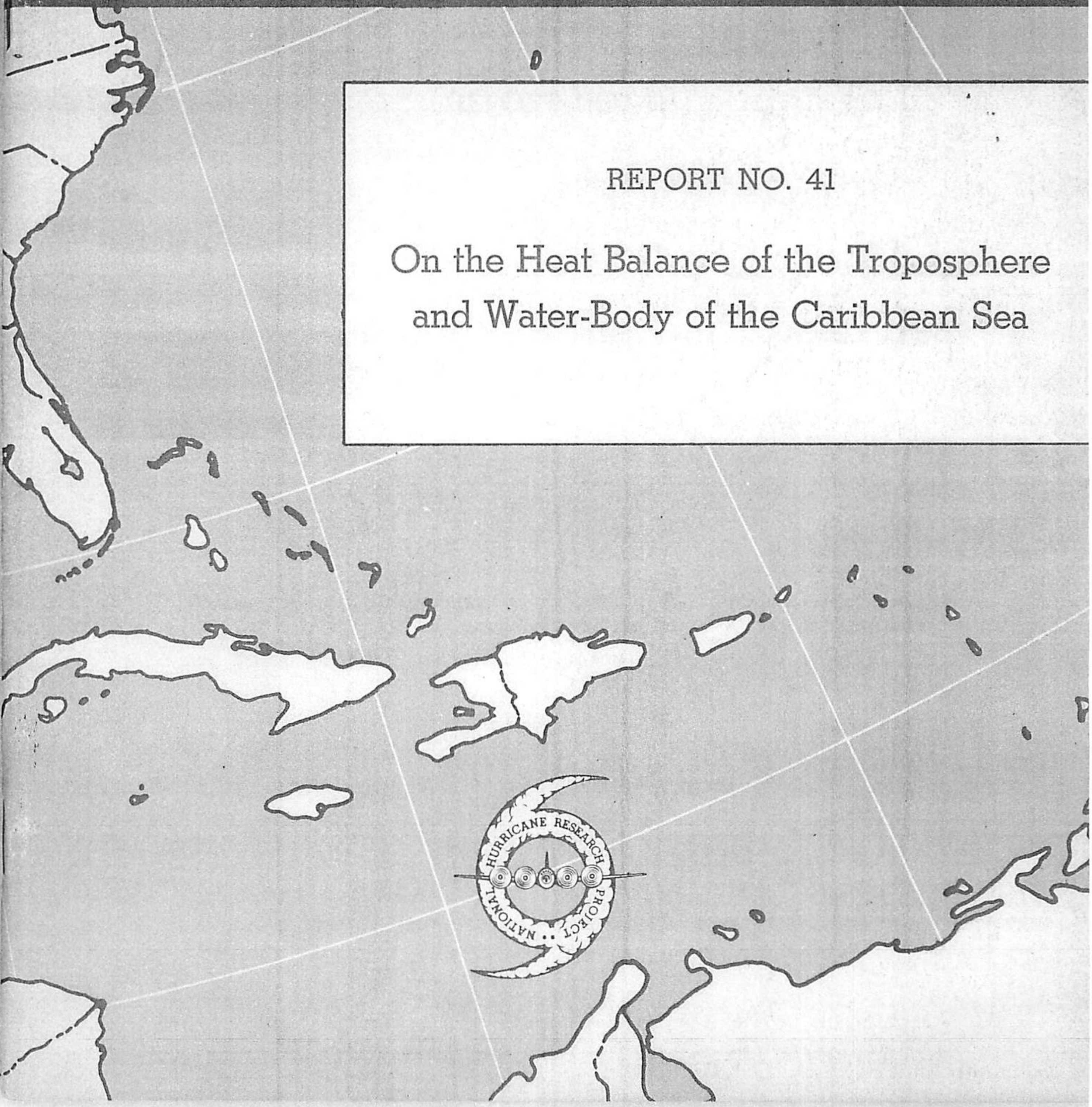
NATIONAL HURRICANE RESEARCH PROJECT

REPORT NO. 41

On the Heat Balance of the Troposphere and Water-Body of the Caribbean Sea



U. S. DEPARTMENT OF COMMERCE
Frederick H. Mueller, Secretary
WEATHER BUREAU
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NATIONAL HURRICANE RESEARCH PROJECT

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On the Heat Balance of the Troposphere and Water-Body of the Caribbean Sea

by

José A. Colón
National Hurricane Research Project, Miami, Fla.

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NATIONAL HURRICANE RESEARCH PROJECT REPORTS

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ON THE HEAT BALANCE OF THE TROPOSPHERE AND WATER BODY OF THE CARIBBEAN SEA

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ABSTRACT

Energy balance calculations for the troposphere above, and the oceanic body of the Caribbean Sea are presented. The atmospheric study is based on mean data for two winter months: December 1956 and January 1957. The oceanic analysis is carried out on a monthly basis with climatological data.

The structure of the atmospheric circulation during the period of study shows a two-layer atmosphere with trade easterlies in the lower layers, and a westerly current of equatorial origin centered near 200 mb. (40,000 feet). The upper current shows, in the monthly picture, horizontal and vertical concentration of kinetic energy, similar to jet streams of middle latitudes. It is a semi-permanent feature of the Caribbean area in winter and seems to be related to the return poleward branch of the tropical meridional circulation cell. The mean structure and properties of the flow are illustrated with vertical wind profiles, space cross-sections, and horizontal charts.

The energy changes in both the trades and the upper current are studied through integration of the energy equations over a volume bounded laterally by aerological stations around the periphery of the Caribbean Sea. Mass and energy fluxes are evaluated, and balance between sources and sinks established for different layers from the surface to the tropopause. The transport of heat from the Caribbean sea to the trades above is

calculated in two ways: as a residual in the water balance, and by means of formulae derived from the theory of turbulent transfer. The seasonal variations in this surface transfer are illustrated and discussed.

The computations in the trade layers verify the well-known role of the trades as accumulators and exporters of latent heat. There is a net export in the form of latent heat, averaging 2.4×10^{18} cal. day^{-1}, which is available for use in other regions of the globe and represents the contribution of the Caribbean atmosphere for maintenance of the general circulation. The balance of sensible heat serves as a basis for a discussion of the role of the heat exchange with the ocean, and the precipitation heating for maintenance of the local circulations and of balance of the radiation losses in the lower half of the troposphere. It is shown that although the Caribbean atmosphere exports latent heat for use elsewhere, at the same time it depends on importation of sensible heat from the equatorial areas to balance the radiation losses, particularly in the upper levels. This importation is achieved through the mechanism of the mean motion, which consists of convergence in the upper and divergence in the lower troposphere.

The total energy balance is illustrated for three main layers: the surface layer, the rain area, and the upper troposphere. The requirements for vertical heat and moisture flow are discussed and it is shown that convective updrafts can account for the required upward transfers.

The analysis of kinetic energy illustrates production inside the volume by a mechanism similar to a pressure head, in which air enters the volume at high and leaves at lower pressure. The production is sufficient to balance the frictional dissipation and export.

LIST OF SYMBOLS

A	Area enclosed by volume boundary
A'	Area occupied by convective updrafts
C	Mean cloudiness
c	Specific heat of water
c_p	Specific heat of air at constant pressure
c_v	Specific heat of air at constant volume
c_n	Component of wind normal to volume boundary
c_s	Component of wind parallel to volume boundary
D_K	Dissipation of kinetic energy
E	Evaporation rate
F	Frictional force
g	Acceleration of gravity
H	Net vertical heat flux across a given level
h	Rate at which heat is added to or subtracted from a system
K	Kinetic energy
k	Frictional drag coefficient
L	Latent heat of condensation
M_h	Horizontal mass flux
M_z	Vertical mass flux
M_{hy}	Horizontal mass flux on water current at Yucatan Channel
M_{he}	Horizontal mass flux on water current at east side of the Caribbean Sea
$\vec{n}$	Unit vector normal to the flow
P	Precipitation
P_K	Production of kinetic energy
p	Pressure
Q	Heat content
Q_a	Net heat transfer from ocean to atmosphere
Q_b	Back radiation with cloudy skies
Q_{bo}	Back radiation with clear skies
Q_e	Latent heat transfer from ocean to atmosphere
Q_o	Mean heat content at the surface
Q_r	Net radiation absorbed by water surface
Q_s	Sensible heat transfer from ocean to atmosphere
Q_t	Heat change in water body due to changes in mean temperature

Q_v	Divergence of heat transport by the water current
q	Specific humidity
q_a	Specific humidity of air at ship's deck level
q_w	Saturation specific humidity of air at the temperature of the sea surface
R_a	Atmospheric radiational cooling
S	Perimeter of lateral volume boundary
s	Horizontal length element around boundary
T	Temperature
T_w	Water temperature
T_b	Water temperature at bottom
T_e	Water temperature at east side of water current
T_y	Water temperature at the Yucatan Channel
t	Time
$\vec{t}$	Unit vector tangent to the flow
V	Resultant wind speed
$\vec{V}$	Velocity vector
$\bar{V}$	Average wind speed irrespective of direction
v	Surface wind speed
w	Vertical velocity
$\hat{w}$	Area average of vertical velocity
w'	Vertical velocity in convective updraft inside clouds
z	Contour height
$\hat{z}$	Area average of contour height
α	Volume element
ρ	Air density
ρ_w	Water density
θ	Potential temperature
σ	Lateral boundary surface
τ	Frictional stress
τ_o	Frictional stress at the surface
$\vec{\Omega}$	Rotational vector of the earth

1. INTRODUCTION

One of the fundamental problems in meteorology concerns the processes by which the radiation from the sun is absorbed and distributed throughout the atmosphere to create and maintain the general circulation. Studies of the radiation balance in the global scale (cf. Simpson [43, 44]; Baur and Phillips [3]; Gabites [13]; Houghton [15]; London [21]) reveal a surplus of insolation absorbed in tropical latitudes and a deficit in middle and high latitudes. A net transfer of heat from the Tropics into higher latitudes must then take place in order to maintain proper balance everywhere. Simultaneously, a northward transport of angular momentum is also required for the maintenance of the extratropical westerlies.

The atmospheric energy cycle is initiated to a large extent over the tropical oceans, where the absorption of solar energy exceeds the emission of long-wave radiation. The excess heat is transferred from the sea surface to the atmosphere above by conduction and, mostly, by evaporation, and energy in the form of latent heat is accumulated in the trades. Due to the presence of the trade inversion it is largely concentrated in the lower atmosphere (Riehl et al. [41]). A portion of this latent heat is ultimately transported to extratropical latitudes to maintain the circulation and offset radiation losses there. Conversion to other forms of energy and transport to the high troposphere takes place mostly along the equatorial trough (Riehl and Malkus [38]).

Studies of the transfer mechanisms in the hemispheric scale have been centered on the existence and role of residual circulation cells in the meridional planes, as compared to eddy motions on the scale of cyclones and anticyclones in horizontal planes. The existence of a mean meridional circulation cell over the Northern Hemisphere tropical belt in winter has by now been definitely established by Riehl and Yeh [39], Priestley [31], Palmén [27], Tucker [48, 49], and Palmén, Riehl, and Vuorela [29]. This cell consists of equatorward motion in the trade regime at low levels, ascent to the high troposphere along the equatorial trough, and a return poleward current near 200 mb. (40,000 feet), followed by descent in the subtropics. The most recent evaluations (Palmén et al. [29]) indicate that the circulation cell of winter can account for heat transport from the equator to the subtropics of the right order of magnitude to satisfy continuity requirements. Riehl and Malkus [38] have shown that the structure and energy properties of the equatorial trough are compatible with that of a simple energy-producing heat engine, with warm ascent in the core of the trough and descent of denser air at some distance, as demanded by the meridional cell machinery.

