Further Studies on the Origin of Hurricanes

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Introduction

While there is no question that release of latent heat of condensation is a major factor in driving mature hurricanes, uncertainty has remained about its role in hurricane formation and the cause of large intensity increases sometimes experienced by a hurricane which had persisted on some plateau for several days. As shown by Riehl (1954) surface pressure cannot fall below about 1,000 mb if the whole troposphere is filled by ascent of air with the properties of the subcloud layer in the mean tropical atmosphere (Jordan 1958) and if the height of the 150-mb surface -- roughly the level where the parcel ascent intersects the mean tropical summer atmosphere curve -- remains constant. Further pressure fall leading to hurricane-type pressures and pressure gradients requires warming of the tropospheric mean virtual temperature. Especially in the early stages this will happen only if the typical energy content of the ascending air, measured say in terms of equivalent potential temperature Θ_e, increases beyond the 350K of the undisturbed subcloud layer.

It has never been easy to see how such an increment can be obtained from internal storm characteristics only. Evidence concerning the importance of downdrafts in reducing the energy of the surface layer in tropical synoptic systems (Riehl 1965) has aggravated the difficulty. The author (Riehl 1948, 1950) showed early that local baroclinity existed in cyclogenetic areas and that part of the baroclinity was released when a hurricane formed. The principle is that put forward by V. Bjerknes et al many years ago (1933) though the actual arrangement as determined from sounding balloons and aircraft traverses is rather different. Sinking of a cold air mass occurs largely in the layer 500-200 mb rather than lower down. Various analysts, noting the disappearance of upper cold lows, have spoken of transformation of cold core into warm core systems. While this statement exaggerates the situation -- the place where the cold mass sinks normally is at least 500 km from the forming surface center -- it nevertheless contains a large element of truth.

Fig. 1 illustrates the process thought to take place within 500-1,000 km from an initial tropical disturbance. Cold advection in the middle and upper troposphere takes place against the outskirts of this disturbance, resulting in a high-tropospheric pressure field as sketched in the illustration. The exact geometry is unimportant; sometimes the cold advection occurs in other positions. In Fig. 1 $\mathbb{V}$ is the wind vector, V_s its component along the isobars, V_n perpendicular to them, f is the Coriolis parameter, k curvature of trajectories, p pressure and ρ density. The flow may be initially balanced; following strengthening of the pressure gradient force through arrival of the upper cold layer the flow becomes subgradient and is accelerated toward lower pressure. In a typical case (Riehl 1959) a small jet stream with 60 knots central velocity was generated during such cross-isobar flow (Fig. 2). An acceleration of circulation in the vertical cross section is initiated which also strengthens the low level inflow toward the disturbance center, perhaps even including the low troposphere. The angle of inflow may become very sharp (Fig. 3), even 90 degrees at the outskirts. The equation for horizontal kinetic energy

$$\frac{dK}{dt} = -\frac{1}{\rho}\mathbb{V}\cdot\nabla p + \mathbb{V}\cdot\mathbb{F} \qquad (1)$$

where $-\frac{1}{\rho}\mathbb{V}\cdot\nabla p = -\frac{1}{\rho} v_n \frac{\partial p}{\partial n}$ is the generation term, $\frac{1}{2}\, dV^2/dt = dK/dt$, the substantial change in kinetic energy per unit mass and $\mathbb{V}\cdot\mathbb{F}$ the frictional dissipation, Rewriting equation 1

$$\frac{dK}{dt}\Big/\mathbb{V}\cdot\mathbb{F} = -\frac{v_n}{\rho}\frac{\partial p}{\partial n}\Big/\mathbb{V}\cdot\mathbb{F} + 1 \qquad (2)$$

where the lefthand ratio indicates the efficiency of the production term in actually generating kinetic energy. In most large-scale circulation systems, notably the trades (Riehl and Malkus 1957), $dK/dt \to 0$; the work done by pressure forces is entirely used up in balancing frictional dissipation. In the mature hurricane, $dK/dt/\mathbb{V}\cdot\mathbb{F} \approx 1$, probably the largest observed ratio in the surface friction layer (Riehl 1963).

If, in a given system, the inflow (cross-isobar) angle α increases, the production term will increase, other quantities remaining equal, since $v_n = V \sin \alpha$. Then the ratio $dk/dt/\mathbb{V}\cdot\mathbb{F}$ must also increase, at least temporarily. The existing pressure gradient is used effectively for actually producing increases in kinetic energy on short trajectories from high to low pressure, as sketched in Fig. 3. There, the dashed trajectory indicates a case where air converging toward an initial center may move inward at constant kinetic energy on a long trajectory where dissipation fully balances generation. When the inflow trajectory changes to that given by the solid line, a marked increase in kinetic energy will take place for the same pressure field. The importance of a change in inflow angle already has been demonstrated for the mature hurricane by Malkus and Riehl (1960); using the dynamical model proposed the maximum wind increases from 55 to 87 m/sec when the inflow angle changes from 20 to 25 degrees.

Fig. 4 illustrates the change in surface inflow from the dashed to the solid curve in Fig. 3 for an actual case (Riehl 1948). A weak initial disturbance was located in the central Gulf of Mexico on 10-11 Sept. 1941. Later on that day, winds over Florida and the Bahamas backed quickly to northeast, and the trajectories assumed sharp clockwise curvature with large cross-isobar component on their way into the Gulf. Very soon the tropical depression in the central Gulf became a tropical storm.

With increased wind speed the transfer of latent and sensible heat from the ocean must also increase assuming validity of the turbulence bulk transfer formulae. Thereby, inward warming in the subcloud layer will contribute to the pressure gradient as shown by Riehl and Malkus (1957) for the trade winds. Further, the ascent of air in cumulus clouds will shift to higher θ_e, and this can lead to the mid-tropospheric warming as observed for the early deepening stage, provided that the ascent does not take place purely in hot towers but that some of the cloud energy is mixed outward by entrainment.

Lacking adequate upper-air observations in 1941, it cannot be proved for Fig. 4 that the origin of the increased cross-isobar flow came from an acceleration of circulation as postulated initially. Falling pressure over the Bahamas does suggest the approach of an upper cold Low from the east, as indicated in Fig. 2. For more definitive computations, it is preferably to resort to later cases, when upper-air data, especially of wind, are more plentiful. Even then, the network still is very marginal for carrying out computations which demand knowledge of delicate parameters from the observational standpoint, notably the field of divergence and cross-isobar wind components. Nevertheless, an attempt was made to analyze these cases qualitatively as far as possible.

Hurricane of October 1949

Attention was drawn early to the role of subsiding cold air during the spectacular period 29 September - 1 October, 1949 (Riehl and Burgner 1950). On 29 September a deep upper trough was situated over the central United States, extending into the Gulf of Mexico (Fig. 5). Two days later, (Fig. 6) only weak remains of this trough are found over the northeastern coast whereas a large ridge has built over the Gulf. A storm center, about to become hurricane, was emerging from the Yucatan pensinsula into the western Gulf. Evidence from other levels published in the 1950 article confirm the impression of a subsiding upper cold dome coupled with tropical cyclogenesis about 1,000 miles farther south.

