A Case Study of an Easterly Jet Stream in the Tropics

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Abstract

An easterly current at 200 mb over the Southern Bahamas and Florida is identified as a jet stream which had originated in middle latitudes and had travelled along the eastern periphery of the dynamic anticyclone situated over the South Eastern United States. The properties of this current are studied and compared with those associated with middle latitudes jet streams. It is found that although this current is comparatively weak (core velocity about 70 kts), horizontal and vertical wind shears near its axis are comparable with those observed near the usually much stronger middle latitudes jet streams.

The mechanism responsible for the concentration of the kinetic energy in this narrow current is shown to be independent of any temperature contrast between two air masses in the lower and middle troposphere. The jet is further shown to be maintained by pressure heads and the continuous production and destruction of kinetic energy involved in its maintenance and motion occur as a result of accelerations in more or less laminar flow rather than by lateral or frictional stresses. Finally, the observed weather distribution is found to be influenced by the resonance and interference of the upper jet stream with waves in the lower easterlies.

1. Introduction

The concentration of kinetic energy in the atmosphere into narrow jets has increasingly engaged the interest of meteorologists in recent years in view of the frequency and wide geographical range of their occurrence, and the important role they play in determining weather conditions. But in spite of the impressive list of studies and some ingenious attempts[1] to account for them, the last word has not been said concerning the nature of the concentrating mechanism. In fact, "the development of an adequate theory for the concentration of momentum into narrow streaks looms as a problem of major importance, perhaps the most important unsolved problem in planetary hydrodynamics" (ROSSBY, 1953).

In order to design a proper framework within which to fit such a theory, the essential features of these currents must be separated from the extraneous or incidental. A good way of weeding out the latter is afforded by comparing different conditions under which wind concentrations occur. The numerous studies of jet streams in various localities and seasons in the middle and high latitudes provide one means for such a comparison. They show, for instance, that while some jet streams seem to be tied with the surface temperature gradient, others, particularly in summer, are limited to the high troposphere. On the other hand, all jets studied to date show near-zero values of absolute vorticity to the right of the jet axis, looking downstream.

The recent introduction of "synoptic" methods of observation in oceanography (FUGLISTER, 1951) which has brought out with clarity the remarkable similarities in the patterns of current concentration in atmosphere and ocean also helps in narrowing down the search for the concentrating mechanism within the right

[1] For a survey of these attempts, the reader is referred to RIEHL, ALAKA, JORDAN and RENARD (1952), Chap. VII.

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framework by relegating to secondary importance the compressibility of the atmosphere and the mode of energy infusion (RIEHL, ALAKA, JORDAN and RENARD, 1954).

Still another method for sifting the essential from the extraneous is provided by controlled experiments in the laboratory. For instance, the importance of dynamic stability generated by the earth's rotation is brought out by the fact that jets injected in a non-rotating fluid sheet tend to mix rapidly with the environment, forming widening wake streams, whereas in rotating fluids, narrow jets appear capable of retaining their concentration for appreciable periods of time (FULTZ, 1951).

The above studies represent different links in the chain of research on jet streams. Another link would be provided by descriptions and analyses of jet streams in low latitudes, especially easterly jet streams, and by comparing their structure and life cycle with that of the extratropical currents. In contrast with middle and high latitudes, however, where the observational network in some areas permits not only locating jet streams with some precision, but also studying in detail their structural properties, life cycle, geographical distribution, and movement, our knowledge of jet streams in tropical regions is fragmentary. Upper air observations are so thinly distributed over large areas that a reliable analysis of the wind field in the upper troposphere is in general all but impossible. Consequently, our information concerning the very existence and geographical incidence of jet streams in the tropics is based in large measure on isolated observations or mean time sections which are individually inconclusive.

In the last few years, however, the amount of raob and rawin observations in the region south of the United States has been increased sufficiently so that this region now contains what may be termed the nearest approximation to a *network* of high-tropospheric reports in lower latitudes. Therefore, in the course of an experiment in tropical analysis and forecasting conducted in Miami during the summer of 1953[1], close watch was kept for indications of easterly jet occurrence south of latitude 30°. Evidence was soon forthcoming; the present paper deals with the best-documented case encountered.

2. Previous observations of low latitude jet streams

In spite of scarce data, several types of jet streams in low latitudes have received brief mention in the literature:

a. *Westerly jets originating at high latitudes and migrating deep into tropical regions, usually in accompaniment with the polar front:* These are chiefly winter phenomena and may be illustrated by the case studied by CRESSMAN (1950) when a westerly jet centered at latitude 40° N on March 20, 1948, drifted southward to latitude 20° N by March 25.

b. *Westerly jets originating in the tropics:* An example of these is the case noted by RIEHL (1954) when a jet stream with a maximum speed over 100 knots passed near Johnston Island (lat. 17° N) on July 24, 1951 and over Hawaii the following day. There was definite indication that it could not have originated in extratropical regions and must have developed in the tropics. As another example, we may cite the case studied by COLÓN (1954), of the high winds over Puerto Rico during the period March 29 – April 2, 1953.

c. *The subtropical westerly jet* (PALMÉN, 1951) which occurs in general during the winter months and which helps account for the fact that the average northern hemispheric westerlies during winter show a maximum around latitude 25° – 30° N (PETTERSSEN, 1950).

d. *Easterly jets originating in the tropics:* There is an abundance of reports of high-tropospheric easterly winds in excess of 50 knots in different localities in the tropics and a tendency by some meteorologists to call these winds jet streams without simultaneous reference to winds in neighboring localities.[2] Clearly, while jet streams are characterized by relatively strong winds, it is fallacy to consider all strong winds as forming part of a jet. Therefore, isolated observations of high winds do not necessarily indicate the existence of jets. And whereas it is undoubtedly our experience in the higher

[1] The work was undertaken under contract with the Office of Naval Research and was conducted under the leadership of Professor H. Riehl.

[2] For a list of articles on this subject the reader is referred to "Meteorological Abstracts and Bibliography": American Meteor. Soc., July, 1953.

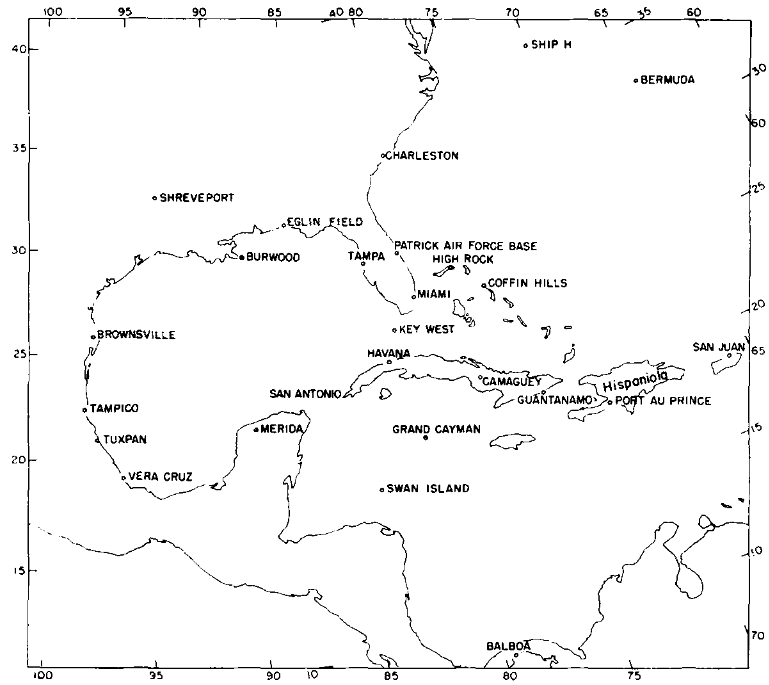

Fig. 1. Map showing some of the stations mentioned in the present study.

latitudes that very strong winds do not usually prevail over very wide belts, thus strengthening the probability that the observed high winds form part of a pattern characterized by a lateral concentration of kinetic energy, this still remains to be shown.

e. *Easterly jets originating in middle latitudes:* Occasionally, during southward motion, a jet stream of the temperate zone splits and one branch travels along the periphery of a dynamic anticyclone as a northerly or northeasterly current. NEWTON and CARSON (1953) have investigated such a case in summer in connexion with the formation of shear lines. Under favorable conditions, such a splinter current may be expected to appear at the equatorward periphery of the subtropical high as an easterly jet.