Since the main source of energy enters the atmosphere by way of the tropical oceans and the trades, it is of the utmost importance to investigate the structure of the circulations and mechanisms for energy transformation and balance along the oceanic trade wind belt. Studies of this nature are usually hampered by the scarcity of adequate observations, and the opportunities for research are limited. With few exceptions the most productive efforts have resulted from observational material gathered during the course of special expeditions, notably by the German vessel Meteor in 1924-26, and by various groups from the Woods Hole Oceanographic Institution in the last 15 years (for references see Riehl [35]; also various publications by Dr. J. Malkus and collaborators since 1954). Apart from expeditions, detailed investigations in the oceanic regions of the trades must wait for suitable opportunities in which data become available under conditions favorable for study. Riehl et al. [41] made use of such an opportunity when surface and aerological observations were made at four oceanic stations along the northeast trades of the Pacific ocean, which, by fortunate coincidence, fell along the trajectory of the flow. They carried out a comprehensive study of the structure and energy balance of the lower layers for a summer period. Their results, in essence, confirmed the role of the trades as energy accumulators and exporters, that had been deduced from the initial studies (Ficker [10]). Riehl and Malkus [37] re-evaluated the energy balance along the Pacific cross-section and emphasized the role of the sensible heat exchange with the ocean surface in the maintenance of the local circulation.

The present report is also concerned with the energy transformations over a sector of the oceanic trade wind region - the Caribbean Sea; and, in some respects, it may be considered as a sequel to the work of Riehl and collaborators on the Pacific trades. Use is made here of the rather convenient location of a group of aerological stations around the periphery of the Caribbean Sea (fig. 1) in order to evaluate the energy balance of the tropospheric volume above this oceanic basin. This project differs from previous attempts in various respects. It covers the entire tropospheric layer and gives a view of the relationships between the trades and the upper-level wind flow. In addition, the availability of oceanographic data has made possible a computation of the heat budget of the water body of the Caribbean Sea, which allows us to associate the balance relationships between the ocean surface and atmosphere. The main purpose of this report is to present the results of these analyses. On the basis of the computations, a discussion is given of the processes by which the energy received from the ocean is distributed by the Caribbean atmosphere to maintain the local circulation, offset the radiation losses inside the area, and provide for export to other regions of the globe. The Caribbean Sea is one of the few regions around the hemisphere where atmospheric and oceanographic data are available in a form and quantity adequate for a study of this type.* The atmospheric research is based on wintertime conditions as depicted by monthly mean data for December 1956 and January 1957; the oceanic study is based on long-term climatological observations.

Interest in the Caribbean region arises, not only because of the developments in the lower layers, but also on account of the upper level circulation regime. The tropospheric flow during the period of study (figs. 3-6) is characterized by a two-layer atmosphere with a steady current in the trade layers moving from east-northeast to west-southwest, and a much stronger westerly current of equatorial origin, centered near 200 mb., blowing in the opposite direction. These equatorial westerlies aloft have been noticed previously (Colon [8]; Hubert and Dagel [16]), and appear to be a characteristic feature of the normal wintertime conditions over this area. There are reasons to believe that this wind regime is related to the return poleward branch of the

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*After completion of this report two papers by S. Manabe [22,23], in which he treated the energy budget of both the atmosphere and water body of the Japan Sea during wintertime regimes by use of methods somewhat similar to those adopted here, came to the attention of the writer.

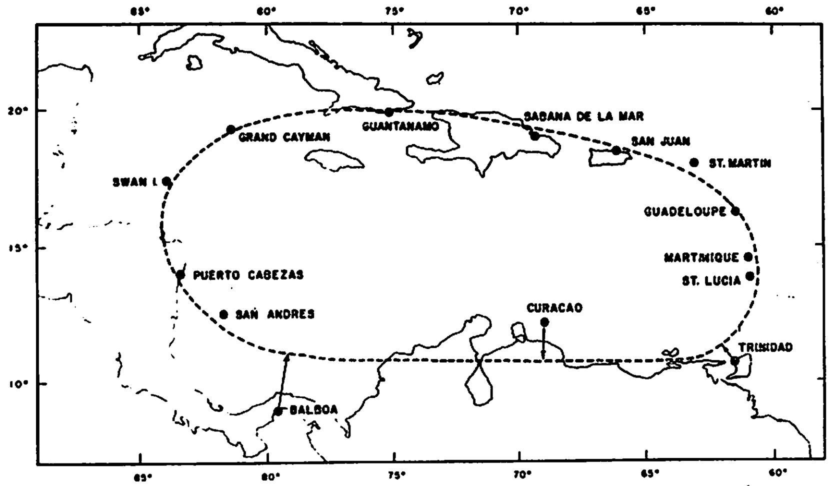


Figure 1. - Location of stations and boundary (dashed curve) used in the evaluation of energy flux over the Caribbean Sea.

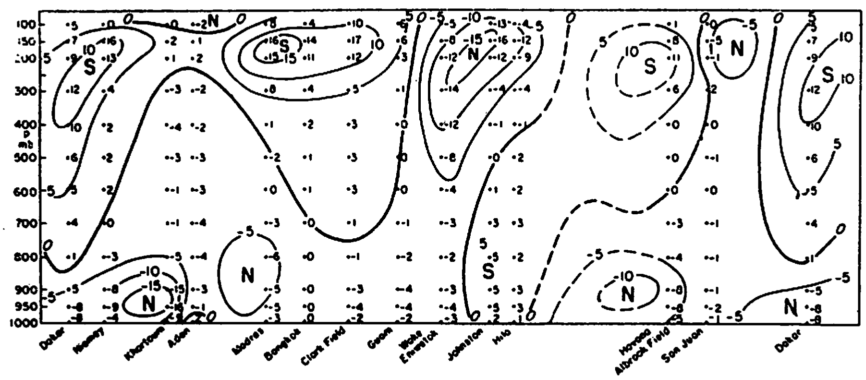


Figure 2. - Regional distribution of the meridional wind component (kt.) around latitude 15°N., December 1955 - February 1956. Dashed lines indicate interpolation in eastern Pacific. (After Palmén et al. [29]).

tropical meridional cell; and, therefore, it plays a significant role in the mechanism for the transport of heat out of the equatorial regions. Recent studies of the mean upper flow over the tropical belt corroborate this idea. Palmén et al. [29] analyzed the distribution of the mean meridional wind component at latitude 15°N. for the period December 1955 to February 1956 (fig.2). Three main channels for northward motion are apparent in the high troposphere with maximum intensity near 200 mb.: (1) over the eastern Pacific Ocean and Caribbean Sea area, (2) over western Africa, and (3) over Southeast Asia. These three regions are related to wave-like undulations of the subtropical jet stream of winter, as revealed by Krishnamurti [19]. Krishnamurti showed that this jet stream is a semi-permanent feature of the subtropical atmosphere during winter. It is centered generally between latitudes 20° and 35° N., with mean position near latitude 27°N. Maximum kinetic energy is concentrated near 200 mb. Around the hemisphere a three-wave pattern is apparent, with troughs located over the eastern Pacific Ocean, the west coast of Africa, and India; ridges are situated over the southeastern United States,East Africa,and

Japan. These studies indicate that the equatorial outflow aloft toward the subtropics does not occur uniformly around the globe, but takes place in a few narrow channels, as suggested by Riehl [34]. One of these preferred channels evidently lies over the western Caribbean Sea and adjacent region of the Pacific Ocean. It is of interest to see in figure 2 that the equatorward flow at low levels also shows concentration in certain longitudinal regions; one of them lies over the western Caribbean Sea. In the present study we shall take a close look at this local circulation, which seems to be of great significance in the evolution of the general circulation.

The work consists essentially of a volume integration of the energy equations, and involves evaluation of energy flux by line integrals around the lateral boundary. For this purpose it is necessary to define a representative boundary around the Caribbean Sea that takes account of the station locations. A quasi-elliptical boundary (fig. 1) defined by the positions of 14 aerological stations around the periphery (12 radiosonde-rawin and 2 pibal stations) has been chosen. The upper and lower volume boundaries are, respectively, the tropical tropopause, located in the mean near 100 mb., and the sea surface. A similar scheme for the study of energy processes over a limited region was used by Riehl [33] over the Gulf of Mexico, and Gangopadhyaya and Riehl [14] over the Bahamas area.

The main source of heat for the atmospheric balance is the flux from the ocean. As mentioned previously, this flux is mostly in the form of latent heat, which is a form of energy inherent in the water vapor because of its physical state, and it is not available for heating the atmosphere, unless condensation and precipitation occur. After condensation and precipitation, a certain amount of heat, usually referred to as precipitation heating, becomes available to the atmosphere to offset in part the infrared radiational losses, which constitute the main heat sink. The flux through the lateral boundary may be a source or a sink depending on the particular properties of the wind flow and energy content. In evaluations of atmospheric energy, as carried out here, it is advisable to separate the latent heat (Lq) and the sensible heat (composed of the sum of potential energy and enthalpy ($gz + c_pT$). This is convenient in the computations, and in addition helps to emphasize the magnitude and role of the latent heat.

The kinetic energy also requires a separate analysis, due to the fact that the magnitudes involved in the kinetic energy balance are negligibly small when compared to those of the other forms of energy. This is one of the paradoxical aspects of atmospheric energy, since the kinetic is usually the one in which we are most interested. A separate evaluation of the balance of kinetic energy illustrates the processes of production and distribution over the Caribbean atmosphere. This analysis is of particular interest when viewed in the light of the possible contribution of the equatorial westerly current aloft to the maintenance of the subtropical jet stream.

In the evaluation of the energy flux from ocean to atmosphere, we come to a very important and difficult problem. It has received considerable attention in the past, but, unfortunately, a final solution is not yet available. Methods for evaluation have followed two general lines of attack: one, by measurements of evaporation in pans at land stations and aboard ships, with empirical

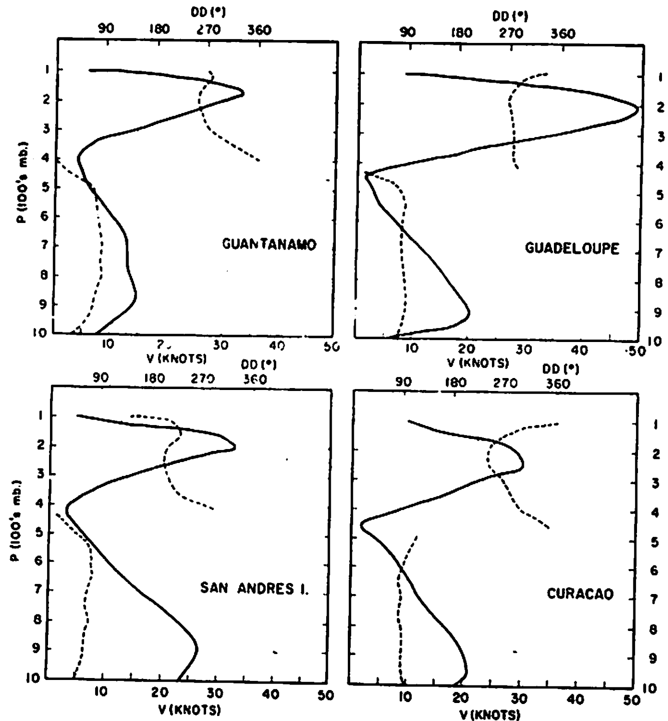


Figure 3. - Vertical profile of the resultant winds at the indicated stations for December 1956.

expressions for their relation to conditions over the open seas; and two, by theoretical development of turbulent diffusion near the surface to obtain formulae in terms of wind, temperature, and humidity measurements made routinely aboard ships. The formulae now available for computation are discussed later. This subject can be approached also from the point of view of the oceanic heat balance. Energy flux to atmosphere can be obtained as a residual in a computation of the energy budget of the ocean, provided that the other terms are determined satisfactorily. The Caribbean Sea offers several advantages for a study of this type. It has definite boundaries, with a very simple oceanic current, and the amount of observational data is generally adequate. In addition to being a natural complement to the atmospheric study, this work has value from the oceanographic point of view. Previous studies concerned with the heat balance of the oceans (Jacobs [17]; Gabites [13]; Pattullo [20]; Fritz [11]) dealt more with the hemispheric picture than with the local conditions over an enclosed basin like the Caribbean Sea. The evaluations have been made on a monthly basis thus giving a rather complete view of the seasonal variations in the energy flux.

a. <u>OBSERVATIONS USED IN STUDY:</u> Until 1956 the rawin stations at Balboa, Canal Zone, and at San Juan, Puerto Rico, were the key stations that supplied information on the equatorial currents aloft over the Caribbean. Since the spring of 1956, through the efforts of the National Hurricane Research Project, a series of additional radiosonde and rawinsonde stations has come into operation. The data gathered by the expanded network (cf. fig. 4) give a more definite picture of the circulation than could be obtained previously, and, in addition, make possible a fairly accurate evaluation of the energy flux into and out of the Caribbean Sea.

