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STUDIES OF UPPER-AIR CONDITIONS IN LOW LATITUDES

PART I

ON THE FORMATION OF WEST ATLANTIC HURRICANES

By

HERBERT RIEHL

The University of Chicago

PART II

RELATIONS BETWEEN HIGH- AND LOW-LATITUDE CIRCULATIONS

By

GEORGE P. CRESSMAN

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A PUBLICATION OF THE DEPARTMENT OF METEOROLOGY OF THE UNIVERSITY OF CHICAGO

VICTOR P. STARR, Editor

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Miscellaneous Reports No. 24

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ABSTRACT[1]

In Part I of this report weather situations in the tropical West Atlantic Ocean which appeared to have cyclogenetic characteristics are studied in an effort to explain why some develop cyclonic circulations while others, apparently similar initially, remain inactive.

The common organized convection zones of the tropics, such as wave troughs and shearlines, have a cold core. In contrast, hurricanes have a warm core, in accord with generally accepted concepts. These observations suggest that air near hurricane centers must have risen from the surface. However, they cannot be explained on the basis of the "convection theory," because of the prevalence of cold-core rather than warm-core conditions in the more common types of convective disturbances. All tropical storms noted during the study developed from pre-existing disturbances, namely, from equatorial shearlines and waves in the easterlies. Intensification of these systems appeared to be the result of superposition of high-level extratropical disturbances on the tropical systems or of formation of unstable shearing waves, or a combination of both effects.

The superposition of an extratropical perturbation on a wave in the easterlies leads to an intensification of the horizontal divergence aloft over and near the surface position of the wave trough. As a result, the surface pressure begins to drop, but the superposition will not be effective unless there is a barotropic stratification of the atmosphere in the subtropics and in the equatorward portion of the belt of polar westerlies.

The formation of vortices generated at equatorial and east-west shearlines is restricted to periods when the latter are intersected by troughs of great amplitude extending equatorward from higher latitudes. Again, barotropic broad-scale conditions in the subtropics and the southern part of the belt of polar westerlies are essential.

Additional factors that restrict the frequency of hurricane occurrence are noted, and several synoptic situations are included as illustrations.

In Part II of this report are described the relations between various perturbations of the circulation in the tropics and perturbations in the circumpolar westerlies.

Troughs in the circumpolar westerlies develop extensions into the subtropical easterlies, provided these troughs are pronounced and

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[1] This report is a result of research in tropical meteorology undertaken jointly by the United States Army Air Forces and the University of Chicago. Manuscript of Part I submitted October, 1946; manuscript of Part II submitted May, 1946.

slow-moving. If they are weak and /or move rapidly, they do not extend far south of the subtropical belt of high pressure. Changes in the intensity and motion of the extratropical systems result in corresponding changes in the tropical extensions. Examples are shown indicating the advantage of the 10,000-foot chart over the sea-level map in analysis of these perturbations.

The importance of the troughs in the circumpolar westerlies in generating and intensifying waves in the easterlies is discussed and illustrated. The superposition of a trough in the westerlies on a wave in the easterlies apparently leads to an intensification of both perturbations.

Various properties of east-west shearlines in the tropics and subtropics are discussed and illustrated. These shearlines, frequently separating flow from the Northern and Southern hemispheres, need not remain near the equator but may move relatively large distances from the equator in a few days with an accompanying large-scale flow of air across the equator.

Finally, the interrelations among the various perturbations and shearlines demonstrated that, in order to understand the events in the tropics and subtropics of one hemisphere, it is necessary to consider the events in the circumpolar westerlies of the same hemisphere and possibly in the tropics and subtropics of the other hemisphere.

PART I

ON THE FORMATION OF WEST ATLANTIC HURRICANES

By Herbert Riehl

A. Introduction

Many authors who have treated the subject of tropical storms sought to ascribe the origin of all such storms to one common cause. Some advocated the convective hypothesis and others the frontal theory. In extratropical latitudes, also, there has been a persistent striving to find a single scheme which satisfactorily explains the development of all cyclones.

As yet, from available evidence it has not been possible to establish the existence of such a single ultimate cause. Synoptically it has become evident that storms may arise in many different ways. In temperate as well as in lower latitudes we have, however, learned to recognize certain distinctive patterns that are favorable for cyclogenesis. For the tropics, such patterns have been described by Gordon E. Dunn (11), among others, for the North Atlantic Ocean, while, in the North Pacific, Deppermann [9], and other works) has noted a variety of flow patterns conducive to the development of typhoons.

Synoptic patterns closely resembling each other do not produce the same developments in all instances. Therefore, forecasting the formation of tropical storms consists principally in keeping a close watch on map situations known by experience to be threatening. The present report represents a study of these potentially cyclogenetic cases in the West Atlantic and Caribbean areas. It seeks to explain why some situations develop, while others, apparently similar, remain inactive.

Note on maps.--The time used for all illustrations is Eastern Standard Time (75th meridian west).

On the upper-wind charts a long barb denotes 10 mph, a short barb 5 mph, and a heavy triangular barb 50 mph. A stands for anticyclonic circulation and C for cyclonic circulation.

Where upper pressures and temperatures are reproduced, the pressure (in mb) is plotted to the right of the station circle and the temperature (in $^{\circ}$C) to the left. Changes of pressure and temperature in 24 hours are plotted below these figures. In Figures 6a, 6e and 17e, 3-km pressures and temperatures have been used.

The "basic charts" are plotted and analyzed as described in an article by H. Riehl and E. Schacht published in the Bulletin of the American Meteorological Society for December, 1946. Winds are for the 5,000-foot level, and numbers to the upper right of the station circles are sea-level 24-hour pressure changes.

In the ship observations (Figs. 4-5 and 12a) pressures are given to the nearest millibar only.

Acknowledgement.--The author wishes to acknowledge the aid rendered by Dr. H. R. Byers, Mr. George P. Cressman, and Mr. Charles Gilman in preparing the report. He is also indebted to Sr. José A. Colón, who assisted most valuably throughout the study with the preparation, analysis, and discussion of the maps.

B. Comparison of the Structure of Tropical Storms with Other Disturbances of the Tropics

Examination of the records of tropical stations in the West Atlantic and Caribbean regions shows that only a minor portion of the total precipitation in those areas can be ascribed to hurricanes. Many parts of the West Indies, for instance, have two mean rainfall maxima, in May and in November, that is, before the start and after the end of the hurricane season. Active convection occurs predominantly in connection with relatively minor disturbances appearing mainly as isallobaric centers, as noted by Dunn (11), who states: "While isallobaric waves move in endless procession through the trade-wind region with about one every three or four days, one or even two months may pass with no development and it is apparent then that most of these waves are relatively stable."

One of the objects of this report is to describe some features of the few unstable waves which distinguish them from the great number of stable systems. In order to determine these differences, it is necessary first to analyze the stable disturbances. The following section is an amplification of earlier studies (29) of this nature.

1. Structure of the Common Organized Convection Zones

Observations.--A pronounced trough of low pressure aloft overlies the zones of organized convection associated with the usual tropical perturbations. In the case of waves in the easterlies this trough tends to attain its greatest strength in the middle troposphere (29). When connected with other types of low-latitude disturbances, such as the shearlines discussed later, the trough may continue to increase upward in intensity to 10 km or even beyond. At fixed upper levels, pressure and temperature falls accompany the arrival of bad weather areas, while at the ground the pressure usually rises during the passage of bad weather. (Temperatures used here are virtual temperatures.) The lowest pressure at the surface very frequently appears on the forward edge of the zone of intense convection.

Table 1 presents the 24-hour pressure and temperature variations observed at Miami, Florida, September 9-10, 1941, during the passage of a strong wave in the easterlies. This wave (mentioned in [29]) is also illustrated in Figures 5-8 of the present report. Surface pressure rises at Miami gave way to falls aloft that were greatest near 3 km and then

TABLE 1

RADIOSONDE OBSERVATIONS AT MIAMI, FLORIDA, SEPTEMBER 9-10, 1941, 0100 EST, AND 24-HOUR CHANGES

H(km)	Sept. 9, 1941, 0100 EST			Sept. 10, 1941, 0100 EST			Changes	
	p(mb)	T(°C)	RH(%)	p(mb)	T(°C)	RH(%)	Δp(mb)	ΔT(°C)
Sfc. . .	1015.2	26.0	M*	1015.9	25.3	94	0.7	-0.7
1. . .	906	22.4		906	19.7	100	0	-2.7
2. . .	808	17.1		807	15.4	100	-1	-1.7
3. . .	718	11.1		716	9.9	73	-2	-1.2
4. . .	636	4.9		635	4.9	72	-1	0.0
5. . .	562	-0.9		561	-0.6	72	-1	0.3
6. . .	496	-6.2		495	-5.7	67	-1	0.5
7. . .	435	-12.5		434	-11.7		-1	0.8
8. . .	381	-19.0		380	-18.7		-1	0.3
9. . .	333	-26.8		332	-26.2		-1	0.6
10. . .	289	-34.9		288	-34.5		-1	0.4
11. . .	249	-43.6		249	-43.2		0	0.4
12. . .	214	-52.4		214	-51.6		0	0.8
13. . .	183	-59.3		183	-58.7		0	0.6
14. . .	156	-65.0		156	-66.1		0	-1.1
15. . .	132	-69.7		131	-72.0		-1	-2.3

*Missing

decreased upward. Considerable precipitation followed the surface wave trough.

Another striking instance of similar pressure variations occurred at San Juan, Puerto Rico, as an intense wave in the easterlies passed over Puerto Rico on July 12, 1944 (Table 2; Figs. 1a and 1b). The wave contained a closed surface low-pressure center when it was over the Lesser Antilles and redeveloped later when reaching the Bahamas. Near Puerto Rico, however, a closed surface center was not in evidence.

TABLE 2

RADIOSONDE OBSERVATIONS AT SAN JUAN, PUERTO RICO, JULY 11-12, 1944, 2200 EST, AND 24-HOUR CHANGES

H(km)	July 11, 2200 EST			July 12,2200 EST			Changes	
	p(mb)	T(°C)	RH(%)	p(mb)	T(°C)	RH(%)	Δp(mb)	ΔT(°C)
Sfc. . .	1015	27.0	87	1015	23.5	90	0	-3.5
1. . .	907	21.5	83	906	20.6	78	-1	-0.9
2. . .	808	16.7	64	807	14.9	75	-1	-1.8
3. . .	717	11.2	41	716	9.1	79	-1	-2.1
4. . .	635	4.1		634	4.5	81	-1	0.4
5. . .	561	-1.4		560	-0.8	81	-1	0.6
6. . .	494	-8.3		494	-6.3	93	0	2.0
7. . .	434	-13.8		434	-11.7		0	2.1
8. . .	379	-21.0		380	-18.8		1	2.2
9. . .	331	-27.0		332	-25.8		1	1.2
10. . .	287	-36.0		288	-34.4		1	1.6
11. . .	248	-44.6		249	-43.1		1	1.5
12. . .	213	-52.9		214	-52.3		1	0.6
13. . .	182	-58.9		182	-60.9		0	-2.0

TABLE 3A

RADIOSONDE OBSERVATIONS AT SAN JUAN, PUERTO RICO, JULY 24-25, 1944, 2200 EST, AND 24-HOUR CHANGES

H(km)	July 24, 2200 EST			July 25, 2200 EST			Changes	
	p(mb)	T(°C)	RH(%)	p(mb)	T(°C)	RH(%)	Δp(mb)	ΔT(°C)
Sfc. .	1015	27.2	87	1015	25.7	89	0	-1.5
1 . .	907	21.6	85	907	20.5	79	0	-1.1
2 . .	808	15.5	54	807	12.4	71	-1	-3.1
3 . .	717	11.2	36	715	9.0	74	-2	-2.2
4 . .	635	5.2	16	633	2.6	84	-2	-2.6
5 . .	561	-0.4	15	559	-2.0	85	-2	-1.6
6 . .	494	-5.9	16	492	-7.6	86	-2	-1.7
7 . .	434	-12.6		432	-13.4		-2	-0.8

TABLE 3B

RADIOSONDE OBSERVATIONS AT SAN JUAN, PUERTO RICO, JULY 29-30, 1944, 1000 EST, AND 24-HOUR CHANGES

H(km)	July 29, 1000 EST			July 30, 1000 EST			Changes	
	p(mb)	T(°C)	RH(%)	p(mb)	T(°C)	RH(%)	Δp(mb)	ΔT(°C)
Sfc. . .	1014	28.7	78	1014	23.8	90	0	-4.9
1. . .	907	21.1	86	905	19.6	68	-2	-1.5
2. . .	807	16.9	72	805	15.3	65	-2	-1.6
3. . .	717	10.9	63	715	8.7	78	-2	-2.2
4. . .	635	4.7	68	633	2.0	93	-2	-2.7
5. . .	561	-1.0	73	558	-3.3	98	-3	-2.3

Very good weather preceded the arrival of the wave trough. Then, as the winds shifted from north to south and the lowest pressure passed at the surface, the sky became overcast, and a prolonged period of thunderstorm activity began (Fig. 1a). The map of July 12, 1930 EST (Fig. 1b), also brings out the asymmetrical weather distribution on the two sides of the wave trough. Unusually clear skies prevailed to the west of Puerto Rico, while over the island and to the east weather conditions were severe.

Tables 3A and 3B give some of the upper pressure and temperature variations at San Juan during a situation which was notable because the pressure and temperature falls remained relatively large up to the high troposphere. A detailed description of this situation will be given later.

It may be argued that the changes shown in Tables 1-3 are so small as to place them within the range of instrumental error of the radiosonde. This objection can be met by pointing out that most cases noted had a similar history. Even more satisfactory, the field of motion provides independent evidence in support of the radiosonde data. When upper pressure falls associated with a disturbance also coincide with temperature falls, the intensity of the system as indicated by upper-wind observations increases upward. The converse holds for warm-core disturbances. It appears to be definitely established that many of the zones of organized convection in low latitudes are cold-core disturbances.

Discussion of the upper pressure and temperature variations.--At

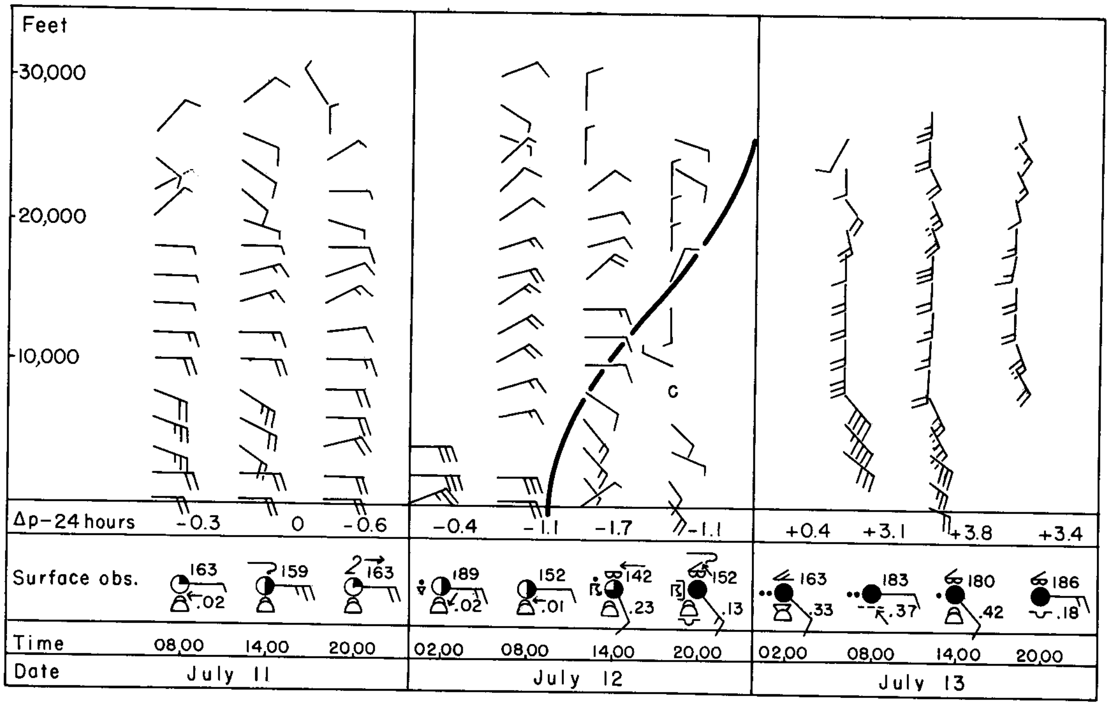

FIG. 1a.--Vertical time-section at San Juan, Puerto Rico, July 11-13, 1944.

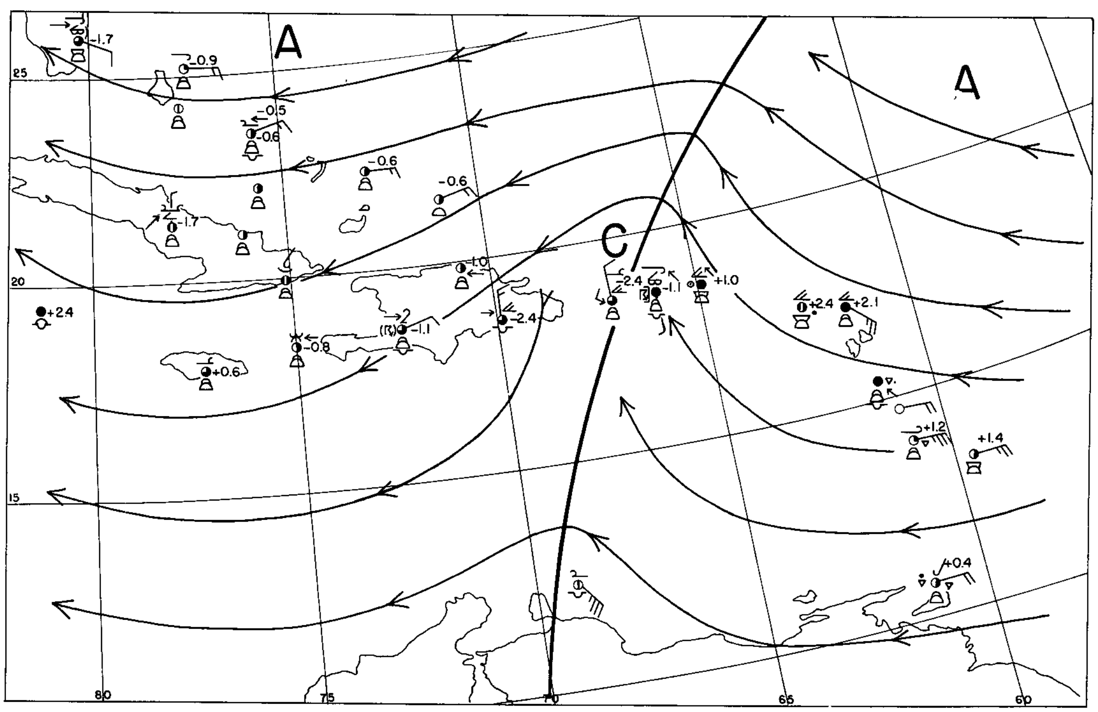

FIG. 1b.--Basic chart of the Caribbean area, July 12, 1944, 1930 EST.

first sight, the upper-air cooling associated with pressure falls at or below 3 km, as brought out by the preceding data, may seem paradoxical. Within the tropical current the temperature lapse rate normally exceeds somewhat the moist adiabatic, and the humidity near the ground is very high. Thus the lifting and convective condensation levels of the surface air are low, as is also indicated by the cloud bases, normally observed between 1,500 and 2,500 feet. Therefore, a large-scale ascent of surface air in which condensation is taking place should lead to upper warming in the broad-scale convection areas. The rise of mean virtual temperature should increase the vertical distance between isobaric surfaces. If the pressure rises at the surface in the convection zone, as is observed, there should also be rises aloft. However, the opposite takes place. As the surface pressure rises rapidly and the rains begin, the upper pressures and temperatures frequently experience their greatest falls. Evaporation from falling rain can hardly account for these maxima of pressure, and temperature falls. It follows that air in the middle troposphere in the organized convection zones cannot be predominantly surface air of local origin. To a large extent it must be air that was aloft before convection set in.[1]

To ascertain the magnitude of upper temperature changes that can result from vertical motion of the upper air itself, the effect of low-level horizontal convergence on a mean tropical atmosphere, such as computed by E. Schacht (33), may be considered (Table 4). Near the ground the mean sounding initially is conditionally unstable but approaches the moist-adiabatic temperature lapse rate above 4 km (Fig. 2). If turbulent mixing distributes the moisture uniformly in the lowest kilometer, the surface air requires lifting of 50-60 mb to reach condensation. The base of the low clouds then appears at approximately 1,600 feet, a reasonable figure for the zones of organized convection.

The dash-dotted curve of Figure 2 represents the mean tropical sounding after the whole column has been lifted 55 mb. The construction of this curve is based on the well-established assumption that ascending air moves through isobaric surfaces and that vertical motion of the isobaric surfaces themselves is negligible compared to that of the air. Mass divergence at high levels must compensate closely for low-level convergence and account for the small vertical movement of isobaric surfaces.

[1] Cf. H. Stommel, "Entrainment of Air into a Cumulus Cloud," Jour. Meteor., IV, No. 3 (1947), 91-94.

TABLE 4

COMPARISON OF A MEAN TROPICAL ATMOSPHERE
FOR THE CARIBBEAN AREA IN SUMMER AT MIDNIGHT
WITH A MEAN HURRICANE SOUNDING FOR THE CARIBBEAN
AREA IN SUMMER, AS COMPUTED BY E. SCHACHT

H(km)	Mean Tropical Atmosphere			Mean Hurricane Sounding			Difference		
	p(mb)	T(°C)	RH(%)	p(mb)	T(°C)	RH(%)	Δ p(mb)	Δ T(°C)	ΔRH(%)
Sfc. .	1013.6	25.3	85	1004.6	25.7	88	-9.0	0.4	3
1 . .	905.2	19.8	83	897.1	19.6	89	-8.1	-0.2	6
2 . .	805.1	14.3	73	798.3	14.7	86	-6.8	0.4	13
3.. .	714.6	9.4	60	708.5	9.7	84	-6.1	0.3	24
4.. .	633.0	4.0	53	628.1	4.7	86	-4.9	0.7	33
5 . .	559.3	-1.6	50	555.5	-0.3	81	-3.8	1.3	31
6 . .	492.6	-7.5	48	489.5	-5.7	73	-3.1	1.8	25
7 . .	432.8	-13.8		430.4	-11.9		-2.4	1.9	
8 . .	378.4	-20.6		377.3	-18.7		-1.1	1.9	
9 . .	329.4	-27.8		328.5	-25.5		-0.9	2.3	
10 . .	286.3	-35.6		286.2	-33.0		-0.1	2.6	
11 . .	247.1	-43.6		247.6	-40.9		-0.5	2.7	
12 . .	212.7	-51.7		213.5	-49.4		0.8	2.3	
13 . .	181.8	-59.9		182.6	-57.8		0.8	2.1	
14 . .	154.4	-67.1		155.6	-65.8		1.2	1.3	
15 . .	130.7	-73.1		132.2	-73.5		1.5	-0.4	
16 . .	110.3	-77.2		111.2	-77.8		0.9	-0.6	
17 . .	92.6	-76.7		94.3	-76.2		1.7	0.5	
18 . .	77.8	-73.5		79.3	-71.1		1.5	2.4	
19 . .	65.9	-67.9							

Owing to the variation of convergence and divergence with height, a lifting of an entire column without changes of cross-section will rarely occur. Nevertheless, the dash-dotted curve of Figure 2 is representative of the magnitude of temperature variations produced by lifting. As also indicated in (a) in Table 5, considerable cooling at standard pressure surfaces of the lower troposphere results from the lifting, and therefore also at standard heights in view of the small vertical displacement of the isobaric surfaces.