3. Scope of study

The case to be presented deals with a situation of the last class described. First intimation of its occurrence came when High Rock, Grand Bahama Island (Fig. 1) reported a 200 mb wind speed of 53 knots from due East on July 31, 1953 at 1500 GCT. Twelve hours later, a wind of 52 knots from the same direction appeared at Miami. By August 2, the wind speed had reached 62 knots at Coffin Hills, Eleuthera Island, and increased to 60 knots at Miami, and to 62 knots at High Rock. Wind directions were bracketed between 80° – 110°. Neither Havana nor Patrick AFB partook in the wind variations to any marked degree. A time section of the wind speeds at nearby stations, projected on the 82° W meridian (Fig. 2), suggested that the above sequence of events was concomitant with the westward advance of a *narrow current similar to jet streams* in lateral concentration of kinetic energy, albeit with less intense maximum velocities than is ordinarily observed in jet stream cores.

Due to lack of detailed and reliable data, none of the jet streams mentioned in the literature could be studied in a manner even remotely comparable to that possible in the middle latitudes. The present paper is a first attempt to make such a study. The network of upper air observations allows a reasonably adequate three-dimensional analysis. Our first objective will be to describe and examine the

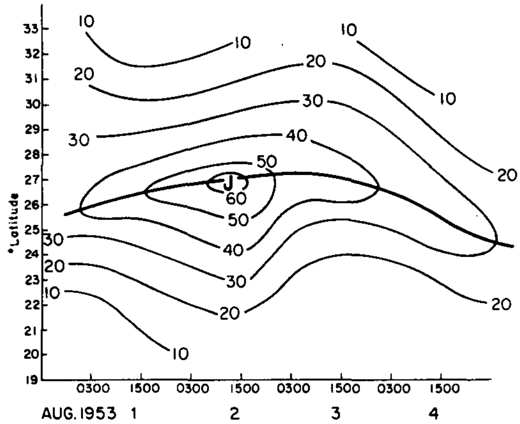

Fig. 2. A time section of the wind speed in knots, at 200 mb along the 82 W meridian, during the period August 1—4, 1953.

thermal and wind structure of the current. Next, an attempt will be made to determine, as far as possible, the mechanism by which the energy of the current is created and destroyed. Finally, the role of the vorticity changes in bringing about the observed weather will be discussed.

4. Analysis

In order to obtain as complete a picture as possible, the following charts and diagrams were prepared for the period July 31 – August 4, 1957:

a) *Time sections* at the stations shown in figure 1. Figures 3 a – c represent a selection of these sections.
b) *Contour maps* at 500 and 200 mb (24-hr continuity); these are not reproduced here.
c) 500 – 200 mb thickness maps (24-hour continuity).
d) *Streamlines and isotachs* at 200 mb (12-hour continuity). The 0300 GCT maps are shown in Figures 4 a – d. Streamlines were drawn from isogons.
e) *Space cross sections* of temperature and height deviations from Schacht's mean tropical atmosphere and of the easterly component of the observed wind at 82° W (24-hour continuity, Figures 5 a – h).
f) *Charts showing distribution of weather and cloudiness* (Figure 6).
g. *Relative vorticity:* We denote the vertical component of the absolute vorticity[1] by ζ_a and write

$$\zeta_a = f + \zeta = f + \frac{V}{r_s} + \frac{\partial V}{\partial n} \qquad (1)$$

where ζ is the vertical component of the relative vorticity,[2] r_s the radius or curvature of the streamlines taken positive for cyclonic circulation and $\partial V/\partial n$ the wind shear normal to the current considered positive when cyclonic.

In most representations of the horizontal wind distribution across jet maxima, whether drawn from observations (cf. Fig. 10) or from theoretical considerations, the cyclonic and anticyclonic limbs of the profile meet in a cusp at the jet axis. This, if correct, implies that the shear term $\partial V/\partial n$ in equation (1) is discontinuous at the jet axis. The 200 mb relative vorticity fields presented in Figures 7 a – d were computed on this assumption. In all cases, the jet axis marks a discontinuity in the relative vorticity.

5. Synoptic Situation

a. *The broadscale setting:* Throughout the period to be discussed, one long wave trough in the westerlies persists in the East Pacific Ocean as shown by 700-mb 5-day mean charts of the Extended Forecast Section, U.S. Weather Bureau; another trough lies near 60° W and the intermediate ridge occupies the Eastern United States. Over the continent the strongest 700-mb westerlies are situated above 40° N everywhere. They pass over the top of the mean ridge between 45° and 50° N. Since, in addition, the limit of the mid-tropospheric easterlies lies north of 30° N, the region on which our interest centers is as well shielded from the influence and interference of extra-tropical weather systems as ever happens. This permits the fullest development of disturbances and patterns of weather evolution that are typical of the tropics in midsummer.

In the lower troposphere a train of waves travels westwards in the easterlies south of 25° – 30° N during the period. Events in the high troposphere are partly connected with the lower waves, and partly independent of them. Thus the jet stream to be demonstrated for the

[1] Hereafter simply called the absolute vorticity.
[2] Hereafter simply called the relative vorticity.

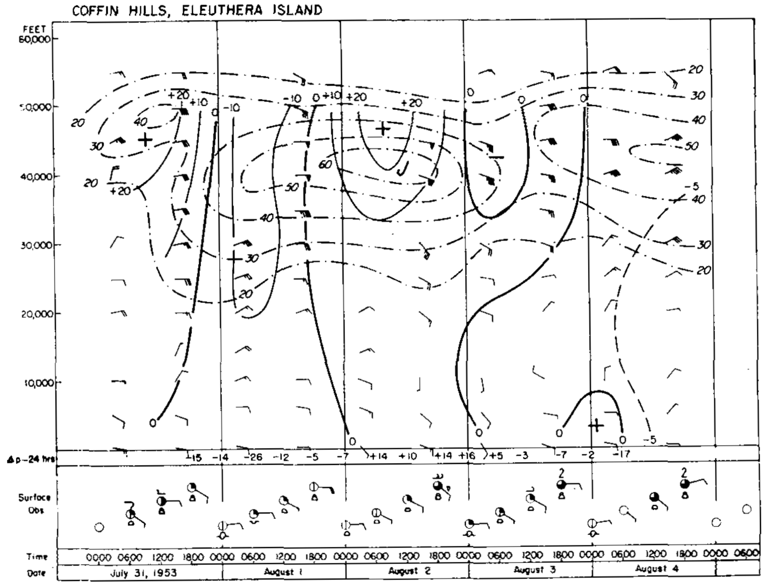

Fig. 3 a. Time section at Coffin Hills, Eleuthera Island during the period July 31—August 4, 1953. Full and dashed lines are isolines of 24-hour height changes in tens of feet; dash-dotted lines are isolines in knots. Twentyhour surface pressure tendencies in tenths of millibars are given in the row labelled Δp-24 hrs.

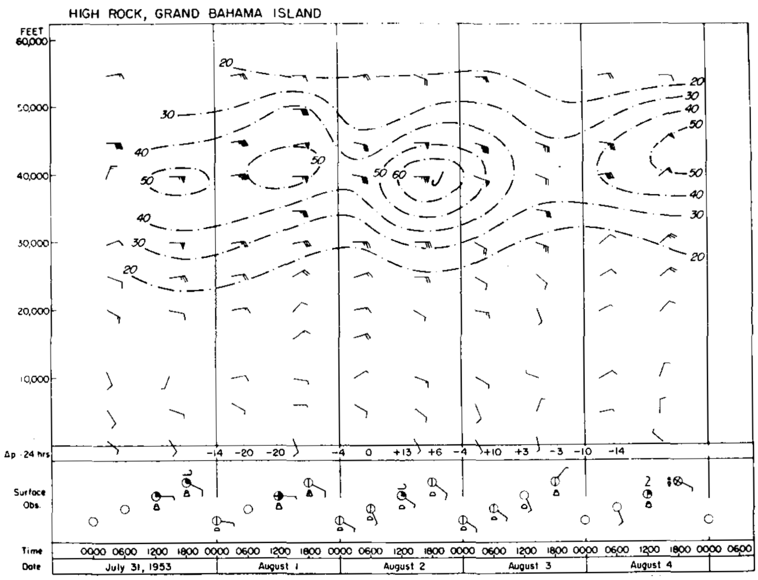

Fig. 3 b. Time section for High Rock, Grand Bahama Island. Lines are isotachs in knots.

high levels has no counterpart in the lower easterlies. On the other hand, we find a westward progression of cyclonic vortices which are closely linked to the wave troughs below.

b. *The 200-mb wind field (Fig. 4 a–d)*: The analysis for August 1 (Fig. 4 a) shows a narrow easterly current advancing over the Bahamas. At 0300 GCT the leading edge of the inner-