Hurricane Audrey (1957)

The antecents of this severe hurricane may be followed from 24 June 1957 when, at 200 mb, a deep upper trough was overlying the United States and the Gulf, a little farther west than in the previous case (Fig. 7). In this instance the trough retrograded and could be found west of the Texas coast on 25 June 1957, 12Z (Fig. 8) while strong cold advection was taking place on its eastern side in the layer 500-200 mb (Fig. 9). In the next 24 hours the trough retrograded even farther at low latitudes, while the hurricane rapidly intensified.

Hurricane Camille (1969)

Various descriptions of this exceptionally severe hurricane have appeared. From the time history (Fig. 10) central pressure had dropped below 1000 mb on 14 August. Then a very sharp fall to near record values ensued on the subsequent two days. The author's attention at the time was attracted to the synoptic situation over the southern United States which was not unlike that of the two cases just shown. There, a closed upper Low had developed from an upper trough with remarkable baroclinity and very low temperatures for the latitude (Fig. 11). The case represents a rare circulation anomaly for the southern United States in August. Two days later, on 16 August (Fig. 12) a very sharp change had occurred. Major sinking of the cold dome had taken place, so that neither the 200-mb Low and cyclonic circulation nor the associated temperature field were still to be found. Only a sharp cyclonic shear zone with very high relative vorticity remained, extending north-south to the vicinity of the approaching center of Camille. It will be observed that all cases so far described have very similar physical characteristics, even though the geometries varied.

Hurricane Edith (1971)

Figs. 13-18 illustrate a development at very low latitudes where a connection with large-scale middle latitude events was not obvious. From the track (Fig. 13) one gathers that principal deepening occurred after 8 September 0600Z. The 200-mb charts on 7 September 12Z and 8 September 00Z (Figs. 14-15) indicate westward progression of two upper anticyclones with a definite trough between them, a frequent feature. At the latter time cold air in the layer 500-200 mb was being advected in relative motion against the western flank of the forming tropical depression (Fig. 16) Twelve hours later, when deepening was accelerating, the upper-air picture had changed completely. The NE-SW oriented trough had disappeared (Fig. 17) together with the cold layer (Fig. 18). Only a northward temperature gradient and a relative wind blowing along the isotherms remained. Again, the baroclinic processes in the early storm phase are synoptically demonstrable.

Tropical Storm Fern (1971)

This circulation was situated near the Texas coast on 8 September just to the right of a high-level trough extending into the Gulf (Fig. 19). The thickness 500-200 mb has been superimposed on the 200-mb flow for the period 12 hours later in Fig. 20. The trough is observed to be very sharp and narrow, with some cold advection on its western side. Fern increased slightly and was called "hurricane" for some time. It remained, however, near minimum intensity.

Tropical Storm Irene (1971)

This storm intensified slowly after traversing the whole Caribbean on its southern edge. On 18 September it lag in a pronounced "delta" of the 200-mb flow (Fig. 21) on a southwesterly course. Actual winds for the layer 500-200 mb indicate strong cold advection from a cold center north of the Antilles (Fig. 22). In relative motion, however, this advection is minimized (Fig. 23) because of the southwesterly path. Slow deepening took place, and the storm may well have reached minimal hurricane intensity had it not entered land. Conditions for intensification, however, were not nearly as favorable as in the case Edith described earlier.

Hurricane Ella (1958)

As a single case of non-development the case of Ella (1958) may be offered. This circulation had been a hurricane in the Caribbean and then weakened to tropical storm strength while traversing most of Cuba on a westward path. Redevelopment was expected as soon as the cyclone entered the Gulf of Mexico. Intensive air reconnaissance, however, clearly denied redevelopment of a hurricane center, although the surface map (Fig. 24) looked very favorable. Ella entered the west coast of the Gulf as a weak center. At 200-mb, a large anticyclone was present indicating the warm-core structure of the surface system (Fig. 25). But the baroclinic processes, stressed above, were absent or situated at too great a distance to be effective in hurricane regeneration. Release of latent heat alone was an insufficient factor.

Conclusions

Many additional cases from older or recent times could be cited in support of the deductions of this article. A baroclinic transformation, sinking of an upper-tropospheric cold dome at a distance of 5 degrees to perhaps as much as 15 degrees of an initial tropical disturbance in the lower and middle troposphere, precedes acceleration of the circulation through the low-level system resulting in a hurricane. There are qualitative forecasting aids: the marked sinking, resulting in strong weakening or disappearance of the upper cold dome, precedes the period of strongest surface pressure fall by about a day: the intensity of the baroclinic release appears to be correlated to the subsequent hurricane strength for given latitude belts, as seen, for instance, in the case of Camille of 1969.

In order to demonstrate the physical transformations quantitatively, all terms in the kinetic energy equation were evaluated for several cases, but the results showed high computational instability from one 12-hour period to the next. The probable cause is poor definition of the field of motion for sensitive terms that depend on knowledge of the ageostrophic flow. The field of motion was generated at 100-mb intervals for the troposphere with hand analysis up to a distance of 20 degrees from the tropical disturbance center whenever possible. Vertical and horizontal smoothing of the data by electronic computer was applied. No improvement of the calculations resulted, so that it is not considered worth while to present them here. Either a much better network of observations or a greatly improved method for determining fields of horizontal divergence and vertical motions must be utilized.

Acknowledgment

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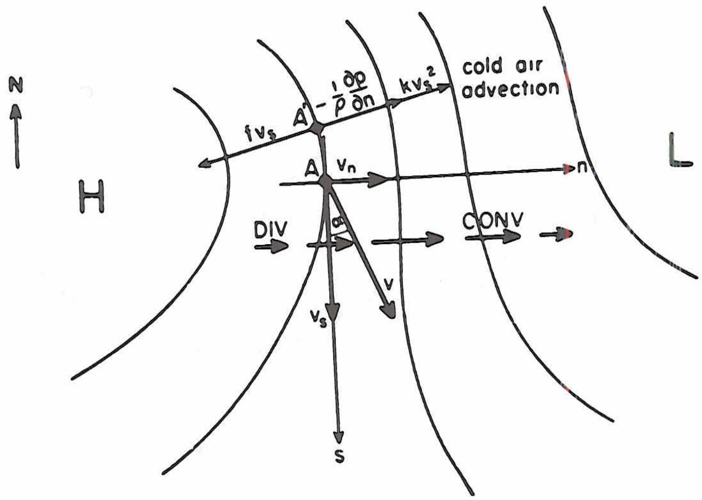

Fig. 1   Sketch of high-tropospheric pressure field and flow in the initial stages of hurricane deepening (Riehl 1950). For explanation of symbols see text.