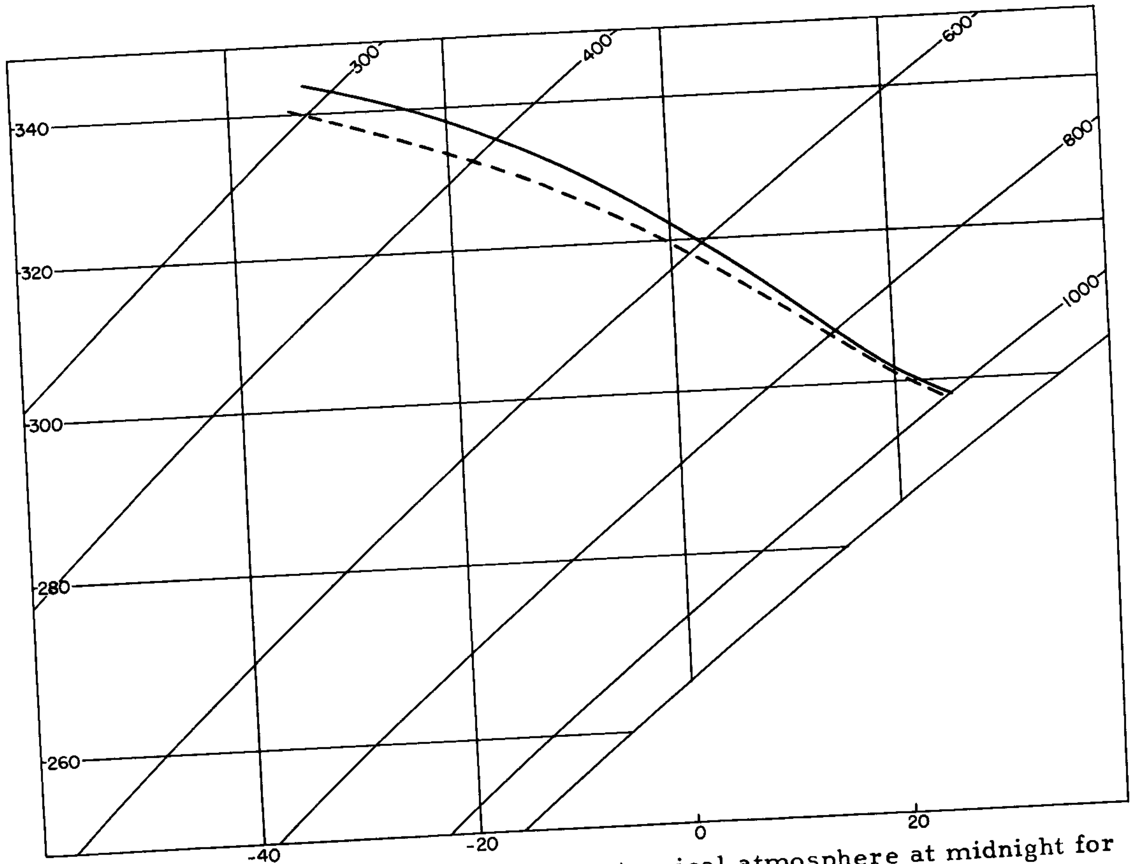

FIG. 2.--Tephigram of the mean tropical atmosphere at midnight for the Caribbean area in summer (full line). Dashed-dotted line represents the mean tropical atmosphere after whole column has been lifted 55 mb. Dashed line represents ascent curve of surface air from lifting condensation level. In diagram vertical lines are isotherms (°C), horizontal lines isentropes (°A), slanting lines isobars (mb).

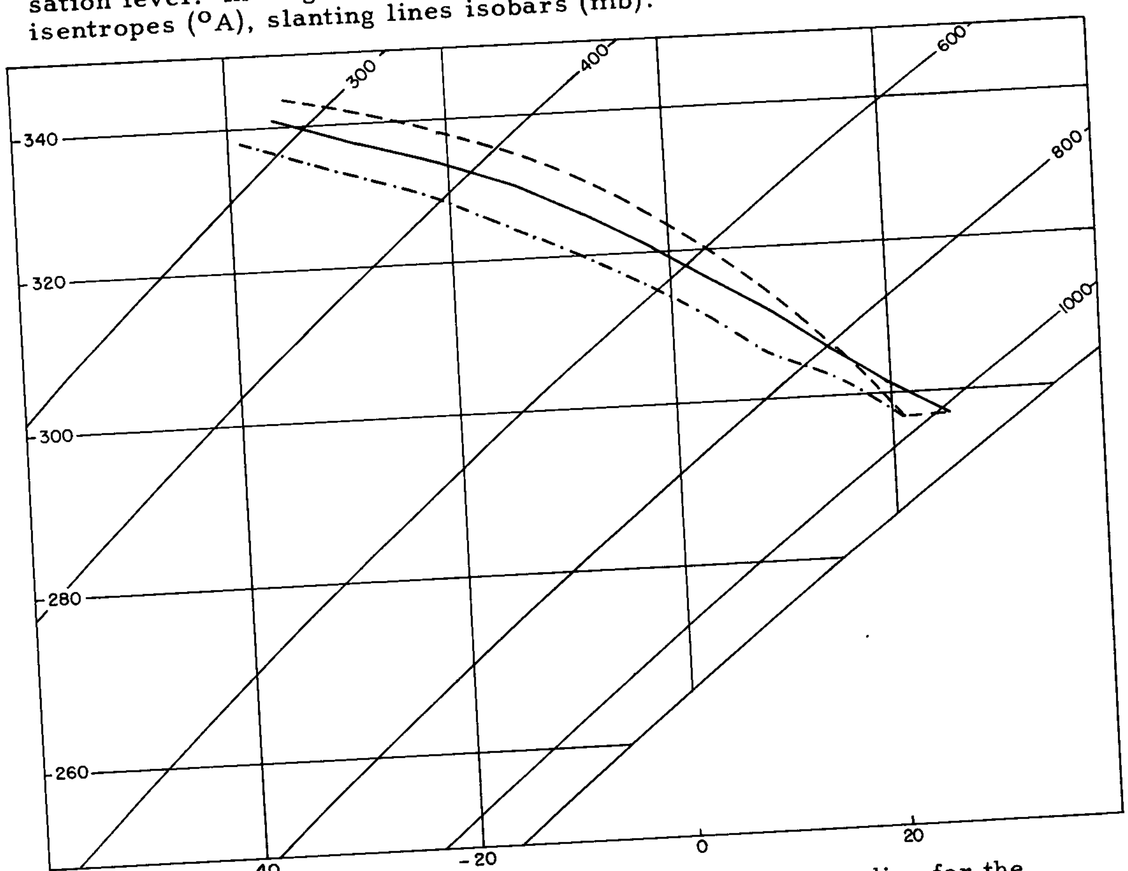

FIG. 3.-- Tephigram of the mean hurricane sounding for the Caribbean area at midnight (after E. Schacht) (solid line), and mean tropical atmosphere (dashed line). For notation on diagram see Fig. 2.

The amount of cooling obtained in individual cases is of course dependent on the initial stability and vertical moisture distribution. But, as may be seen from Tables 1 and 2, and a large number of other representative instances, cooling does occur in the lower troposphere in the areas of convection, and there are pressure falls aloft which damp out downward.

The dashed curve in Figure 2 represents the ascent of the surface air from the lifting condensation level. After passing this level, the surface air soon becomes buoyant,[2] and a considerable upward transport through the surrounding body of air should take place. Some of the surface air will ascend only a short distance, say, 1 km, while other air will penetrate upward much farther. If the transport is sufficiently large to produce mixing of surface air with the upper parts of the column in equal weights (50 mb each in a layer of 100 mb), net cooling as indicated in (b) in Table 5 will result. This cooling is almost uniform throughout the lower levels and then gradually disappears with height. Such a vertical distribution of temperature changes is consistent with many observed cases.

TABLE 5

TEMPERATURE CHANGES AT CONSTANT PRESSURE SURFACES THAT RESULT IF (a) MEAN TROPICAL ATMOSPHERE IS LIFTED 55 MB; (b) IF SUFFICIENT SURFACE AIR IS TRANSPORTED ALOFT TO MIX WITH THE UPPER AIR IN EQUAL AMOUNTS

p(mb)	(a) $\Delta T(^{\circ}C)$	(b) $\Delta T(^{\circ}C)$
900.	-2.0	-1.5
800.	-2.5	-1.0
700.	-3.0	-1.5
600.	-2.5	-1.0
500.	-3.0	-0.5
400.	-3.0	0.0

[2] Careful temperature measurements over the sea have established that the lapse rate below the cloud base is nearly dry-adiabatic and that the "subcloud" layer is almost entirely mixed. In that case no negative area exists near the ground.

Such measurements were made recently by Woods Hole Oceanographic Institution, as reported in "Vertical Motion and Exchange of Heat and Water between the Air and the Sea in the Region of the Trades" (to be published).

In the preparation of Table 5 it was assumed that extensive mixing occurs aloft. Ordinarily, in connection with calculations of the ascent of buoyant surface air, it is postulated that mixing between the rising parcels and their surroundings is very small. Such ascents, represented by the formation of individual cloud towers, last only a short time, rarely exceeding an hour.[3] The upper pressures and temperatures characteristic of a synoptic convection zone as a whole relative to its surroundings in a large-scale sense, however, result mainly from prolonged general convergence at lower levels and divergence in the middle and upper troposphere. While this divergence is operative on the same mass of air, the building and dissipation of many individual cumuli take place within this mass. Under these conditions the assumption that no mixing occurs can no longer apply. While individual clouds within the general convective zone may be warmer than their immediate surroundings, the convection zone as a whole is colder than its outskirts in a broad-scale sense.

It will be seen that the problem of upper temperature changes derived from vertical motion as discussed here differs somewhat from that investigated by Byers (7), who considered the effect of complete vertical overturning on the vertical temperature distribution. Since organized convection zones of the tropics normally correspond to a large-scale area of low-level convergence, the convective downward mass transport of air initially aloft cannot equal the upward mass transport. Lateral divergence in the middle and upper troposphere approximately balances convergence near the ground. Nevertheless, some convective downward transport will take place in many situations to compensate for ascent of buoyant surface air. This descending vertical motion further modifies the initial tropical atmosphere, in which the wet-bulb potential temperature decreases upward to about the 600-mb level (see Table 4). If continued evaporation from falling rain is taking place so that the sinking air moves moist adiabatically, additional cooling results in the lower troposphere. This was shown by Byers (7). In the present case the vertical gradient of the wet-bulb potential temperature is less in the convection areas than in their outskirts but does not vanish entirely as assumed in Byers' example.

The observed structure of the waves also indicates that descent of the air initially aloft cannot compensate for the buoyant ascent of surface air except to a minor degree, because such descent is likely to be

[3] As indicated by Stommel's work (op. cit.) and other more recent observations, it appears probable that the parcel hypothesis must be entirely eliminated in studies of convection, even during the initial building of cumulus.

predominantly dry-adiabatic and therefore would lead to warming. If for some reason large-scale dry-adiabatic descent develops, the structure of the waves will be altered profoundly. In this respect the present treatment differs from that of other investigators who have considered the moist-adiabatic ascent of air through a dry-adiabatically descending environment, and who have assumed that in this process mass compensation exists within the region of convection. In order to explain the upper pressure and temperature changes shown in Tables 1 and 2, it is necessary to assume that the entire lower and perhaps middle troposphere experiences a net lifting.

TABLE 6

RADIOSONDE OBSERVATIONS AT STA. LUCIA, B. W. I.
OCTOBER 9-10, 1943, 2200 EST, AND 24-HOUR CHANGES

H(km)	Oct. 9, 2200 EST			Oct. 10, 2200 EST			Changes		
	p(mb)	T(°C)	RH(%)	p(mb)	T(°C)	RH(%)	Δp(mb)	ΔT(°C)	ΔRH(%)
Sfc. .	1014	24.3	89	1012	25.7	89	-2	1.4	0
1 . .	906	20.6	82	904	19.6	96	-2	-1.0	14
2 . .	806	15.3	84	804	15.4	100	-2	0.1	16
3 . .	716	10.8	74	715	11.0	90	-1	0.2	16
4 . .	634	4.7	74	635	5.2	90	1	0.5	16
5 . .	559	-1.6	89	559	-1.6	99	0	0.0	10
6 . .	493	-6.0	85	493	-5.6	100	0	0.4	15
7 . .	434	-12.5		434	-11.0		0	1.5	
8 . .	379	-19.0		380	-17.0		1	2.0	
9 . .	331	-26.2		332	-24.4		1	1.8	
10 . .	287	-34.5		289	-31.6		2	2.9	
11 . .	248	-43.0		250	-39.8		2	3.2	
12 . .	214	-52.0		216	-48.2		2	3.8	
13 . .	184	-60.0		185	-55.0		1	5.0	

TABLE 7

RADIOSONDE OBSERVATIONS AT MIAMI, FLORIDA, OCTOBER 5, 1941, 2300 EST, OCTOBER 6, 1941, 0200 EST, AND 3-HOUR CHANGES

H(km)	Oct. 5, 2300 EST			Oct. 6, 0200 EST			Changes	
	p(mb)	T(°C)	RH(%)	p(mb)	T(°C)	RH(%)	Δ p(mb)	Δ T(°C)
Sfc. .	1012.5	26.1	90	1009.1	26.4	91	-3.4	0.3
1 . .	903	20.6	98	901	20.6	95	-2	1.0
2 . .	804	16.9	59	802	16.0	98	-2	-0.9
3 . .	715	11.4	74	714	12.3	85	-1	0.7
4 . .	634	6.2	76	633	9.1	74	-1	2.9
5 . .	562	2.1	51	560	1.2	82	-2	-0.9
6 . .	496	-5.1	51	495	-3.7	89	-1	1.4
7 . .	435	-13.6		436	-8.9		1	4.7
8 . .	381	-21.0		383	-15.4		2	5.6
9 . .	332	-28.2		335	-22.3		3	5.9
10 . .	288	-35.2		292	-29.7		4	5.5
11 . .	248	-43.3		253	-36.9		5	6.4
12 . .	214	-51.7		218	-44.1		4	7.6
13 . .	183	-61.2		187	-52.7		4	8.5
14 . .	156	-69.1		160	-61.5		4	7.6
15 . .	131	-76.7		136	-68.9		5	7.8
16 . .	110	-77.8		114	-76.3		4	1.5
17 . .	93	-72.3		96	-73.8		3	-1.5

In summary, the following picture presents itself: In the common zones of organized convection the entire low and middle troposphere experiences lifting, and this lifting is increased through lateral momentum transfer from jets of buoyant surface air. Thus the observed cumuli consist largely of air that was aloft before the beginning of convection.

The zones of organized convection are denser than their surroundings and thus possess greater potential energy. They are maintained presumably by the kinetic energy of the converging air currents.

2. The Hurricane Structure

Observations.--In pronounced contrast to the foregoing, the accumulated evidence regarding hurricanes reveals that these systems are most potent near the ground and weaken upward, although, as noted by Haurwitz (19), the circulation of mature storms must extend to great heights. In view of hydrostatic requirements, this implies that the core is warm relative to its surroundings.

The events at Sta. Lucia, B.W.I., on October 9-11, 1943 (Table 6), illustrate the upper pressure and temperature variations associated with tropical storms. According to the best information available, a disturbance of small diameter was forming late on October 10 just east of the Lesser Antilles and passed over Sta. Lucia early on October 11. Pressure falls which decreased upward preceded the system. Above these falls, in the upper troposphere, were pressure rises. The temperature increased with time at almost all levels, as did also the relative humidity.

Special hurricane soundings at Miami, Florida, taken during the passage of a tropical storm on October 5-6, 1941 (Table 7), also showed a decrease of the pressure falls with height and rises at the higher levels. The magnitude of the changes in the 3-hour interval at Miami was far greater than the magnitude of the variations recorded at Sta. Lucia on October 10, 1943. The center at Miami, however, was of hurricane strength, while the Sta. Lucia storm was very small and just forming. In both instances the pressure falls decreased from the ground upward and were accompanied by warming. Thus Tables 6 and 7 present observational evidence supporting early theories (summarized in [39]) regarding the warm-core nature of hurricanes and the consequent decrease of intensity of the storms with height. In the upper troposphere over the surface centers the pressure is high relative to the surroundings.

In view of the many observational difficulties with individual radiosonde ascents, it may be of interest to compare a sounding representing average conditions in hurricanes with the average tropical atmosphere in which the hurricanes form. E. Schacht (33) has computed the latter using the best available data. Table 4 gives his results, and Figure 3 contains a plot of the mean hurricane sounding. It is clear that Table 4 may be open to certain objections. The number of soundings used is small, and the observations were taken under adverse conditions in different parts of different storms. But a comparison of Tables 6 or 7 with Table 4 shows that the differences between conditions observed outside of and near hurricane centers are similar to those between the mean tropical and mean hurricane soundings. In consequence it is suggested that Table 4 may be considered indicative of conditions in many individual instances throughout the area of heavy precipitation near hurricane centers. The difference between mean tropical and hurricane soundings appears even greater when virtual temperature is utilized.

Table 4corroborates the idea that hurricanes are warm, low-pressure centers. The pressure difference between the hurricane and mean tropical soundings is largest at the ground and gradually decreases upward. At 11 km the hurricane pressures become higher than those of the surroundings. Temperatures and relative humidities throughout the troposphere are higher within the hurricanes than in their surroundings. Thus Figures 2 and 3 and Table 4 corroborate Byers' schematic diagram of hurricane structure (6; p. 432) but provide for dry-adiabatic ascent of surface air to its lifting condensation level.

The tropopause, remarkably, has practically no pressure or temperature variations. The high tropospheric vertical compensating downdrafts, believed to lower the tropopause over the eye of hurricanes, thus cannot extend far toward the outskirts of the hurricane circulations. It should be added that a sounding, released in the eye of a mature storm during October, 1944, at Tampa, Florida, also failed to record an unusually low tropopause, although it exhibited extraordinary stability in the middle troposphere. The invariance of the tropopause refutes the supposition, advanced by some, that the low pressure in hurricanes may be due in part to superposition of a "warm arctic stratosphere" over the lower tropical current. Decrease of pressure falls from the ground upward also demonstrates that this contention is not tenable.

It is well to point out that some evidence is apparently not in accord with Tables 6 and 7. Especially when upper isotherms are drawn on a map with a network of widely spaced radiosonde stations, the highest temperatures (i.e., at 10,000 and 20,000 feet) will frequently appear in the anticyclones flanking the storms. In these systems, owing to subsidence, there are large positive departures of temperature from the mean tropical atmosphere. Because such highs have a great horizontal extent, a sounding from their interior will usually be available. The warm zone within hurricanes, however, extends only over a very restricted area. Therefore, the likelihood that an observation will be taken in the inner storm area is relatively small, especially if the violence of the weather and very adverse release conditions are also considered. Reports usually will be received only from the warm high and the intermediate zone between this high and the storm, where temperatures are near or even below normal. This creates the impression that the temperatures drop quite generally toward the hurricane center, while the few soundings taken near the inner part of the storm show that the temperature gradient actually reverses there.

It should be noted, however, that in mature storms of tropical origin, which are connected with a very large high-latitude trough, the temperature distribution at times actually appears to differ from that just described, especially late in the hurricane season. Under such circumstances the wind variation with height also is dissimilar from that normally encountered, and conditions exist that are no longer typical of those prevailing during the earlier stages of the life-history of tropical cyclones.

Discussion of the upper pressure and temperature variations.-- The question arises: How do the high temperatures throughout the body of the hurricanes originate? As indicated in Figure 3, the mean hurricane sounding is nearly moist adiabatic; it follows closely the ascent of the surface air given by the dashed line of Figure 2. It therefore appears probable that the interior region of hurricanes, excepting, perhaps, the eye, is filled with air that has risen from the vicinity of the surface, that is, from the "subcloud" layer, in contrast with the common tropical perturbations.

It is not at all obvious how the surface air first penetrates a deep atmospheric layer over a relatively wide area. It is easier to understand the maintenance and growth of the warm-core vortex once it has formed, even though the rotational motion may as yet be slight. Hann (17; p. 620) suggests a close analogy with a tank experiment carried out by Helmholtz (24). Following Hann's reasoning, the air, crossing the isobars toward lower pressure, accelerates radially. As it approaches the center of rotation, however, the centrifugal force acting on it increases rapidly because of the development of a large tangential velocity, initiated by the effect of the earth's rotation. Thus the radial motion subsequently decreases. Eventually the pressure-gradient force will no longer exceed the centrifugal force. Cyclostrophic balance then exists with respect to the air column considered, or gradient balance, if the Coriolis force is also taken into account.

Hann also states with reference to Helmholtz's experiment: "Especially instructive in this experiment is the fact that because of a rotational motion which is initially weak the distribution of energy in the fluid will become entirely different from what it would be without the introduction of this motion [concentration of energy toward the center of the vortex]."

An exception to the establishment of cyclostrophic balance occurs in the friction layer. Here, as suggested by Hann, motion toward the center is still possible because of the frictional loss of momentum near the ground. Shaw (35) and Willett (42) mention the same point. Haurwitz (23), using certain assumptions, shows that the gradient wind level is lower in cyclones than in anticyclones and also that it lowers as the gradient wind speed increases. In hurricanes the gradient wind level may be depressed to an elevation of only a few hundred feet above sea level.

The foregoing shows that the convergence near hurricane centers becomes limited to the layer nearest the ocean surface, and mostly surface air ascends in the interior of a storm. In this manner the vertical temperature distribution as given in Table 4 can be explained.

The problem also arises as to how the upper outflow takes place. As indicated in Table 4, the pressure gradient decreases with elevation and reverses its sign above 10 km. In the upper troposphere, therefore, the air rising near the center is accelerated outward.

The actual outflow, however, will begin at much lower elevations. Durst and Sutcliffe (13) have considered the effect of the angular momentum of rising surface air on the outflow aloft. Because of the decrease of pressure-gradient force with height, the rising air seeks to draw away from the center as soon as the centrifugal force exceeds the pressure-gradient force. Values indicating the variations of the pressure-gradient force with height are tabulated below, based on Table 4 (cf. also [35]). The values in the right-hand column are the ratios of the pressure-gradient force at different elevations to that at the surface, expressed in per cent. The variation with height is surprisingly regular except at 9 and 16 km. The departures from the regular trend at these two levels cannot be considered significant.

H(km)	Per Cent	H(km)	Per Cent
Sfc.	100	10.	3
1	98	11.	-17
2	92	12.	-31
3	90	13.	-35
4	80	14.	-60
5	70	15.	-84
6	63	16.	-59
7	54	17.	-131
8	28	18.	-142
9	25		

Assuming that the ascending current retains its angular momentum, withdrawal from the central area would start approximately at the 3-km level. At this elevation the magnitude of the pressure-gradient force directed toward the hurricane center begins to drop significantly with height. In reality, outflow will begin at even lower heights, owing to decrease of the frictional force above the ground.

Durst and Sutcliffe also postulate an increase of the vertical velocity with height and show that such an increase may contribute to the radial outflow of the air aloft. In view of the preceding discussion, however, which restricts the convergence to the vicinity of the surface, it is unlikely that the upward vertical motion increases with height except near the ground.

3. First Formation of Tropical Low-Pressure Centers

In the cases of hurricane formation noted in the course of the study, deepening began, without exception, in pre-existing perturbations such as were described in the first part of this section. There was no evidence of spontaneous formation due to convection over an overheated tropical ocean.

The suggestion that hurricanes of the West Indies form from upper perturbations is not at all new. Even in 1915 Hann (17; pp. 620-21), following Bigelow (1), notes a connection between cyclogenesis in the eastern Caribbean and the mean seasonal upper-wind field over that area. Hann states:

"The cloud observations in the West Indian region have pointed toward another region of origin of the cyclones. During summer, the easterly trades predominate over the Greater and Lesser Antilles in the low troposphere to the altocumulus level (5 km). Higher up, however, there is another wind system, from the altostratus to the cirrus level (6-10 km). Over the Lesser Antilles there are southeast and south winds, which presumably belong to the Atlantic high-pressure region. In contrast, northerly winds predominate in the western Caribbean (Jamaica, Cuba). They appear to be connected with the eastern border of a western high-pressure cell at those levels. Between these two wind systems there exists a region of counter currents, and therefore probably also an area of lower pressure. Here there is an opportunity for the formation of a cyclone in the upper levels, when the pressure distribution just described and the corresponding currents are intensified."

As brought out by Bigelow and by Hann, the mean trough noted by them increases in intensity upward. Yet, though tropical storms originate in cold troughs of this kind, the deepening to hurricane strength, as borne out by available data, takes place at first near the ground. Gradually, as a surface vortex develops, the deepening extends into the upper air. The problem then arises how a cold-core disturbance, increasing upward in intensity to at least 3 km and often to the high troposphere, can generate a warm-core perturbation which is strongest near the ground and weakens upward.

Very few suggestions as to possible processes of cyclogenesis applicable to the tropics have appeared in the literature. Of these, the "convective hypothesis" in its simplest form must be discarded, as brought out by many authors (cf. [35]). As an alternative, the possibility of development of unstable waves along a surface of velocity discontinuity has been mentioned (4), as well as the generation of unstable waves in narrow but finite zones of wind shear (18, 23). It has also been stressed by various sources, especially by the adherents of the Frankfurt school of meteorology (33), that the superposition of a high-level disturbance on a lower perturbation can result in deepening. The superposition produces changes in the surface pressure field, and rotational motion develops as a result of these changes.

In the following two sections it will be considered under what circumstances a transition from cold- to warm-core systems takes place. Broadly speaking, it may be stated that the two suggested processes of cyclogenesis, noted above, appear to account satisfactorily for the vortices studied, particularly if they occur simultaneously. However, unless certain other processes are active in the area of cyclogenesis, the vortices remain feeble. These additional factors will be brought out in the ensuing discussion.