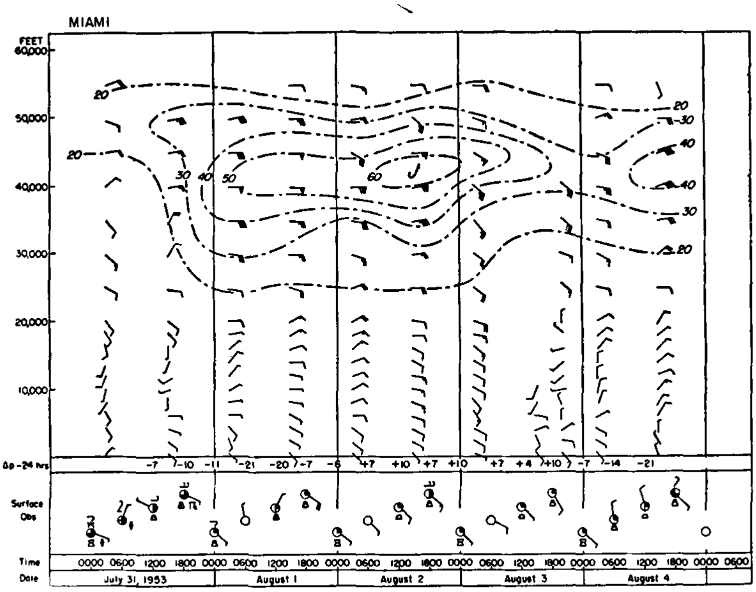

Fig. 3 c. Same as Fig. 3 b, for Miami, Florida.

most isotach, with a value of 50 knots, is just west of Miami which reports a 53 knot wind. A wind maximum is indicated northeast from there on the assumption (discussed below) of the continuity of the current with the winds reported by Ship H and Bermuda, together with the time sections for Coffin Hills (Fig. 3 a), High Rock (Fig. 3 b), and Miami (Fig. 3 c). Each of these stations shows a wind maximum subsequent to this time. If the isotach field is divided into four quadrants by the axis of the current, denoted by the thick line, and a line perpendicular to it drawn through the point of maximum wind speed (not entered on diagram), the cyclonic vortex noted in the 200-mb contour field can be located in the left forward quadrant.

The downstream propagation of the wind maximum can be seen from the successive positions of the leading edge of the 50-knot isotach, shown in Figure 8. The positions are fixed on the basis of a 43-knot wind at Burwood on August 3, 0300 GCT and an 48-knot wind at Shreveport on August 4, 1500 GCT and the assumption of a nearly constant rate of propagation between these observations. On this basis the leading edge of the 50-knot isotach shows a movement of about 3° latitude during the next 24 hours. This is about the same as the distance travelled by the cyclonic vortex, which thus retains unchanged its spatial correspondence with the wind maximum. This correspondence seems to persist throughout the period.

The maximum winds remain about the same or increase slightly up to August 2, 1500 GCT but show an unmistakable tendency to diminish thereafter as is evidenced by the progressively smaller area which can be enclosed within the 50 knot isotach after this date.

On August 3, 0300 GCT (Fig. 4 c), the fact

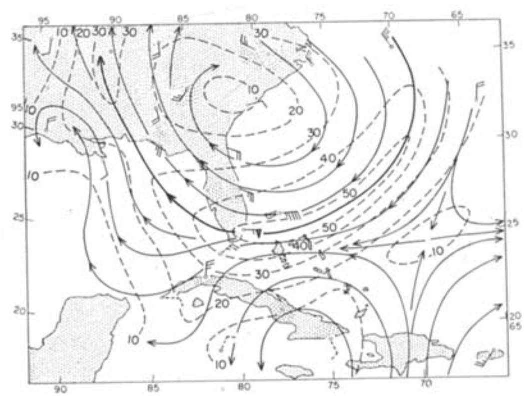

Fig. 4 a. 200-mb streamlines (full lines) and isotachs in knots (dashed lines) at 0300 GCT, August 1, 1953. Thick line denotes axis of current.

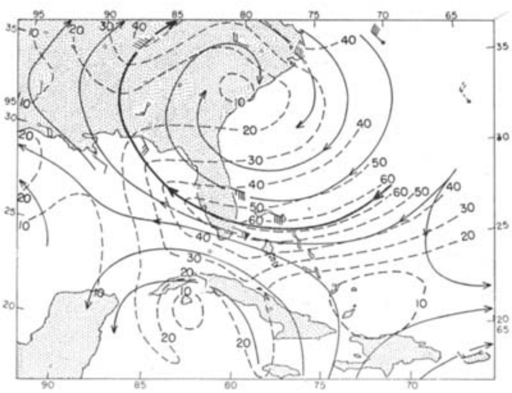

Fig. 4 b. 200-mb streamlines and isotachs, August 2, 0300 GCT.

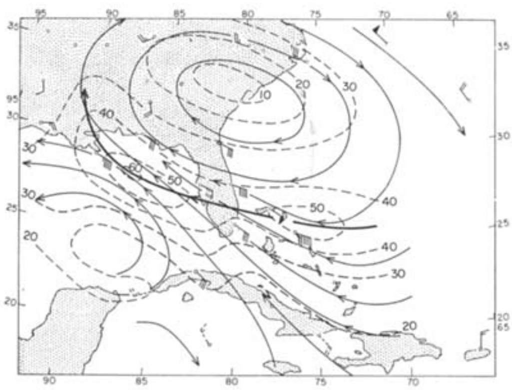

Fig. 4 c. 200-mb streamlines and isotachs, August 3, 0300 GCT.

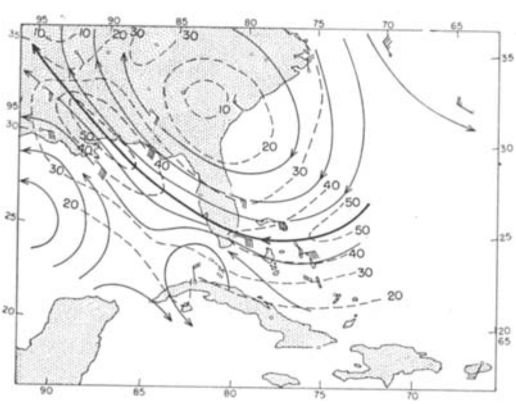

Fig. 4 d. 200-mb streamlines and isotachs, August 4, 0300 GCT.

that the winds at Miami and Tampa are weaker than that reported by either High Rock to the east or Burwood to the west, suggested a split of the current into two maxima. The new cyclonic vortex at 200 mb, drawn near Havana 24 hours later, is also located in the left front quadrant with respect to the rear maximum.

6. The Origin of the Current

The wind analysis in Figure 4 a is based on the inference that the easterly current which appeared over the Bahamas, originated in the middle latitudes as a branch of a split westerly jet which curved around the eastern periphery of the sub-tropical high cell centered over the eastern United States. The following evidence supports this inference:

a. On July 29, a jet stream with core velocities exceeding 130 knots was situated over the United States at about latitude 45° N. In the southerly fringe of the jet, the 200 mb wind directions changed from W—WNW over the continent to NW—N near the east coast, suggesting southward turning. Figure 9 shows the 200 mb winds at Bermuda and Ship H (latitude 36.8° N, longitude 69.6° W) during the period July 29 – 31. At Bermuda the wind direction fluctuated between N and NE. These observations together with the 45-knot northerly wind recorded at Ship H on July 30, suggest that a southern limb of the middle latitude jet stream was being injected into the tropics between the east coast and Bermuda.

b. On August 1, 0300 GCT, the 200 mb wind was northeasterly at Bermuda and easterly over the Bahamas and Miami. The 200 mb contours (not reproduced here) drawn for these winds and the reported heights show a channel extending from latitude 35° N to the Bahamas. By contrast, the contour field to the east of the Bahamas drawn to fit the southwesterly wind at San Juan and the southeasterly wind at Guantanamo, make it improbable that the current over the Bahamas was derived from the eastern or central tropical Atlantic.