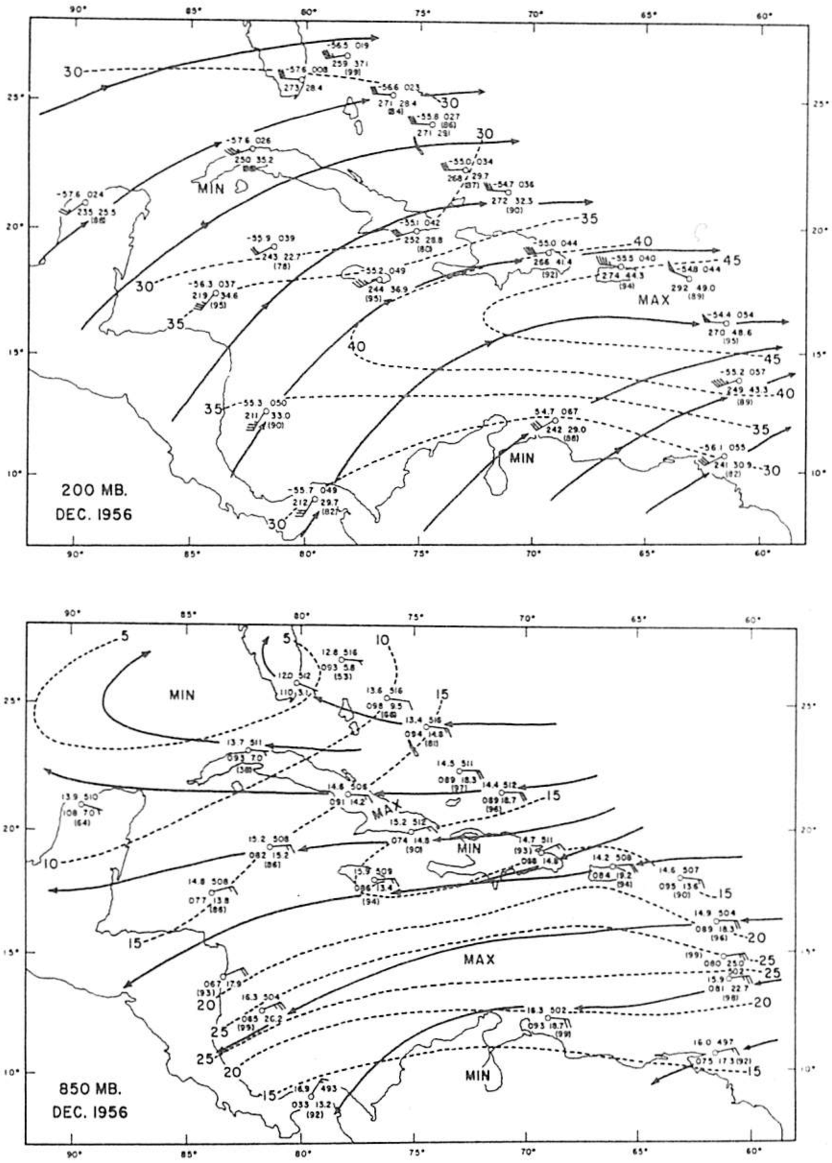


Figure 4. - Mean flow, December 1956. Solid lines are streamlines, dashed lines isotachs. Station symbols are as follows: at the top, temperature (°C.) and contour heights (10's of feet, first digit omitted); at the bottom, resultant wind direction (degrees) and speed (knots). Number in parentheses stands for wind steadiness in percent.

The first two winter months after the establishment of the new stations - December 1956 and January 1957 - were selected for study. Aerological data in the form of monthly averages of temperature, relative humidity, contour heights of isobaric levels, and monthly resultant winds were obtained at 50-mb. intervals from the surface to the top of the soundings for all stations in the area. These data were processed by machine methods at the National Weather Records Center, Asheville, N. C. Only the observations taken at 0300 GMT (10:00-11:00 p.m. LST) were used in the study, except in the case of the data for San Andres in December 1956, and for Curacao, which were available only once a day at 1500 GMT. Tabulations of the 1500 GMT observations at all other stations were also prepared and used occasionally as reference.

The data for the stations around the boundary, which entered directly in the evaluations of mass and energy flux, were examined in great detail. All parameters were plotted in vertical profile form and examined for homogeneity and internal consistency. As is usually the case, the number of observations in the monthly samples decreased with height; the decrease at some stations was quite noticeable above 200 mb. Nevertheless, a representative amount of data, completely adequate to define the character of the flow, was available in all cases.

The mean heights of the isobaric levels for most stations around the boundary were re-evaluated from the mean monthly soundings. These heights entered in the measurement of potential energy. The new values differed little from those obtained from the average of the daily heights, but generally allowed for a more consistent analysis. The contour heights obtained at Trinidad, and to a lesser extent those at Balboa, still appeared doubtful in that they were generally lower than demanded by the geostrophic relationship in the vicinity. For example, values at Trinidad at high levels combined with those farther north at St. Lucia showed in general less gradient than the winds would call for. The data for 1500 GMT were checked and showed the same tendency. This possible bias in the contour heights was critical only in computations such as geostrophic wind, which depend on small differences in the actual values at a given level. In the evaluations of energy content and and energy flux the computational procedure (to be described later) was such that possible inaccuracies or bias in the observations at any one station, either in height, temperature, wind or moisture, did not impair the final results.

In addition to the aerological data a representative sample of surface ships' observations over the Caribbean Sea for December 1956 and January 1957 was obtained from Data Tabulations published by the U. S. Weather Bureau. A total of over 500 observations, taken at 1200 GMT (0700-0800 LST) of surface winds, air temperature, water temperature, and dewpoint was available for each month. These data were utilized as an analysis aid in the lower layers, and in the computation of the energy flux from ocean to atmosphere.

In the evaluation of the oceanic heat balance and in some aspects of the atmospheric balance, data from climatological sources were used. The main part of the oceanic work, namely, the evaluation of the heat storage by the water body, was based on the analysis of a collection of bathythermograph data compiled at the Woods Hole Oceanographic Institution. Publications consulted frequently were the Marine Climatic Atlas of the World - Atlantic Ocean (U. S. Navy [50]); Climatic Charts of the Ocean (U.S.Weather Bureau [51]), and papers by Fuglister [12] and London [21]. Other sources of data used occasionally are mentioned later in the text.

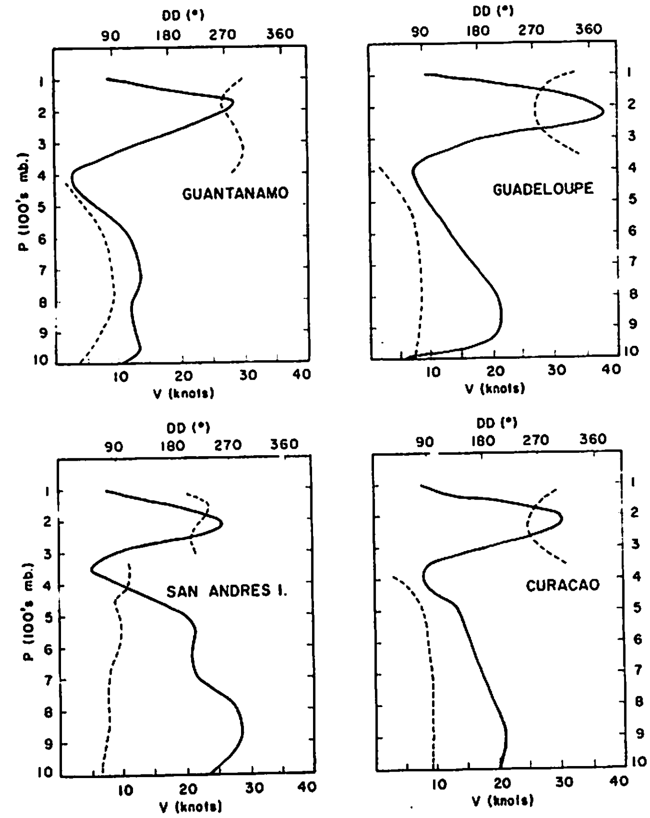


Figure 5. - Vertical profile of the resultant wind at the indicated stations for January 1957.

2. MEAN TROPOSPHERIC CIRCULATION FOR DECEMBER 1956 AND JANUARY 1957

The mean flow over the Caribbean area for December 1956 and January 1957 is illustrated in figures 3-6. Figures 3 and 5 show the vertical distribution of the resultant wind for selected stations, two on each side of the Caribbean Sea; figures 4 and 6 show the horizontal flow at two selected levels: 850 mb. and 200 mb.

The vertical wind distribution illustrates a two-layer atmosphere with lower easterlies and upper westerlies. The trade current is essentially from east-northeast at all stations with greatest strength of 20 to 25 knots at 900 mb. In the high troposphere, the flow is mainly from southwest at San Andres Island and Curacao, and more westerly at the northern stations. A striking velocity concentration exists near 200 mb. with large vertical shear above and below the level of maximum wind. Between 600 and 300 mb. the flow is generally light; this layer marks the transition from the lower easterly to the upper westerly flow regime.

On the average the base of the tropopause lies near 100 mb. The circulation above this level will not be discussed. It is characterized by a renewed shift in wind direction to east; speeds increase with height in the lower

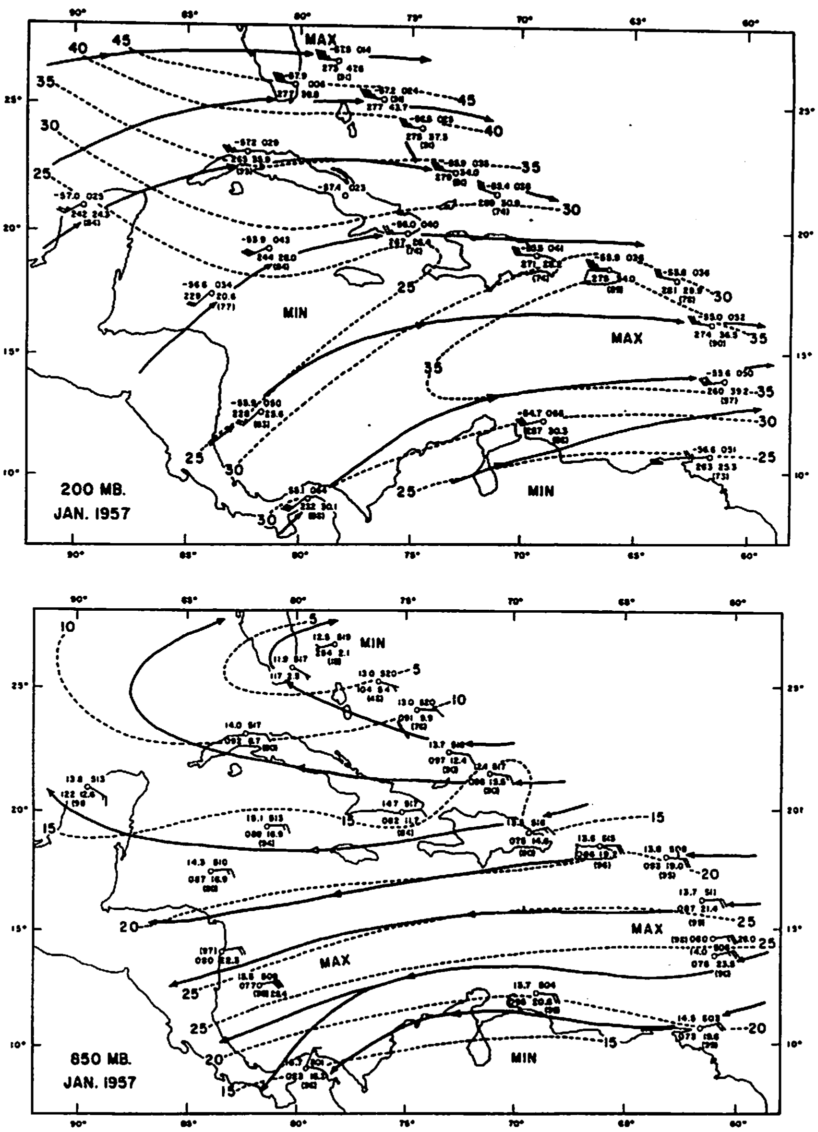


Figure 6. - Mean flow January 1957. Symbols as in figure 4.

stratosphere. Thus, in spite of the season, the so-called Krakatao easterlies are found. They have been discussed extensively in the literature (for references, see Riehl [35].