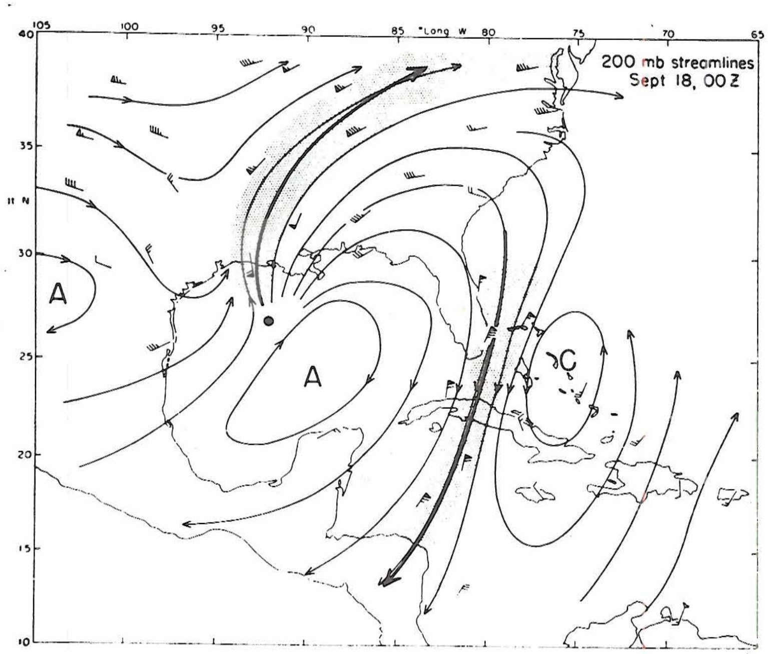

Fig. 2   Streamlines at 200 mb, 18 Sept. 1957, 00Z (Riehl 1959)

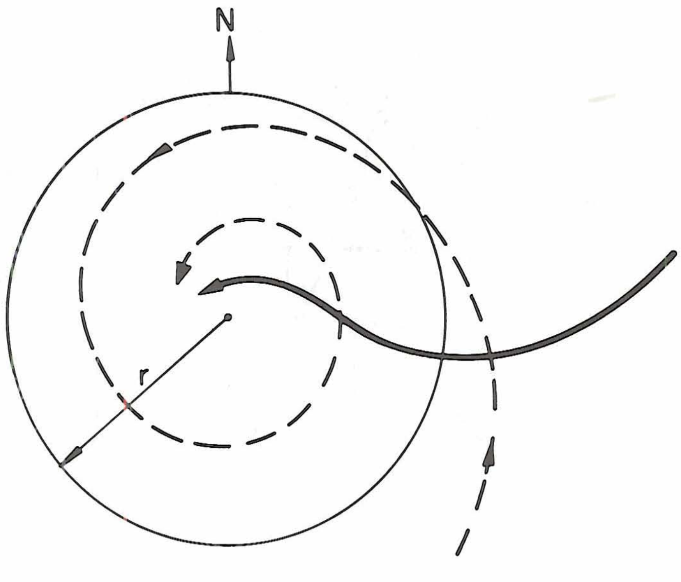

Fig. 3 Sketch of surface inflow into initial disturbance at small and large crossing angles at outer periphery.

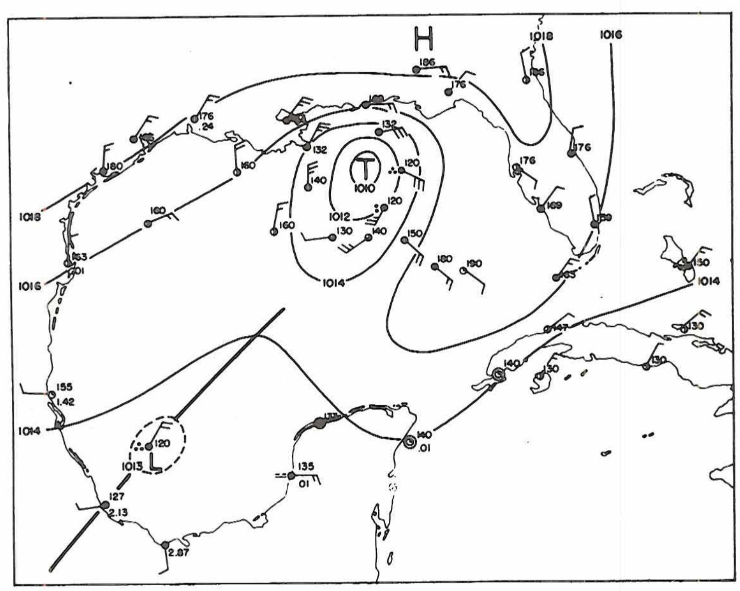

Fig. 4 Surface map, 11 Sept. 1941, 1200Z (Riehl 1948)

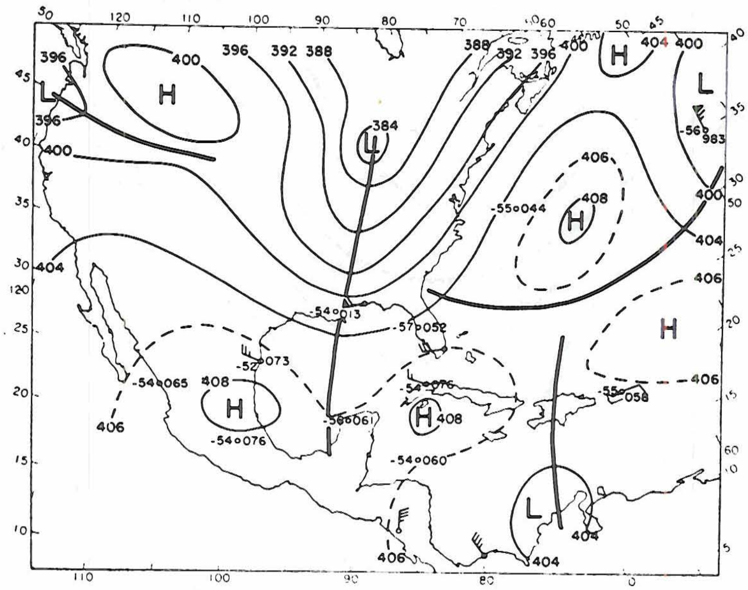

Fig. 5 200-mb chart, 29 Sept. 1949, 0300 Z
(Riehl and Burgner 1950)