C. CYCLOGENESIS DUE TO SUPERPOSITION OF EXTRATROPICAL HIGH-LEVEL PERTURBATIONS ON WAVES IN THE EASTERLIES

1. Effect of Superposition on Field of Divergence

Even in midsummer, high tropospheric perturbations of the field of pressure that are connected with disturbances in the belt of polar westerlies occasionally overlie the tropical easterlies of the West Atlantic trade-wind belt. Waves progress westward in the lower trade current, while in the upper troposphere disturbances move from west to east. Because of this motion in opposite directions, an upper perturbation at times becomes superimposed on a lower wave in the easterlies. When this occurs, the tropical wave usually intensifies and very frequently also the middle-latitude part of the extratropical disturbance (cf. Part II). Such a superposition produces favorable circumstances for tropical cyclogenesis (30).

Ordinarily, low-level divergence, partially counterbalanced by upper convergence, precedes a wave that moves westward without change in shape (29). The wave trough slopes eastward with elevation. Conversely, an eastward-moving wave in the westerlies usually slopes westward with height. East of the wave low-level convergence is accompanied by upper divergence. The presence of low-level convergence in middle latitudes is demonstrated synoptically by the fact that the area of cloudiness and precipitation associated with the wave lies predominantly to its east. Since eastward propagation of the wave takes place at the same time, it follows that mass divergence which exceeds the lower convergence must be present at upper levels. To the rear, the reverse must be true.

The extratropical eastward-moving perturbation usually extends to more southerly latitudes in the high troposphere than in lower levels. This is most readily understood when the west wind increases in intensity to the tropopause and when the field of divergence may be explained following the reasoning of Bjerknes and Holmboe (3). The base of the westerlies then rises only gradually toward the equator. In the trade-wind zone lower easterlies are associated with upper westerlies, and the superposition of the upper trough on the lower wind field frequently results in a cyclonic deformation of the lower easterlies (29). Eastward propagation of high-level disturbances, however, also occurs on many occastions when the mid-latitude westerlies are weak or do not strengthen with height. The equatorward slope of the base of the westerlies then is usually steep, and the

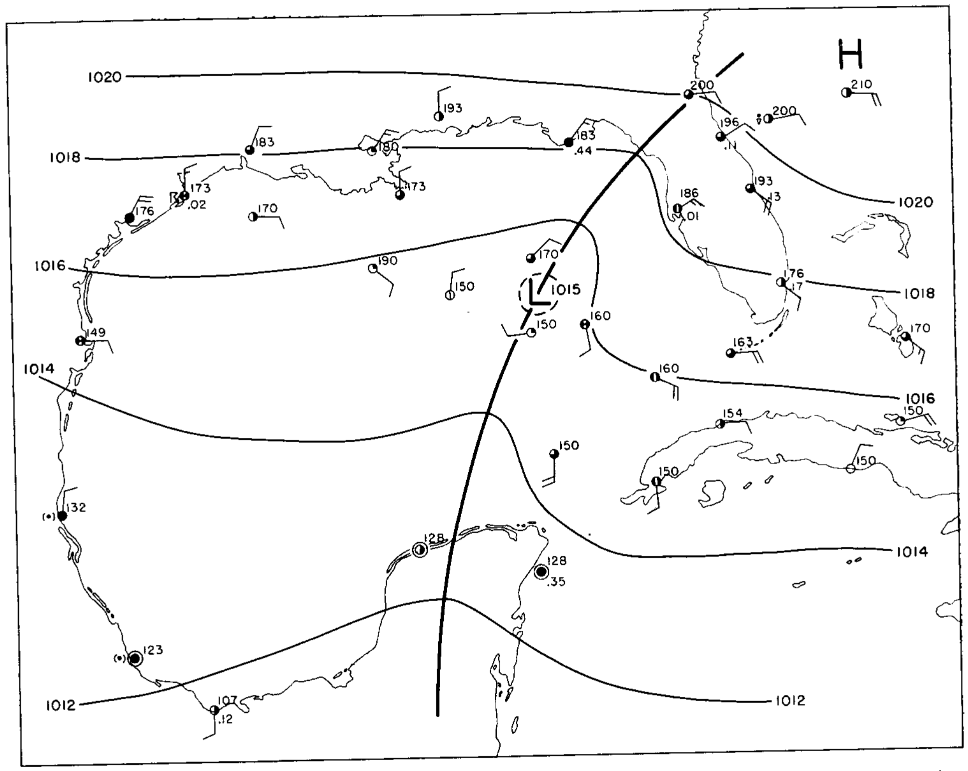

FIG. 4a.--Surface chart of the Gulf of Mexico and vicinity, September 10, 1941, 0730 EST.

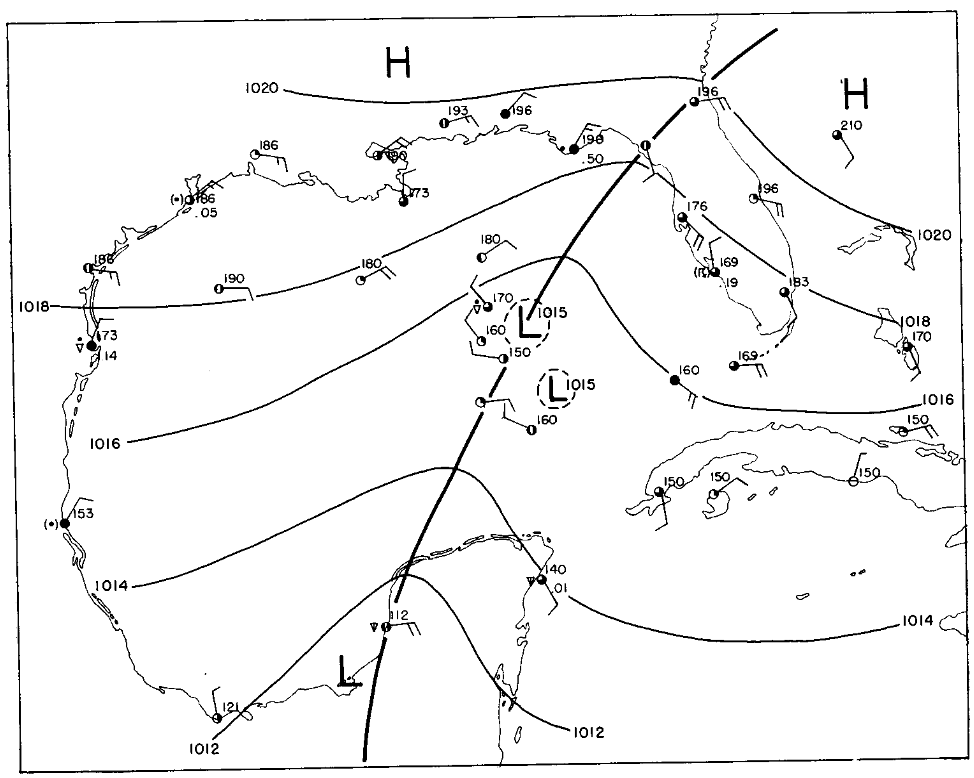

FIG. 4b.--Surface chart of the Gulf of Mexico and vicinity, September 10, 1941, 1330 EST.

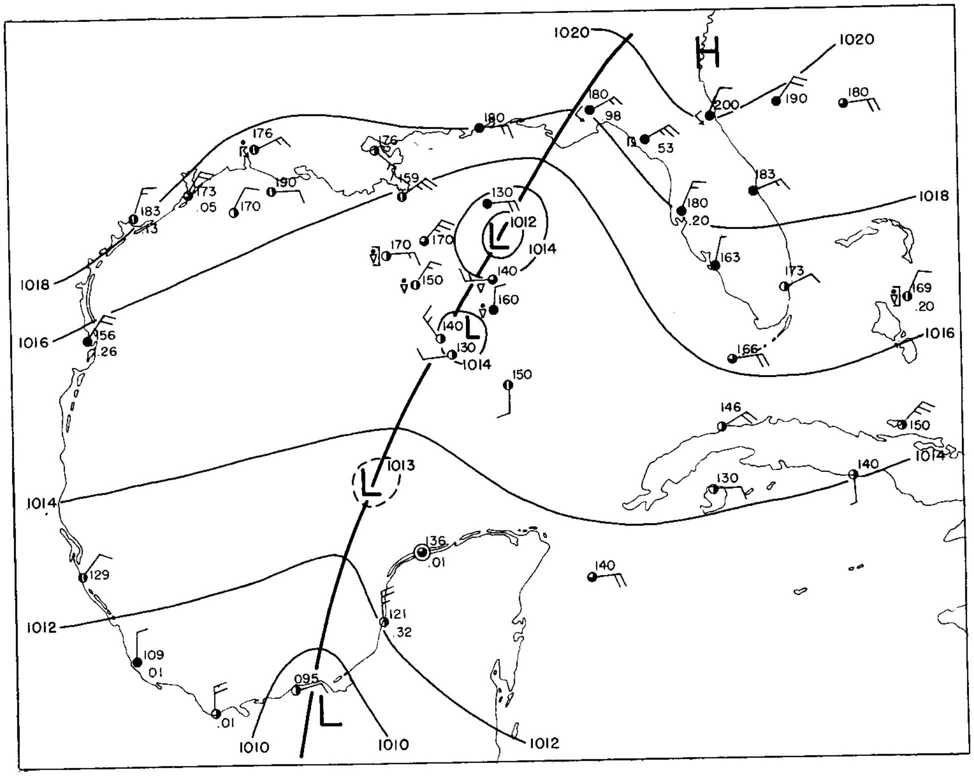

FIG. 4c.--Surface chart of the Gulf of Mexico and vicinity, September 10, 1941, 1930 EST.

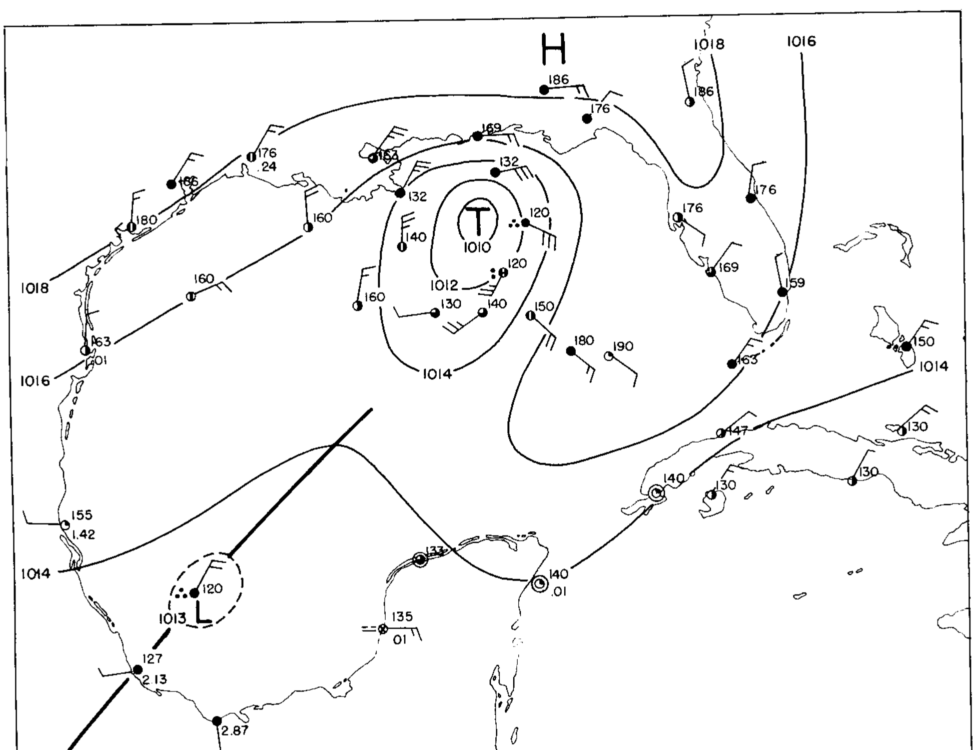

FIG. 5.--Surface chart of the Gulf of Mexico and vicinity, September 11, 1941, 0730 EST.

eastward-moving perturbation in the pressure field at high levels, for instance, at 10 km, may extend across the subtropical col into the easterlies at that level. An explanation of these remarkable yet frequently observed eastward displacements of high-level pressure troughs has not yet been given.

When a wave in the easterlies approaches the southern, high atmospheric portion of a perturbation in the westerlies, the region of divergence west of the wave in the easterlies will become superimposed on the zone of divergence in advance of the upper perturbation. Divergence then extends through a deep atmospheric layer, and this leads to deepening at the ground as the two systems approach each other. This deepening tends to accentuate the surface isobaric field especially near the surface position of the wave in the easterlies where cyclonic curvature of the isobars is already present. The tropical wave then begins to alter its shape, and the field of rotation, postulated by Hann (17), is established.

Since the strengthening of the field of divergence near a wave trough in the easterlies begins before the wave has merged with the extratropical disturbance, the deepening near the ground should also begin before the upper troughs have completed their juncture. This is often observed. Among many representative situations, the period September 9-11, 1941, has been selected for illustration. Cyclogenesis took place in the northeastern Gulf of Mexico, therefore in an area much closer to an adequate network of upper-wind reports than is usually available during periods of hurricane formation.

A tropical low-pressure center developed on a strong wave in the easterlies which had drifted westward across Florida into the Gulf of Mexico from the Antilles region. Prior to September 10, 0730 EST (Fig. 4a), there was no evidence of a low-pressure center at the surface. The first development therefore must have begun only a short time before the 0730 map. Later, on September 10 (Figs. 4b and 4c), ship reports indicated the presence of several small vortices along the wave trough. The reliability of some of these reports may be open to question, but it should be noted how well they fit together in Figure 5. As shown by this figure, the northernmost vortex, situated just south of the subtropical ridge line at high levels, grew rapidly, while the disturbance to its south was suppressed. Another small vortex appeared to be approaching the Mexican coast on September 11.

Prior to the beginning of cyclogenesis (Figs. 6a-6c) a pronounced wave in the westerlies had started to advance eastward from the Rockies, while the wave in the easterlies crossed Florida going westward. Deepening aloft over the Gulf coast took place on September 9 (Figs. 6b-6c), as a pressure trough developed at 10,000 and 15,000 feet between the disturbances in the windfield. This trough, located west of the surface position of the wave in the easterlies on September 9, 2300 EST, moved eastward on the 10th in conjunction with the wave in the westerlies. Deepening along the wave in the easterlies began as soon as the effect of superposition made itself felt during the morning hours on September 10. Later, at midnight of September 10, the northern and southern waves joined to form an almost continuous troughline.

2. Effect of Structure of Belt of Polar Westerlies

The characteristics of the equatorward portion of the belt of polar westerlies fall in summer, broadly speaking, into two essentially different classes:

1. Those connected with baroclinic westerly currents. The speed of the west wind increases with height to the upper troposphere, and the base of the westerlies slopes upward gradually toward the equator.

2. Those connected with quasi-barotropic westerly currents. The speed of the westerlies does not vary appreciably with elevation, and the slope of the base tends toward the vertical.

If the westerly current is baroclinic, or if the entire belt of westerlies extends to an unusually southern latitude, conditions exist which are very unfavorable for the generation of tropical storms. The streamline amplitude in the equatorward portion of the belt of westerlies then is small, and a zone of strong horizontal anticyclonic shear in the upper troposphere overlies the tropical easterlies. Waves in the upper westerlies, and also waves in the lower easterlies, move rapidly and have a small amplitude. They pass each other with little effect on their intensity when they become superimposed. In the anticyclonic shear zone at high levels the magnitude of the pressure changes connected with the eastward-moving disturbances decreases rapidly southward and the high-level systems are only weakly reflected in the lower atmosphere. In addition, the waves both in the easterlies and in the westerlies often move rapidly. Dunn (11) and Durham (12) previously noted that the likelihood of cyclogenesis in the tropical waves decreases if their speed is greater than average (15 mph).

Many months of data can be cited to demonstrate that, during periods in which this baroclinic westerly current predominates, a pressure and

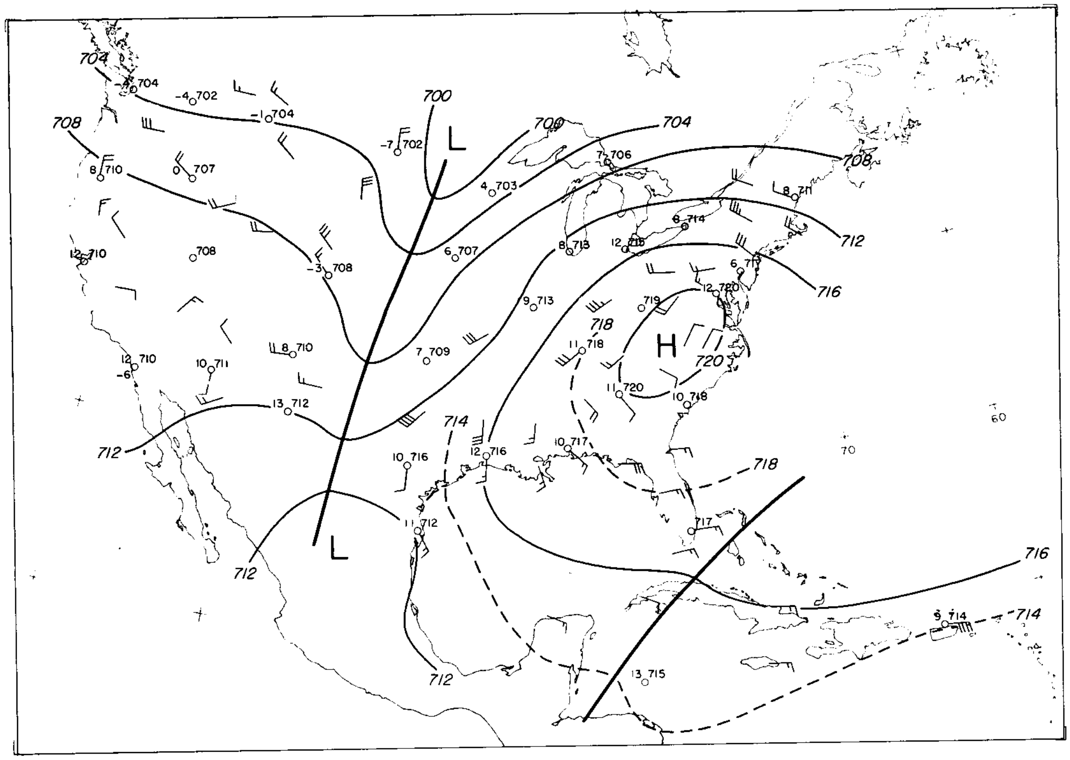

FIG. 6a.--Pressures and winds at 10,000 feet for the United States and West Indies, September 8, 1941, 2300 EST.

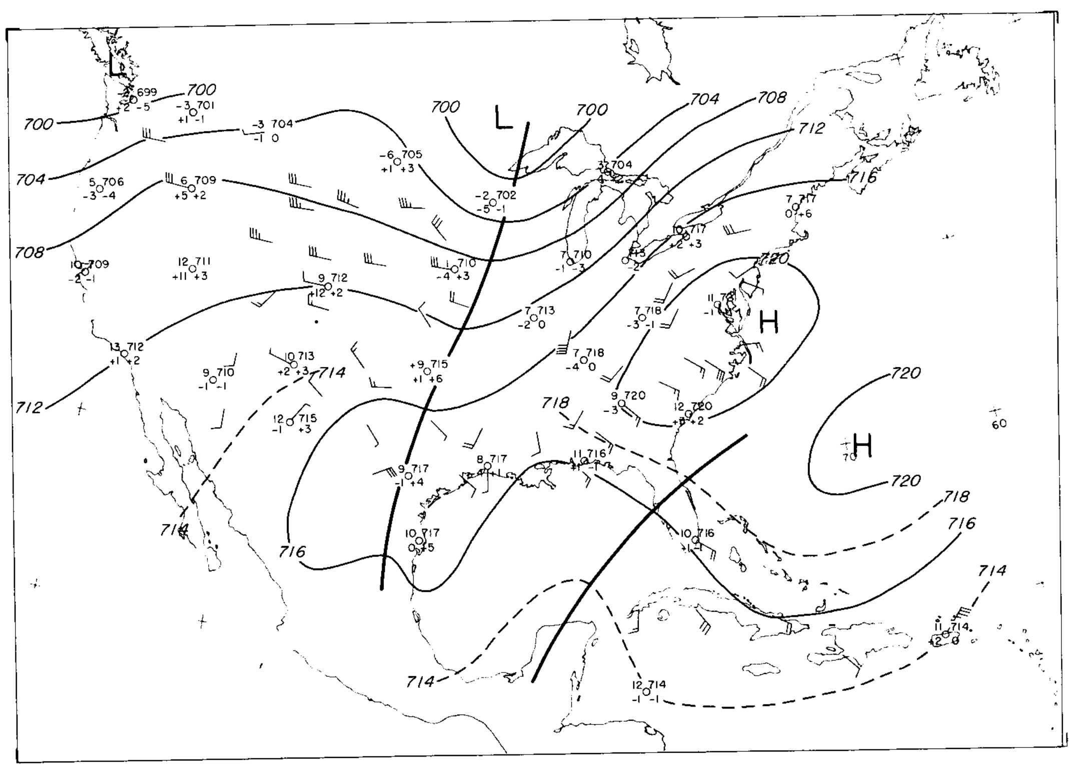

FIG. 6b.--Pressures and winds at 10,000 feet for the United States and West Indies, September 9, 1941, 2300 EST.

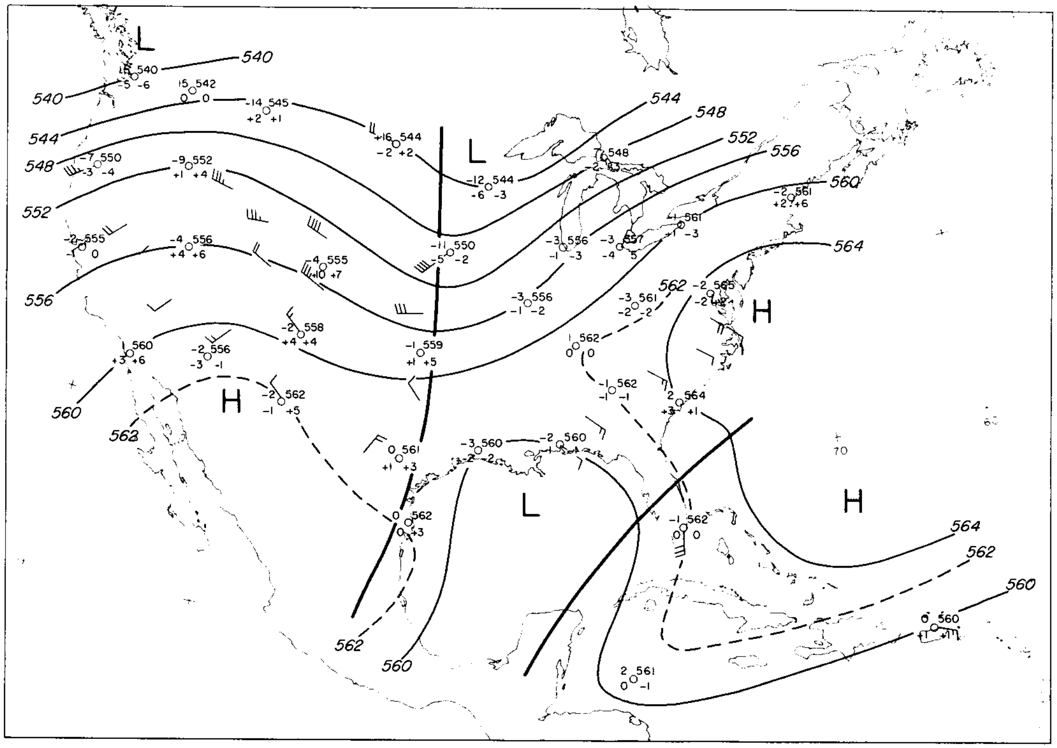

FIG. 6c.--Pressures and winds at 15,000 feet for the United States and West Indies, September 9, 1941, 2300 EST.