Thus, unless the strong winds developed very close to the Bahamas, the inference is inevitable that the current appearing east of Florida must have been propagated from the middle latitudes. This statement, however, should not

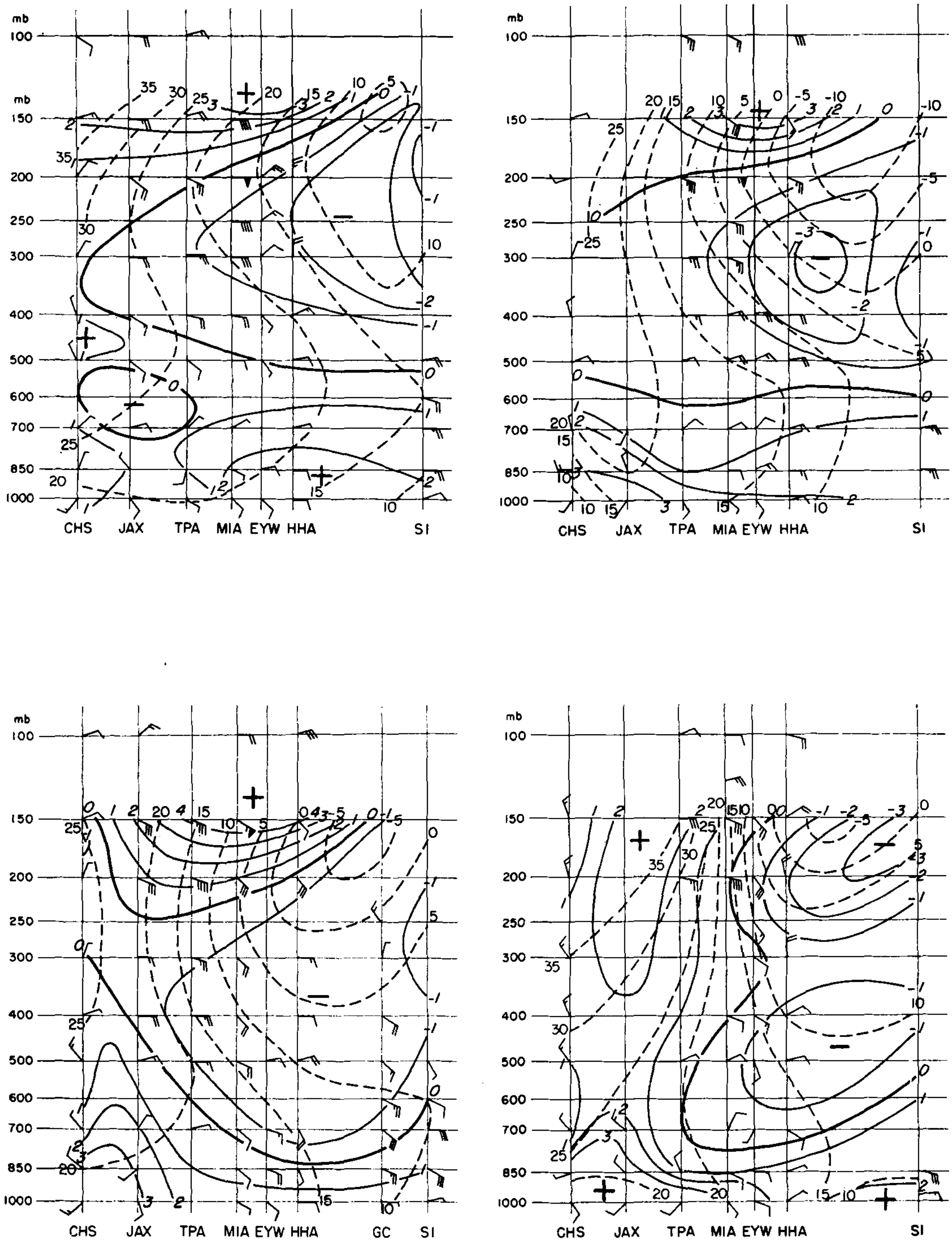

Fig. 5 a—d. Meridional cross sections showing temperature anomalies in degrees centigrade (full lines), and height anomalies in tens of feet from Schacht's mean tropical atmosphere at 0300 GCT on: (a) August 1, (b) August 2, (c) August 3, (d) August 4, 1953.

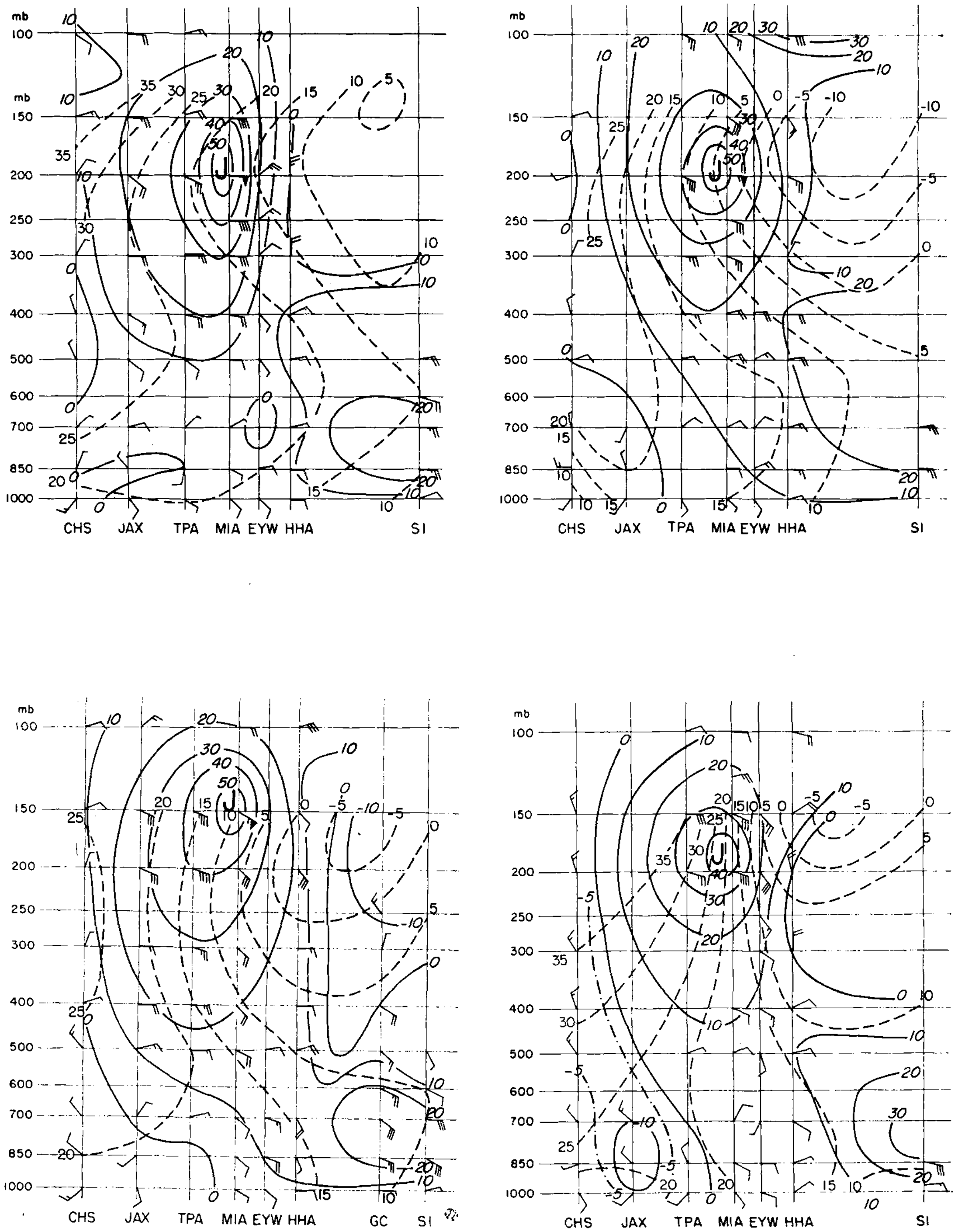

Fig. 5 e—h. Height anomalies in tens of feet from Schacht's mean tropical atmosphere (dashed lines) and isotachs of the observed zonal wind component in knots (full lines, easterly wind positive) at 0300 GCT on (e) August 1, (f) August 2, (g) August 3, (h) August 4, 1953.