The horizontal flow at all levels from the surface to about 700 mb. shows essentially the same features as figures 4 and 6 (850 mb.). Easterly flow predominates over the entire area. The isotach field contains a center of relatively high speeds oriented from east-northeast to west-southwest across the Caribbean. A secondary wind maximum extends from northeast to southwest over central Cuba, probably associated with the mean position of the polar front at that time of year. Cold fronts moving southeastward across the Bahamas usually become quasi-stationary with an E-W or ENE-WSW orientation north of Hispaniola. Strong winds occur along and to the rear of these fronts, and this may be reflected in the monthly wind averages.

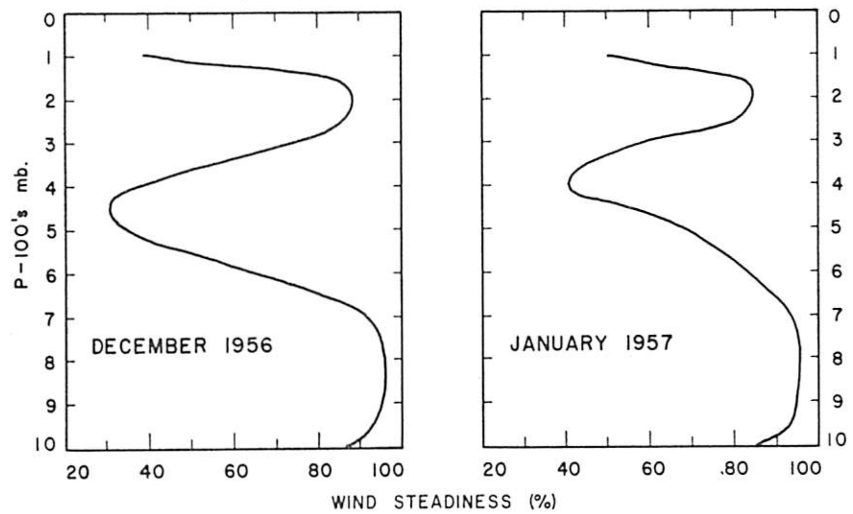


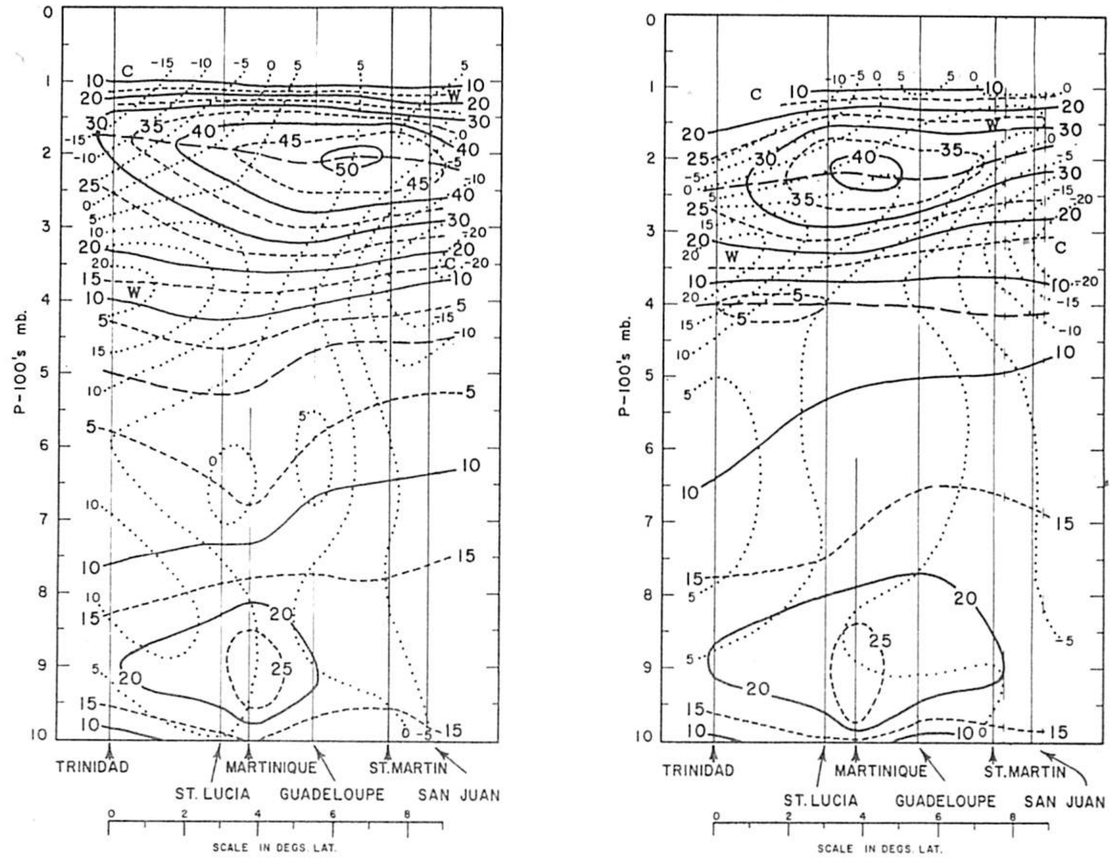
Figure 7. - Vertical distribution of wind steadiness (percent) averaged for the four stations shown in figures 3 and 5.

It is not possible to determine the precise shape of the isotach field over the central Caribbean, but probably the wind maximum observed over the Lesser Antilles continues to the western border. This supposition is supported by surface wind data over the water (not illustrated), which show a definite tendency for highest speeds to occur in a belt oriented from east to west across the central Caribbean. This strong wind belt can be followed across Nicaragua to the eastern Pacific Ocean south of Central America. Such banding of the trades is also found in long-term climatic means presented in the Atlas of Climatological Charts of the Oceans (U. S. Weather Bureau [51]).

The flow at 200 mb. (figs. 4,6) shows a southwesterly current coming out of the equatorial Pacific and South America and turning clockwise over the Caribbean. A band of highest speeds is present, oriented WSW to ENE; along this band the wind speed increases downstream to about 50 knots over the Lesser Antilles. Wind speeds decrease laterally toward South America and toward the Greater Antilles, with a region of minimum speeds over Cuba. Still farther north, wind speeds again increase toward the mean position of the subtropical jet stream and the higher latitudes.

The major features for January 1957 (figs. 5-6) differed very little from those for December 1956. The intensity of the upper current was somewhat less and it was centered a little farther south than in December.

a. STEADINESS OF THE FLOW: Wind steadiness was computed with the formula $S = V/\overline{V}$ (in percent) where V is the monthly vector resultant wind and $\overline{V}$ the average wind speed irrespective of direction. This formula measures only directional stability of the winds and has considerable shortcomings, as described, for instance, by Riehl [35]. Nevertheless, the fact that very high values were obtained (figs. 4, 6, 7) is informative. Steadiness was near 96



A
December 1956

B
January 1957

Figure 8. - Space cross-sections perpendicular to the upper flow along the Lesser Antilles. Isotachs in solid and short-dashed lines; heavy long-dashed lines show the axes of maximum and minimum speeds. Temperature anomalies (tenths of °C.) from the mean at each isobaric level in dotted curves; C stands for colder, W for warmer.

percent in the trade current during both months. The vertical profile, with a pronounced minimum of steadiness in the transition zone between lower easterlies and upper westerlies, reflects mainly the course of the resultant speed. Near 200 mb. the wind direction was subject only to very small time fluctuations, while at 100 mb. the wind direction oscilliated again more strongly in the transition zone toward the stratospheric circulation.

b. <u>VERTICAL SPACE-SECTION ACROSS THE CURRENT:</u> Figure 8 is a vertical cross section through the eastern Caribbean, oriented normal to the upper current. It illustrates the vertical and horizontal structure of the monthly wind field. Maximum speeds above 25 knots are observed in the trades. The upper current, with central speeds above 50 knots, is centered slightly north of the trade wind maximum. During January 1957 (fig. 8B) the two cores are situated nearly along the vertical.

Figure 8 (A and B) also indicates the lateral temperature distribution. In the lower half of the atmosphere a normal meridional temperature gradient

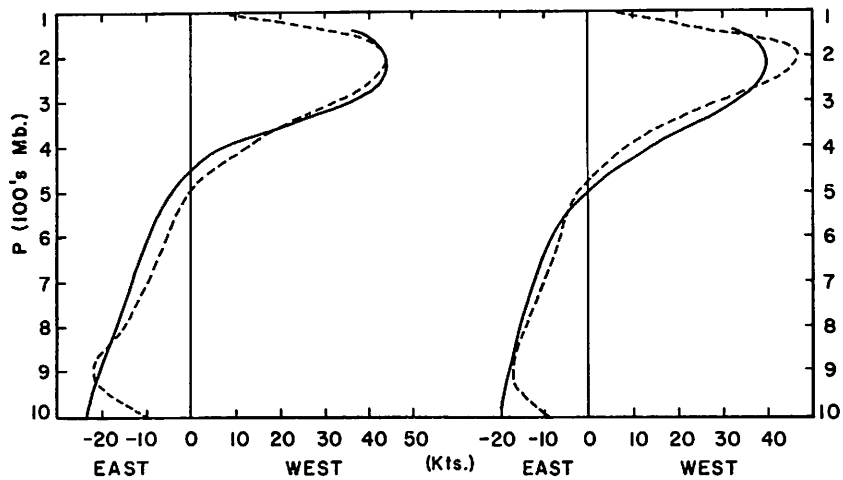


Figure 9. - Comparison of geostrophic winds (solid curve) and actual winds (dashed) between St. Lucia and Guadeloupe, on the left; and between Guadeloupe and St. Martin, on the right; for cross-sectional data of figure 8A. Speeds in knots.

is evident. Above 500 mb. there is a drastic change. At first the gradient increases rapidly upward to a zone of maximum near 400 mb., where the southern end of the section is about 4°C. warmer than the northern end. This cannot be explained on the basis of radiation. Between 400 and 200 mb. the gradient decreases again until, at about 200 mb., it reverses. Above this level it is warmest in the north, and this accords again with previous findings on the temperature field above the level of maximum wind (Riehl and Malkus [38]; Krishnamurti [19]).

From the geostrophic thermal wind relationship the observed meridional temperature gradient calls for a decrease of east wind, or increase of west wind, with height throughout the troposphere to 200 mb., and the reverse above. The geostrophic wind normal to the section was evaluated by taking the height difference between stations from Trinidad to St. Martin. The average actual wind (u component) between stations was used for comparison. The agreement was quite good in the sector from St. Lucia to St. Martin (fig. 9). The computed and actual wind are about equal throughout the troposphere; the differences observed at some levels are within the limits of accuracy of the data. The only zone where the shear of the two winds differs is near the surface, where the trades increase upward to 900 mb., while the geostrophic wind decreases. This increase in the actual winds, noted previously in all studies of the trades, may be explained readily as due to the effect of the frictional drag on the earth's surface against the trade flow.

In the sector between Trinidad and St. Lucia (not illustrated) the agreement was poor, particularly at high levels. The computed geostrophic wind between 300 and 150 mb. was only about one-fifth of the actual wind. The difficulty lies essentially in the temperatures at Trinidad and St. Lucia, which show very small differences above 400 mb.; therefore, small values are computed for the thermal shear. The observations for 1500 GMT were examined

Table 1. - Percentage frequency distribution of wind direction and speed at the 200-mb. level during winter (Dec., Jan., Feb.) at Balboa, C. Z. A 5-year summary taken from [50].

	Wind speed (knots)					
	0-9	10-29	30-49	50-74	75-99	Total
N		+	1			1+
NE		3				3
E		2	1	+		3+
SE		5	3	+		8+
S		10	11	1	+	22+
SW		10	15	8	1	34
W		6	11	3	1	21
NW		1	1			2
Total	6	37+	43	13	2+	100

and found to show the same variations. The evaluation of the January data is similar to that of December in all respects.