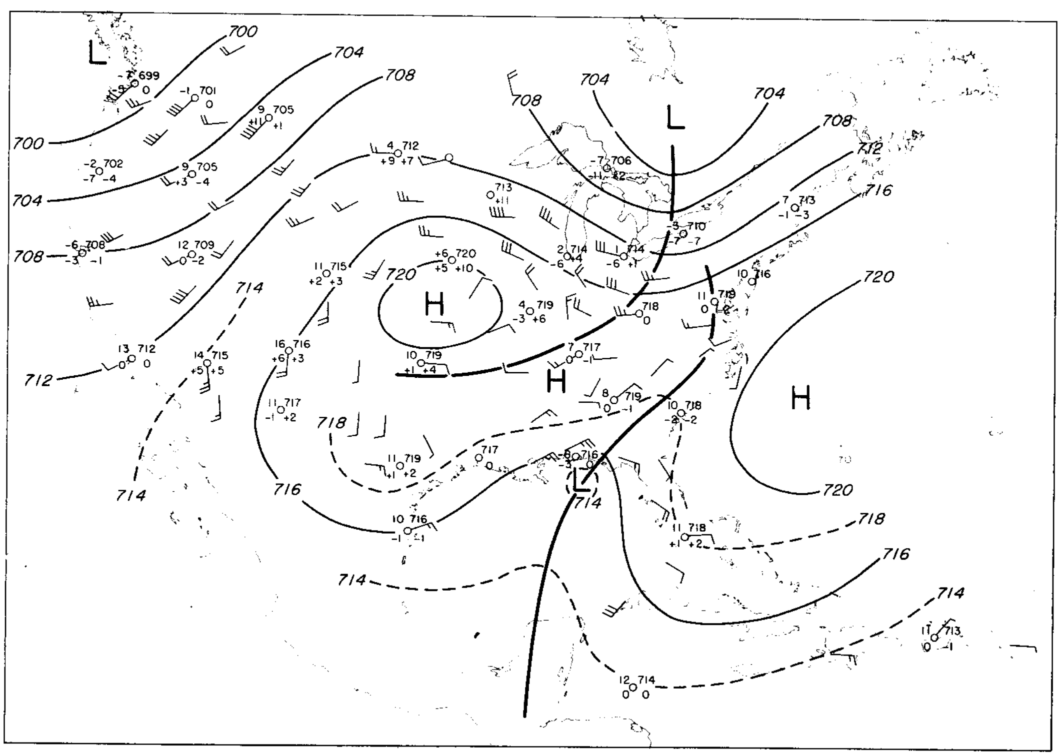

FIG. 6d.--Pressures and winds at 10,000 feet for the United States and West Indies, September 10, 1941, 2300 EST.

wind field that is potentially cyclogenetic usually does not even develop in the lower atmosphere. It is only when the speed of the upper westerly winds over the subtropics begins to decrease and when the subtropical ridgeline in the upper troposphere shifts poleward that the amplitude and intensity of tropical perturbations begin to increase and that major subtropical break-throughs develop.

The seasonal variation of hurricane frequency itself indicates that absence of broad-scale polar westerlies over the subtropics is necessary for the formation of tropical storms. Hurricanes originate most frequently in late summer and early fall. It is precisely at that time of year that the subtropical atmosphere approaches the barotropic state most closely. A barotropic state, once attained, usually persists for a few days and occasionally for as long as two weeks. Then there is an interruption, and a baroclinic westerly current predominates again. Barotropic conditions may return intermittently for several intervals during one hurricane season. Depending on the state of the general circulation, they may prevail for a total duration of 2 months or more in some years, while in other years they hardly appear at all.

Among all factors that seem to have a major bearing on the formation of hurricanes, the state of the general circulation during summer has the largest interannual variability. In fact, it seems to be the only factor that is variable to a high degree, not only from one summer to the next, but also from summer to winter. The writer considers it impossible to attribute the absence of hurricanes during winter to the slight seasonal temperature and humidity changes at the ground.[4] The persistent baroclinic structure of the polar westerlies during winter must be responsible, as reflected in changes of temperature- and humidity-lapse rate from summer to winter. During summer, barotropic conditions appear to last only a few weeks on the average, although statistics are still lacking. Therefore large-scale conditions favorable for formation of hurricanes usually exist only for a short part of each hurricane season and with sufficient breaks to make the actual generation of a tropical storm a rare event.

The foregoing indicates that for forecasts of hurricane formation it is of foremost importance to study the strength of the zonal currents throughout the troposphere and changes of the intensity of these circulations with time. The superposition of a high-level eastward-moving disturbance on a low-level wave in the easterlies will become significant only when the rapid

[4] Note also the occasional formation of tropical storms in midwinter over the Western Pacific Ocean.

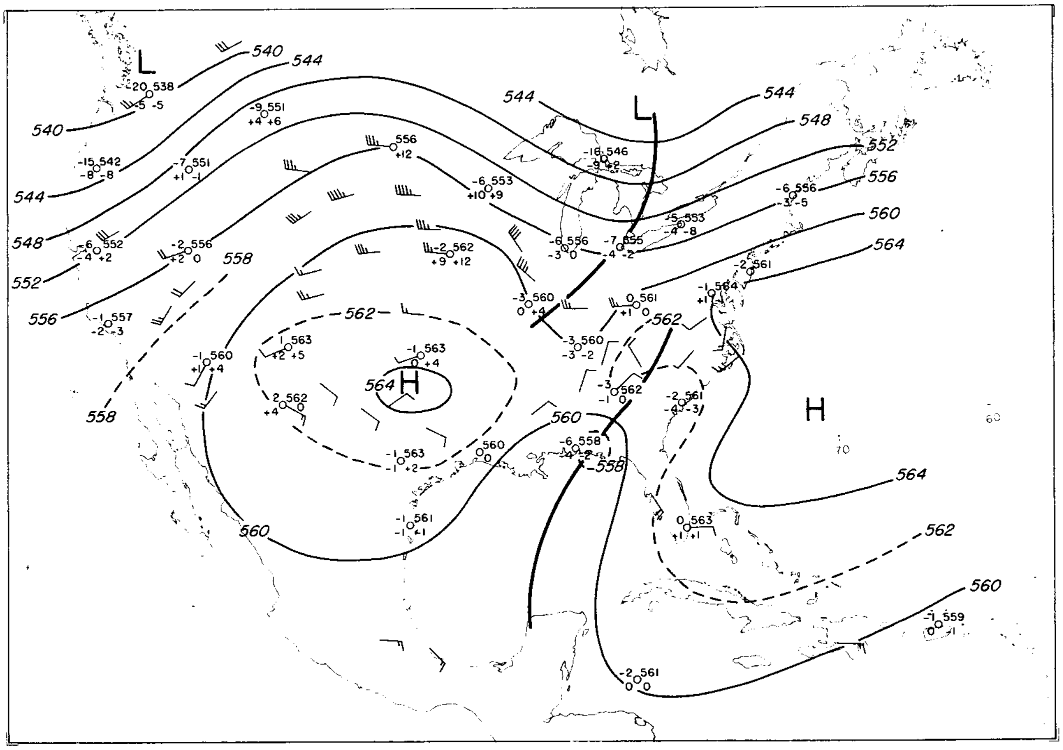

FIG. 6e.--Pressures and winds at 15,000 feet for the United States and West Indies, September 10, 1941, 2300 EST.

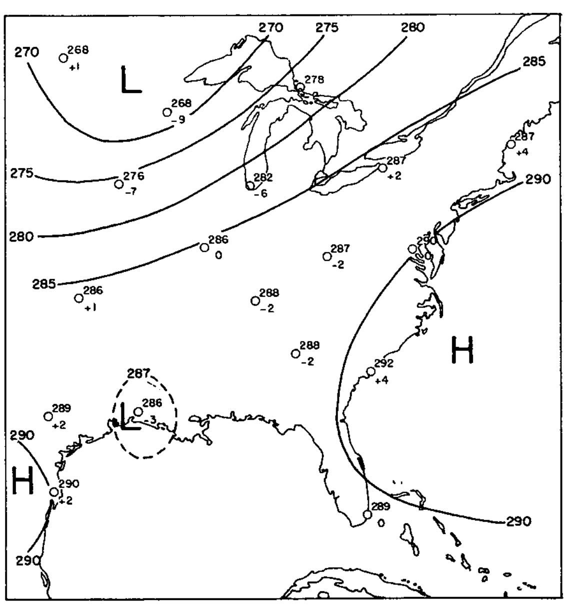

FIG. 7a.--Pressures at 10 km for the central and eastern United States, September 9, 1941, 2300 EST.

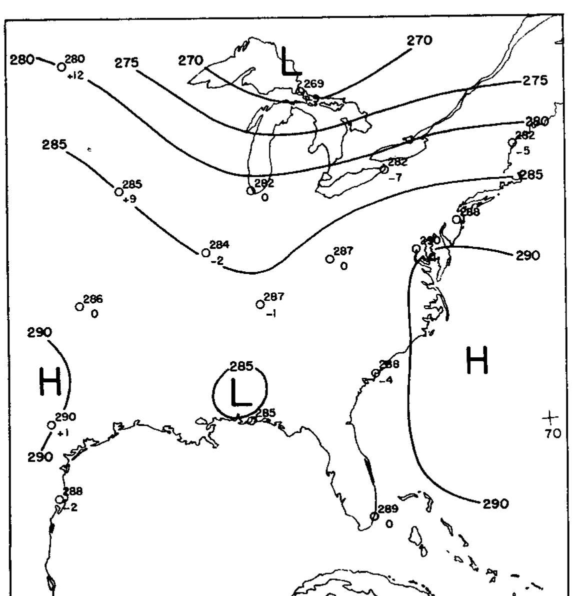

FIG. 7b.--Pressures at 10 km for the central and eastern United States, September 10, 1941, 2300 EST.

polar westerlies at high levels over the subtropics disappear or are not in existence initially. It may be seen from Figures 6 and 7 that during the period September 9-11, 1941, westerly winds were absent over the southeastern United States up to high levels.

While a decrease of the westerly circulation aloft is favorable for hurricane formation, the appearance of strong westerlies after a period of inaction, coupled with a southward displacement of the subtropical ridge line, tends to suppress any cyclogenesis that may be in progress in the trade-wind belt. The situation of July 18-19, 1944, over Cuba and the Bahamas furnishes a good example (not reproduced here). A strong wave in the easterlies met with an extratropical trough to the north. In spite of very favorable low-level conditions, deepening did not result, because the upper westerlies intensified and the subtropical ridge line at 20,000 feet and higher was displaced to the south of the position of the incipient center.

Figures 10a-10c of this report provide another example. They also demonstrate that it is very important to interpret correctly the type of west winds appearing in the high troposphere over the trade-wind belt. These westerlies may not be a part of the polar westerlies. Frequently they are separated from them by a band of deep easterlies. When this happens, the broad-scale pattern is highly favorable for vortex development in the tropics. Thus it is necessary to ascertain whether any observed high-level westerlies are a continuous portion of the polar westerlies or whether they are "equatorial" westerlies; whether they are indicative of the large-scale basic currents present or whether they are associated with perturbations superimposed on the basic current.

It has been suggested by some that pressure rises in the upper troposphere north of a wave in the easterlies indicate impending cyclogenesis. This is not consistent with the example shown in Figures 7a and 7b. Pressures at 10 km dropped in the central and eastern United States on September 9 and 10. The only rises on September 10 occurred west of the eastward-moving pressure trough, which was situated directly north of the region of cyclogenesis on September 10, 2300 EST (Fig. 7b). High-level pressure rises of this kind appear on practically all days of the year.

It is clear, however, that the 10-km pressures must rise over a wide area prior to tropical cyclogenesis if baroclinic westerlies prevail aloft initially, but, after the transition from a baroclinic to a barotropic broad-scale state in the subtropics, hurricane formation need not follow. The barotropic broad-scale state merely establishes a favorable basis for their generation.

3. Special Characteristics of Extratropical Disturbances

The amplitude of eastward-moving extratropical perturbations largely depends on the structure of the zonal current in which these disturbances travel, as may be expected. If the southern portion of the belt of polar westerlies is baroclinic, the amplitude of middle-latitude troughs south of the region of strongest westerlies becomes very small and the intensity of the troughs diminishes rapidly toward the subtropical ridgeline. As such a trough passes a wave in the easterlies, a subtropical col develops in the lower troposphere which is characterized by elongated lows and by highs of small latitudinal extent. As shown by Petterssen (27; p. 253, Fig. 113D, F) such a col is likely to possess anticyclonic vorticity. In most cases its existence is only temporary.

If the circulation in the subtropics tends toward a barotropic state, an increase of the amplitude of the equatorward part of the extratropical troughs usually occurs. When a meeting between such a trough and a wave in the easterlies takes place, the subtropical col that develops in the lower atmosphere will be marked by narrow lows and elongated highs of great latitudinal extent. Following Petterssen (27; p. 253, Fig. 113, G-I), that is characteristic of a col with cyclonic vorticity. Moreover, the cyclonic vorticity is usually observed to be localized in a small area. Such a col is likely to persist so that the juncture of the northern and southern perturbations completes a subtropical break-through.

It is evidently not sufficient to consider only the streamline amplitude to judge the intensity of a trough. If the winds are very weak, large changes of wind direction across a trough are not necessarily an indication of a strong system. It is estimated, without statistical verification, that extratropical troughs must be accompanied by a 90-degree turning of the wind at 10,000 feet in the belt of strongest westerlies, and a wind speed of at least 20 mph, if the trough is to become a significant factor for deepening in the tropics.

As an extratropical trough increases in amplitude under the circumstances here considered, it frequently tends to develop a long, equatorward bulge (cf. Figs. 6b-6c). When it becomes superimposed on a wave in the easterlies, westerly winds, at times at an elevation as low as 10,000 feet, may appear over the tropical system. Thus there is a third type of west wind that may appear over the trade-wind belt during summer, further emphasizing the need for a careful interpretation of low-latitude upper-wind observations. The westerlies just mentioned are characterized by a very limited east-west extent. They are not a part of the general circumpolar westerlies, but appear only over a restricted area along the extended trough. In the Antilles only one station may report a wind with a west component, while 200 miles to the east and to the west deep easterlies prevail and the subtropical highs are centered farther north than normal. Along the trough line, the westerly winds may be continuous to high latitudes. However, there is little, if any, increase of the wind speed northward. On numerous occasions, the westerly component decreased north of the Antilles, suggestive of a closed cyclonic circulation aloft.

If such westerly winds of very limited zonal extent appear aloft over the tropical portion of an extended trough, the possibility of cyclogenesis should always be considered. One statistical study (12) attempted to show, although with limited material, that the likelihood of vortex formation increases as the base of these westerlies lowers.

The more pronounced the extratropical upper trough, the stronger the reaction which may be expected in the tropics. High-level deepening north of the subtropical col is followed by tropical cyclogenesis with surprising frequency. The period October 11-14, 1944, in the United States and the Caribbean is an excellent example (not reproduced). Due-south winds of 50 mph and more in the United States appeared east of a 10,000-foot trough of extreme intensity, and due-north winds of the same strength to its west. A dynamic high was present over New England, and the trough rotated counterclockwise. Rapid cyclogenesis in this trough in the United States coincided with hurricane formation on an equatorial shearline in the western Caribbean between Jamaica and Cuba.

4. Concentration of Vorticity

The growth of the field of rotation brought about by the superposition of an extratropical disturbance on a wave in the easterlies is also affected by the instantaneous distribution of relative vorticity (22), which becomes concentrated in convergent areas. Concentration of a field with anticyclonic relative vorticity reduces the effectiveness of convergence in producing increases of cyclonic vorticity, especially near the equator, where the deflecting force of the earth's rotation is small. Conversely, concentration of a field with cyclonic vorticity should be favorable for cyclogenesis.

Because of lack of data, computations of vorticity over the oceans are almost never possible. Therefore, important aspects of the problem of hurricane formation, especially in connection with flow across the equator, could not be studied, except in a very qualitative manner. In the situations analyzed it appeared that only feeble vortices resulted from superposition if the low-level winds to the north and east of an area of deepening were subnormal. Conversely, an increase of the wind speed, especially to the northeast of a forming center, contributed to intensification, presumably largely because of the generation of an area of cyclonic wind shear and its subsequent concentration near the vortex core. The increase of the strength of the easterlies usually coincided with surface pressure rises and intensification of the subtropical high at the surface northeast of the area of deepening. These pressure rises in summer can be explained readily if the broad-scale circulation changes from a baroclinic to a barotropic state.

Surface pressure rises of the kind described took place in the West Atlantic High that was flanking the area of deepening, September 9-11, 1941. The rises occurred mainly on September 8-9, as may be seen from the maps reproduced in an earlier report (29). On September 10 there was an appreciable surface pressure gradient to the northeast of the developing vortex.

Conversely, the wind field north and northeast of a cyclogenetic zone east of Florida, September 19-21, 1944, was very weak (not reproduced). In most other respects the situation was very favorable for strong deepening. Nevertheless, only a weak cyclonic center developed. Several days later this center drifted westward into the Gulf of Mexico, where its intensity increased as the pressure gradient to the north tightened.

On several occasions a zone of strong cyclonic wind shear in the low troposphere also developed south of latitude 20 to the east of waves in the easterlies that became part of an extended trough. The formation of such a shear zone contributed to the instability of the tropical waves. A rapid transition from light southeasterly winds of 5-15 mph to strong southeasterly or east-southeasterly winds of 30-50 mph marked the position of the shearline which in the horizontal plane was oriented nearly parallel to the wave trough itself. In the lower levels this shearline was situated 100-300 miles east of the principal troughline and coincided with the area where light upper westerly and southwesterly winds near the main troughline gave way rapidly to deep easterlies. Thus the shearline separated the area dominated by the tropical wave trough and that under the influence of the succeeding ridge. Entry of the shearline into the cyclogenetic area near the wave trough accelerated the deepening (cf. also [17], pp. 763 ff.) Its arrival, however, was favorable for deepening only within the framework of the broad-scale flow pattern described in this part of the report. Under other circumstances, weakening of the wave in the easterlies frequently resulted.

Formation of a shearline will occur only if an unusually pronounced ridge or high-pressure center follows or develops to the rear of the wave in the easterlies. The formation of such a dynamic high in lower latitudes occurs when an extended trough of great amplitude is present, that is to say, during the prevalence of the broad-scale circulation pattern considered necessary for cyclogenesis. For such situations the subtropical high east of the extended troughline spreads far to the north and south and, as suggested in an earlier report (28), tends to break up into smaller "meridional highs." Owing to this splitting, a separate center of high pressure may develop near latitude 20. As this new center strengthens to the east and even southeast of the most active portion of the wave in the easterlies, pressure gradients there intensify, causing an acceleration of air toward lower pressure and producing convergence. The cyclonic shear east of the tropical wave intensifies, and after some time a well-defined shearline characterized by rapid transition from weak to strong winds can be distinguished.

Another common feature of the low-latitude anticyclone is its small east-west dimension and the close proximity of the next trough to the east. Such a small zonal extent of the high is in agreement with the broad-scale pattern of troughs and ridges of great amplitude.

The West Atlantic High present on September 10-11, 1941, was characterized by a similar small east-west extent. Upper winds at Miami turned from south to east late on September 10 (Figs. 6d-6e). Six hours later the wind direction was northeast. The approach of a trough from the east was also indicated by a pronounced change in the surface isobaric field south and east of Florida (Figs. 4c, 5).

In general, the period September 9-11, 1941, demonstrates that the various conditions suggested as necessary or favorable for cyclogenesis were satisfied to a reasonable degree. On the basis of the very strong indications on September 9, 2300 EST, it would seem plausible to forecast a cyclonic development.

D. Cyclogenesis Near Equatorial Shearlines

1. The Bjerknes Theory

Well before the turn of the century meteorologists had observed that tropical storms in the West Pacific Ocean form regularly in the shearing zone between the trade winds and the surface equatorial westerlies. This observation led Bjerknes (4; pp. 756 ff.) to extend the theory of cyclogenesis on the polar front to equatorial regions:

"The tropical cyclone probably exhibits initially wave characteristics analogous to extratropical cyclones, but does not attain great intensity during this stage and is not recognized as a hurricane. The mature tropical cyclone, the tropical hurricane, is always a vortex, therefore rather symmetrical and cannot show marked fronts. . . . The problem regarding the great energy of the tropical storm has not yet been solved on the basis of direct aerological measurements. There are, however, strong indications that the atmosphere is conditionally unstable near the place of formation but that a release of a special kind is necessary to cause vertical overturning on a large scale with gain of kinetic energy. If the intertropical-front surface is dynamically unstable, and if at the same time the conditional instability is sufficiently great, conditions for large-scale overturning are present."*

The development of shearing waves is dependent on the amount of shear present and, as shown by Haurwitz (20,23), on the width of the shearing zone. Therefore, it becomes of interest to determine to what extent intense, narrow shear zones, comparable to the West Pacific equatorial convergence zone, appear over the western Atlantic. Such shear zones are characterized by a flow of air parallel rather than normal to the axes of the systems and thus do not possess a wave structure.

2. Equatorial Shearlines in the Western Tropical Atlantic

Except for the Panama region, a surface equatorial convergence zone is seldom present over the waters east of Central America. Occasionally, along the Guiana coast of South America, surface west winds appear which do not seem to be of a local character. Aloft, however, the situation is very different. The prolonged presence of a cyclonic shearing zone at high

*Writer's translation.

levels over the western tropical Atlantic has been noted by Bigelow (1) and by Georgii and Seilkopf (14). Their findings have been summarized by Stone (37). The existence of high-level shearlines, oriented mainly east-west and therefore located between meridional highs, has also been mentioned by the writer (28). At the surface there often is no indication of these shearlines, while aloft they appear with a cyclonic turning of the wind that may attain almost 180 degrees. East winds then prevail to the north of the line and west winds to its south.

Upper-wind observations over the Atlantic and the Caribbean have confirmed the ideas stated above. Frequently, in summer, westerlies appear above 20,000 feet over the Caribbean area which are not a part of the belt of polar westerlies. They are equatorial westerlies, since there are deep easterlies farther north. As a check, the zonal wind directions at Bermuda and San Juan, Puerto Rico, were compared for the periods June 15 - October 15, 1944 and June 15 - October 14, 1945. The results for days when simultaneous measurements were made at high levels at both stations are given in Table 8. All days with light winds aloft or very small zonal wind components were omitted.

TABLE 8

Group	Direction of Wind		Heights of Observations				
	San Juan	Bermuda	20,000 ft.		30,000 ft.		40,000 ft.
1. . . .	East	East	53	31%	37	28%	4
2. . . .	West	West	15	9%	19	14%	7
3. . . .	East	West	93	53%	51	39%	4
4. . . .	West	East	12	7%	24	19%	9
Total number of observations.			173	100%	131	100%	24

Groups 1 and 3 predominate at 20,000 feet, in correspondence with the normal flow pattern of the middle troposphere. But the picture changes considerably between 20,000 and 30,000 feet. Group 1 loses fewer observations on a percentage basis than might be expected from some accounts of the general circulation. The percentual drop of Group 3 is not surprising, but the increase of observations in Group 2 is far smaller than would be expected. Westerly winds at both San Juan and Bermuda occurred mainly when the polar westerlies temporarily reached wintertime strength, and this happened only infrequently. The tendency for westerly winds at San Juan to be associated with easterly winds at Bermuda increases greatly between 20,000 and 30,000 feet. At 40,000 feet Group 4 is largest, but, owing to the small number of observations at that level, the figures there may not be representative.

The persistence of equatorial shearlines and of the anticyclonic circulations to the north and south, once established, is great; and they may dominate the tropical weather for a week or more without interruption. It has been noted (29; 4; p. 737) that the formation of meridional highs in winter is connected with quasi-periodic outbreaks of polar air that produce a transformation of the subtropical highs. The widespread existence of meridional highs during summer indicates a general circulation problem which has not yet been solved. For the present, it must suffice to note that upper highs will split when they attain the dimensions of surface subtropical highs (28). If the belt of polar westerlies is displaced to the north of its mean position and its southern portion is quasi-barotropic, this splitting will almost always occur.

If there are polar westerlies over large parts of the trade-wind belt at upper levels, equatorial shearlines cannot be present at a latitude sufficiently removed from the equator to permit cyclogenesis. Nevertheless, traces of the shearlines are observed under such circumstances, even in midwinter. In this season the easterlies north of the line may be replaced by weak westerlies, with stronger westerlies to the south and much stronger westerlies farther north. Such a nonlinear distribution of shear of the zonal wind component in the upper troposphere seems to be a semi-permanent feature of the high-level wind field over large areas of the tropics throughout the year. Although tropical cyclogenesis is infrequent in winter, the influence of the upper shear zones on tropical rainfall should remain considerable in areas subject to its periodic and aperiodic variations.

3. Structure and Formation of Equatorial Shearlines

According to classical concepts an intertropical front forms when a warmer air mass from the summer hemisphere and a colder air mass from the winter hemisphere converge in low latitudes. This front should slope upward toward the winter hemisphere.