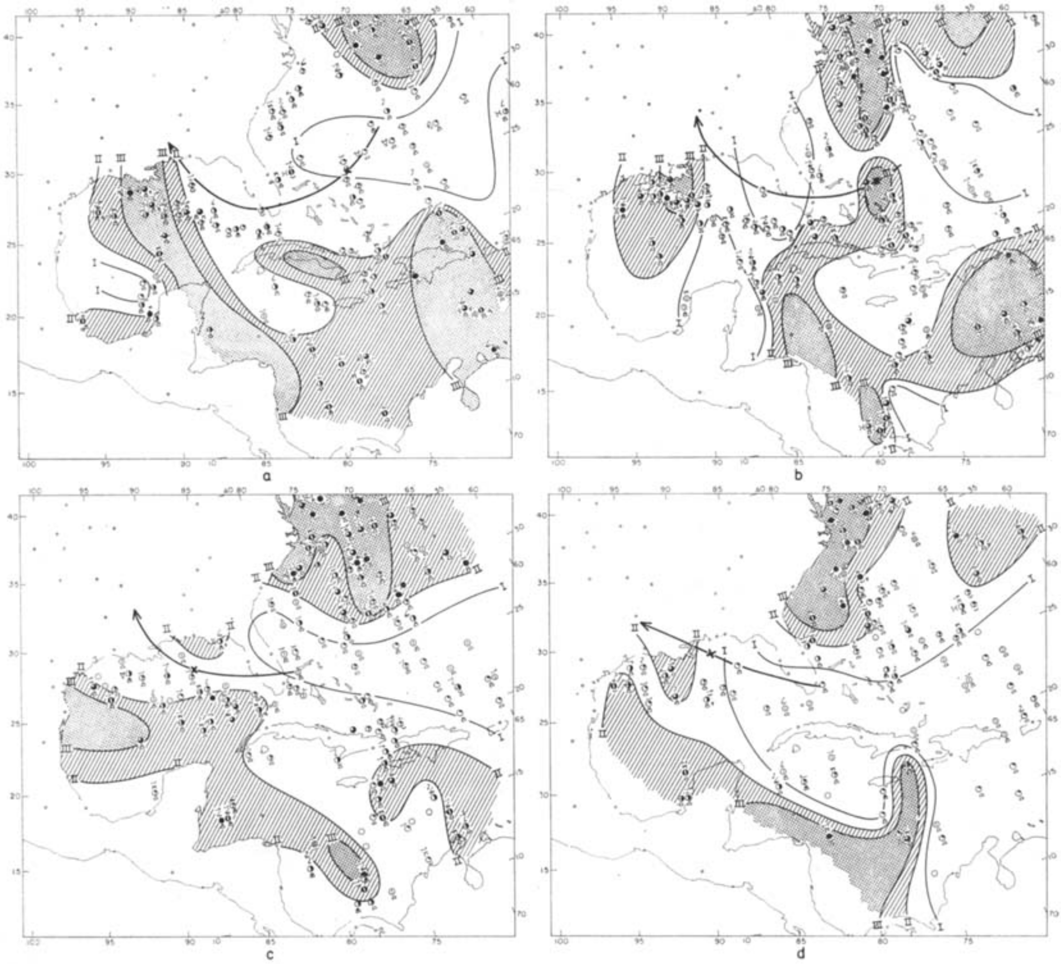

Fig. 6. Maps showing distribution of cloud cover on (a) August 1, (b) August 2, (c) August 3, (d) August 4, 1953. Lines marked III separate areas (cross hatched), where low cloud cover exceeds 5/10, from areas (hatched) where total cloud cover exceeds 5/10. Lines marked II separate latter from areas with total cloud cover less than 5/10 and more than 3/10. Lines marked I enclose areas where total cloud cover is less than 3/10. Maps are constructed from 6-hourly ship reports during 24-hour periods beginning 1830 GCT (see text).

be taken to imply that merely an injection of high kinetic energy from the north is taking place. As shown by Figure 8, the 50 knot isotach moves at only 7.5 knots, about an order of magnitude less than the wind speed itself. This involves a continuous production of kinetic energy in the entrance zone as the air particles overtake the speed pattern, and a corresponding dissipation in the exit zone ahead of the speed maximum while the jet travels through the tropics.

7. Kinetic and Thermal Structure: a Comparison with other Jet Streams

a. *Is this current a jet stream?*: In order to answer this question, we have to define the term "jet stream". When first introduced into meteorology this term referred to a narrow current with high westerly winds embedded in a relatively quiescent ambient atmosphere and meandering eastward around the hemisphere. Since that time, studies have revealed a marked "streakiness" in the kinetic texture of

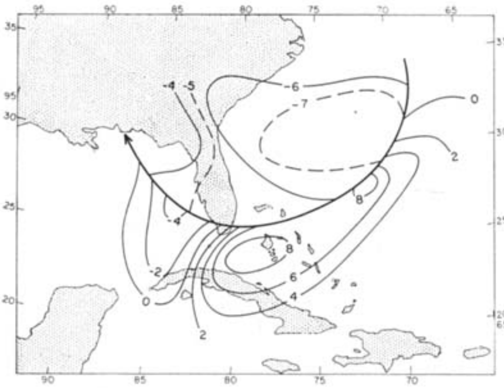

Fig. 7 a. Relative vorticity (10^{-5} sec^{-1}) at 200 mb on August 1, 1953 at 0300 GCT.

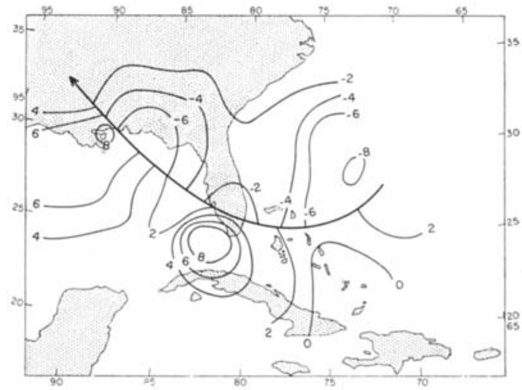

Fig. 7 d. Same as Fig. 9 c, for August 4.

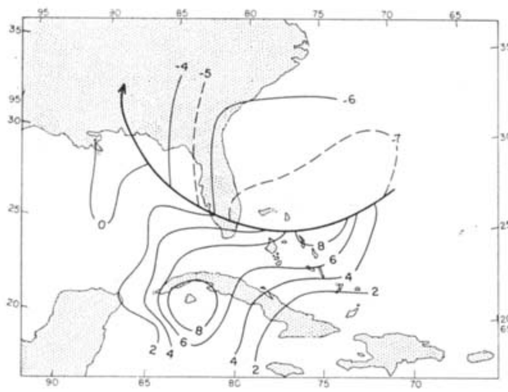

Fig. 7 b. Same as Fig. 9 a, for August 2.

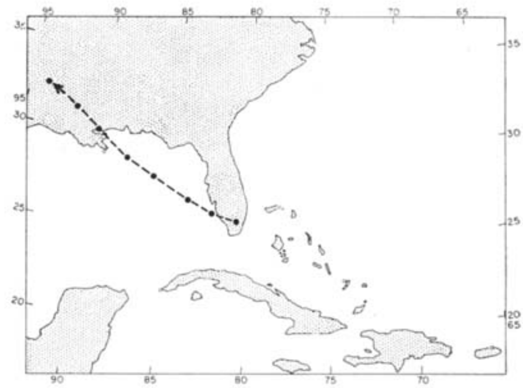

Fig. 8. Successive 12-hourly positions of the leading edge of the 50-knot isotach starting on August 1, 1953, 0300 GCT.

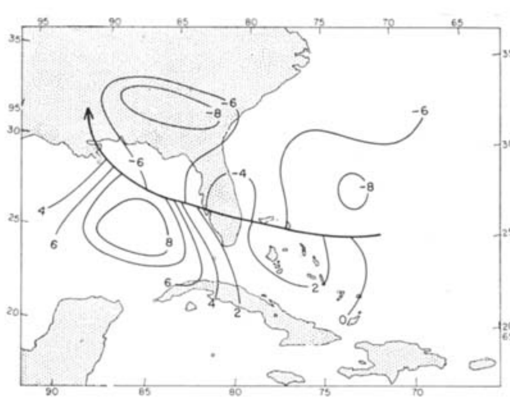

Fig. 7 c. Same as Fig. 9 b, for August 3.

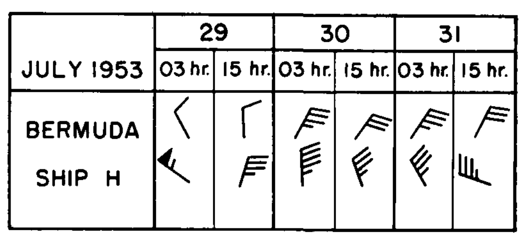

Fig. 9. 200-mb wind at Bermuda and Ship H (lat. 36.8° N; long. 69.6° W).

this general current. The energy concentration occurs along multiple axes of limited geographical extent along which wind maxima alternate with regions of lower wind speed. The maxima are characterized by a vertical as well as a lateral concentration of momentum. They retain their identity for short periods of time but undergo constant modification as they propagate downstream with a speed considerably less than that of the wind of which their cores are composed. Following recent usage, we shall call a jet stream each individual axis along which the wind speed is concentrated laterally and vertically and which features one or more velocity maxima with the characteristics of life cycle and motion described above.

In the present case, the lateral and vertical concentration of the current is clearly revealed by Figures 4 a – d and 5 e – h. The wind distribution pattern can be traced over a period of four or more days as the leading edge of the current moves slowly from the Bahamas westward through the Gulf of Mexico, then northwestward to the United States. The successive 12-hourly positions of the leading edge of the 50-knot isotach (Fig. 8) show an average rate

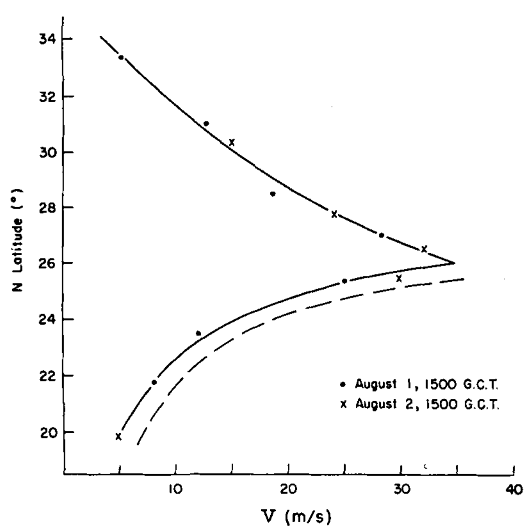

Fig. 10. A composite wind profile (solid line) along the 82° W meridian drawn from wind observations on August 1 and 2, at 1500 GCT by projecting the observed winds at nearby stations on the meridian. Points are plotted according to their distance from the wind maximum on each day. The dashed curve shows the distribution of wind velocity, drawn for the appropriate latitude belt, when the ratio of the absolute vorticity to the wind velocity is a constant with a value of (150 km)$^{-1}$.

of motion of about 10 knots or about $^1/_5$ of its denomination.