An attempt to prepare similar cross-sections for the western side of the Caribbean was made, but it was not successful due to the unfavorable location and spacing of stations. However, as seen from the horizontal charts and vertical velocity profiles, the structure there has the same characteristic features as in figure 8, except that wind speeds are lower and less concentrated laterally.

c. COMPARISON WITH NORMAL DATA: Inspection of wind summaries for longer periods indicates that the flow patterns for December 1956 and January 1957 are representative of normal winter conditions over the area. Summaries of the wind flow at selected levels over the Caribbean for all seasons are presented in the Marine Climatic Atlas of the World (U.S.Navy [50]). The picture for the winter season at 200 mb. is essentially the same as illustrated here. Table 1 presents the distribution of 200-mb. winds at Balboa for December, January, and February 1946 to 1954, extracted from the above publication. Winds from the southwest quadrant predominate; over 56 percent of the observations are from south and southwest. Speeds above 50 knots, mostly from southwest, are recorded about 14 percent of the time; readings above 75 knots also occur. Summaries for San Juan (not presented here) show a predominance of westerly winds, with extreme readings above 100 knots. These statistics suggest a climatic flow pattern similar to that of figures 4 and 6.

3. HEAT BUDGET OF THE CARIBBEAN SEA

As discussed previously, the main object in the study of the oceanic heat balance is to evaluate the heat flux from the ocean to the atmosphere. This flux can be computed also with transfer formulae expressed in terms of meteorological variables in the vicinity of the ocean surface. The formulae that have been proposed and used in the past, and results obtained through their application, will be discussed at the end of this section. The oceanic heat analysis serves also to illustrate the processes of energy absorption and distribution operating in the oceans and should be of interest also from the oceanographic viewpoint.

Let Q_a be the total heat flux, Q_e the latent, and Q_s the sensible heat flux from sea to air. Then,

$$Q_a = Q_e + Q_s.$$

The condition of heat balance for an oceanic body is expressed by

$$Q_e + Q_s = Q_r - Q_v - Q_t, \qquad (1)$$

where Q_r is the net absorption of radiation by the water, Q_v is the divergence of heat transport by the water currents, and Q_t is the time rate of change of the heat content of the water. This term vanishes when annual balance is considered, since the annual mean temperature of the oceans remains essentially constant. On a monthly or seasonal basis, however, the heat storage can become important, and changes sign seasonally. Q_v is usually considered to be small, although in some oceanic areas, like the North Atlantic Ocean, it makes a significant contribution to the meridional transfer of heat (Sverdrup et al. [47]).

In the following sections the different terms in equation (1) will be evaluated for the Caribbean Sea on a monthly basis. In some respects this water body is favorable for this purpose. It is relatively small, with definite boundaries, and contains one well-defined water current with entrance in the Lesser Antilles and exit in the Yucatan Channel. The Caribbean has always attracted the attention of oceanographers. Various expeditions have combed the waters there to make observations. As a result, the accumulation of data and knowledge is as great, perhaps greater, than for any other tropical oceanic area. In spite of this, the data are not adequate for a rigorous evaluation of all terms in the heat balance equation. Several quantities, particularly those connected with the radiation, had to be adapted from what has been worked out on a hemispheric basis. Further, even in the computations for which the amount of observations is large, the accuracy is such that the results have to be viewed with restraint. Hence precise balances are not to be expected.

a. <u>NET ABSORPTION OF RADIATION</u> (Q_r): The incoming radiation received and absorbed by the earth's surface has been evaluated for different latitudinal belts and seasons by various writers (cf. Simpson [44], Baur and Phillips [3], Houghton [15], London [21]). One of the most recent and thorough treatments of the problem is that by London [21], whose results are used extensively in the present study. The radiation absorbed at the earth's surface consists of two parts: absorption of short-wave insolation, and long-wave radiational exchange with the atmosphere. Most short-wave radiation reaching the sea surface is absorbed, and constitutes the primary source of energy. The long-wave exchange results in a net flux to the atmosphere.

The evaluation of the radiation absorbed by the Caribbean Sea is outlined in table 2. Monthly values of the incoming radiation at the top of the atmosphere for the latitude belt 10° - 20° N. were adapted from the Smithsonian Tables (List [20]). These are based on a solar constant of 1.94 ly. $min.^{-1}$, and were adjusted slightly to conform to a value of 2.00 ly. $min.^{-1}$ suggested by more recent determinations (Johnson [18]). Solar radiation is depleted in the atmosphere on account of absorption, scattering reflection, etc. by the atmospheric constituents including clouds. Monthly values of net percent transmissivity of the atmosphere under mean cloudiness conditions were adjusted from London's seasonal results; the insolation reaching the surface of the earth was then determined (line 3, table 2).

Absorption of insolation by the surface depends essentially on its composition. For an average oceanic surface the absorptivity has been estimated at about 94 percent, or the albedo at 6 percent, except at very low solar altitudes (Houghton [15]). The amount of radiation absorbed by the Caribbean Sea was calculated (line 4, table 2) on the basis of a 6 percent albedo constant throughout the year. The results are valid not only for the Caribbean, but also for any oceanic area around the hemisphere in the belt 10°- 20°N. The only parameter discussed so far that is subject to longitudinal differences around the hemisphere is the atmospheric transmissivity, which may vary on account of differences in the distribution of mean cloudiness, water vapor, and ozone content. Regional variations in mean cloudiness probably are more important than those of other factors. However, the seasonal course of cloudiness in the latitude belt 10°- 20°N. was considered to be sufficiently homogeneous to warrant the use of London's transmissivity coefficients.

The long-wave radiational exchange between the surface of the earth and the atmosphere, also known as the back or nocturnal radiation, depends on the temperature and radiative properties of the underlying surface, as well as on the amount of water vapor and clouds present in the atmosphere. Sverdrup et al. [47] published a graph for estimating the back radiation from an oceanic surface under clear skies as a function of water temperature and relative humidity of the air at ship's deck level. Computations were made from this graph using the mean temperatures and humidities of the Caribbean.

In the presence of clouds this back radiation is reduced. The magnitudes for cloudy skies (Q_b) were obtained with use of the relation (Sverdrup et al. [47]): $Q_b = Q_{bo}(1 - 0.83C)$, where Q_{bo} is the radiation for clear skies and C is the mean cloudiness on a scale from 0 to 1. Monthly values of cloudiness were adopted on the basis of the average seasonal cloudiness around the globe in the belt 10° - 20°N., published by London [21]. These are somewhat lower than those of Brooks [4]. They ranged from about 50 percent in summer to about 35 percent in winter. The computed back radiation for cloudy skies is given in line 7, table 2. The net radiation absorbed is then the difference between

Table 2. - Evaluation of radiation absorbed by water body of the Caribbean Sea (latitude 10°- 20°N.).
Units: ly. day^{-1}.

	J	F	M	A	M	J	J	A	S	O	N	D	Ann
1. Radiation at top of atmosphere (List, 1951) [20]	716	797	874	926	936	933	931	926	893	824	744	692	850
2. Percent transmissivity of atmosphere (London [21])	59	59	58	57	56	53	51	51	52	53	55	57	55
3. Radiation reaching the surface	422	470	507	528	524	494	475	472	464	437	409	394	467
4. Radiation absorbed by water (6 percent albedo)	397	442	477	496	493	464	446	444	436	411	384	370	439
5. Back radiation (clear skies)	242	242	242	240	238	236	236	236	236	236	238	240	238
6. Mean percent cloudiness (London, 1957) [21]	36	35	38	42	45	47	48	50	50	48	44	40	44
7. Effective back radiation	170	172	166	156	149	144	142	138	138	142	151	160	151
8. Net absorbed radiation	227	270	311	340	344	320	304	306	298	269	233	210	288
9. Total radiation absorbed by Caribbean Sea (Unit: 10^{18} cal. day^{-1}	4.97	5.91	6.81	7.45	7.53	7.01	6.66	6.70	6.52	5.89	5.10	4.60	6.31

the absorbed short-wave radiation and the back radiation emitted (line 8, table 2).

The absorbed short-wave radiation (line 4, table 2) is greatest in April and May, and least in December. The range between the largest and smallest monthly values is about 30 percent of the annual mean. Seasonal differences result mainly from variations in the altitude of the sun and the mean cloudiness. Of these, the latter is especially important in the warm season. During June, July, and August, the altitude of the sun is about the same as in April and May, but the cloudiness is much greater.

The back radiation for clear skies (line 5, table 2) is relatively constant throughout the year. A slight seasonal variation is apparent for cloudy skies (line 7, table 2) with maximum in winter and minimum in summer; that is, opposite to the march of cloudiness. The seasonal variations in net absorbed radiation (line 8, table 2) depend essentially on the short-wave absorption. It ranges from about 340 ly.day $^{-1}$ in April-May to 210 ly. day^{-1} in December. The range is quite large; about 42 percent of the annual mean. It may be noted that the values computed in table 2 differ only slightly from those given by London [21] for the whole latitude belt 10°- 20° N.

The rate of heat gained by radiation over the entire Caribbean Sea (Q_r) is obtained by multiplying line 8 of table 2 by the area of the region considered, in our case the area enclosed by the ellipse of figure 1 (2.19×10^{16} cm.^{2}).

b. SEASONAL ENERGY STORAGE IN THE WATER (Q_t): Evaluation of seasonal changes in heat content within the water body requires knowledge of the variation of water temperature down to the depth where an annual temperature cycle is no longer noticeable. The monthly variations of surface water temperature are known to a fair degree of accuracy; data for the Caribbean are available in tabulated averages for 5° latitude-longitude squares (U.S.Weather Bureau [51]) and in the form of isotherm maps (Fuglister [12]). Both of these publications used the same original source of data. The method of analysis used by Fuglister involved a more refined smoothing of the inhomogeneities in the data; after some consideration, an area average over the entire Caribbean obtained from his maps was adopted here. The seasonal curve of surface-water temperatures (fig. 10) shows a maximum of 83.2°F. in September and a minimum of 78.6°F. in February; the annual range is 4.6°F. The data in the Climatic Charts of the Ocean give essentially the same annual variations with an annual range of 4.4°F.

At lower depths, temperature measurements are scarce. Until recently, data of this nature were obtained only in the course of expeditions, which were conducted mostly during the spring months. During the last two decades, however, depth soundings with bathythermograph have been taken more frequently, and data for all months of the year have accumulated gradually. Depth soundings for the Caribbean now exist in a number sufficient for reasonably accurate conclusions about the temperature variations below the surface. A tabulation of monthly temperature soundings, averaged for one-degree squares

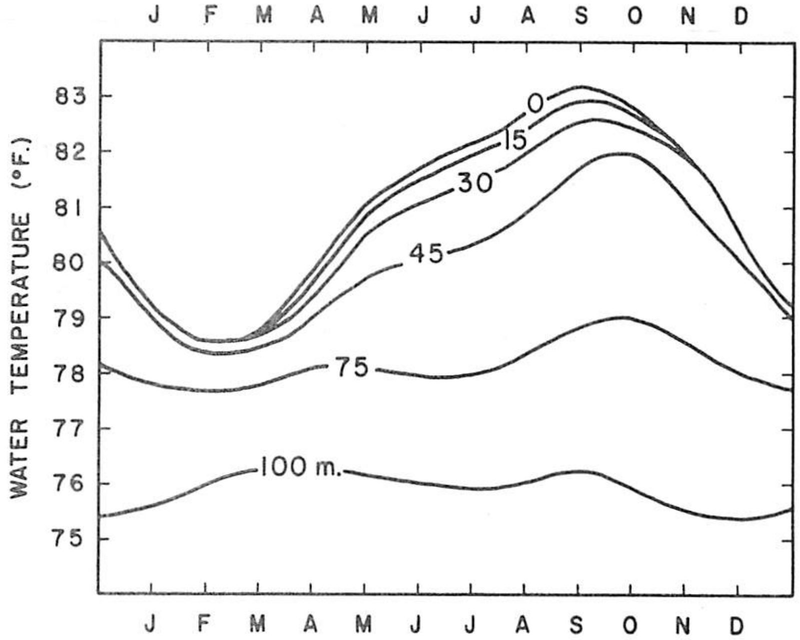


Figure 10. - Annual variation of water temperatures at the indicated levels (meters) averaged over the Caribbean Sea.

of the Caribbean, has been compiled at the Woods Hole Oceanographic Institution. This tabulation was obtained* and analyzed.

About 8,000 soundings were available in the sample. The number of observations was greatest during late winter and spring, with over 1,000 soundings in March and April, and least during summer and early fall, with only about 200 soundings in August. Observations were not evenly distributed over the Caribbean; the great majority was taken in the north-central portion. There was a good sample also from the eastern part, but very few from the southwestern and western sections. The number of observations decreased with depth; below 100 meters this decrease was very rapid.