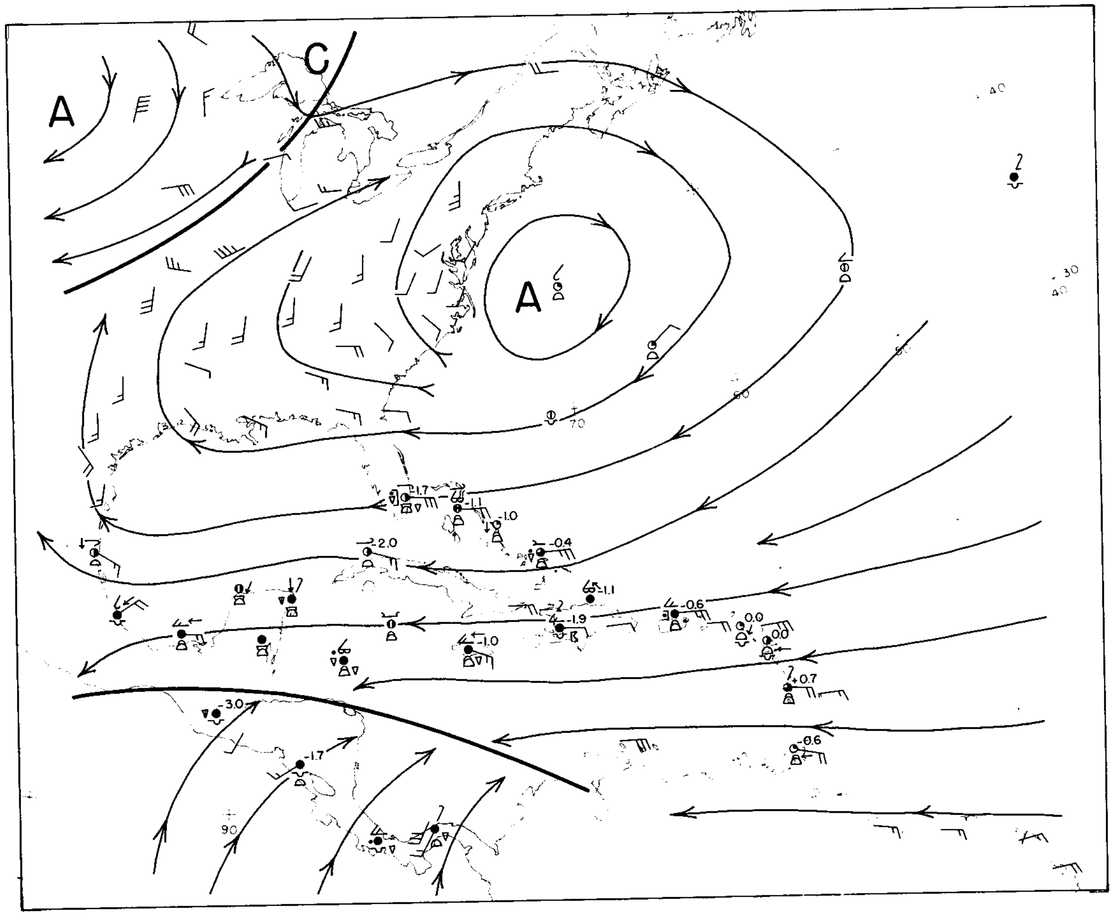

FIG. 8.--Basic chart for the Caribbean area, September 27, 1945, 1700 EST.

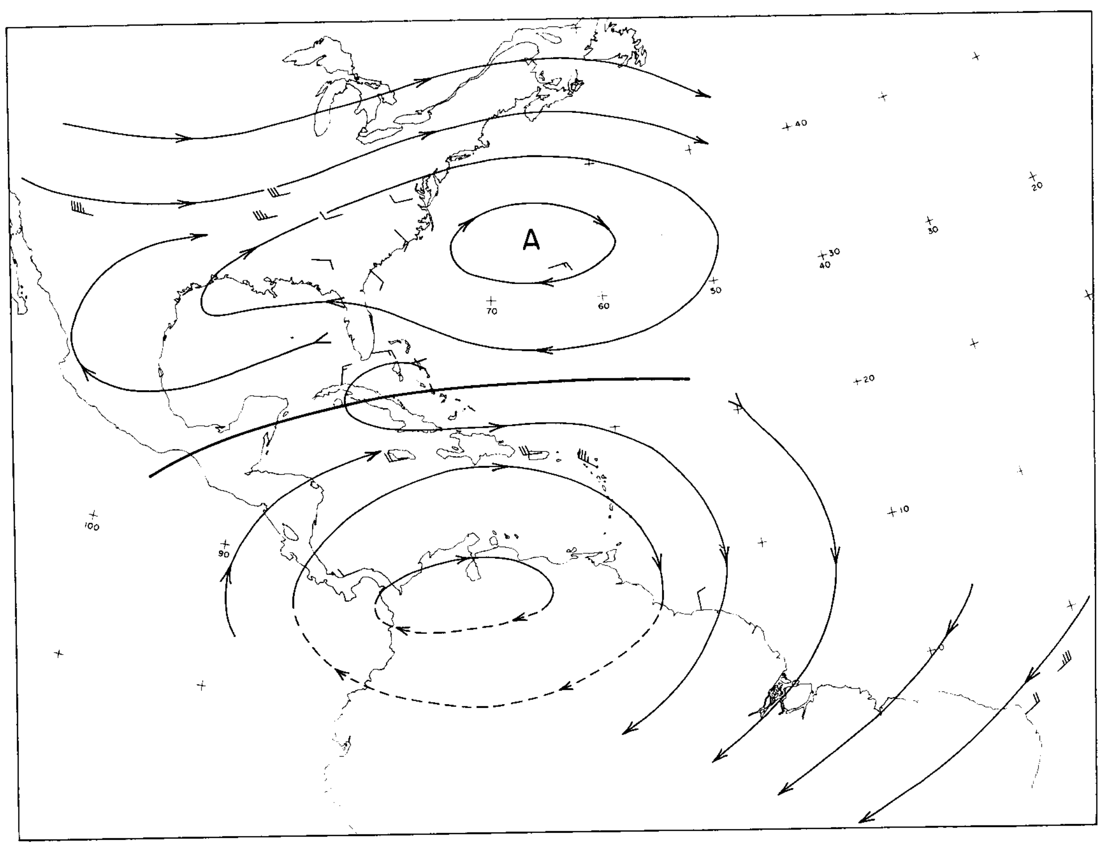

FIG. 9.--Winds at 30,000 feet, September 27, 1945, 1700 EST (including winds within 12 hours of map time).

During the three Northern Hemisphere summers studied, the slope of equatorial shearlines observed in the Caribbean was vertical or northward - into the summer hemisphere. Figures 8 and 9 show a representative case of a shearline which persisted in an almost invariant position for about 10 days. Therefore, many more observations were available for the analysis than appear in the illustrations. In the lower levels (Fig. 8), westerlies in the western Caribbean outlined the equatorial convergence zone In the eastern Caribbean there was only a weak cyclonic shear in the easterlies at 5,000 feet. At 10 km (Fig. 9) the shearline was extremely well marked. Its position at that level was far to the north of the 5,000-foot position, and its intensity was greater at the upper level. The westerly current south of the line most probably came from the Southern Hemisphere.

Although the radiosonde network was too sparse to ascertain accurately the temperature distribution across the shearlines noted in the case of the study, characteristic changes of temperature and pressure at one station, as shown in Table 9, occurred in several instances. During the period October 4-7, 1945, the shearline illustrated in Figures 8 and 9 moved northward from the Caribbean across the Greater Antilles into the Atlantic (cf. Figs. 10a and 10b). Intensification in the low troposphere had taken place between September 27 and October 4. At first, temperatures and pressures aloft fell at San Juan, during approach of the shearline, and convection became very intense. At 10 km the troughline passed near midnight, October 5 (Fig. 10a). As the Southern Hemisphere air arrived, temperatures and pressures began to rise in the upper troposphere. The rises then appeared at successively lower elevations as the base of the equatorial westerlies lowered.

These data indicate that at upper levels the coldest air was concentrated near the troughline, which sloped northward with height. Warmer temperatures prevailed outside the system, to the north and south. This observation suggests a dynamic rather than an advective origin of the temperature gradients. If active, low-tropospheric divergence took place north of the surface position of the northward-moving shearline and low-level divergence to its south, the observed temperature changes could be explained along the lines discussed in the initial part of this report. The weather sequence at Puerto Rico, shown in Figure 10a and verified by personal observation of the writer, confirms that such a distribution of convergence and divergence must have existed. The most violent convection took place during the night of October 4-5. At that time strong easterlies

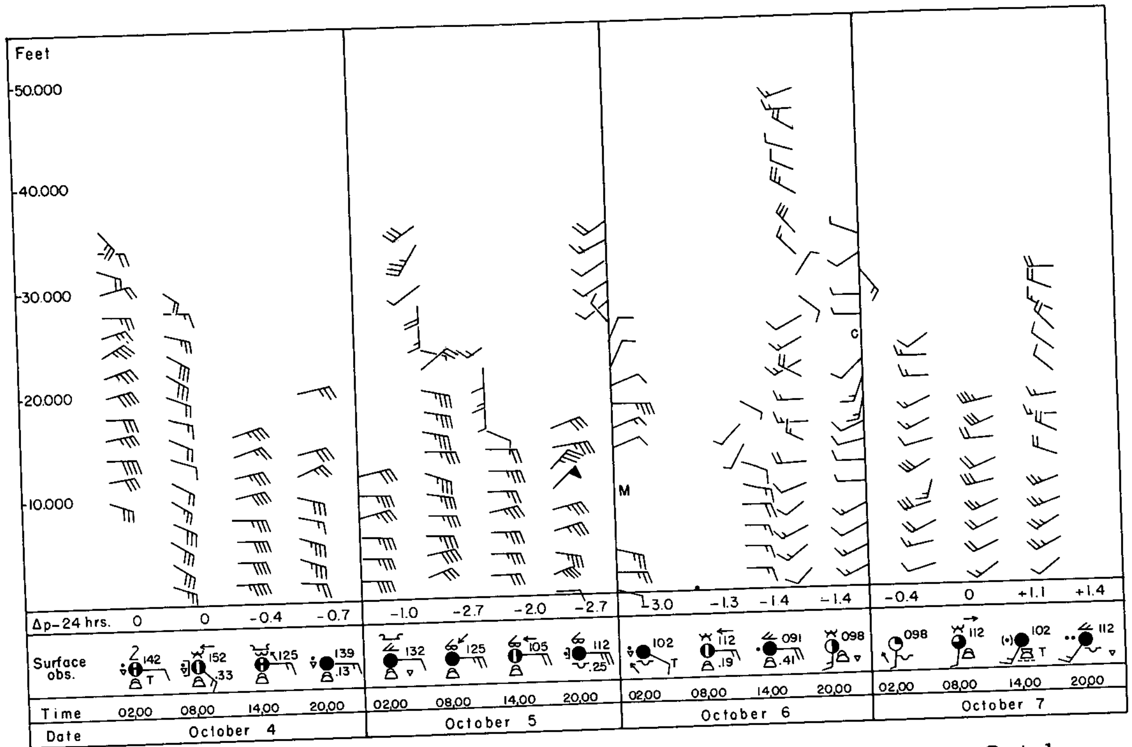

FIG. 10a.--Vertical time-section at San Juan, Puerto Rico, October 4-7, 1945.

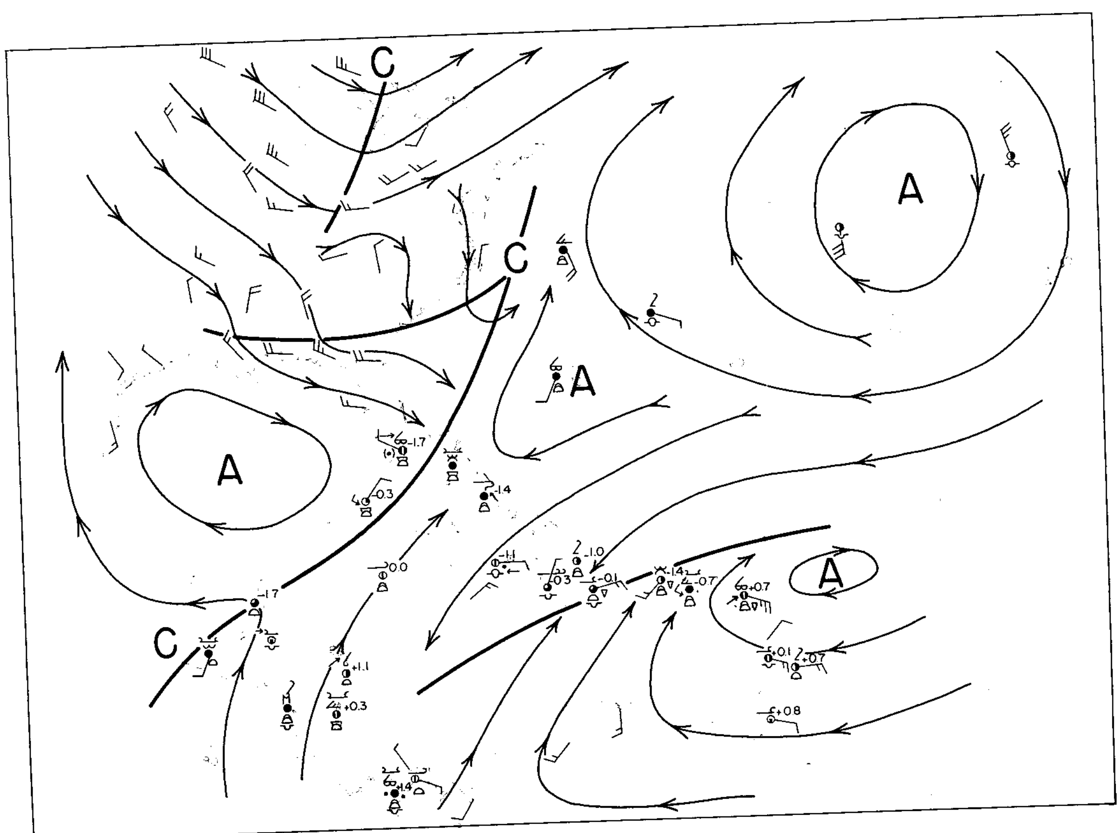

FIG. 10b.--Basic chart for the Caribbean area, October 6, 1945, 1930 EST.

through a deep layer still prevailed, and there was no overrunning of Southern Hemisphere air. The clearing on October 6 was spectacular.

It was the object of the preceding remarks to establish the dynamic origin of the temperature contrasts and weather changes. The weather distribution with respect to a shearline may differ in other cases from that described here. Deppermann (9) finds active convection in the "southwest monsoon" but also a different vertical structure of that current. In this instance the observations also point to a dynamic origin of the rainfall. It seems untenable to consider horizontal density advection as the factor primarily responsible for the existence of the West Atlantic equatorial shearlines. This may be confirmed also by noting how intense the shearlines are at 10 km, while at 5,000 feet an analyst often may be unable to trace them. Warm air ascending along a frontal surface is said to diverge and turn anticyclonically (4). Vertical motion is assumed to decrease with height above the frontal surface, and therefore the frontal wind shear must decrease from lower to upper elevations. It should therefore be very difficult to observe a front in the high troposphere where the normal meridional temperature gradient is very small. Yet it is precisely at high elevations that the equatorial shearlines attain their greatest intensity.

The streamline pattern in the equatorial zone (Fig. 9) suggests another approach to the problem of the formation of equatorial shearlines. This flow pattern persisted for several days so that the streamlines represent trajectories. Southern Hemisphere air enters the Northern Hemisphere as a southwesterly current and leaves it again after some time on a persistent clockwise path. Such motion may be explained by means of vorticity considerations (15; 31; 32); on this basis the position of the equatorial shearline is given by the maximum latitude which the Southern Hemisphere air can reach in the Northern Hemisphere. It should be noted that, owing to accelerations in the equatorial region, the path described by the equatorial westerlies in the Northern Hemisphere will differ somewhat from that prescribed in the case of constant absolute vorticity. The curvature of the trajectories should be smaller, and therefore the penetration into the Northern Hemisphere greater.

In the foregoing explanation we assume that an asymmetry exists between the trade currents of both hemispheres. Such asymmetry is usually present, because of seasonal differences between the hemispheres and because of synoptic changes in the subtropics. At an equatorial shearline air masses from widely distant regions of the world are brought in close proximity. In this sense the term "front" is not inappropriate. However, it is not the temperature difference but the vorticity contrast that is the characteristic and important feature of equatorial shearlines.

TABLE 9

UPPER 24-HOUR PRESSURE AND TEMPERATURE VARIATIONS AT SAN JUAN, PUERTO RICO OCTOBER 4-6, 1945

	Height	Δ p(mb)	Δ T(°C)
Oct. 4-5, 1000 EST	5,000 ft.	-3	-2
	10,000 ft.	-3	-2
	15,000 ft.	-3	-1
	20,000 ft.	-2	0
	10 km	-1	1
Oct. 4-5, 2200 EST	5,000 ft.	-4	0
	10,000 ft.	-1	1
	15,000 ft.	0	1
	20,000 ft.	1	1
	10 km	1	1
Oct. 5-6, 1000 EST	5,000 ft.	0	2
	10,000 ft.	1	4
	15,000 ft.	3	2
	20,000 ft.	2	1
	10 km	2	1
	13 km	3	2
Oct. 5-6, 2200 EST	5,000 ft.	1	3
	10,000 ft.	1	3
	15,000 ft.	3	1
	20,000 ft.	2	2
	10 km	3	3

4. Formation of Vortices at Equatorial Shearlines

The horizontal wind shear across equatorial and east-west shearlines frequently is similar to that shown in Figures 8-10: there is a reversal of wind direction of almost 180 degrees, and the shear is concentrated in a very narrow zone. Within this zone the field of motion is so nearly discontinuous that very favorable conditions for formation of shearing waves exist. The great majority of tropical storms noted during the study formed in connection with such shear zones.

Vortex development begins when the horizontal wind shear across shearlines is intensified due to interaction of another disturbance, especially an extratropical trough. Such occurrences are quite similar to those described previously in this report and will be discussed further subsequently.

As dynamic instability develops, there is a tendency toward simultaneous appearance of several distinct centers in many cyclogenetic situations. The spacing between these vortices may be only 50-100 miles, while the diameter of individual systems can be as small as 10-15 miles initially. Formation of such vortices may take place over a distance of several hundred miles along shearlines and along well-marked waves in the easterlies. According to Frolow, of the French Meteorological Service at Martinique, F.W.I., this fact explains the great discrepancies between various weather services in locating incipient storm centers. As time goes on, one vortex grows, and then there is agreement as to position of the hurricane. Numerous instances of the kind mentioned by Frolow can be found on historical weather maps. G. Norton, in charge of the hurricane forecast center of the United States Weather Bureau at Miami, also has pointed out to the writer that great care is necessary in determining the location of the principal disturbance. During the period September 8-10, 1944, for example (cf. Fig. 27), a weak cyclonic circulation drifted into the northeastern Caribbean. Winds aloft at that time did not hint at the presence of a second center. Subsequently, however, weather reconnaissance flights found another more intense storm northeast of Antigua, B.W.I., This disturbance grew rapidly to become a hurricane of great proportions, while the low inside the Caribbean Sea vanished.

Figures 4-5 suggest that a corresponding development occurred in the Gulf of Mexico on September 10-11, 1941. Intensification of the horizontal wind shear across the wave in the easterlies at that time in the Gulf is likely to have contributed to the cyclonic development. In this instance the northernmost center became the dominant circulation, a representative occurrence in situations when instability develops at waves in the easterlies or along shearlines which lie at a considerable angle to the latitude circles. The northward increase of the Coriolis acceleration helps the northern disturbance to develop its cyclonic circulation most rapidly. In addition, the effect of the superposition of upper extratropical troughs weakens southward. It should be stressed, however, that the northernmost position referred to here is the highest latitude south of the subtropical ridgeline in the middle and upper troposphere.

Equatorial shearlines, once formed, tend to be stagnant and influence tropical weather steadily for a week or more. Thus "families of vortices" will appear along these lines, analogous to cyclone families of higher latitudes. At the beginning of August, 1945, for example, and during the middle of September of the same year, a series of three vortices approached the Lesser Antilles from the Central Atlantic. Only one vortex of each family developed to great strength. This tendency toward the production of vortex trains has already been noted by Hann (17; p. 617).

5. On the Origin of Certain Vortices

According to the perturbation theory (4), an initial disturbance is postulated which causes the air at the shearline to withdraw from the central zone. As a result, the surface pressure drops, and thus the rotational field is initiated which was described in Section B and which is also mentioned by Bjerknes (4; pp. 757-58). As the wind shear initially is greatest at some distance above the ground, the tropical perturbations, in contrast to frontal waves, must originate well above the surface layer. This leads to ascent of air below the layer in which the shearing wave forms. Depending on the lapse rate and moisture content of the rising air, low-level compensation for surface pressure falls can and frequently does follow. The surface low-pressure center then fills again before assuming major proportions. Under certain circumstances compensation does not occur. The observations contain an indication regarding the development of the initial disturbance at waves in the easterlies and shearlines with a marked asymmetry of weather (cf. Figs. 1a, 1b, 10a, and 15b). In these cases a forming low-pressure center was situated at the boundary between a very clear and a very cloudy area, that is, between an area of downward and an area of upward vertical motion. Such an initial position of the surface depression is very frequent and suggests a connection between the impulses for cyclogenesis and the presence of a strong horizontal shear in the vertical motion. This connection may be established by considering the interaction between horizontal and vertical components of vorticity.

The contribution of convergence and divergence to changes in the vertical components of relative vorticity brought about by vertical velocity gradients is usually neglected. This contribution is

$$\frac{\partial u}{\partial z}\frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}\frac{\partial w}{\partial x} ,$$

where u, v, and w are the wind components defined in the usual manner. This term is not considered negligible by Hesselberg and Friedman (25), who point out that it may attain an appreciable order of magnitude. For considerations of changes of broad-scale flow patterns it is perhaps justified to neglect this effect. But it is the small-scale circulations that are involved here.

Along a wave in the easterlies, as shown in Figures 1a and 1b, $\partial w/\partial x$ is positive. Thus, when the south wind decreases and the wave slopes eastward with height, the term $-\frac{\partial v}{\partial z}\frac{\partial w}{\partial x}$ makes a contribution to cyclonic vorticity. Map experience confirms that vortex formation is unlikely if a wave slopes westward with height.

If the east component of the broad-scale flow is constant with height near a deepening wave in the easterlies, $\frac{\partial u}{\partial z}\frac{\partial w}{\partial y}$ will not affect the vorticity about the vertical axis. If the east component decreases upward, the term will give a positive contribution only if the area of greatest convection lies to the north of the region of deepening. Such a distribution of the vertical velocity along a meridian is unusual, at least in the tropics. It occurs mainly when an extended trough is present. Normally the disturbed weather diminishes toward the subtropical ridgeline. It is therefore again the case of the extended trough which offers the most favorable conditions for cyclogenesis.

In order to compute the approximate magnitude of the vertical motion term, the gradient of the vertical velocity must be postulated. Restricting the discussion to $-\frac{\partial v}{\partial z}\frac{\partial w}{\partial x}$ and referring to the 3-km level, a downward motion of only 2 cm/sec just ahead of the wave trough and the same amount of upward motion to its rear at that level will be assumed. This corresponds to a vertical displacement of about 1700 m/day, if the same air were to remain in the zone of convergence or divergence for 24 hours, which usually is not the case. The horizontal convergence or divergence necessary to produce this vertical motion may be calculated from the equation of continuity, which, for an incompressible atmosphere, can be written

$$w = - \int_0^h div_2 \mathbb{V} \, dz \quad .$$

This equation, although approximate, will at least indicate whether the order of magnitude of the assumed vertical motion is reasonable. If h=3 km as specified before, the mean value of divergence or convergence in the layer between sea level and 3 km necessary to produce the postulated vertical motion will be 6.7×10^{-6} sec^{-1}, or 2.4 mph/100 miles. Certainly this represents a minimum value in narrow zones of convection of the type here considered.

If transition from downward to upward vertical velocity occurs within a distance of 100 km and $\frac{\partial v}{\partial z}$ = 10 mps/4 km (in reasonable agreement with Fig. 1), the magnitude of $-\frac{\partial v}{\partial z}\frac{\partial w}{\partial x}$ will be 4.3×10^{-5} sec^{-1} per 12 hours. If this vorticity change is expressed as shear only, the cyclonic shear will increase (or the anticyclonic shear decrease) 16 mph/100 miles in 12 hours. At latitude 20 the convergence in the term $-f \, div_2 \mathbb{V}$ of the vorticity equation has to reach a magnitude of 7.5 mph/100 miles, or 2.1×10^{-5} sec^{-1} in order for this term to produce an equal change of the relative vorticity. Therefore, in the problem under discussion, convergence and divergence have a greater effect on changes of the relative vorticity of air parcels through generation of rotation around a horizontal axis than through the latitude term which is ordinarily considered.

The foregoing suggests that the following sequence of events may take place during the initial formation of cyclonic vortices under the conditions here considered. At first, the field of horizontal convergence and divergence in the area of cyclogenesis is strengthened through interaction with a middle-latitude disturbance. A quasi-discontinuous weather pattern results, featured by an abrupt transition from extreme convection to clear skies. The resulting rotation in vertical planes produces rotation around a vertical axis. The rapidity with which this rotation grows depends on the initial distribution of cumulonimbi. Where large clouds are present at the outset, cyclonic whirls and therewith also surface low-pressure centers form with greatest speed. In this manner the initial disturbances can be generated and upward funneling of large quantities of surface air within a small area initiated. Further growth and maintenance of the vortices evidently must depend on release of latent heat of condensation. But interaction of a middle-latitude disturbance of finite, even large, amplitude seems to be a necessary mechanism for the initiation of organized release of latent heat.