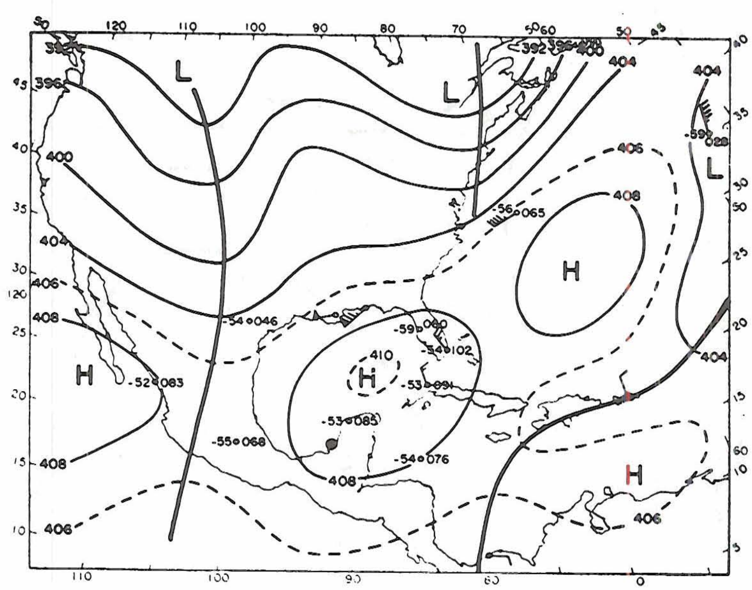

Fig. 6 200-mb chart, 1 October 1949, 0300 Z
(Riehl and Burgner 1950)

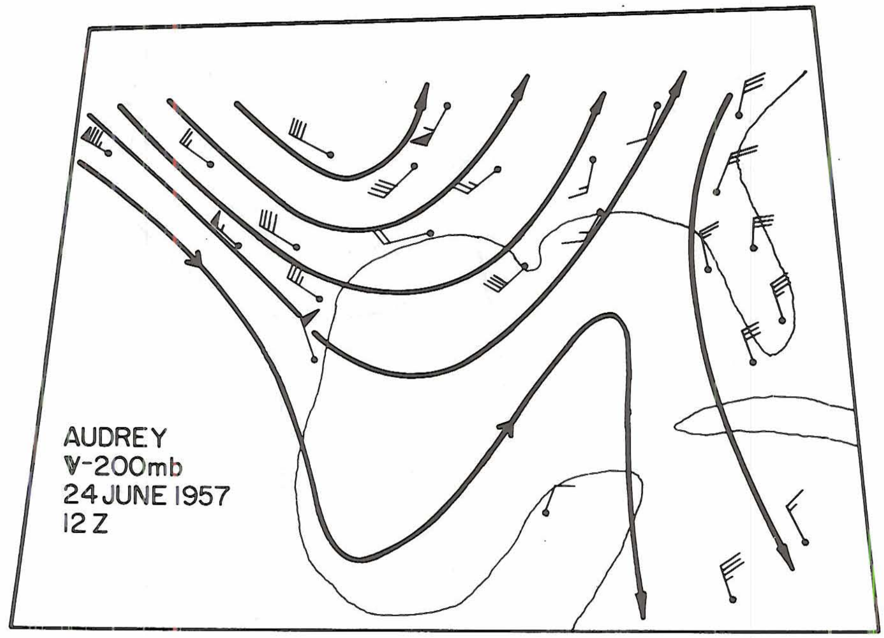

Fig. 7 200-mb streamlines, 24 June 1957, 1200 Z

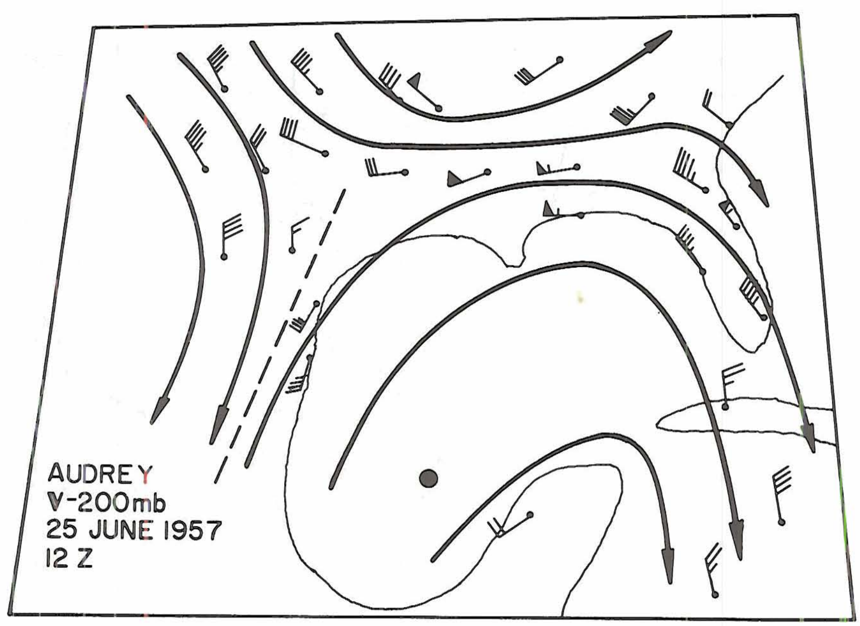

Fig. 8 200-mb streamlines, 25 June 1957, 1200 Z

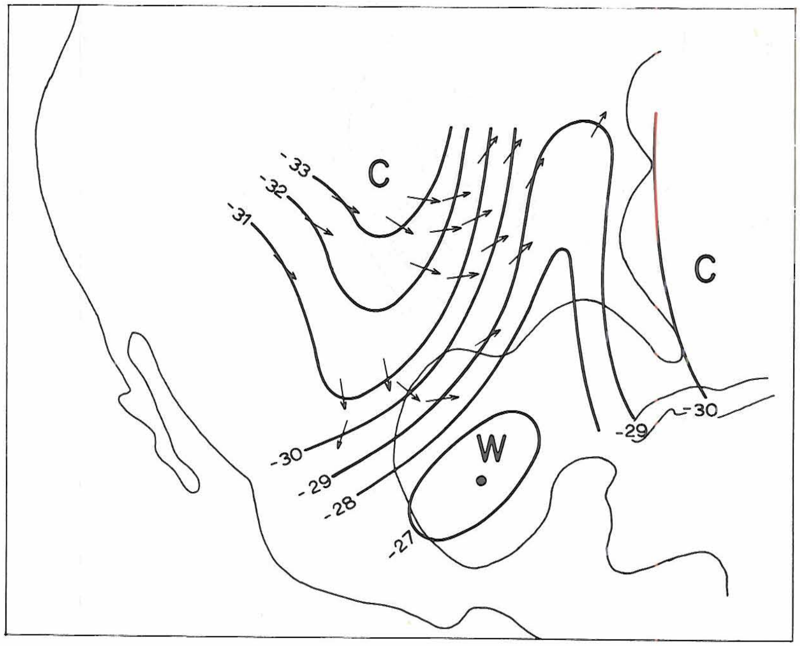

Fig. 9 Mean isotherms (^{O}C) for the layer 500-200 mb and arrows indicating mean wind direction of the layer, 25 June 1957, 1200 Z

