No. 7, Sept. 1950, pp. 244-253.
27. H. Riehl, F. Baer, and K. Veigas, "Hurricane Formation in the Gulf of Mexico," Third Technical Report prepared for the American Petroleum Institute (Panel on Hurricanes), 1962, 25 pp.
28. J. S. Sawyer, "Notes on the Theory of Tropical Cyclones," Quarterly Journal of the Royal Meteorological Society, vol. 73, 1947, pp. 101-126.
29. M. Shinoda, "Some Aspects of the Incipient Stage of Typhoons Generated in the Seas Adjacent to the Mariana Islands," Journal of Meteorological Research of Japan, vol. 10, No. 5, May 1958, pp. 429-436.
30. S. Synōno, "On the Formation of Tropical Cyclones," Tellus, vol. 5, No. 2, May 1953, 179-195.
31. M. Yanai, "A Detailed Analysis of Typhoon Formation," Journal of the Meteorological Society of Japan, vol. 39, No. 4, Aug. 1961, pp. 187-214.
32. M. Yanai, " Dynamical Aspects of Typhoon Formation," Journal of the Meteorological Society of Japan, vol. 39, No. 5, Oct. 1961, pp. 282-309.