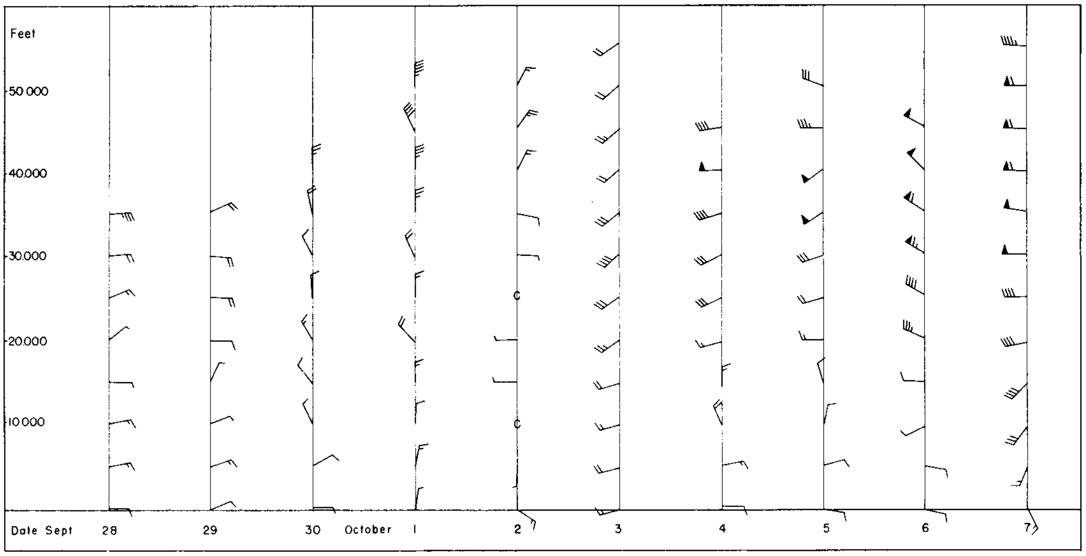

Fig. 10c.--Vertical time-section of upper winds at Bermuda, September 27-October 7, 1945. Observations at 1100 EST.

6. Influence of Extratropical Troughs

Intense shearlines may persist over the West Indies for long periods of time without vortex development. For instance, the shearline shown in Figure 9 remained in a relatively steady state for more than a week, as the broad-scale flow pattern remained unbroken.

On September 29-30 the Bermuda winds at last changed decisively, backing from east to north (Fig. 10c). One day later the southwesterly flow over the southeastern United States had broken down. The trough shown situated over the Rocky Mountains in Figure 9 moved eastward into the Mississippi Basin. Another trough developed over the southeastern United States and extended rapidly into the Caribbean. Hurricane formation near the intersection of this trough and the equatorial shearline near Swan Island followed immediately. This example shows that, as for waves in the easterlies, it is necessary for an extratropical trough to intersect an equatorial shearline if a hurricane is to form. Such a point of intersection has been referred to as "triple point" by various authors.

The remarks made earlier regarding special characteristics of extratropical troughs that are conducive to hurricane formation along waves in the easterlies also apply to troughs that intersect equatorial shearlines. Another case that brings out the importance of the presence of an extratropical trough of considerable amplitude is presented in Figures

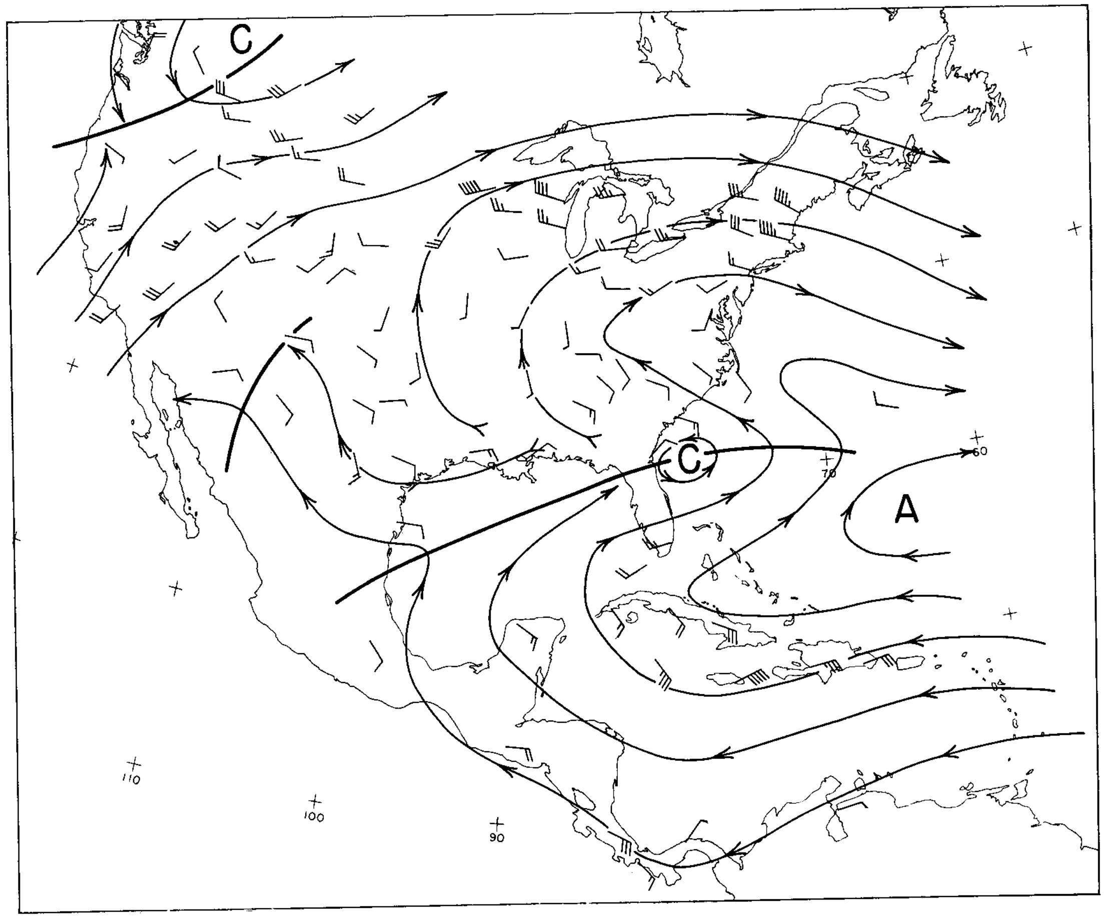

FIG. 11a.--Winds at 10,000 feet for the Caribbean area and the United States, June 25, 1943, 2300 EST, and June 26, 0500 EST, combined.

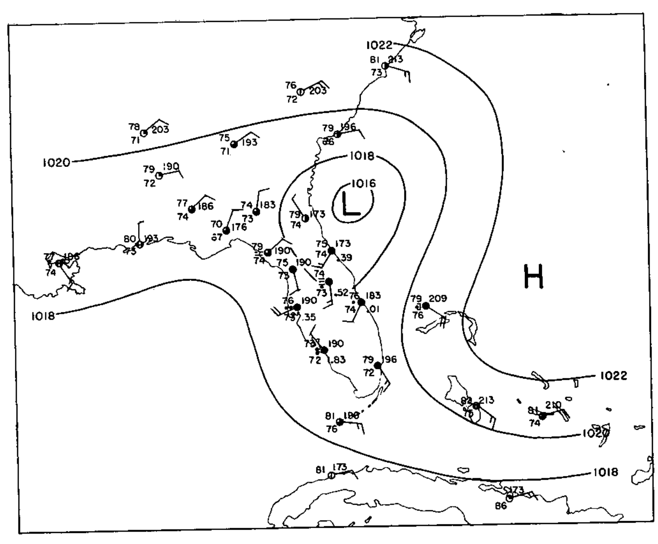

FIG. 11b.--Surface chart for the southeastern United States, June 26, 1943, 0730 EST.

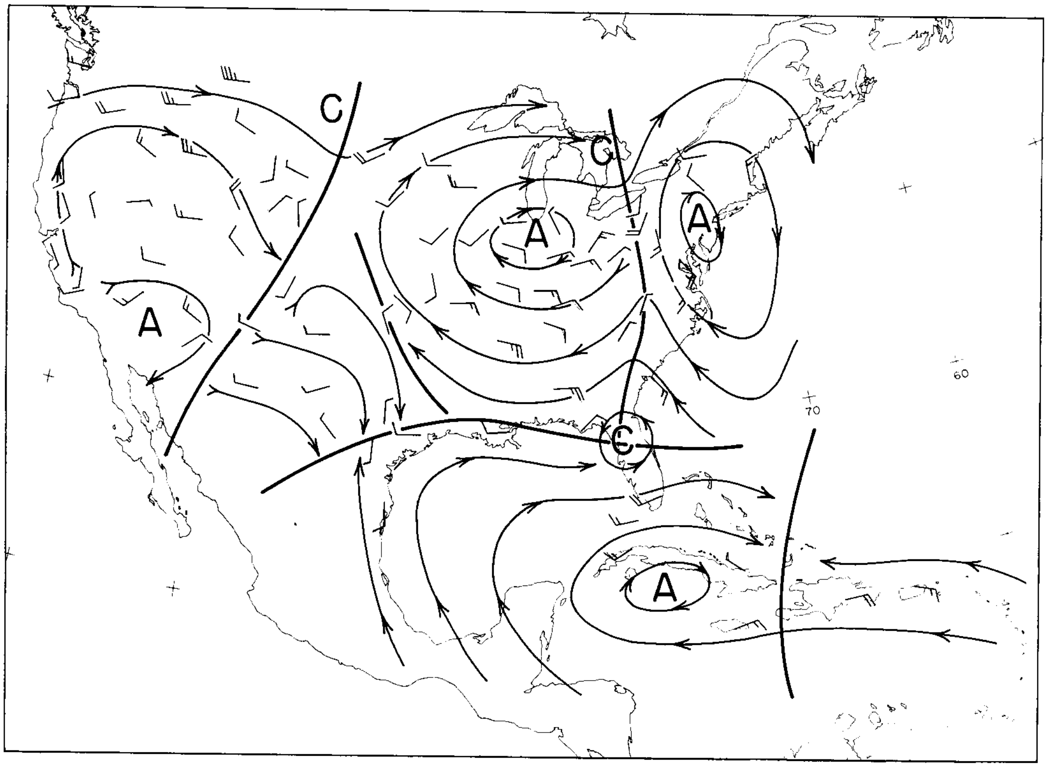

Fig. 11c.-- Winds at 10,000 feet for the United States and the Caribbean, August 2, 1940, 2300 EST.

11a-11c. A strong east-west shearline persisted for more than a week over the Gulf of Mexico during the latter part of June, 1943. Two days before the map shown a pronounced low-latitude high-pressure center had arrived from the Atlantic moving westward to the rear of an active shearline. Prior to this, the shearline had stagnated between Central America and southern Florida, oriented east-northeast--west-southwest. As the speed of the trade winds increased, an upper vortex and a weak surface low-pressure center developed near southern Florida. The whole system gradually advanced northward, strengthened slowly, and was attended by severe weather. Figures 11a and 11b show it at a stage when hurricane formation would seem imminent. But nothing happened. Throughout the critical period, as shown in Figure 11a, no active trough arrived to the north of the shearline. Gradually the promising surface center, located just east of Florida, disintegrated.

In contrast note Figure 11c. The low-latitude circulation and the shearline on August 2, 1940, corresponded closely to the picture of Figure 11a. North of the line, however, a pronounced trough extended southward, and the large amplitude of the northern system was further emphasized by

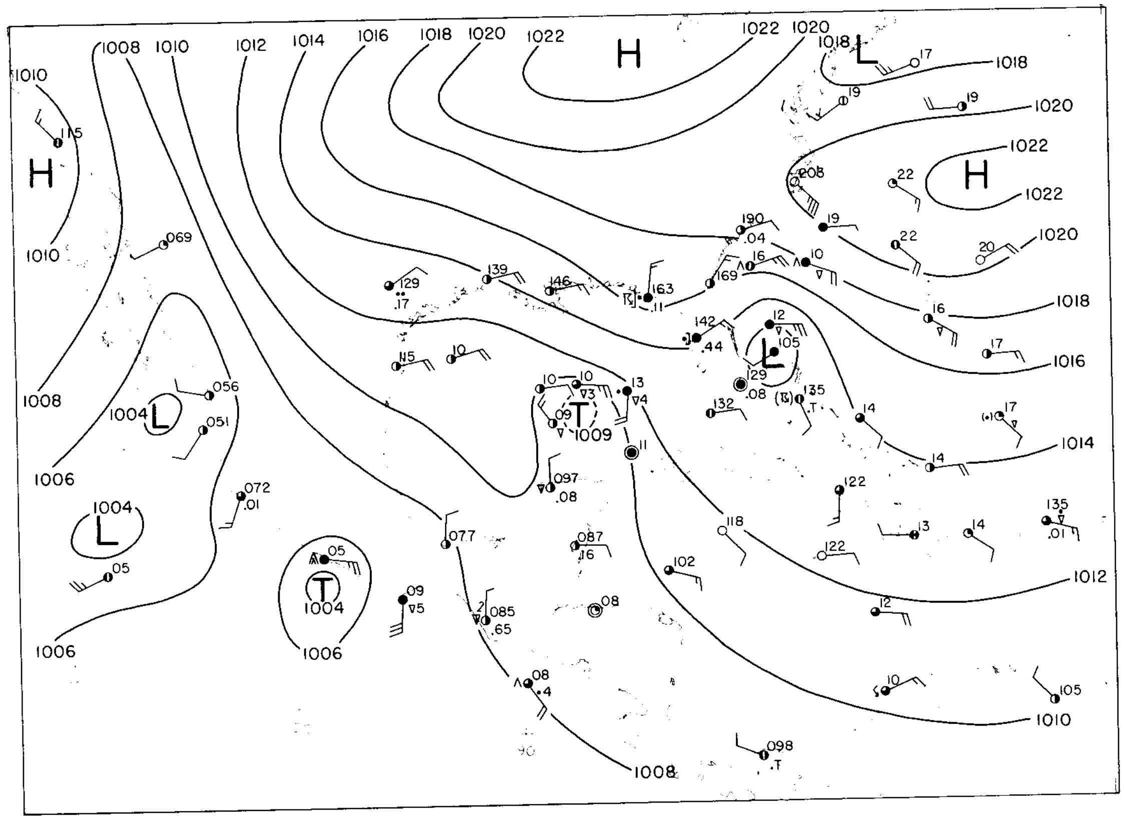

FIG. 12a.--Surface chart for the western Atlantic and eastern Pacific, September 17, 1941, 1930 EST.

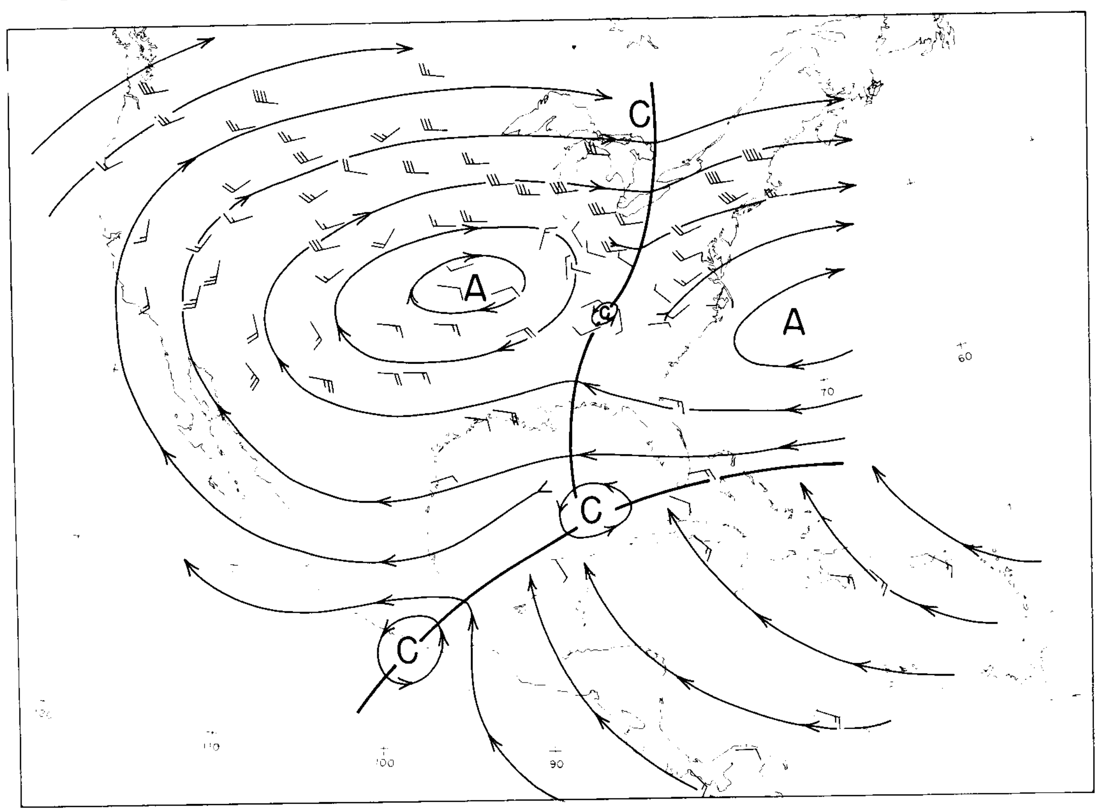

FIG. 12b.--Winds at 10,000 feet for the United States and Caribbean area, September 17, 1941, 0500 EST.

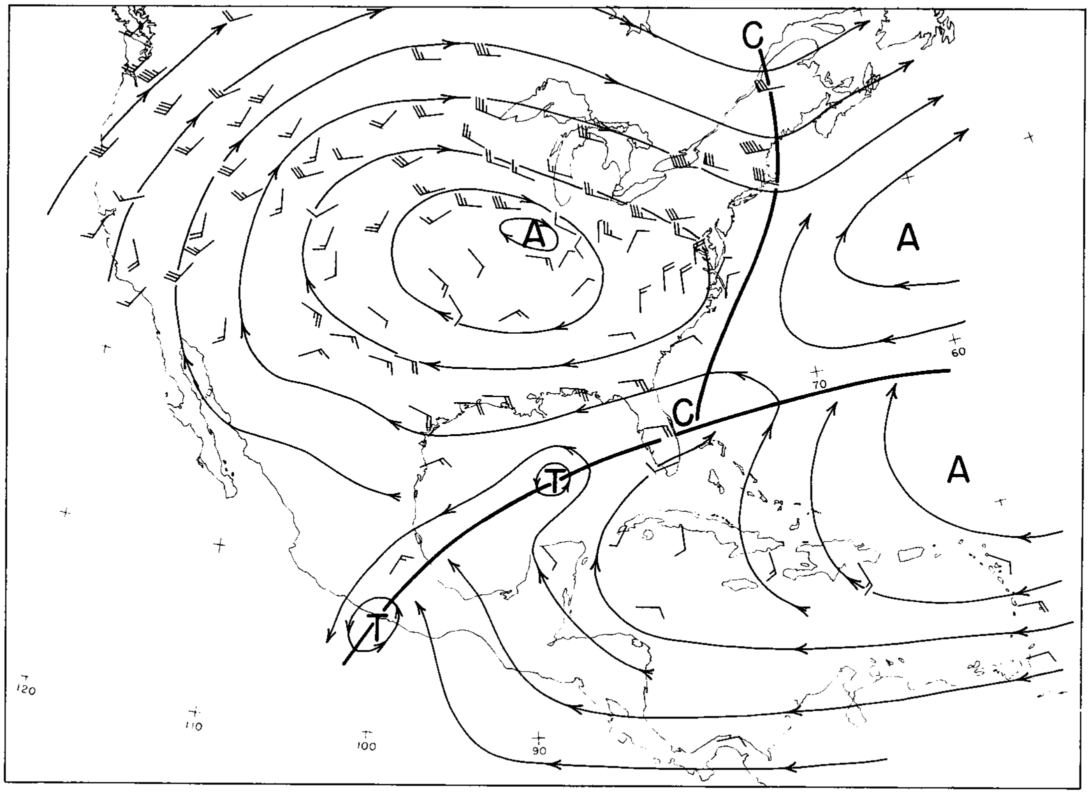

FIG. 12c.--Winds at 10,000 feet for the United States and Caribbean area, September 17, 1941, 1700 EST.

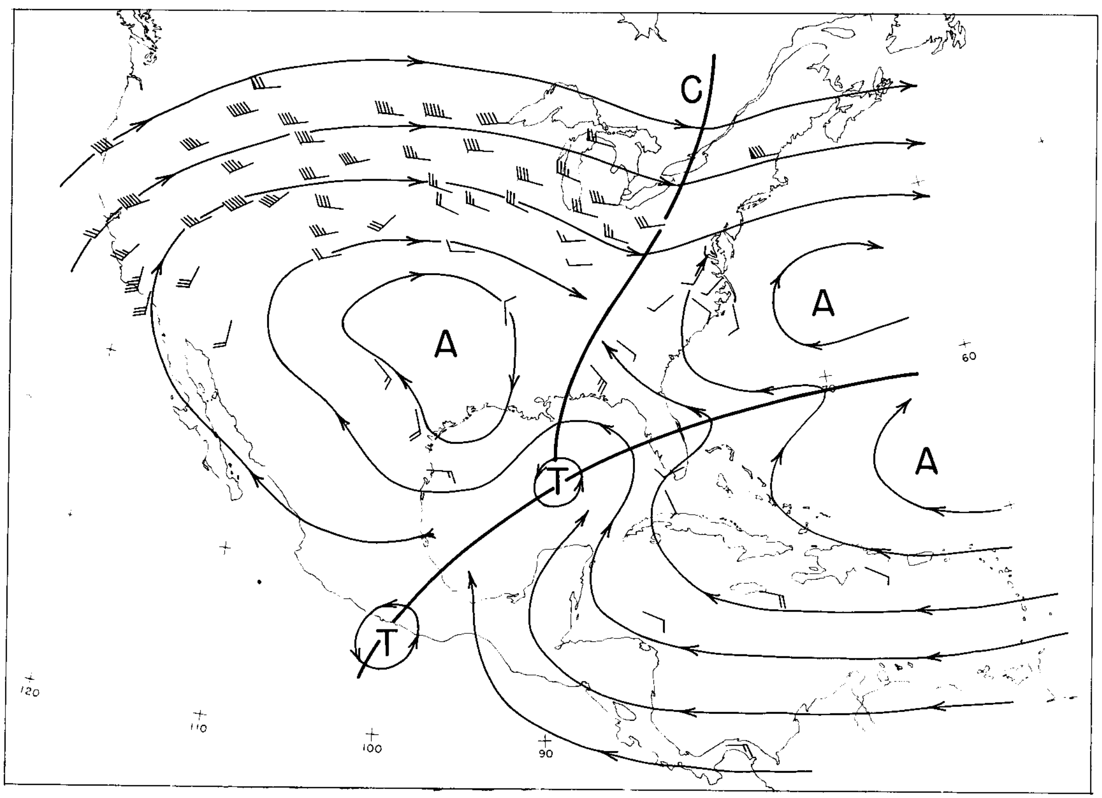

FIG. 12d.--Winds at 20,000 feet for the United States and Caribbean area, September 17, 1941, 1700 EST.

its short wave length. The cyclonic center located over northern Florida in Figure 11c acquired hurricane intensity as it drifted into the Gulf of Mexico on the following day.

The maps of September 17, 1941(Figs. 12a-12d), contain the best instance of almost simultaneous formation of several cyclones in recent years. Three growing centers appear in Figure 12a, all of which reached hurricane strength.

An east-west shearline aloft extended over the cyclogenetic region (Figs. 12b-12d). In the Gulf of Mexico and in the Atlantic, two vortices began to grow as a trough passed eastward across the United States, followed by a large dynamic high. The amplitude of this trough, at first slight, increased rapidly as it approached the East Coast.

7. Effect of Polar Westerlies

If upper baroclinic westerlies overlie the trade-wind region, or if the entire west-wind belt is displaced unusually far south, the subtropical ridgeline aloft is located over the latitudinal band in which hurricanes usually develop. The formation of equatorial shearlines over that area then is not possible. Correspondingly, if the westerlies strengthen after a period of inaction, any shearlines that may have developed or moved over the trade-wind belt are placed in the anticyclonic shearing zone at the southern edge of the polar westerlies. In consequence, the cyclonic shearing zone is destroyed by the effect of the higher latitude circulation, and the equatorial west winds merge with the polar westerlies.

This interaction between low and middle latitudes was of decisive importance in the situation reproduced in Figures 8-10. As noted earlier, a hurricane formed on October 1-2, 1945, near Swan Island on the shearline shown in Figure 9, as the streamline amplitude to the north increased and a large upper-air trough entered the eastern and southeastern United States. Eastward passage of this trough across Bermuda on October 4 (Fig. 10c) activated the shearline in the eastern Caribbean. Widespread bad weather with severe easterly squalls broke out between San Juan and Antigua, B.W.I., and the shearline began to move northward. A closed surface isobar could be drawn in the eastern Caribbean Sea, and at least one weather service began to issue advisories.