Thus, in accordance with the definition stated above, the easterly current is a jet stream. An interesting feature of the present jet stream is its comparatively weak core velocities. Jet streams – or at least those of middle latitudes – have always been associated with high wind speeds. We note here the tendency of the currents in the upper troposphere to concentrate into narrow bands irrespective of the amount of maximum speed.

In spite of the weak core velocities, the present easterly jet reveals some structural features which appear to be determined by the same properties characteristic of the more intense westerly jet streams of the middle latitudes. Of these we cite the following:

b. *Cyclonic wind shear:* In a recent study, RIEHL and MAYNARD (1954) have noted that on the low pressure side of some jet streams the ratio of the absolute vorticity to the wind velocity is a constant (k) with a value of (150 kms)$^{-1}$. By integrating the differential equation describing this relation for straight currents they found that

$$V = \frac{f}{k} + \left(V_0 - \frac{f}{k}\right) e^{-kn} \tag{2}$$

where V is the wind speed expressed in meters per second and is equal to V_0 at the reference point, and n is the distance normal to the current from the reference to the computed point.

Figure 10 represents a composite wind profile drawn from wind observations on August 1 and 2 at 1500 GCT projected on the 82° W meridian. The points are plotted according to their distance from the maximum wind on each day. The lower branch represents the wind profile on the low pressure side of the current. The dashed curve shows the meridional distribution of the wind velocity according to equation (2), drawn for the appropriate latitude belt. The fit between the latter and the observed wind profile is remarkable.

c. *Anticyclonic wind shear:* Computations based on both observed (RIEHL and MAYNARD, 1954) and geostrophic winds (PALMÉN and NEWTON, 1948) have revealed that on the high-pressure side of jet streams, the anticyclonic shears are such that the absolute vorticity has small positive values, with a small core of zero

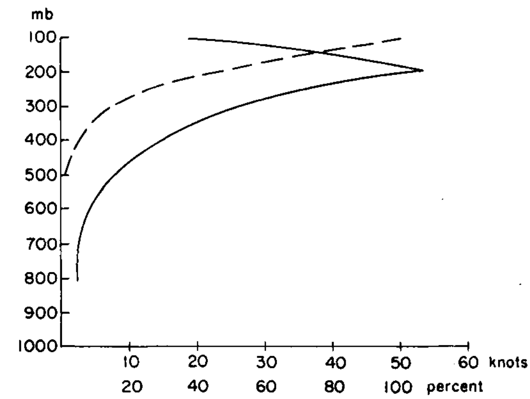

Fig. 11. Vertical profile (solid line) of easterly wind component through jet axis at 82° W meridian on August 1, 1953, 0300 GCT. Dashed line represents curve of cumulative percentage of kinetic energy from 1,000—100 mb.

or negative absolute vorticity within 150 kms from the axis. Our vorticity computations presented in Figure 7 bear out this fact. The negative relative vorticities to the right of the jet axis are everywhere numerically slightly less than, or equal to, the Coriolis parameter, except in narrow regions where they are numerically slightly higher than the Coriolis parameter, indicating negative absolute vorticity.

d. *Vertical concentration:* A majority of jets, especially those occurring at lower latitudes and in summer are restricted to a shallow layer in the upper troposphere. Often conditions at 500 mb give no indication of those higher up. The present case is no exception in this respect, as is readily revealed by an inspection of figures 3 a – c and 5 e – h. The vertical concentration of the kinetic energy is graphically portrayed by Figure 11 in which the full line represents the vertical profile of the easterly wind component[1] on August 1, 0300 GCT at the latitude of strongest easterly wind along the 82° W meridian (see Fig. 5 e). The average shear between 300 and 200 mb is about 4×10^{-3} sec^{-1}. The dashed curve shows the cumulative percentage of the kinetic energy from 1,000 to 100 mb, corresponding to the wind profile. Only a negligible percentage of the kinetic energy is found below 400 mb and fifty-four percent of the total is concentrated in the shallow layer 100 – 200 mb.

The vertical shear of 4×10^{-3} sec^{-1} is of the same order of magnitude as that observed in the much more intense middle latitude westerly jets in winter. But a distinguishing feature of the latter is the co-existence of a frontal layer beneath the jet core which contributes an appreciable percentage of the total vertical wind shear. For example, the mean cross-section drawn by PALMÉN and NEWTON (1948) from 12 cases in December, 1946, shows that fully one-third of the total increase in the troposphere occurs in the shallow frontal layer. No such layer exists in the present case.

e. *The thermal structure:* The thermal field shown by the full lines of figure 5 a – d reflects the independence of the concentrating mechanism of any temperature contrast between two air masses in the lower and middle troposphere. In fact, on August 1 and 2 (Fig. 7 a, b) the layers below 500 mb show a "reversed" temperature gradient, indicating a decrease of wind with height at the latitude of the jet core. On August 3 and 4 (Fig. 5 c, d) the northward temperature gradient underneath the jet core extends nearer to the surface but it is quite certain that no air mass contrast is involved.

8. The Balance of Forces

The continuous production and destruction of kinetic energy in jet stream regions is associated with a distinctive pattern of particle acceleration and deceleration. The manner in which this pattern is related to the structure and motion of jets is best studied by considering the laws of motion valid for particles near jet stream axes.

Aside from the vertical pressure force which is balanced by the force of virtual gravity, the balance of forces on a particle in relative horizontal motion is conveniently described by two equations: A tangential equation,

$$\frac{dV}{dt} = -g\frac{\partial h}{\partial s} + F_s \qquad (3)$$

and a normal equation,

$$\frac{V^2}{R} + fV = -g\frac{\partial h}{\partial n} + F_n \qquad (4)$$

In the above equations V is the horizontal wind speed, h the geodynamic height, s and n

[1] This profile closely approximates that of the total wind in view of the small deviation of the latter from the easterly direction.

unit distances respectively along and to the left of the direction of motion, f the Coriolis parameter, R the radius of curvature of the trajectory, and F_s and F_n the respective frictional forces in the s and n directions.

The frictional forces in the free atmosphere are in general assumed negligible in comparison with the other terms in equations (3) and (4). The validity of this assumption is not *a priori* evident in jet stream regions where strong lateral and vertical wind shears occur. We shall, therefore, attempt to determine to what extent such an assumption is justified in the present case. We thus write

$$\frac{dV}{dt} = -g\frac{\partial h}{\partial s} \tag{5}$$

or

$$V_2 - V_1 = -g\int_{t_1}^{t_2}\frac{\partial h}{\partial s}dt \tag{6}$$

and find out whether this equation is balanced.

The integration of equation (6) is performed along a trajectory traversed during the time interval $t_1 - t_2$. The subscripts 1 and 2 respectively refer to quantities at the initial and terminal points of the trajectory.

The computation of the right-hand side of equation (6) in its present form involves measurement, along the trajectory, of the angles between contours and wind. In view of the smallness of these angles the error in their measurement may be percentually large. In order to circumvent this difficulty, we write,

$$\frac{dh}{dt} = \frac{\partial h}{\partial t} + V\frac{\partial h}{\partial s} + w\frac{\partial h}{\partial z}$$

and if we neglect vertical motion, equation (5) becomes

$$V\frac{dV}{dt} = -g\left(\frac{dh}{dt} - \frac{\partial h}{\partial t}\right)$$

Multiplying both sides by dt, integrating over the period $t_1 - t_2$, and denoting the heights at t_1 and t_2 repectively by h and h' we have

$$\frac{V_2^2 - V_1^2}{2} = -g\left[(h_2' - h_1) - \int_{t_1}^{t_2}\frac{\partial h}{\partial t}dt\right]$$

The integral on the right-hand side of the above equation expresses the local height change occurring over the trajectory during the time $t_2 - t_1$. We assume a linear distribution of this quantity; this assumption is justified if no extreme values of local height changes occur between the initial and terminal points of the trajectories. The integral then becomes

$$\int_{t_1}^{t_2}\frac{\partial h}{\partial t}dt = \frac{h'_1 - h_1}{2} + \frac{h'_2 - h_2}{2} \tag{7}$$

Substituting the above value and defining $\overline{V} = \frac{V_1 + V_2}{2}$, equation (6) reduces to

$$V_2 - V_1 = -g/\overline{V}\left[\frac{(h'_2 + h_2)}{2} - \frac{(h_1 + h'_1)}{2}\right] \tag{8}$$

Putting $\frac{h_2 + h_2'}{2} = \bar{h}_2$, the mean height at the terminal point of the trajectory during the time interval $t_1 - t_2$, and $\frac{h_1 + h_1'}{2} = \bar{h}_1$, the mean height at the initial point of the trajectory during the same time interval we have,

$$V_2 - V_1 = -g/\overline{V}(\bar{h}_2 - \bar{h}_1) \tag{9}$$

The terms of equation (9) are readily computable from contour and wind maps. Table 1 shows the result of six such computations performed during the period August 1 – 3. The computations were made along 12-hour trajectories starting at Miami and High Rock at 0300 GCT. These stations were selected because of their positions near the axis of the jet, and also because the ends of the 12-hour trajectories were located over or near the southern United States where height and wind observations at nearby stations made the computations easier to perform and subject to less error. Furthermore, the passage of the wind maximum across these stations makes the computations representative of conditions both ahead and behind the jet.