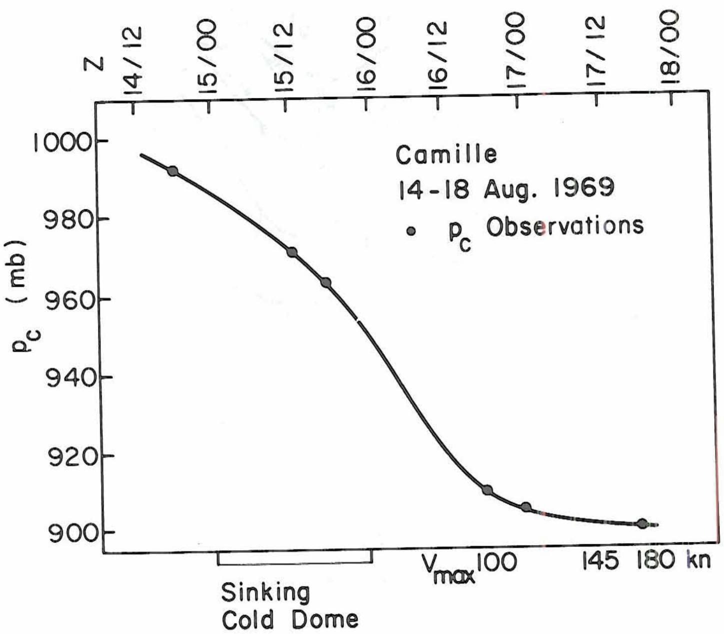

Fig. 10 Surface pressure profile of hurricane Camille in the Gulf of Mexico, 14-18 August 1969

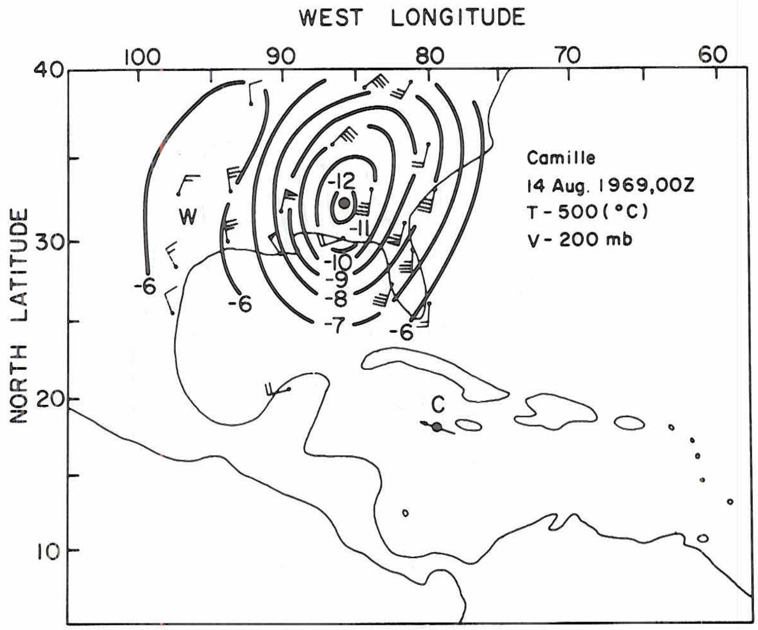

Fig. 11 500-mb isotherms (^{O}C) and 200-mb wind arrows, 14 August 1969, 00Z

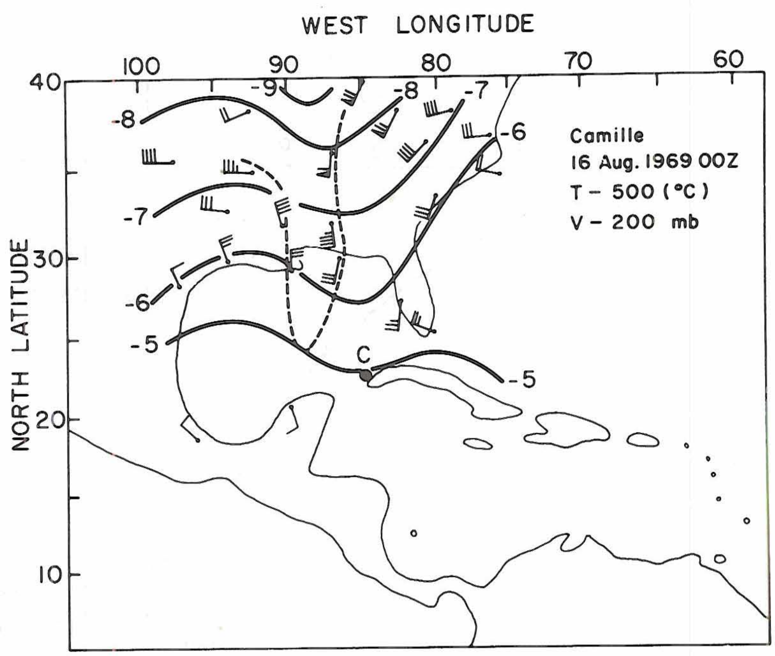

Fig. 12 500-mb isotherms (^{O}C) and 200-mb wind arrows, 16 August 1969, 00Z; dashed line outlines residual high-tropospheric vorticity maximum

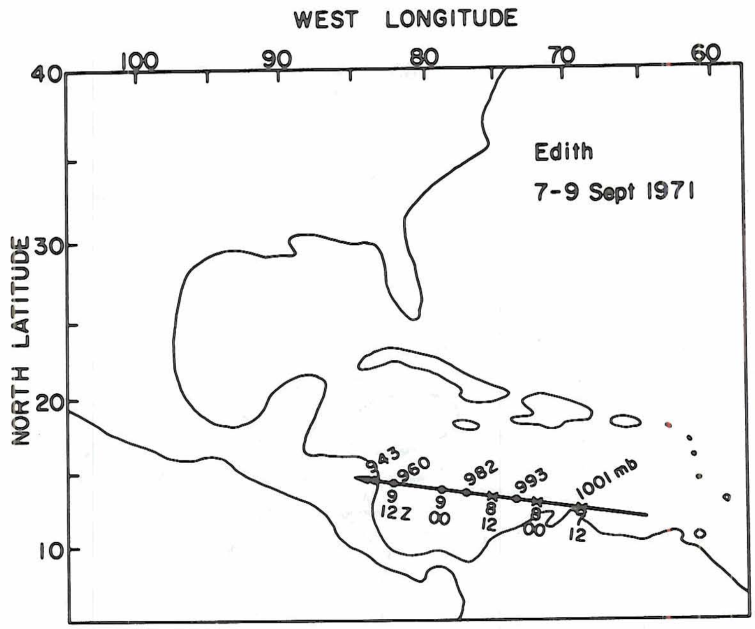

Fig. 13 Track and central pressures of hurricane Edith in the Caribbean, 7-9 Sept. 1971