The temperature data were analyzed in various ways in order to obtain the best possible representation and to smooth out inhomogeneities. None of the approaches chosen, however, proved entirely satisfactory. Finally, all observations were averaged at each level without area weighting and plotted on graph paper, then curves were drawn and smoothed by hand. Fortunately, very little smoothing was required. Only one adjustment proved necessary. The surface temperatures obtained from the Woods Hole sample indicated warmer temperatures in summer (when the amount of data was least), and a greater seasonal range at the surface (5.9°F.) than those shown by Fuglister's normal charts and by the Climatic Charts of the Ocean. Probably the sample is not entirely representative of average conditions. The monthly range at the different levels was adjusted proportionately so as to make the surface temperature variation

* Courtesy of Dr. Joanne S. Malkus, Woods Hole Oceanographic Institution.

equal to that adopted from the Fuglister charts. Results are shown in figure 10. The annual range, 4.6° F. at the surface, remains essentially unchanged at a depth of 15 m.; at 30 m. it is reduced to about 87 percent of the surface value; at 45 m. to about 76 percent; and at 75 m. to 26 percent. At a depth of 100 m., a reliable seasonal cycle could no longer be detected; this was perhaps largely due to the rapid decrease in sounding frequency and a consequent increase in the inhomogeneity of the data. At the next depth for which temperatures were tabulated, 150 m., the number of observations was only 30 percent of that at the surface.

In addition to the reduction in seasonal range, the temperature variation with depth was characterized by a seasonal lag in the time of maximum temperature. At a depth of 45 m. and lower, the highest temperature was recorded in October. A similar lag was not evident for the minimum temperature.

From the foregoing, only the temperature cycle in the upper 100 meters was considered necessary for computing Q_t; this can be considered to be given by figure 10. Actually, it was assumed that the seasonal variation becomes zero at a depth of 90 m. At this level the seasonal cycle was found to be smallest from plots of all monthly soundings in vertical profile form. Lower down the cycle, if anything, changes phase. Average monthly temperatures for the layer 0-90 m. were determined from the profile for each month by the method of equal areas. The annual range of mean temperature averaged over this layer was 2.7° F., with highest temperature in September - October and lowest temperature in February.

The rate of change of heat content in the oceanic layer is given by the relation

$$Q_t = c\int_{\alpha} \frac{\partial T_w}{\partial t} \rho_w \, d\alpha \qquad (2)$$

where c is the specific heat of water, ρ_w its density, and α denotes volume. Both c and ρ_w were taken as unity. The local temperature change, $\frac{\partial T_w}{\partial t}$, was approximated by taking 2-month overlapping temperature differences centered at each month. The volume is given by the product of the area of the ellipse (fig. 1) and the depth of the layer.

The rate of change of heat content (table 3) is zero in late September and late February. There is very rapid decrease in late fall and early winter, but the increase during spring and early summer is more gradual. The maximum rate of cooling amounts to -141 ly.day^{-1} in December; the maximum warming rate is +82 ly.day^{-1} in April. There is also a secondary maximum of +82 ly.day^{-1} in August. These results compare favorably with those available in the literature. Gabites [13] obtained storage values for the oceanic areas over latitude belts with the assumption that the temperature variation observed at the surface held for the first 25 meters, with a steady decrease from there on to zero seasonal variation at a depth of 125 meters. His values for the latitude belt 10°- 20° N. show maximum rates of -145 ly.day^{-1} in December and about +85 ly. day^{-1} in March and April.

An evaluation of the oceanic storage for a sector of the eastern Caribbean Sea, latitudes 15°- 20°N., longitudes 63°- 68°W., was made by Pattullo [30]. She worked with bathythermograph data compiled at Woods Hole Oceanographic Institution, the same source used here. Her values are considerably higher than those shown in table 3. She gives maximum rates of change of -254 ly.day^{-1} in December and +218 ly.day^{-1} in April. The reason for the large differences lies most probably in the method of analysis. She evidently used the temperatures as given by the bathythermograph sample, which, as explained previously, give a much larger annual range than the normal charts. In addition, since she was concerned with a more limited area, her data may have indicated a larger value for the level of zero seasonal change than the one adopted here. Fritz [11] also analyzed the storage over the oceans using bathythermograph data and computed weighted means over latitude belts. His values for the latitude belt 10°- 20°N. are in the same range as those presented here; maximum cooling of approximately -140 ly.day^{-1} is indicated in December, and maximum warming of +20 - 60 ly.day^{-1} in the spring and early summer (it is somewhat difficult to estimate exact values from his published charts).

c. <u>DIVERGENCE OF HEAT</u> (Q_v): Evaluation of the divergence of heat transport by the water currents in the Caribbean Sea is simplified by the fact that there is one main basic current, very well defined, with entrance in the Lesser Antilles and outlet in the Yucatan Channel. However, the temperature data were generally inadequate and the evaluation of the heat divergence falls short of the desired goal.

The divergence of heat transport is given by

$$Q_v = c \int_\sigma M_{hy} T_y \, d\sigma - c \int_\sigma M_{he} T_e \, d\sigma \qquad (3)$$

where M_{hy} and T_y are the horizontal water-mass flow per unit area and temperature across the Yucatan Channel, and M_{he} and T_e denote the same parameters at the east side of the Caribbean; the symbol σ denotes the lateral area across the respective current boundaries. The first integral represents the heat flux outward on the west side, the second represents the flux inward on the east side. For a rigorous evaluation of equation (3) we need to consider the correlations between T and M_h over the different sections of the boundaries. Analytically, the total flux across the western boundary, for example, can be expressed in the form of the contributions by the mean and "eddy" motions using the standard notation. Let

$$\begin{aligned} M_{hy} &= \tilde{M}_{hy} + M'_{hy} \\ T_y &= \tilde{T}_y + T'_y , \end{aligned} \qquad (4)$$

where the symbol $\sim$ denotes an average over the current boundary, and the primes denote deviations from that average. The mean value of the product $T_y M_{hy}$ becomes

$$\widetilde{T_y M_{hy}} = \tilde{T}_y \widetilde{M_{hy}} + \widetilde{T'_y M'_{hy}} \tag{5}$$

A similar expression will hold also for the east side and equation (3) is reduced to

$$Q_v = c \int_\sigma [\tilde{T}_y \widetilde{M_{hy}} + \widetilde{T'_y M'_{hy}}] \, d\sigma - c \int_\sigma [\tilde{T}_e \widetilde{M_{he}} + \widetilde{T'_e M'_{he}}] \, d\sigma. \tag{6}$$

Since we are working here with time averages, one would have to consider also variations in time. The time average of the product can be reduced to expressions equivalent to (4) and (5), and a third term, referred to as the "time eddies" term, would appear in each of the integrals in equation (6). However, since the data are adequate only for a rough estimate of the contribution by the mean motion we need not go into additional refinements. In the estimation of the heat divergence presented here, correlations in space and time between the temperature and mass flux were disregarded, not because they are considered to be insignificant, but because there is no way to estimate their effect. Equation (6) can then be reduced to

$$Q_v = \tilde{M}_h c \, (\tilde{T}_y - \tilde{T}_e) \tag{7}$$

where $\tilde{M}_h$ is now the integrated mass transport, assumed to be the same on both sides of the current.

In order to evaluate equation (7) the mean temperature integrated over the depth of the layer considered here must be computed. It was not possible to determine the seasonal temperature variations below the surface in the areas of the Yucatan Channel and the Lesser Antilles separately on account of inadequacy of the data. The procedure followed was to determine the seasonal variations in surface temperature (fig. 11) and assume that the reduction in temperatures with depth at both ends followed the same pattern and proportion obtained for the entire Caribbean Sea.

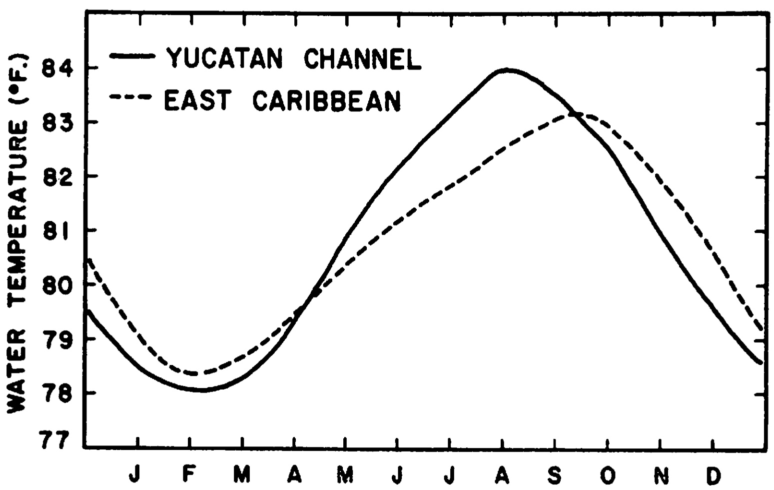


Figure 11. - Annual variation of surface water temperatures at the east and west sides of the Caribbean current.

In evaluating the temperatures at the eastern and western ends of the Caribbean, it was discovered that the seasonal changes follow somewhat different patterns so that $(T_y - T_e)$ changes sign with the seasons. Figure 11 shows the seasonal course of surface temperature in the Lesser Antilles and the Yucatan Channel as determined from Fuglister's charts. In the Yucatan Channel the annual range is larger and the maximum temperature occurs earlier than in the entrance zone. As a result, from May to September, it is warmer in the Yucatan Channel, while from October to April it is colder there than at the eastern end of the Sea, although by a small amount only. This same effect is shown also by the temperature data contained in the Climatic Charts of the Oceans and the new Marine Climatic Atlas of the World (U. S. Navy [50]). If figure 11 can be accepted as realistic, and representative of the temperature variations below the surface, then there is a net export of heat by the ocean current from June to September and a net import during the dry season.

The mass transport of water across the Florida Straits has been estimated as 26×10^6 m.^{3}sec.$^{-1}$ (Sverdrup et al. [47]), of which about one-third takes place in the upper 100 meters. The transport across the Yucatan Channel should be about the same. Although Sverdrup's estimate has been subject to debate (Stommel [46]), it has nevertheless been used in this study with the additional assumption that the transport in the upper 100 m. of the Caribbean Sea was also equal to one-third of the total transport. Further, the mass transport was taken as constant throughout the year; another simplification that results from the shortage of measurements.

With the foregoing assumptions, monthly temperature profiles were prepared for both sides of the Caribbean Sea and mean temperatures for the layer from the surface to 100 m. then determined by the equal area method. Equation (7) was evaluated; the results are presented in table 4. The magnitudes of the heat divergence are quite small compared to those of table 2; hence an error of as much as 100 percent in estimating the mass transport or $(T_y - T_e)$ would not be important. Significant values were obtained only for June to September, when export of heat amounted to 4-9 percent of the radiation, and for November to December, when an import amounting to 5-6 percent of the radiation was indicated. The annual average amounted to only 1 percent of the net radiation received, hence was well within the range of error in determining this quantity.