The above method, aside from the fact that it does not depend on the measurement of the small cross-contour angles of the wind, has the added advantage that the acceleration of a particle along a trajectory is measured even though the acceleration is zero at the initial and terminal points of the trajectory. Consider, for instance, the trajectory of the particle between two maps, and assume that on both these maps, the streamlines and the contours coincide. Nevertheless, the particle is accelerated during the time interval between the two maps and this acceleration can be computed by evaluating the right-hand side of equation (9).

The computations are, of course, subject to error. The uncertainty in drawing the trajectories and the assumption of linear variation of height changes in time and space constitute the main sources of error. However, because of the low wind speeds and the comparatively short distances involved, the error in locating the terminal points of the trajectory is smaller than would be the case in fast-moving westerlies.

A simple test of the accuracy of the trajectories and also of the consistency of the analysis of the wind and pressure fields can be made with the help of equation (4). Again neglecting friction, we solve for the radius of curvature of the trajectory:

$$R = -\frac{V^2}{fV + g\dfrac{\partial h}{\partial n}} \qquad (10)$$

Table 2 compares the radius of curvature thus computed with that measured from the trajectories at Miami and High Rock on August 2 – 4. The correspondence between the two quantities is very close.

Tables 1 and 2 indicate that the jet is maintained by pressure heads and that the continuous production and destruction of kinetic energy occurs as a result of accelerations in more or less laminar flow rather than by lateral or vertical frictional stresses.

The unbalance between wind and pressure is further illustrated in Figure 12 in which the easterly components of winds observed at 0300 GCT during the period August 1 – 4 are plotted against the corresponding geostrophic winds. Both the observed and computed winds were averaged along the 82° W meridian over a belt 4° of latitude in width, centered at the latitude of maximum winds. The shift from super- to sub-geostrophic winds attending the passage of the wind maximum across the 82° W meridian between August 1 and 4 is clearly discernible. The fact that on August 3 after the passage of the maximum, the winds are still supergeostrophic is associated with the rapidly weakening core winds after August 2.

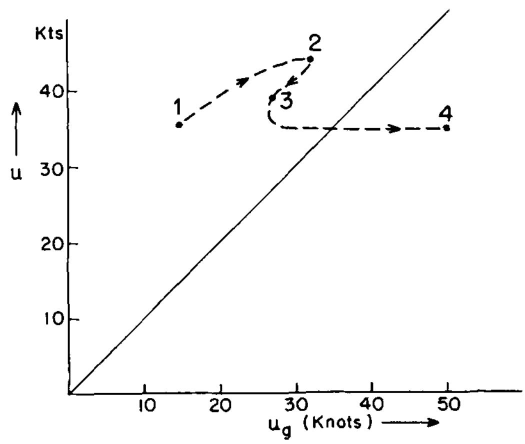

Fig. 12. 200-mb geostrophic vs. observed easterly component averaged over 4° L zone centered at latitude of maximum wind, at 0300 GCT during the period August 1—4, 1953.

Table 1. A comparison of the left hand side (column A) and the right hand side (column B) of equation 8 made from measurements along 12-hour trajectories at 200 mb starting at Miami and High Rock at 0300 GCY (see text).

Station	Date	A (mps)	B (mps)
Miami	August 1	12.0	10.6
High Rock	August 1	7.5	7.4
Miami	August 2	15.0	10.7
High Rock	August 2	1.5	.7
Miami	August 3	—3.0	14.1
High Rock	August 3	3.0	1.9

Tabel 2. A comparison between the measured and computed radii of trajectory curvatures at 200 mb on August 1—3, 1953, 0300 GCT at Miami and High Rock. All curvatures are anticyclonic.

Station	Date	Measured (°latitude)	Computed (°latitude)
Miami	August 1	9.0	7.5
High Rock.....	August 1	9.0	7.5
Miami	August 2	15.0	12.5
High Rock	August 2	8.5	7.5
Miami	August 3	35.0	35.0
High Rock	August 3	20.0	15,0

9. Divergence and Vorticity; Distribution of Weather Activity

The distribution of upper tropospheric divergence can also be determined by considering the distribution of vorticity which, in the present case, depends mainly on the wind shear normal to the jet axis.

a. *The distribution of vorticity:* Figure 7 shows that with the exception of August 1, 0300GCT, the relative vorticity is positive to the left and negative to the right of the axis indicating the preponderance of the shear over the curvature term in equation (1).

The distribution of the positive relative vorticity to the left of the axis is such that a vorticity maximum is associated with every speed maximum and the cyclonic vortex which accompanies it. The movement of these vorticity maxima in response to the movement of the jet center is easy to follow by comparing Figures 4 a–d and 7 a–d. The relative vorticity in general increases downstream behind the wind maxima and decreases ahead of them. The maximum relative vorticities computed do not exceed twice the Coriolis parameter – a value much less than that observed on the low pressure side of some vigorous middle latitude jet streams in winter.

To the right of the axis the relative vorticity decreases downstream behind the wind maxima and increases ahead of them. Here, the negative relative vorticity is numerically less than or equal to the earth's vorticity except in a small region some distance away from the velocity maxima where its value numerically exceeds the Coriolis parameter, indicating small negative values of the absolute vorticity. Such small values are within the limit of computational error but should not *ipso facto* be discounted as spurious. A discussion of the vorticity equation will help elucidate this point. Stripped of the small frictional and solenoidal terms and also of the terms relating to the curvature of the earth's surface, the vorticity equation is as follows:

$$\frac{d\zeta_a}{dt} = -\zeta_a \operatorname{div}_2 \mathbf{v} + \frac{\partial V}{\partial z}\frac{\partial w}{\partial n} \qquad (11)$$

The first term shows that divergence reduces and convergence increases the vorticity of a particle and the last term describes the influence of vertical motion in changing vorticity about the vertical. In general, the latter term is considered insignificant compared to the divergence term and the vorticity equation is often written

$$\frac{d\zeta_a}{dt} = -\zeta_a \operatorname{div}_2 \mathbf{v} \qquad (12)$$

There is little doubt that equation (12) is accurate enough in most cases, especially in regions of strong cyclonic vorticity as, for instance, to the left of jet stream axes. In such cases particle vorticity changes are very sensitive to horizontal divergence, or convergence. Air subject to sustained convergence will increase its cyclonic vorticity without any theoretical limit. On the other hand, when ζ_a approaches zero the vorticity changes become insensitive to horizontal divergence. Indeed, equation (12) does not allow for the decrease of the absolute vorticity below zero. Accordingly, if this equation always represented the true condition of the atmosphere negative absolute vorticities would be dynamically inadmissible.

But, as pointed out by Bjerknes (1951), as ζ_a approaches zero, the divergence term becomes less important than the last term of equation (11) and the vorticity equation becomes

$$\frac{d\zeta_a}{dt} = \frac{\partial V}{\partial z}\frac{\partial w}{\partial n} \qquad (13)$$

This equation does not place any restriction against the occurrence of negative absolute vorticities and lends credence to the existence of such values in limited regions to the right of jet stream axes, as indicated by figures 7 a–d, and as has also been found by previous investigators (Riehl and Maynard, 1954; Palmén and Newton, 1948).

b. *The distribution of divergence:* The distribution of divergence can be discussed by applying equations (12) and (13), respectively, to the left- and right-hand portions of jet streams. Considering the left side of a steady jet stream with its axis oriented east—west, moving down stream with constant speed c, we may write from equation (12),

$$(V - c)\frac{\partial \zeta}{\partial s} = -\zeta_a \operatorname{div}_2 \mathbf{v} \qquad (14)$$

neglecting the vertical vorticity transport. Upstream from the wind maximum $\partial\zeta/\partial s$ is positive indicating convergence. In the downstream quadrant $\partial\zeta/\partial s$ is negative; here divergence prevails. If the 200-mb chart is located above a level or surface of non-divergence located somewhere in the middle troposphere, it follows from mass continuity reasoning that the air must ascend upstream and descend downstream from the wind maximum.