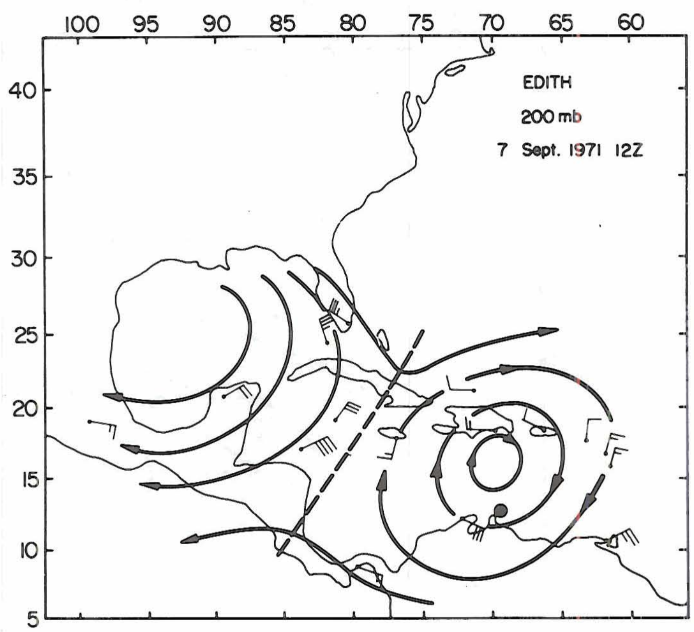

Fig. 14 200-mb streamlines, 7 Sept. 1971, 12 Z for the Caribbean. Heavy dot marks tropical disturbance center

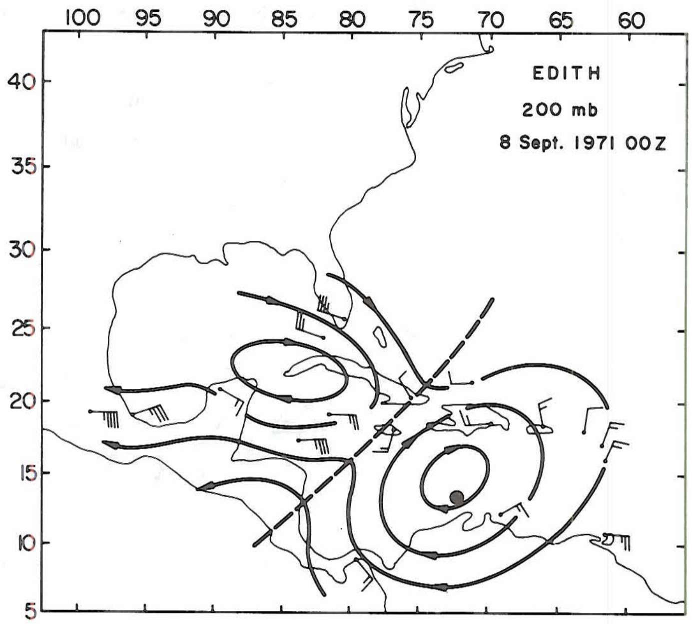

Fig. 15 200-mb streamlines, 8 Sept. 1971, OOZ for the Caribbean

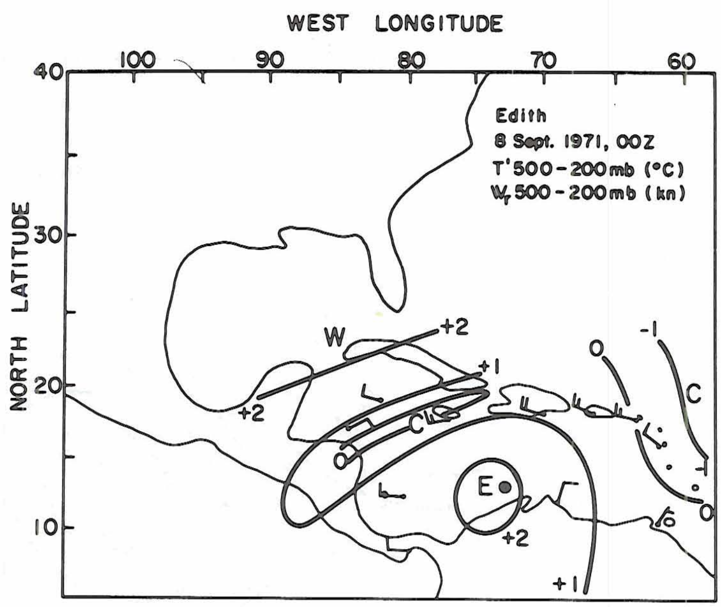

Fig. 16 Temperature anomaly from mean tropical atmosphere in ^{o}C for the layer 500-200 mb, and mean wind for the layer, relative to incipient tropical storm, 8 Sept. 1971, OOZ