Stommel [46] has suggested that a very slow ascending motion with an order of meters per year may be taking place over large parts of the tropical oceans. This is an hypothesis and no data exist for a well-founded computation. Nevertheless the order of magnitude of the effect of such upwelling on the heat budget should be estimated if possible. The vertical mass transport is $M_z = \rho_w wA$, where w is the vertical velocity component and A the horizontal area of the water body. Then the heat divergence due to vertical motion is $c(\widetilde{\overline{T}}_y - \overline{T}_b)M_z$ in the first approximation, neglecting all correlations; the bar denotes horizontal averaging. The difference $(\widetilde{\overline{T}}_y - \overline{T}_b)$ varies from 1°- 3°C., and $W \cong 1$ cm./day (3.65 m./year). The heat export due to this effect will be then 0.2 - 0.6

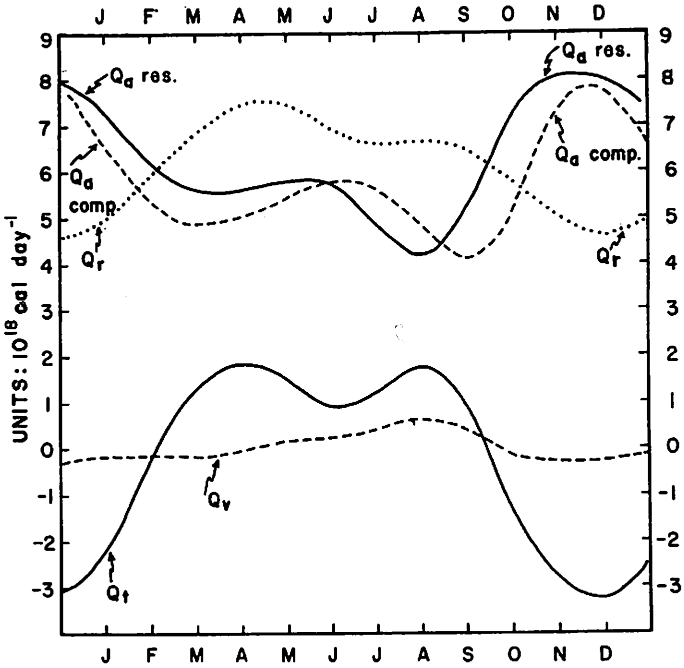


Figure 12. - Annual variations in heat transfer parameters from Caribbean Sea to atmosphere. Solid curve on top: net transfer (Q_a), obtained as residual in oceanic balance; dashed curve on top: net transfer (Q_a), computed with turbulent transfer formulae for section north of the Canal Zone; dotted curve: net radiation (Q_r) absorbed by Caribbean Sea. Solid curve on bottom: change in heat content of the water due to temperature changes; dashed curve on bottom: net heat divergence by Caribbean water current.

x 10^{17} cal./day; hence in general it is about one order of magnitude less than Q_v.

d. HEAT TRANSFER TO ATMOSPHERE (Q_a): With the data in tables 2, 3, and 4, equation (1) can be evaluated to yield value of Q_a as a residual. The results (table 5 and fig. 12) show maximum values of total heat transfer of about 8.0 units* in November and December and minimum of 4.2 units in August. The seasonal variations in the heat transfer seem to be determined mostly by the variation in Q_t.

Separate values of the sensible and latent heat can be obtained by a suitable assumption about the Bowen ratio, Q_s/Q_e. Following results obtained by Jacobs [17] and Bunker [6], we adopted a value of 10 percent. It should be mentioned, however, that recent work by Houghton [15] on the hemispheric heat balance, and by Riehl and Malkus [38], on the heat balance of the Equatorial Trough zone, indicate that the sensible heat transfer may be as much as 40

* The term unit used throughout this report stands for the quantity 10^{18} cal. day^{-1}.

Table 3. - Rate of change of heat content in water body of Caribbean Sea

	J	F	M	A	M	J	J	A	S	O	N	D	Ann
Change per unit area (ly.day^{-1})	-101	- 9	+ 59	+ 82	+ 66	+ 42	+ 59	+ 82	+ 42	- 59	-125	-141	0
Total for Caribbean (10^{18} cal.day^{-1})	-2.21	-0.20	+1.29	+1.80	+1.45	+0.92	+1.29	+1.80	+0.92	-1.29	-2.74	-3.09	0

Table 4. - Divergence of heat by upper 100-meter layer of Caribbean current. Unit: 10^{18} cal.day^{-1}.

J	F	M	A	M	J	J	A	S	O	N	D	Ann
-0.13	-0.13	-0.13	-0.04	+0.21	+0.29	+0.41	+0.62	+0.41	-0.13	-0.25	-0.29	+0.07

Table 5. - Transfer of heat from Caribbean Sea to atmosphere. Units: 10^{18} cal.day^{-1}.

	J	F	M	A	M	J	J	A	S	O	N	D	Ann
$Q_s + Q_e$	7.31	6.24	5.65	5.69	5.87	5.80	4.96	4.28	5.19	7.31	8.09	7.98	6.24
Q_s	0.66	0.57	0.51	0.52	0.53	0.53	0.45	0.39	0.47	0.66	0.74	0.73	0.57
Q_e	6.65	5.67	5.14.	5.17	5.34	5.27	4.51	3.89	4.72	6.65	7.35	7.25	5.67
Evaporation (cm.day^{-1})	0.52	0.45	0.40	0.41	0.42	0.41	0.36	0.31	0.37	0.52	0.58	0.57	0.44

percent of the latent heat transfer. As discussed by Riehl and Malkus, the greater amount of sensible heat transfer to the atmosphere could be the result of the warming of relatively cold downdrafts that reach the earth's surface in the center of thunderstorm cells. Such an effect should certainly be insignificant during the winter season in the Caribbean Sea, but would suggest a larger value of the Bowen ratio in summer than in winter. It is not possible to allow for this effect here, and the Bowen ratio was assumed constant all year around. However, one can see that if the value were larger in summer, then Q_e would be less than shown in table 5 and the seasonal variation in evaporation might be greater than indicated here.

The values of Q_e as obtained range from 7.35 units (336 $ly.day^{-1}$) in November to 3.89 units (178 $ly.day^{-1}$) in August. The evaporation rates are 0.58 $cm.day^{-1}$ in November and 0.31 $cm.day^{-1}$ in August. The annual evaporation rate is 0.44 $cm.day^{-1}$, or 161 $cm.year^{-1}$.

e. HEAT TRANSFER FROM OCEAN TO ATMOSPHERE BY TURBULENT DIFFUSION: Direct computation of the energy transfer from ocean to atmosphere can be obtained from the theory of turbulent diffusion. This subject is discussed at length by Jacobs [17] and Priestley [32]. Jacobs introduced a set of formulae for evaluating the net transfer Q_e and Q_s from observations of water temperature, and of wind, temperature, and moisture content of the air in the vicinity of the ocean surface. Riehl et al. [41] used the following, somewhat different, set of formulae for computation of heat transfer:

$$Q_e = LE = 1.71 \times 10^{-6} L (q_w - q_a) v \tag{8}$$

$$Q_s = 4.16 \times 10^{-7} (T_w - T_a) v \tag{9}$$

The symbols T_w and q_w represent the temperature and saturation specific humidity of the water; T_a, q_a, and v represent the temperature, specific humidity, and wind speed at ship's deck level. L stands for the heat of vaporization and E for evaporation rate. All units are in the cgs system; i.e., if L, q, T, and v are in cgs, Q_e and Q_s come out in $cal.sec^{-1}$.

The above expressions were used to evaluate the heat transfer with climatological data given in the Marine Climatic Atlas of the World (Atlantic Ocean) (U.S.Navy,[50]) for a small sector of the Caribbean north of the Canal Zone, and with synoptic data for the entire Caribbean Sea for December 1956 and January 1957.

The data were adequate for evaluating the latent heat transfer only; an estimate of the total transfer was obtained by assuming a 10 percent value for the Bowen ratio. The results (fig. 12) agree fairly well with the residual values. The annual variation is about the same with maximum in December and minimum in September, and a secondary maximum in June. The magnitudes month by month differ on the average by around 13 percent, the only large differences

(30-40 percent) are in October and November; the values for November were disregarded in drawing the annual curve in figure 12. The corresponding evaporation rates range from 0.56 cm. day^{-1} in December to 0.30 cm. day^{-1} in September. The annual mean is 146 cm. year^{-1}. The agreement between the two sets of values is certainly within the limits of accuracy of the data, particularly since it is not expected that the climatological normals for the region north of the Canal Zone be exactly equal to those for the entire Caribbean Sea.

The data used in evaluations for December 1956 and January 1957 consisted of the averages of over 500 surface observations taken in all areas of the Caribbean Sea during the given months. The results are listed below:

	1956 December	1957 January	(Units: 10^{18} cal. day^{-1})
Latent heat transfer (Q_e)	7.64	7.81	
Sensible heat transfer (Q_s)	0.75	0.76	
Net (Q_a)	8.39	8.57	

Values of transfer of heat somewhat greater than the previously evaluated normal values were obtained in both months. The magnitude of the Bowen ratio indicated by these data is approximately 10 percent.

f. COMPARISON WITH PREVIOUS RESULTS: Jacobs [17] computed heat transfer from ocean to atmosphere in the North Atlantic and North Pacific Oceans. His equations are equivalent to (8) and (9), differing only in the values of the constants. The climatological compilations contained in the *Atlas of Climatic Charts of the Oceans* (U. S. Weather Bureau [51]) were used as a source of data. The results for the evaporation rates in the Caribbean Sea give averages of about 0.40 cm.day^{-1} for the winter season (December, January, February) and 0.32 cm.day^{-1} in summer (June, July, August), with intermediate values for spring and autumn. The annual average is 0.34 cm.day^{-1}, for an annual total of 124 cm.

Wüst [53] computed annual evaporation rates for latitude belts of the North Atlantic Ocean based on energy balance relationships, and obtained 146 cm.year^{-1} for the latitude belt 10°- 20°N. Not much difference should exist between the annual average for the Caribbean Sea and the average for the whole ocean at the same latitude. The magnitudes obtained in the present study based on the energy balance relations show, therefore, annual evaporation rates of the order of 15-30 percent higher than the previous estimates available in the literature. However, it is felt that the analysis carried out here has more closely followed the local conditions over the Caribbean Sea than the previous studies have.

4. ENERGY BALANCE OF THE ATMOSPHERE

The energy equations for the atmosphere can be readily derived from the equations of motion and the first law of thermodynamics. Discussions on the applicability of the energy principles to atmospheric motions are available in textbooks in dynamic meteorology and other meteorological publications (for instance Starr [45] and Miller [25]). If the equation of motion, expressed in standard vector notation in the form:

$$\rho \frac{d\vec{V}}{dt} = - \nabla p - 2 \rho \vec{\Omega} x \vec{V} + \rho \vec{g} + \rho \vec{F} \tag{10}$$

is multiplied vectorially by the velocity vector $\vec{V}$, then

$$\rho \vec{V} \cdot \frac{d\vec{V}}{dt} = \rho \frac{1}{2} \frac{dV^2}{dt} = - \vec{V} \cdot \nabla p - \rho w g + \rho \vec{V} \cdot \vec{F}. \tag{11}$$

All symbols follow the usual meteorological notation; for details, see the list of symbols. Rearranging (11) we obtain

$$\rho \frac{d}{dt} (\frac{V^2}{2} + gz) = - \vec{V} \cdot \nabla p + \rho \vec{V} \cdot \vec{F}, \tag{12}$$

which is an expression for the total change in kinetic and potential energy in terms of the rate of work done by pressure forces and the frictional dissipation.

The first law of thermodynamics can be expressed in the form:

$$h = c_p \frac{dT}{dt} - \frac{1}{\rho} \frac{dp}{dt} , \tag{13}$$

where h denotes the rate at which heat is added or subtracted from the system, and the other symbols are conventional. Subtracting equation (12) from equation (13), and making use of the relation

$$\frac{dp}{dt} = \frac{\partial p}{\partial t} + \vec{V} \cdot \nabla p \tag{14}$$

we obtain

$$\rho h = \rho \frac{d}{dt} (\frac{V^2}{2} + g z + c_p T) - \frac{\partial p}{\partial t} - \rho \vec{V} \cdot \vec{F}, \tag{15}$$

which constitutes an expression for the total energy changes. The first term on the right denotes the change in kinetic energy per unit mass, $\frac{V^2}{2}$; the change in potential energy, gz; and the change in enthalpy, $c_p T$, which in turn is equal to the internal energy, $c_v T$, plus the work by pressure forces, $\frac{p}{\rho}$.

With the use of the equation of continuity, equation (15) becomes

$$\rho\, h = \frac{\partial}{\partial t}\,\rho\left(\frac{V^2}{2} + g\,z + c_p T\right) + \nabla\cdot\rho\vec{V}\left(\frac{V^2}{2} + g\,z + c_p T\right) - \frac{\partial p}{\partial t} - \rho\vec{V}\cdot\vec{F}. \qquad (16)$$

Computations to be made will be over time intervals of a month, so that local time changes as given by the first and third terms on the right-hand side of equation (16) can be neglected; the atmospheric heat storage is negligible compared to that of the oceans. Integrating over the atmospheric volume, denoted by α, and making use of the divergence theorem,

$$\int_\alpha \rho h\, d\alpha = \iint c_n\left(\frac{V^2}{2} + g\,z + c_p T\right)\rho ds dz - \int_\alpha \rho\vec{V}\cdot\vec{F}\, d\alpha \qquad (17)$$

where c_n is the wind component normal to the lateral boundary (positive outward) and ds is a horizontal length element around the periphery. The flux from the ocean is included in the heat term on the left, and no heat transfer is considered across the tropopause.