Proceeding to the right of the jet stream axis and applying equation (13)

$$(V-c)\frac{\partial\zeta}{\partial s}=\frac{\partial V}{\partial z}\frac{\partial w}{\partial n}, \quad \text{or}$$

$$\frac{\partial w}{\partial n}=\frac{(V-c)\,\partial\zeta/\partial s}{\partial V/\partial z}$$

In the right rear quadrant the vorticity decreases downstream and therefore the term $\frac{\partial V}{\partial z}\frac{\partial w}{\partial n}$ is negative. At 200 mb, $\partial V/\partial z$ is positive as can be seen from Figures 3 and 5. Therefore $\partial w/\partial n$ must be negative, i.e., w increases to the right. The order of magnitude of this rate of increase can be estimated. Since

$$\frac{\partial V}{\partial z}\sim 10^{-3}$$

$$\frac{\partial w}{\partial n}\sim\frac{10^{3}.10^{-12}}{10^{-3}}\sim 10^{6}$$

or 1 cm sec^{-1}/10 km.

The horizontal shear of the vertical motion normal to the current is thus very sharp near the jet axis. In view of the small vertical velocities in the atmosphere, the subsidence prevailing to the left of the axis must soon give way to rising motion, active weather conditions, and upper air divergence to the right of the axis. But this rapid increase in vertical motion will not persist a long distance from the axis since the advective term $(V-c)\ \partial\zeta/\partial s$ decreases rapidly away from the jet stream axis.

A similar argument can be applied along the leading edge of jet streams to show that active weather conditions with upper air divergence in the left front quadrant rapidly give way to settled conditions and upper air convergence across the jet axis.

The above discussion deals with the simple case of a jet with its axis oriented east—west. When the latter has a north—south component, the above model is modified in the following manner: where the flow is directed toward lower latitudes, conditions of organized convection are accentuated and vice-versa.

In summary, we may state that the distribution of vorticity visualizes, under model conditions, upper divergence, and therefore upward motion and active weather conditions in the front left and rear right quadrants; and upper convergence and therefore descending motion and inactive weather conditions in the front right and rear left quadrants.

c. *A discussion of the weather situation:* An instance when the weather distribution closely follows model conditions, is that occurring to the right of the jet axis during the period studied. The area or organized convection situated west of Bermuda on August 1, and whose leading edge reaches the north-west coast of Florida by August 4, follows the movement of the jet maximum and is all the time located in the rear right quadrant of the jet stream. The prevalence of suppressed convection over Florida on August 2 is similarly attributable to the downstream increase of vorticity in the right front quadrant of the jet stream.

Often conditions are more complicated when the weather is determined by a combination of two or more weather producing systems, each with its own typical weather distribution. Certain weather features are then intensified while others are damped, depending on the phases in which the different systems meet. The weather distribution to the left of the jet axis during the period August 1–4 provides an interesting illustration of this fact. Three different weather producing systems, each ordinarily capable of occurring indepedently of the other two, and each characterized by a distinctive weather pattern co-exist:

1. Waves in the lower easterlies. The weather distribution, associated with these systems differs from case to case depending on the slope of the trough and ridge lines with height and the vertical variation of the basic current (RIEHL, 1954). More often than not, however, there is a tendency for organized convection to occur to the east of the surface trough line and suppressed convection to the west of it.

2. Zones of wind shear in the upper troposphere. Such zones frequently occur over the

Caribbean in summer. Typically, they have a west—east orientation with easterly winds to the north and westerly winds to the south. They slope northward with height with unsettled weather often occurring beneath this slope (RIEHL, 1954).

3. A jet stream with a model weather distribution as described at the beginning of this section.

The actual weather distribution can be discussed from the viewpoint of net particle vorticity changes resulting from the resonance or interference of the above systems.

These changes are brought about in a different manner near and away from jet streams. Near jet stream axes, vorticity is determined mainly by the horizontal wind shear. Vertical motion is determined chiefly by the horizontal variation of this quantity near the level of strongest winds, while the vertical transport plays only a secondary role. Away from jet streams, on the other hand, vorticity is determined mainly by the curvature of the streamlines, and the variation of this property with height may be as important as the horizontal transport of vorticity.

The above fact is illustrated by the area of organized convection situated west of Puerto Rico, and that occurring over Cuba on August 1 (Fig. 6). Over the former area, the 15 knot wind at San Juan is greater than the rate of westward propagation of the trough line. Air particles in this region are therefore approaching the trough and gaining cyclonic curvature. This indicates that convergence predominates at least up to this level. Particles rising through this level lose cyclonic curvature as they arrive into the anticyclonically curved flow at 200 mb where divergence, therefore, prevails.

Over Cuba, on the other hand, the curvature is more cyclonic at 200 mb than at 500 mb; nor does the cyclonic curvature decrease downstream at the former surface. The organized convection is due to the downstream decrease of vorticity occurring at 200 mb due to the strong horizontal wind shear associated with the jet stream.

The above case also illustrates the interference between the easterly wave trough and the jet stream. The area of disturbed weather occurs to the west of the surface position of the trough line[1] which is ordinarily characterized by suppressed convection. Air particles are moving away from the trough line at 500 mb, thereby losing cyclonic curvature. At 200 mb, particles flowing into the region are, if anything, gaining cyclonic curvature. This situation would be conducive to upper air convergence and descending motion except for the presence of the jet stream which is so situated that the downstream decrease of vorticity at 200 mb associated with it (Fig. 7 a) produces a net upper air divergence.

The area of organized convection over the western part of the Gulf of Mexico on August 3 is one in which the jet stream and the easterly wave trough meet in the same phase. The downstream decrease of vorticity at 200 mb (Fig. 7 c) due to the advance of the jet stream north of this region occurs east of the surface position of the surface wave trough as determined from the time section of Brownsville. Both are favorable for organized convection.

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[1] As deduced from time-sections.

REFERENCES

RIEHL, H., ALAKA, M. A., JORDAN, C. L., and RENARD, R., 1954: The Jet Stream. *Meteorological Monographs.* Amer. Meteor. Soc., 2. No. 7, 100 pp.

BJERKNES, J., 1951: Extratropical Cyclones. *Compendium of Meteorology*, pp. 577—598.

COLÓN, J., 1954: A Case Study of the High Tropospheric Wind Field Over the Caribbean Region in Spring. Mimeo. report, Dept. of Meteor., Univ. of Chicago, prepared under Contract N° N6ori-02036, Project NR 085003, Office of Naval Research.

CRESSMAN, G. P., 1950: Variations in the Structure of the Upper Westerlies. *J. Meteor.*, 7, 39—47.

FULTZ, D., 1951: Experimental Analogies to Atmospheric Motion. *Compendium of Meteorology*, pp. 1235—1248.

FUGLISTER, F. C., 1951: Multiple Currents in the Gulf Stream System. *Tellus*, 3, 230—233.

NEWTON, C. W., and CARSON, J. E., 1953: Structure of Wind Field and Vorticity in Summer. *Tellus*, 5, 321—339.

PALMÉN, E., 1951: The Role of Atmospheric Disturbances in the General Circulation. *Q. J. Met. Soc.*, **77**, 337—354.

—, and NEWTON, C. W., 1948: A Study of the Mean Wind and Temperature Distribution in the Vicinity of the Polar Front in Winter. *J. Meteor.*, **5**, 220—226.

PETTERSSEN, S., 1950: General Circulation. *Cent. Proc. Roy. Met. Soc.*, 120—155.

RIEHL, H., 1948: Studies of Upper Air Conditions in Low Latitudes, I. *Misc. Report N° 24*, Dept. of Meteor., Univ. of Chicago.

—, 1954: *Tropical Meteorology*. New York: McGraw-Hill Book Co., 392 pp.

RIEHL, H., and MAYNARD, H., 1954: Exploration of the Jet Stream by Aircraft During the Winter of 1953. Mimeo. Final Report prepared under Contract N° 189s-88360 between Project AROWA of the Bureau of Aeronautics, Department of the Navy and Dept. of Meteor., Univ. of Chicago.

ROSSBY, C.-G., 1953: A Comparison of Current Patterns in the Atmosphere and in the Ocean Basins. *Proc. U.G.G.I., 9th Gen. Assembly, Brussels, Autumn 1951* (publ. 1953).

SCHACHT, E. J., 1946: A Mean Hurricane Sounding for the Caribbean Area. *Bull. Am. Meteor. Soc.*, **27**, 324.