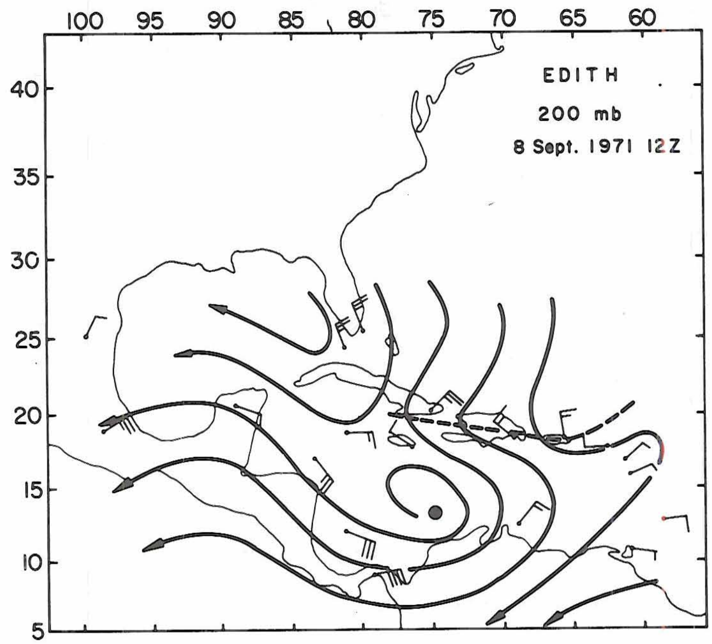

Fig. 17 200-mb streamlines, 8 Sept. 1971, 1200Z for the Caribbean

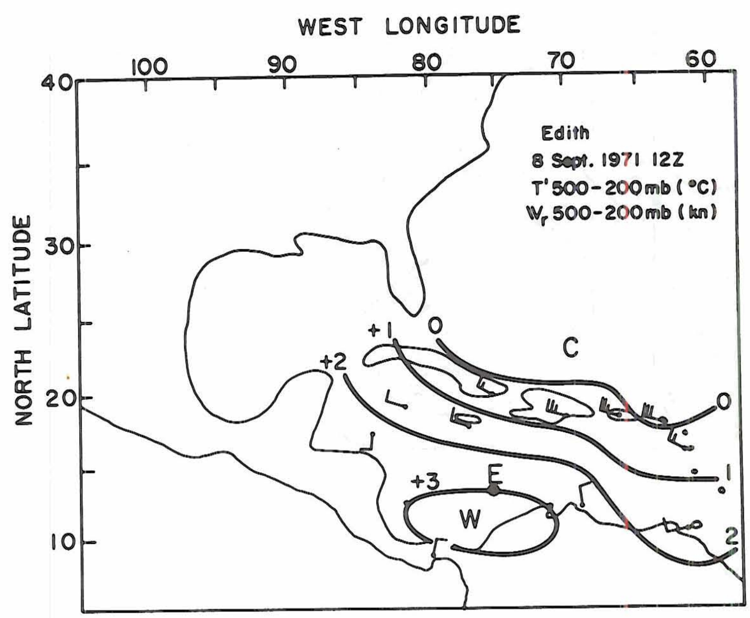

Fig. 18 Temperature anomaly for the layer 500-200 mb as in Fig. 16, and mean wind relative to tropical storm center, 8 Sept. 1971, 1200Z

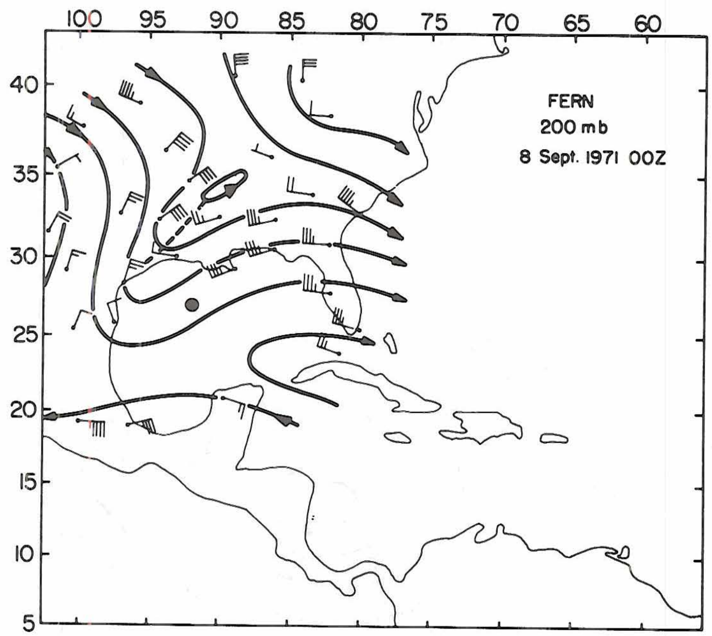

Fig. 19 200-mb streamlines, 8 Sept. 1971, for the United States and Gulf of Mexico

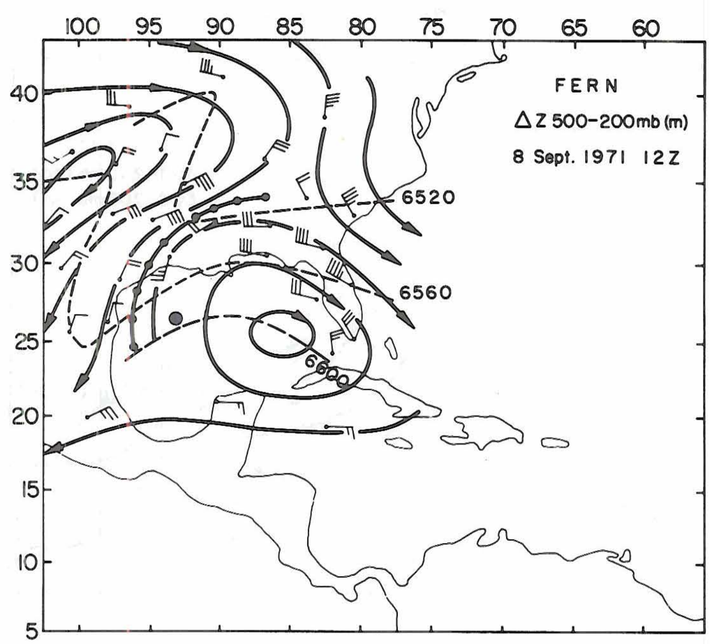

Fig. 20 200-mb streamlines and thickness 500-200 mb (m), 8 Sept. 1971, 1200 Z. Heavy dot marks position of tropical storm Fern, heavy dotted line shows axis of 200-mb shear