As seen from Figure 10a, the Caribbean shearline passed San Juan almost 2 days earlier above 25,000 feet than at the surface. The coldest temperatures in the upper troposphere were observed on October 4. Subsequently pressures and temperatures at the upper levels began to rise (Table 8), while nearer the ground temperature increases coincided with pressure falls that weakened upward. The presence of extensive cloudiness and high relative humidities shows that the warming could not have resulted from subsidence. Therefore, the distribution of pressure and temperature changes with elevation, especially at 2200 EST, October 5, is very similar to that experienced in connection with incipient hurricanes.

Why did this very favorable situation pass without hurricane development? The changes in the broad-scale circulation over the West Atlantic that took place during the first week of October provide an answer. The arrival of the polar trough to the north of the Antilles coincided with the appearance of wintertime westerlies at Bermuda (Fig. 10c). These westerlies, previously light, increased persistently between October 4 and 7 in the higher troposphere, finally attaining 60 mph. Thus the equatorial westerlies to the south of the shearline became continuous with the polar westerlies, while the band of deep easterlies in evidence on October 4 (Fig. 10a) vanished. The subtropical ridgeline was displaced southward over the Caribbean. Under these circumstances the large pressure rises in the high troposphere at San Juan on October 6 must be ascribed to this displacement. It is suggested that the foregoing sequence of events prevented hurricane formation.

The importance of the circulation in the subtropics for the growth of low-latitude vortices is brought out further by the weather conditions prevailing over the western Atlantic during the last 10 days of July, 1944. At the beginning of this period, the winds at Bermuda (Fig. 13a) had only small westerly components that did not increase upward or even decreased. An equatorial shearline was attempting to form over the Caribbean but was not yet well developed. The shear across the line consisted mainly in a drop from strong to light easterlies, and the width of the shearing zone was at least 5 degrees latitude. At that time the system appeared to be better developed in the middle troposphere than at 30,000 feet and higher, where fairly uniform easterlies prevailed.

To the east of the Lesser Antilles, the shearline was probably more developed, as several small vortices traveled from the central equatorial Atlantic westward into the Caribbean. No independent developments took place inside the Caribbean Sea. The two most important vortices passed Sta. Lucia, B.W.I., on July 24 and 27 (Fig. 13b). The first of these centers had a closed circulation near the ground as shown by the Sta. Lucia and

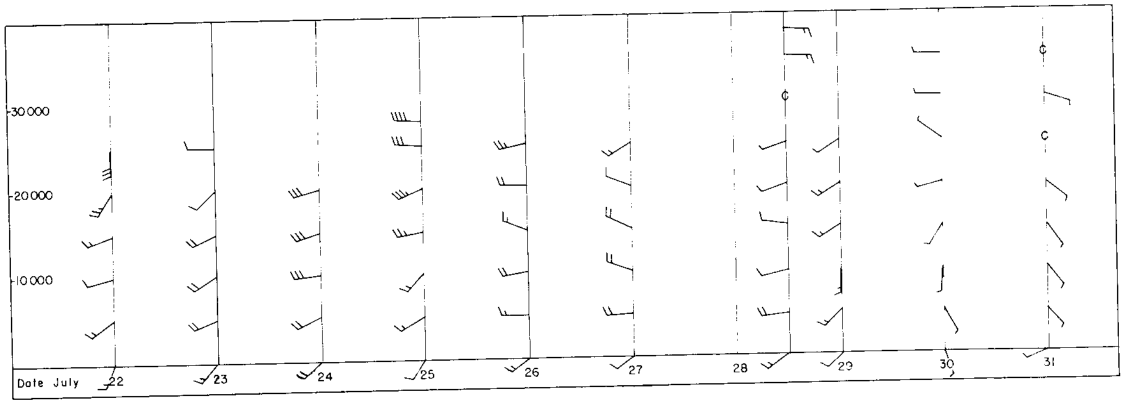

FIG. 13a.--Vertical time-section of the upper winds at Bermuda, July 22-31, 1944. All winds are 1100 EST except July 28, where the wind at 2300 is plotted.

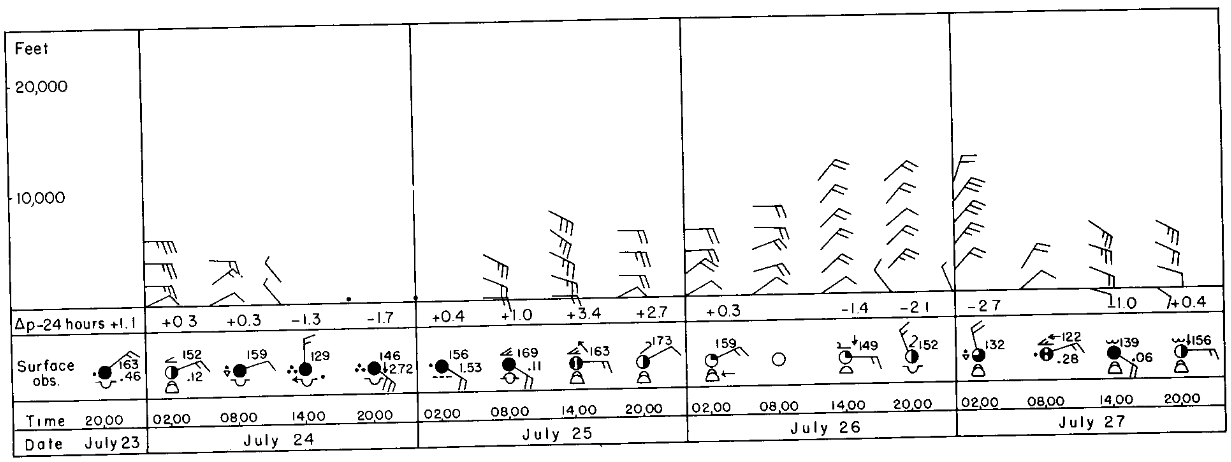

FIG. 13b.--Vertical time-section (EST) at Sta. Lucia, B.W.I., July 23-27, 1944.

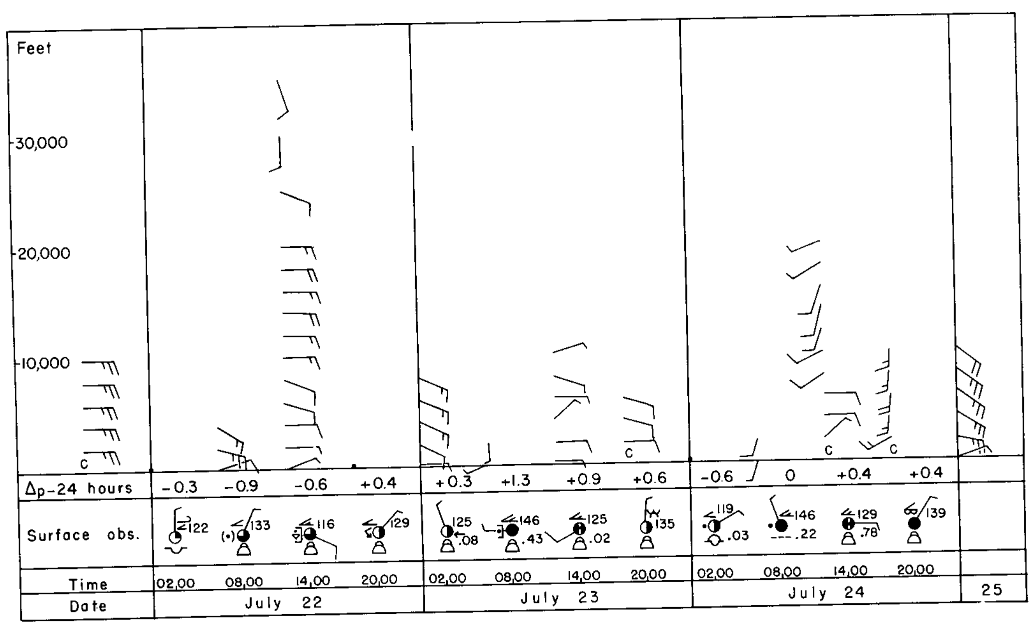

FIG. 13c.--Vertical time-section (EST) at Trinidad, B.W.I., July 22-24, 1944.

TABLE 10

RADIOSONDE OBSERVATIONS AT STA. LUCIA, B.W.I., JULY 23-26, 1944, 2300 EST, AND 24-HOUR CHANGES

	H	p(mb)	T(°C)	Δ p(mb)	ΔT(°C)
July 23, 2300 EST	Sfc.	1015	24	1	-2
	5,000 ft.	851	15	-1	-1
	10,000 ft.	708	6	-2	-3
	15,000 ft.	587	-2	-3	-1
	20,000 ft.	483	-9	-3	-1
	10 km	285	-36	-1	-2
	13 km	181	-55	-2	-1
July 24, 2300 EST	Sfc.	1015	24	0	0
	5,000 ft.	852	16	1	1
	10,000 ft.	710	7	2	1
	15,000 ft.	589	0	2	2
July 25, 2300 EST	Sfc.	1015	27	0	3
	5,000 ft.	853	17	1	1
	10,000 ft.	711	9	1	2
	15,000 ft.	590	0	0	0
	20,000 ft.	486	-9		
	10 km	287	-35		
July 26, 2300 EST	Sfc.	1014	26	-1	-1
	5,000 ft.	852	16	-1	-1
	10,000 ft.	710	8	-1	-1
	15,000 ft.	589	0	-1	0
	20,000 ft.	485	-9	-1	0
	10 km	286	-38	-1	-3

Trinidad time sections (Figs. 13b-13c). It was also attended by heavy precipitation. At the approach of the perturbation, Sta. Lucia recorded pressure and temperature falls that increased from the surface to 10,000 feet and then remained nearly steady with height (Table 10). Corresponding

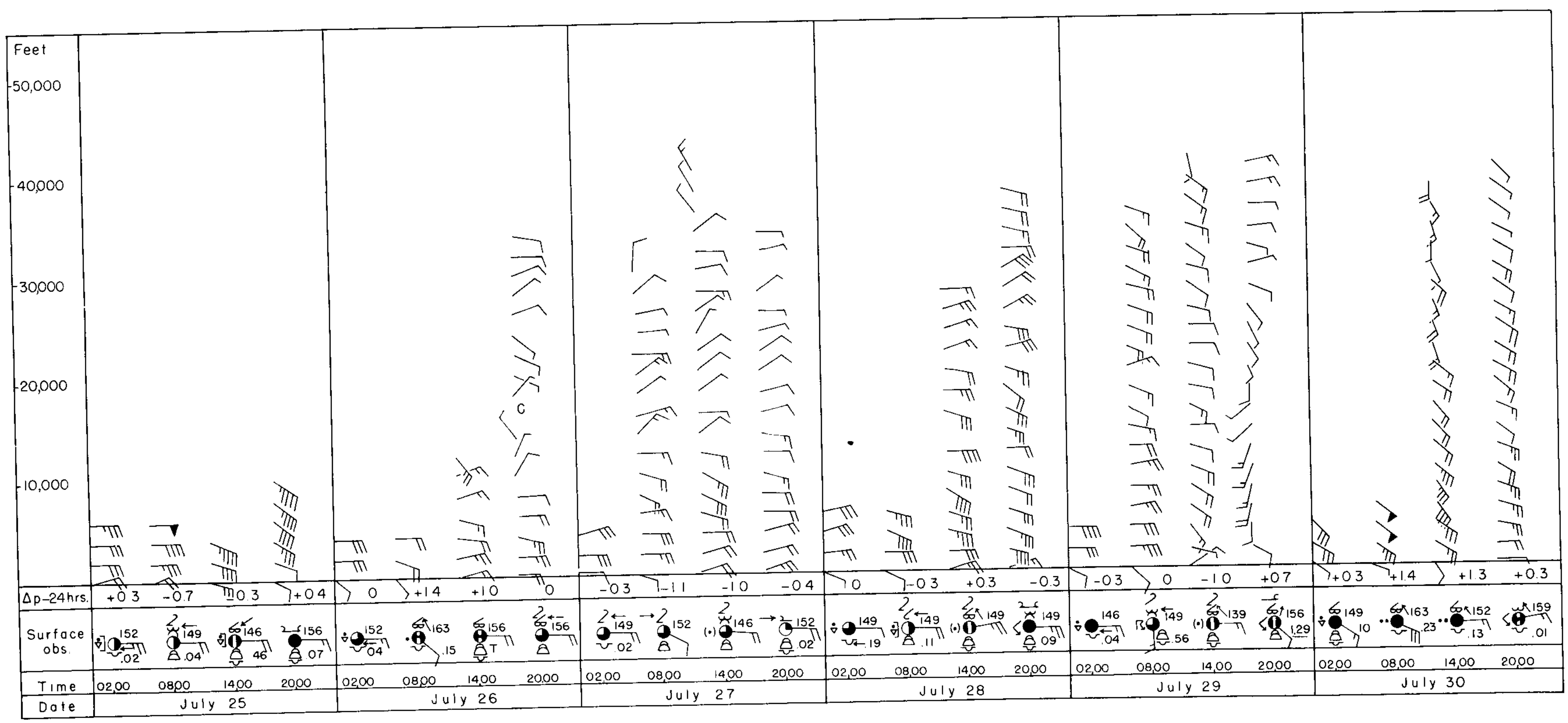

FIG. 13d.--Vertical time-section (EST) at San Juan, Puerto Rico, July 25-30, 1944.

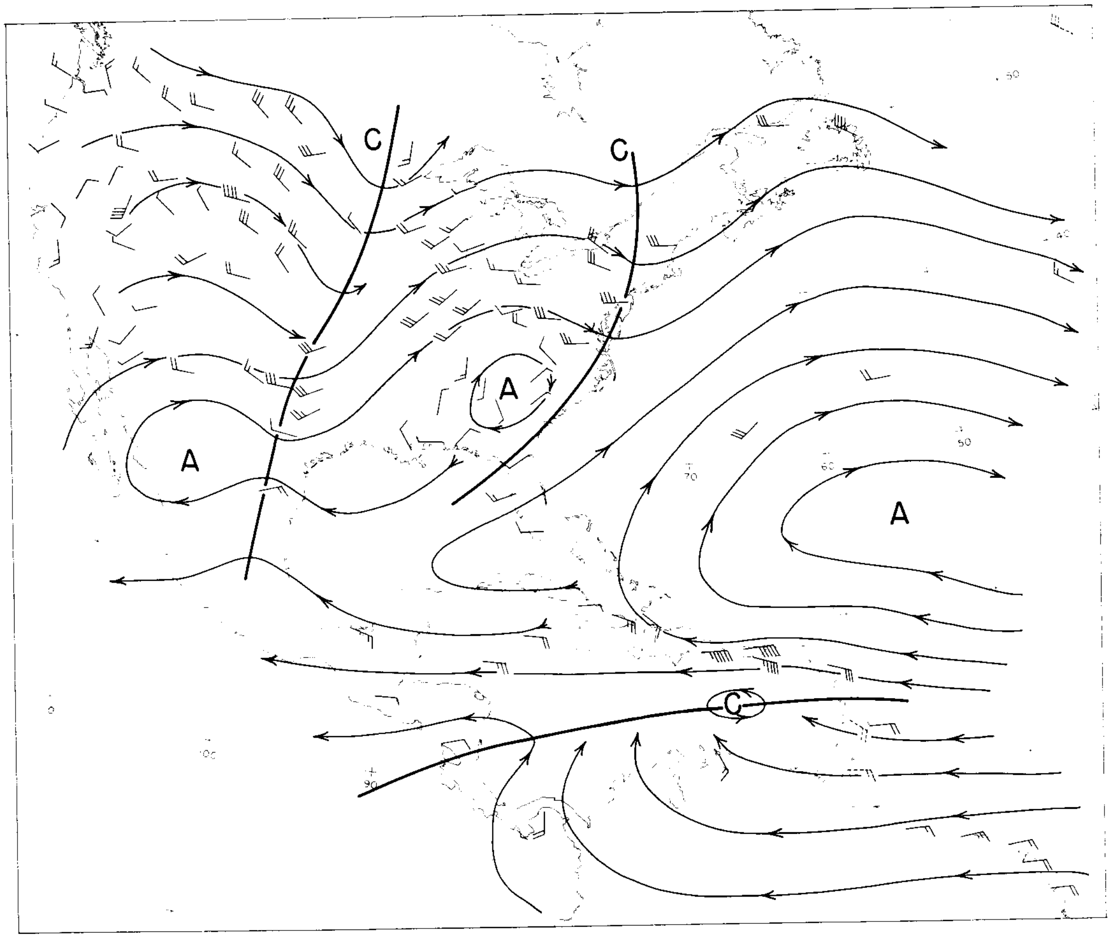

FIG. 13e.--Winds at 10,000 feet for the United States and Caribbean area, July 25, 1944, 1700 EST.

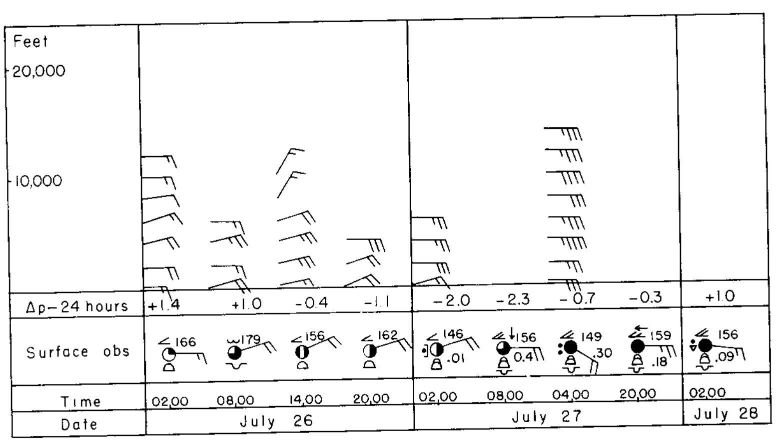

FIG. 13f.--Vertical time-section (EST) at Antigua, B.W.I., July 26-28, 1944.

TABLE 11

RADIOSONDE OBSERVATIONS AT ANTIGUA, B.W.I., JULY 23-27, 1944, 2300 EST, AND 24-HOUR CHANGES

	H	p(mb)	T(°C)	Δ p(mb)	ΔT(°C)
July 23, 2300 EST	Sfc.	1013	26	0	-1
	5,000 ft.	853	18	0	2
	10,000 ft.	711	9	-1	-2
	15,000 ft.	590	-1	-1	0
	20,000 ft.	486	-9	-1	-1
	10 km	286	-38	-1	-2
July 24, 2300 EST	Sfc.	1013	26	0	0
	5,000 ft.	853	18	0	0
	10,000 ft.	712	11	1	2
	15,000 ft.	590	1	0	2
	20,000 ft.	488	-7	2	2
	10 km	288	-34	2	4
	13 km	183	-58		
July 25, 2300 EST	Sfc.	1015	27	2	1
	5,000 ft.	855	17	2	-1
	10,000 ft.	713	10	1	-1
	15,000 ft.	590	0	0	-1
	20,000 ft.	488	-8	0	-1
	10 km	287	-36	-1	-2
July 26, 2300 EST	Sfc.	1013	27	-2	0
	5,000 ft.	852	15	-3	-2
	10,000 ft.	711	9	-2	-1
	15,000 ft.	590	-1	-2	-1
	20,000 ft.	485	-10	-3	-2
	10 km	285	-37	-2	-1
	13 km	180	-61		
July 27, 2300 EST	Sfc.	1013	27	0	0
	5,000 ft.	852	16	0	1
	10,000 ft.	711	9	0	0
	15,000 ft.	590	0	0	1
	20,000 ft.	486	-10	1	0
	10 km	287	-35	2	2
	13 km	181	-60	1	1

pressure rises with warming followed the disturbance. The center passed into the Caribbean near Sta. Lucia during the early afternoon of July 24, so that a measurement at the time of passage was not available. Nevertheless, the data of Table 10 permit the conclusion that the pressure and temperature variations were not those representative of incipient hurricanes and that low-level compensation for the upper pressure falls was taking place. Note also that the 24-hour surface pressure falls lasted for only 12 hours at Sta. Lucia (Fig. 13b).

As the disturbance entered the Caribbean, the westerlies at Bermuda intensified and began to increase with height. Although the balloon runs did not reach the high troposphere, at least one west wind of 40 mph (at 28,000 feet) was recorded. The increased westerly circulation seemed not only to prevent the further growth of the disturbance that was moving westward from the Lesser Antilles but apparently damped it out. At San Juan (Fig. 13d), passage of the disturbance on July 25 could be inferred only by slight 24-hour surface pressure falls, a freshening of the easterly wind, and some deterioration of the weather. The upper pressure and temperature variations (Table 3), however, closely corresponded to those of Antigua and Sta. Lucia one day previously. On the 10,000-foot chart (Fig. 13e) a closed cyclonic center has still been entered. Its existence even at this time, however, was doubtful, and it was not heard of thereafter.

The second vortex arrived at Sta. Lucia on July 27 with somewhat greater surface pressure falls than the first depression, but without any evidence of westerly winds over the Lesser Antilles. At Antigua (Fig. 13f) the surface pressure falls also were fairly well pronounced, and disturbed weather that extended well to the east of the troughline attended the system. The pressure and temperature changes aloft were similar to those connected with the perturbation of July 24 (Tables 10 and 11). It might have been suspected that the future history of this disturbance would take the same course as that of its predecessor.

The circulation in the belt of westerlies, as indicated by the Bermuda rawin (Fig. 13a), was weakening again at this time. High-level winds became light and variable, and east components appeared on some observations after July 29. On July 31 the upper anticyclone appeared to be centered over Bermuda as indicated by very light and variable winds to the top of the balloon runs. Thus the disturbance over the Lesser Antilles did not vanish but continued its motion toward the west-northwest, with increasing intensity aloft. Its passage at San Juan was well marked, and the upper winds over Puerto Rico on July 29 suggested the presence of a small cyclonic center near 15,000 feet. The pressure and temperature falls at San Juan, however, still continued to increase with height, opposite to the variations characteristic of tropical storms. The disturbance was still a cold-core center, and the surface pressure variations were slight, even smaller than over the Lesser Antilles, in spite of increased intensity of the system aloft. Weather experienced over Puerto Rico was not that associated with waves in the easterlies. Intermittent precipitation took place throughout the period. At the approach of the perturbation from the east, the bad weather intensified.

Coincident with weakening of the westerly circulation over the western Atlantic, the trade freshened along a line that moved from east to west and reached San Juan on July 30, a day after the disturbance. As the upper cyclonic center continued to advance west-northwestward toward the Bahamas, accompanied by this surgeline, it became more and more pronounced. When it finally came under the influence of an extratropical trough north of the Bahamas, surface deepening began and the vortex attained hurricane strength on August 2, 1944.

The Situation, July 22-23, 1943.--As a final example, this series shows details of the wind field and the pressure and temperature distribution during the formation of a subtropical cyclonic vortex which originated within the network of observing stations over the United States. The series therefore is not entirely representative of tropical conditions. The center formed over land and was generated along a shearline which previously had been a front and which had lost its frontal characteristics only a brief time prior to the cyclonic development. Nevertheless, the series brings out some of the features of nonfrontal formation of low-pressure centers which have been discussed in the foregoing and which are difficult to illustrate over the tropical oceans because of scarcity of reports.

A large extended trough had been stagnating near the United States East Coast for several days before the beginning of the series. On July 22 a cold front was moving southward from Canada in the northerly current west of the extended trough. This front could be located distinctly in the wind field at 4,000 feet (Fig. 14b). It was also marked by a trough of low pressure at 1,500 m (Fig. 14a), and a closed upper isobar could be drawn around the area where the front joined the extended trough. The temperature field at 1,500 m was less favorable for frontal interpretation of the system. Although low temperatures were reported in the vicinity of the front, the temperature gradient was located mainly to its south.

While the front continued to advance gradually southward on July 23, the weather situation in the southeastern United States changed radically during the afternoon of that day. Pronounced convergence developed along the southern part of the extended trough. At the same time this trough started to move westward across Florida, and extreme convection ensued over the whole state of Florida, well after the hour of diurnal thunderstorm maximum. The surface map of July 23, 1930 EST (Fig. 15b), indicated the presence of a weak low-pressure center over the peninsula in the middle of the thunderstorm area. The principal region of low pressure was situated in Georgia and Alabama, representing the remainder of the cold front that had traveled southward from Canada. An air-mass contrast across the trough at the surface was no longer noticeable. The troughline, however, formed the boundary between clear skies to its north and the extreme convection area that had just developed to its south. At 2,000 feet (Fig. 15a) a well-marked shearline was still in evidence at that time.

Figure 16a shows that the good weather north of the troughline was due to subsidence in the southward-moving polar air mass as might be expected. The figure indicates the vertical stretching and shrinking experienced by three air columns during July 23. These columns were followed as closely as possible by the trajectory method, and individual air parcels were identified by their potential temperature. In view of the clear skies and the subsiding motion indicated by Figure 15b, the identification should have been reliable except possibly on the trajectory PT-PPL. The air moving on this trajectory entered the area of severe convection as it approached the Gulf coast.

All three trajectories started well to the rear of the cold front on July 22, 2300 EST. The downward vertical motion shows that sinking and horizontal divergence of the polar air mass near the ground took place. It is of interest to note that the trajectory PT-PPL which passed through the low-pressure trough, was exposed to considerable convergence in spite of the downward displacement of the lower part of this column.

The occurrence of downward vertical motion north of the troughline can be inferred also by noting that at the 1,500-m level (Fig. 16b) warming took place on July 23 in the northerly current relative to the troughline while the sky was clear. In contrast to the previous evening the warmest air was located mainly to the north and west of the troughline on July 23, 2300 EST. Cold temperatures prevailed along the East Coast, in conjunction with the extended trough. Thus the isotherms at 1,500 m were oriented normal rather than parallel to the troughline, and any previous cold-front characteristics of the system had disappeared.

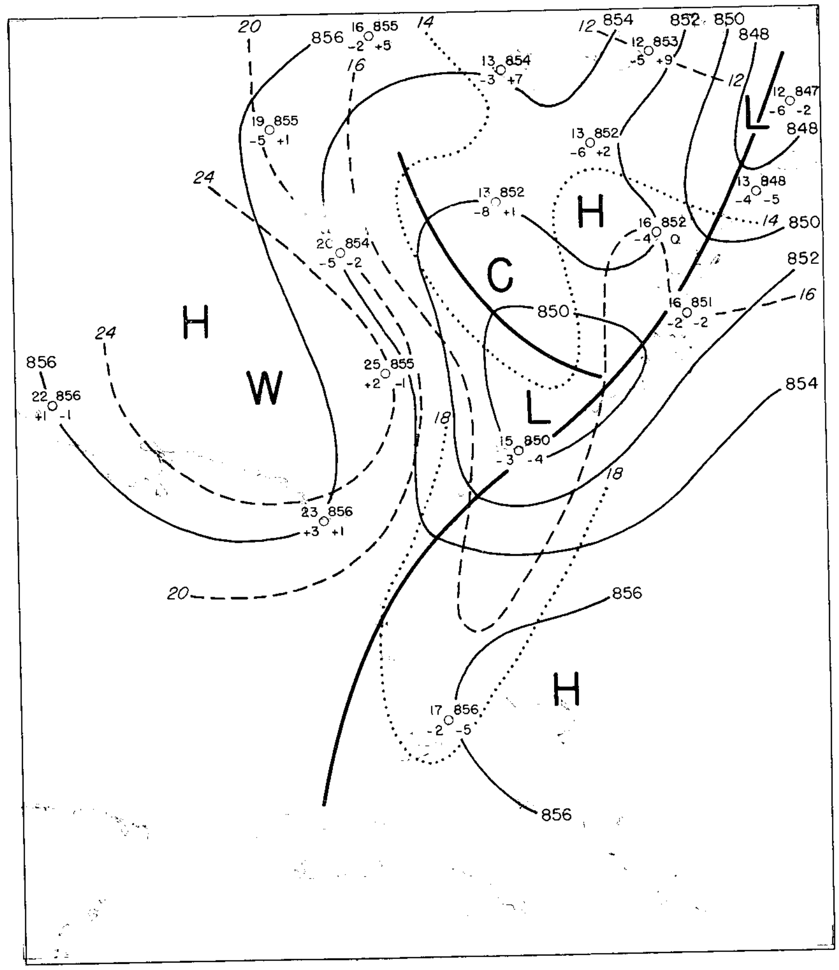

FIG. 14a.--Pressures and temperatures at 1500 m for the southeastern United States, July 22, 1943, 2300 EST. Solid lines are isobars and dashed lines isotherms. W stands for warm and C for cold.

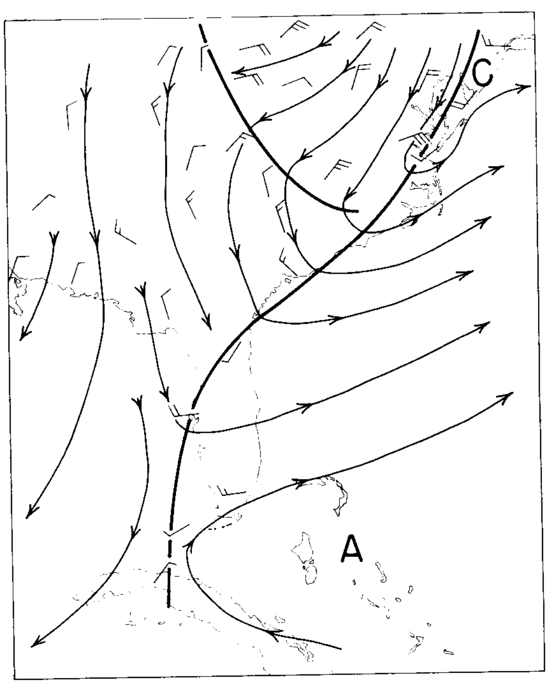

FIG. 14b.--Winds at 4,000 feet for the southeastern United States, July 22, 1943, 2300 EST.

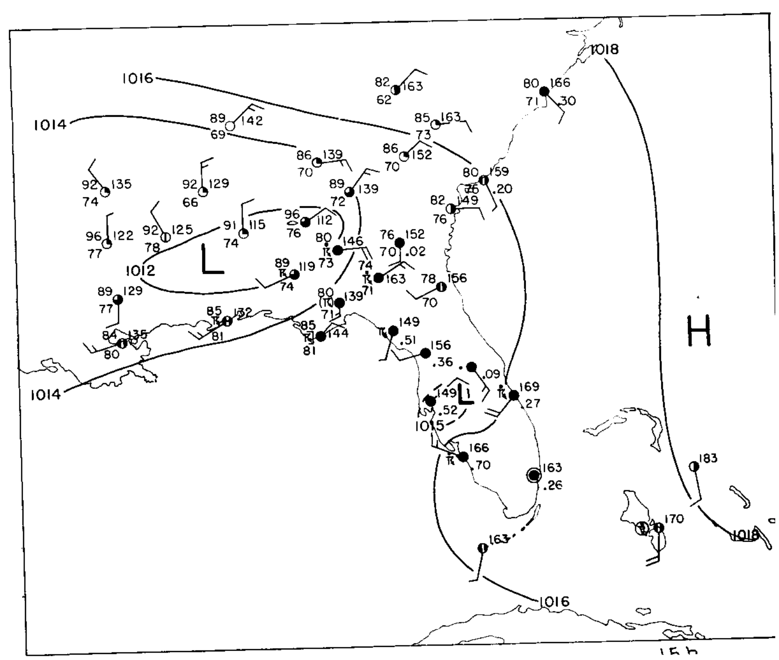

FIG. 15b.--Winds at 2,000 feet for the southeastern United States, July 23, 1943, 1700 EST.

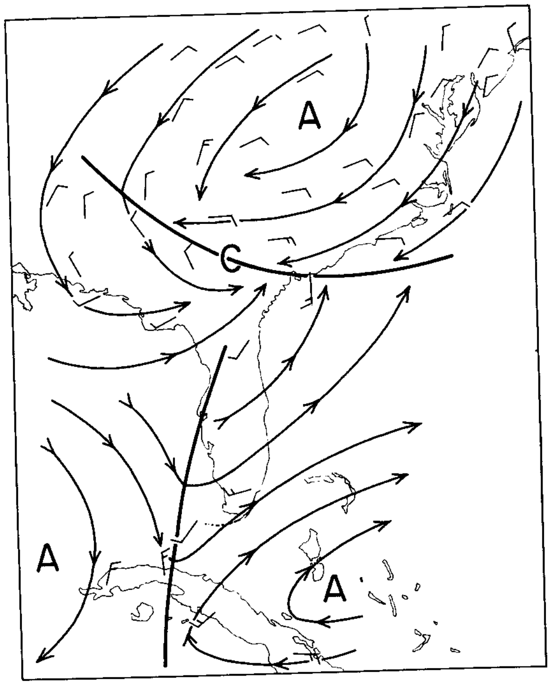

FIG. 15a.--Surface chart for the southeastern United States, July 23, 1943, 1930 EST.

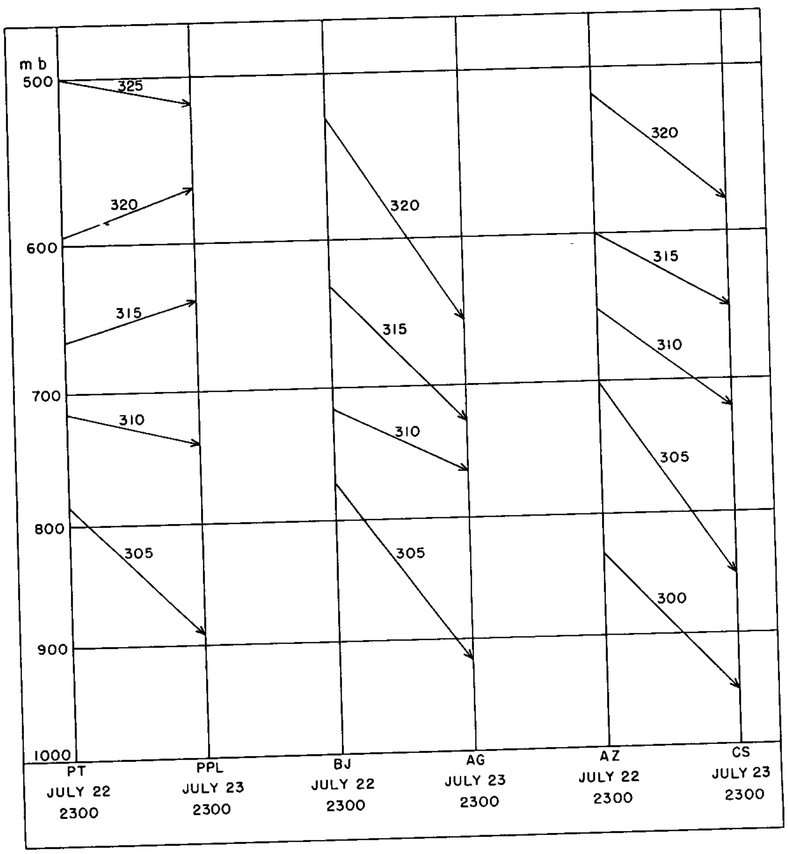

FIG. 16a.--Pressure variation at selected isentropic levels on approximate air trajectories, July 22-23, 1943: Pittsburgh, Pa., to Appalachicola, Fla. (PT - PPL); Buffalo, N.Y., to Atlanta, Ga. (BJ - AG); Albany, N.Y., to Charleston, S.C. (AZ - CS).

As shown by the foregoing, the observed temperature field on July 23, 2300 EST, was dynamically and not advectively produced. At that time, therefore, the system was no longer a cold front. It had become a non-frontal shearline. Subsidence to its north and ascending motion to its south brought about the observed reversal of the temperature field from the preceding day.

Coincident with this reversal the vertical slope of the shearline also changed during July 23 as seen from Figure 16c. This figure represents a time-section of the upper winds approximately 200 miles north and south of the surface position of the shearline. At first, the shearline sloped northward with height. Later, as the temperature field changed, the slope straightened, and finally the shearline sloped upward toward the south. Coincident with this reversal the initial increase in the west component of the wind with height vanished, and at the end of the period there was a decrease of this component with height in the lowest levels. Thus the vertical shear of zonal wind and the north-south gradient of vertical velocity were at least in the correct sense for the generation of cyclonic vorticity. As may be inferred from Figure 15a, the entire gradient of vertical velocity must have been concentrated in a very narrow zone near the troughline.

The downward motion north of the troughline may have lasted all day, although it was presumably intensified as the line reached the lee side of the Appalachian Mountains. But the strong upward motion to its south developed only during the late afternoon of July 23. At 1330 EST violent thunderstorms were not yet reported, and the 1330 EST map (not reproduced) had the appearance of a surface chart on an ordinary summer afternoon.

As the contrast of weather across the shearline intensified, so that it finally became almost discontinuous, rapid vortex development ensued. Whereas Figure 15b showed only the presence of a well-marked zone of cyclonic wind shear at 2,000 feet, a pronounced vortex was in evidence at that level 6 hours later (Fig. 17a). The wind distribution at that time was perfect for a tropical cyclonic center, with the strongest winds located to the north and east of the center of rotation.

The vortex decreased upward in strength and consequently must have been a warm low-pressure center. This is shown very clearly by Figures 17b-17e. Already at 4,000 feet (Fig. 17b) the rotational circulation was less than at 2,000 feet; at 6,000 feet (Fig. 17c) it had vanished. At 10,000 feet (Fig. 17e) even the east-west troughline had disappeared. While the vortex

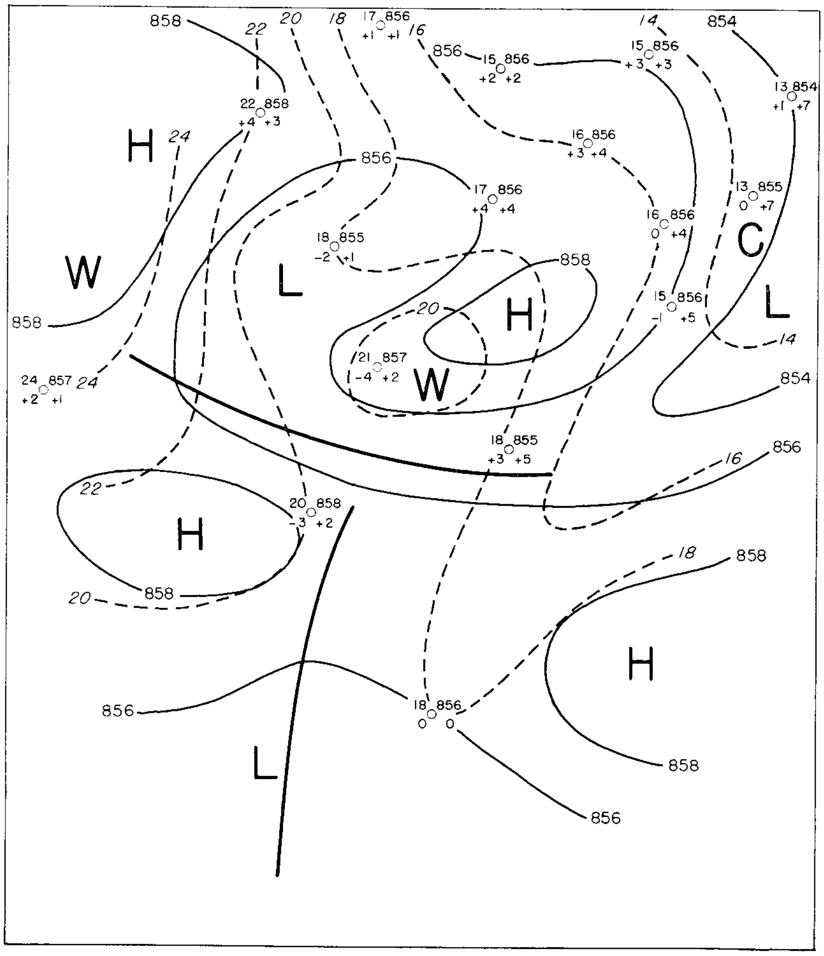

FIG. 16b.--Pressures and temperatures at 1,500 m for the southeastern United States, July 23, 1943, 2300 EST. Lines and notation as in Fig. 14a.

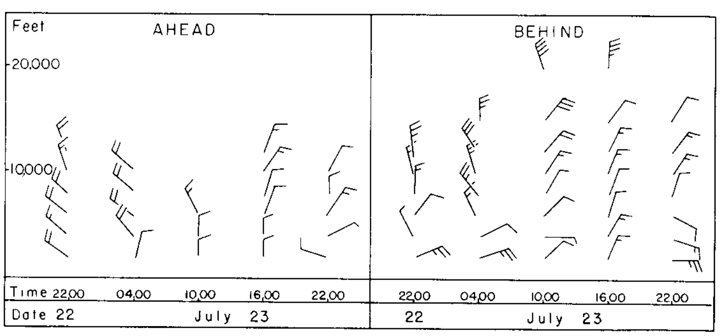

FIG. 16c.--Vertical time-section of the upper winds 200 miles ahead and behind surface position of shearline moving southward in the southeastern United States, July 22-23. 1943.

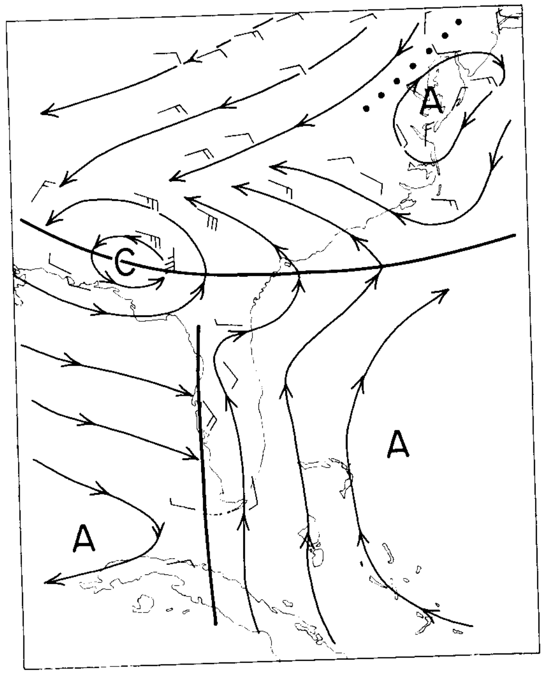

FIG. 17a.--Winds at 2,000 feet for the southeastern United States, July 23, 1943, 2300 EST.

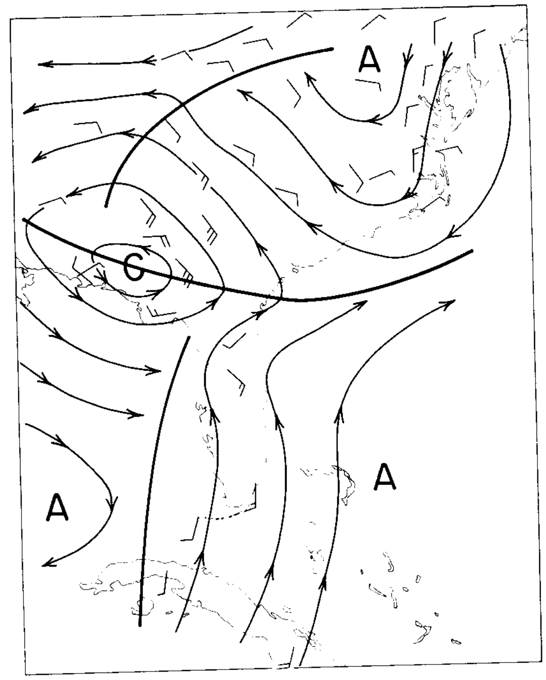

FIG. 17b.--Winds at 4,000 feet for the southeastern United States, July 23, 1943, 2300 EST.

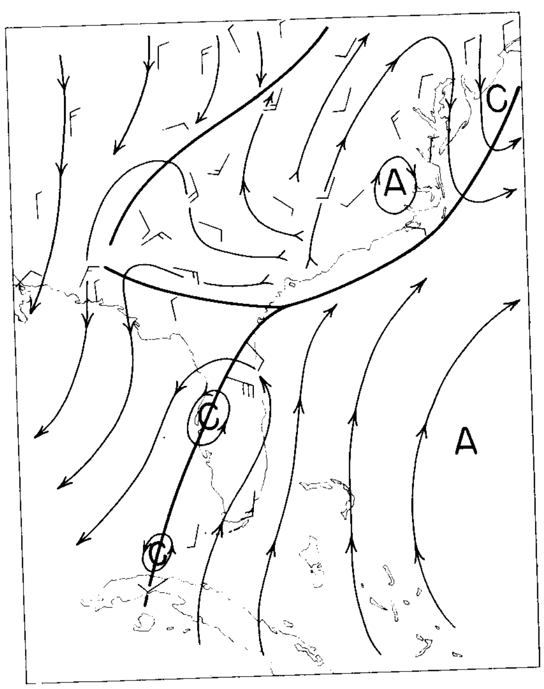

FIG. 17c.--Winds at 6,000 feet for the southeastern United States, July 23, 1943, 2300 EST.

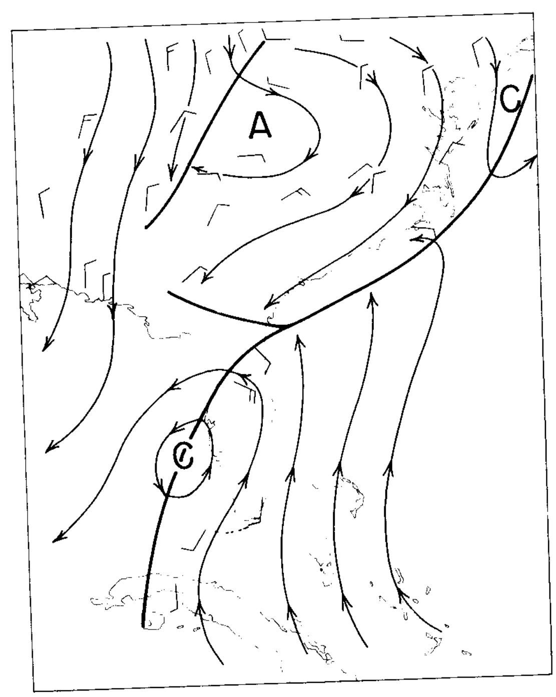

FIG. 17d.--Winds at 8,000 feet for the southeastern United States, July 23, 1943, 2300 EST.

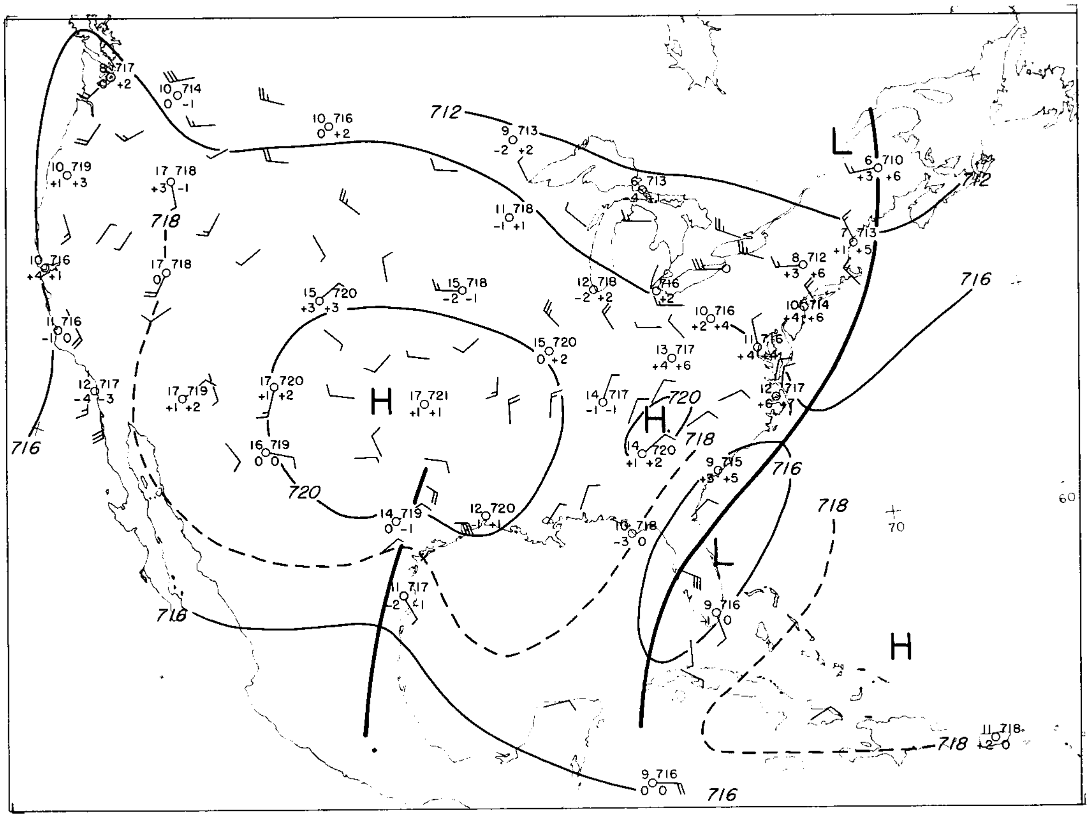

FIG. 17e--Pressures and winds at 10,000 feet for the United States, July 23, 1943, 2300 EST.

north of the Gulf coast rapidly weakened with height, the cyclonic circulation located over the central part of the Florida peninsula strengthened upward and became the dominant feature of the flow pattern in the southeastern United States above 6,000 feet. This increase of intensity with height suggests that this system was a cold-core vortex, in spite of its location in the middle of a zone of extreme convection. A comparison between both vortices emphasizes once again that it is often not the conditions within an area of violet convection but those at its boundary that are most favorable for hurricane formation.