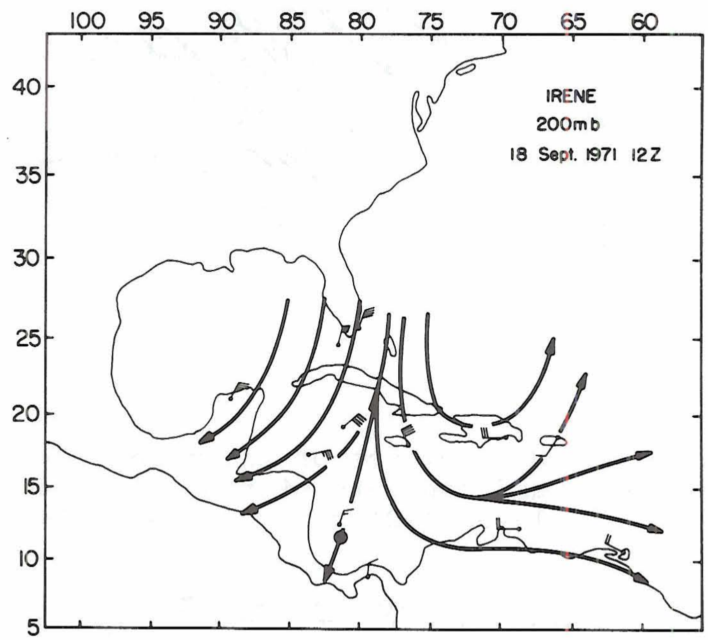

Fig. 21 200-mb streamlines, 18 Sept. 1971, 1200 Z for the Caribbean

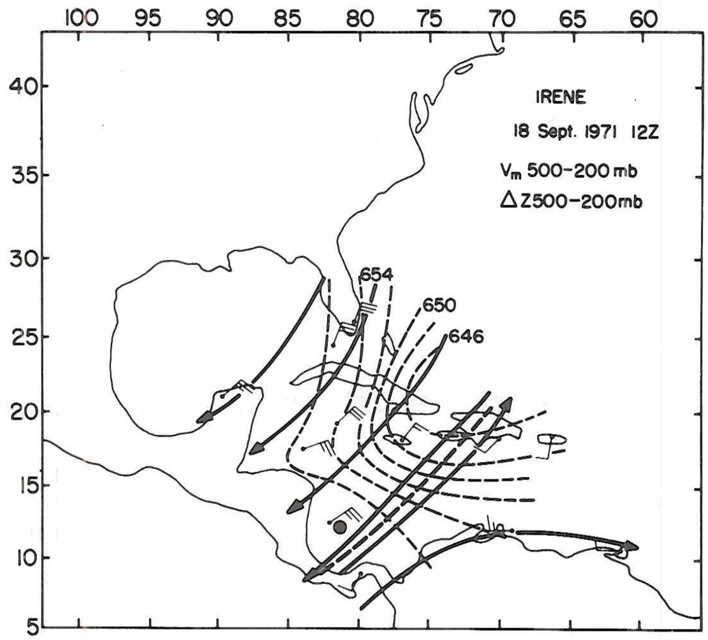

Fig. 22 Streamlines of the mean flow 500-200 mb and contours of the thickness of this layer (10' s m), 18 Sept. 1971, 1200 Z

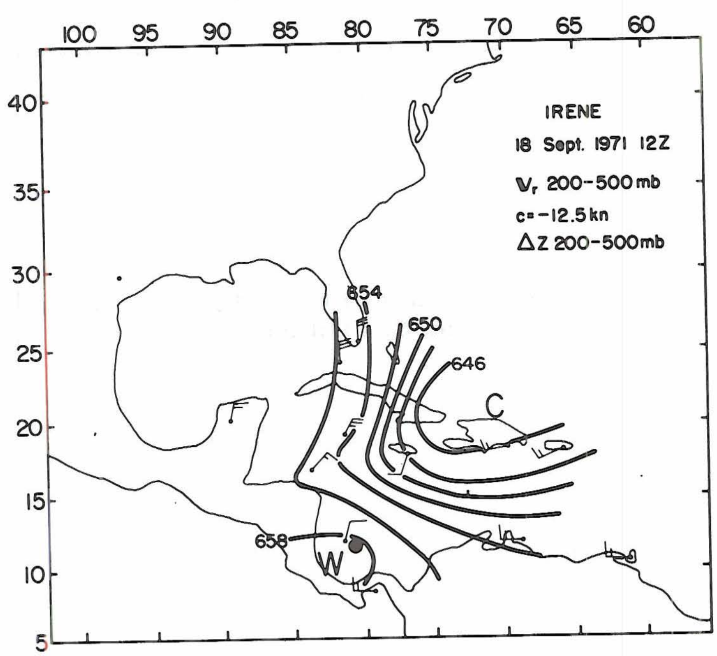

Fig. 23 Same as Fig. 22, but the winds are relative to the moving storm center Irene

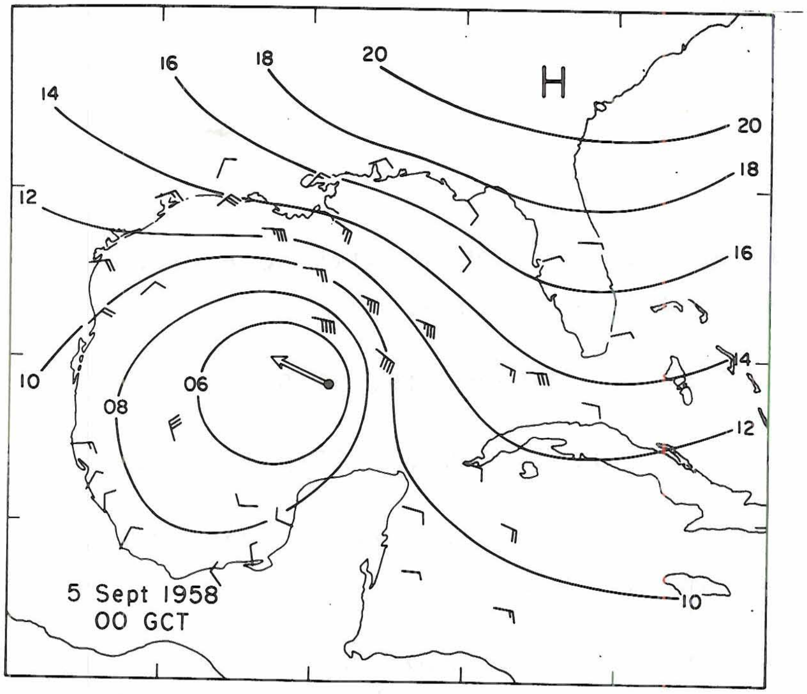

Fig. 24 Surface map for the Gulf of Mexico, 5 Sept. 1958 with tropical storm Ella, 00Z

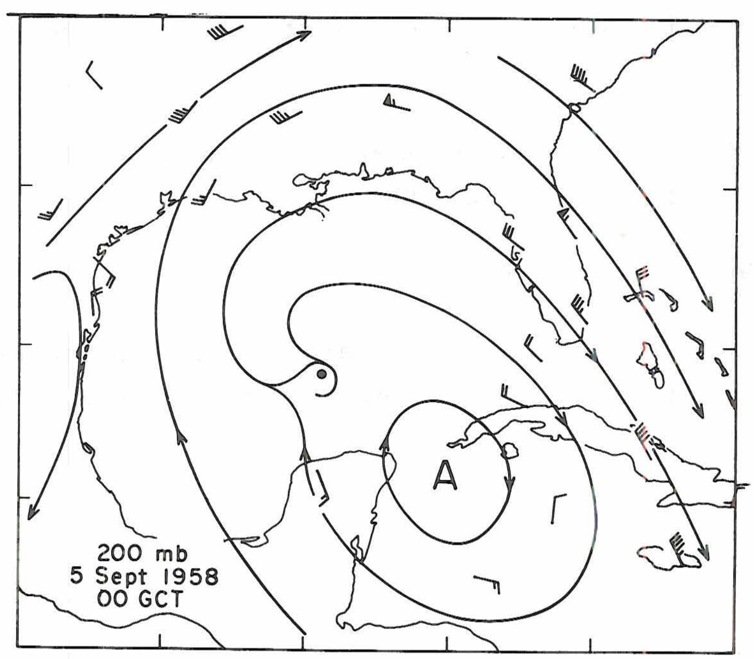

Fig. 25 200-mb streamlines for 5 Sept. 1958, 00Z

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15. Supplementary Notes

16. Abstracts

It has previously been hypothesized by the author that hurricane initial intensification is often related to a one to two day prior increase of upper tropospheric baroclinicity 500-1000 km around the pre-cyclone disturbance. This prior increase of upper level baroclinicity is usually associated with a traveling middle-latitude westerly cyclone. A number of recent year examples in the West Indies are presented.

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West Indies

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17c. COSATI Field/Group

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