The integral on the left side of the equation represents the heat sources and sinks. The two integrals on the right denote, respectively, the energy flux across the lateral boundary and the kinetic energy dissipation inside the volume by friction. Considering heat sources and sinks, the net radiation, R_a, is the sink; the sources are the sensible heat conduction from the ocean (Q_s) and precipitation heating (LP), where P denotes precipitation rate and L is the heat of condensation of water vapor. If the symbols R_a, Q_s, and LP denote values integrated over the volume, equation (17) becomes

$$R_a + Q_s + LP + \iint c_n\left(\frac{V^2}{2} + gz + c_p T\right) ds\,\frac{dp}{g} + \int_\alpha \rho\vec{V}\cdot\vec{F}\, d\alpha = 0 \qquad (18)$$

where the vertical coordinate in the flux integral has been changed to a pressure coordinate by means of the hydrostatic equation.

Equation (18) states the sensible heat balance requirement. Latent heat of water vapor can be incorporated into this equation by interpreting it as if it were part of the internal energy. It is more informative, however, to establish a separate latent heat balance equation. This equation has the same form as equation (18); it consists of a balance between the flux across the lateral boundary, the transfer from ocean to atmosphere by evaporation, and the transfer from atmosphere to ocean by precipitation. Thus

$$Q_e - LP + \iint Lq\, c_n\, ds\,\frac{dp}{g} = 0 \qquad (19)$$

where q denotes specific humidity.

Equations (18) and (19) are the formulae to be evaluated. The terms Q_s and Q_e were discussed in the section on the oceanic heat balance. Results obtained for December 1956 and January 1957 will be used here. Evaluation of the flux terms and of the atmospheric radiation cooling are discussed below.

Precipitation heating could not be computed directly due to lack of rainfall data over the oceans and must be treated as a residual. However, it can be determined in two ways, from either equations (18) or (19); comparison of these determinations will serve as a check on the validity of the calculations.

a. EVALUATION OF ENERGY FLUXES: The flux integrals above involve the product of c_n and the energy parameters around the boundary and may be expanded into the contributions of mean and eddy motions in a manner similar to equations (4) and (5). The contributions by eddy motions arise from variations of the winds and other variables around the boundary of integration (space correlations) and/or from fluctuations with time (time correlations). The analysis here will be with monthly mean data and the question will be asked, to what extent do monthly data satisfy the balance requirements of equations (18) and (19); i.e., time fluctuations of the variables on scales less than a month will be omitted. For our purpose, then, the mean of the monthly values averaged around the ellipse at one layer will be denoted with bars, and deviations from this mean with primes. With these considerations the flux terms in equations (18) and (19) can be expanded as follows:

$$\iint\left(\frac{V^2}{2} + gz + c_pT\right)c_n ds\frac{dp}{g} = \iint \bar{c}_n\overline{\left(\frac{V^2}{2} + gz + c_pT\right)}ds\frac{dp}{g} + \iint \overline{c_n'\left(\frac{V^2}{2} + gz + c_pT\right)'}ds\frac{dp}{g} \tag{20}$$

$$\iint Lq\, c_n\, ds\, \frac{dp}{g} = \iint L\, \bar{q}\, \bar{c}_n\, ds\frac{dp}{g} + \iint L\, \overline{q'c_n'}\, ds\frac{dp}{g} \tag{21}$$

Vertical integration in equations (20) and (21) was performed by averaging the different parameters over layers of 100-mb. thickness from 1000 to 100 mb. The 100-mb. level coincides well with the base of the tropopause, while the 1000-mb. level is located a short distance above the surface. The layer between the surface and 1000 mb. was ignored.

i. Computational Procedures: The first step was to compute the mass transport across the boundary of the ellipse. At 50-mb. intervals from 1000 mb. to 100 mb. the wind was broken into a c_s-component (parallel to the ellipse of figure 1, positive counterclockwise) and a c_n-component (normal to the ellipse, positive outward). Values of c_n at each station were plotted against pressure on a linear scale and profiles were drawn with minor hand smoothing. Mean values of c_n for layers of 100-mb. thickness were read from these profiles: between 300 and 100 mb., values for 50-mb. layers were used in some cases of exceptionally large gradients.

These layer-mean values were plotted and analyzed on vertical cross-sections, with pressure as ordinate and peripheral distance around the boundary as abscissa (fig. 13). Horizontal charts, drawn at 100-mb. intervals, were used as aids in the areas of scarce data. Another aid used occasionally in analysis was to plot the layer values of given parameters in a graph with

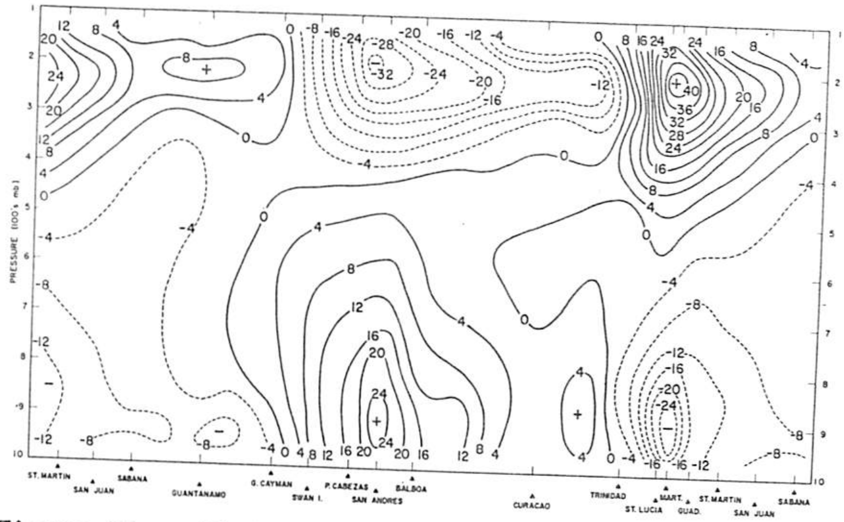


Figure 13. - Distribution of c_n (kt.) around the lateral boundary of the Caribbean Sea for December 1956. Solid curves show positive (outward) values; dashed curves indicate negative (inward) values.

magnitude as ordinate and peripheral distance around the boundary as abscissa and then draw smooth curves by hand. The South American coast, particularly from Balboa to Curacao, was the only part of the ellipse that presented major analysis problems on account of unrepresentative local circulations and the wider gap between stations. Fortunately, the flow in that area was nearly parallel to the boundary, so that c_n was small. Minor adjustments, however, proved necessary in the low levels at Balboa, Panama Canal Zone. Much weaker flow was measured there than was evidently taking place slightly north of the station, as seen from a sample of surface wind observations from ships. Local breaking of the general trade in the Panama Canal Zone is a well-known feature. It should be added that errors introduced through the analysis in this area made, at most, a small contribution to the total mass balance.

After completion of the analysis, 20 equally-spaced points around the perimeter of the ellipse were selected and values of c_n for each 100-mb. layer were read from the analysis at each point. The 20-point technique was adopted in order to weight all sections around the boundary equally, regardless of the spacing of the stations. These 20 values were averaged horizontally to obtain the mean normal component, $\overline{c_n}$, around the boundary for each layer. Deviations from the layer mean were also computed at each of the 20 points.

A similar procedure was used with the energy parameters. From the monthly data of temperature, humidity, and contour heights, the quantities $gz + c_pT$, Lq, $V^2/2$, etc., were computed at 50-mb. intervals from 1000 to 100 mb., then plotted vertically against pressure and mean 100-mb.-layer values obtained (the total sensible energy, $gz + c_pT$, was first treated jointly as a sum; later

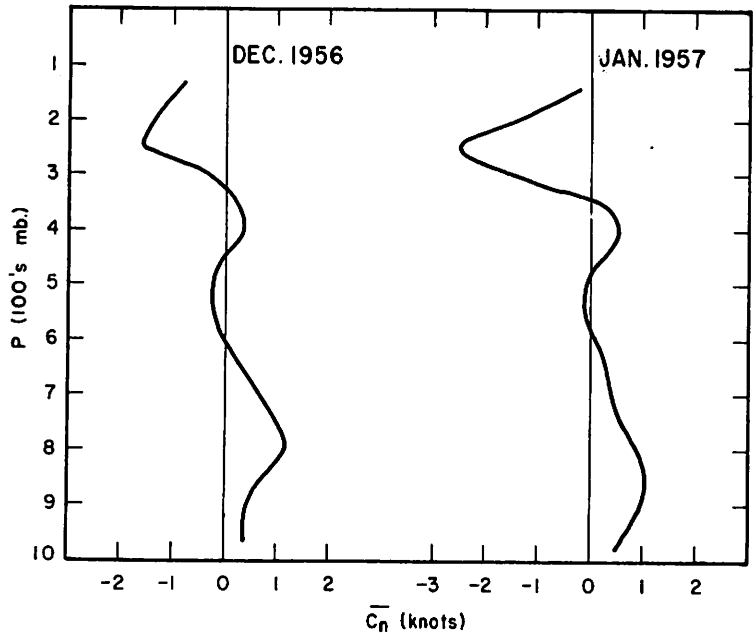


Figure 14. - Vertical profile of the mean values of c_n.

gz was evaluated separately and then c_pT obtained as a residual). These values were plotted in the boundary cross-section, and analyzed. The magnitudes at the 20 selected points were read back from the analysis and their layer means and deviations computed (figs. 15, 16). The individual deviations at each of the 20 points were multiplied by the corresponding values of c_n to obtain the space correlations required for the evaluation of the "eddy" terms in equations (20) and (21). Before presenting the results of the flux calculations, it may be of interest to examine the mean mass circulation and energy content of the Caribbean troposphere revealed by the data.

ii. Mass Balance: The c_n-cross section for December (fig. 13) shows mainly a zone of inflow in the Lesser Antilles with outflow over Central America in the lower atmosphere,and the reverse situation aloft, so that there is a tendency for local mass flow compensation everywhere.

If no vertical mass flow takes place through the upper or lower boundary, and the pressure inside the volume remains (nearly) constant, then the net mass flux around the closed surface integrated from the surface to the tropopause should be equal to zero. This constraint must be met; it also serves as a useful check on the accuracy of the data and analyses employed. Fortunately, mass balance was obtained readily for both months.

The vertical distribution of $\overline{c}_n$ (fig. 14) shows outflow (divergence) in the lower layers, inflow (convergence) in the high troposphere. This is consistent with previous results (Riehl et al. [41]), which indicate divergence in the trade winds layers. The magnitude of the velocity divergence is about 1.4×10^{-6} sec.$^{-1}$ in the low levels and over -3.0×10^{-6} sec.$^{-1}$ near 200 mb.

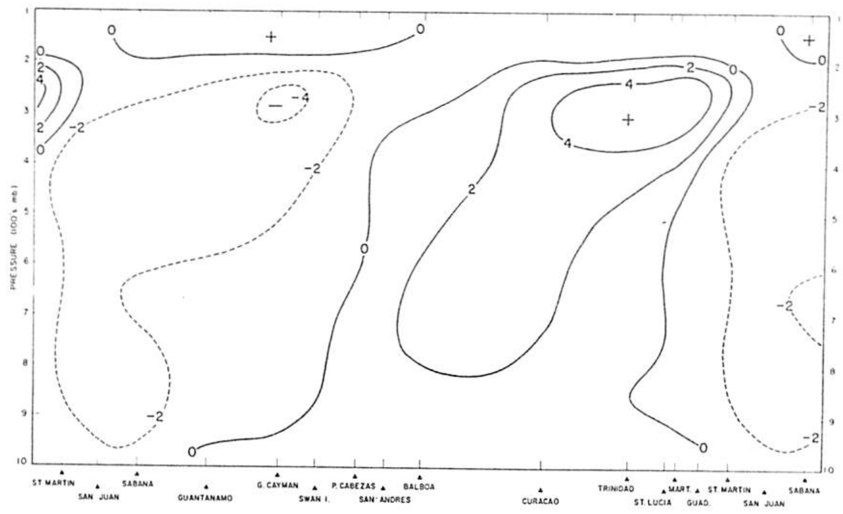


Figure 15. - Distribution around the lateral boundary of the anomalies of $gz + c_p T$ (tenths of cal.gm.$^{-1}$) from the mean values at each 100-mb. layer. Data for December 1956.

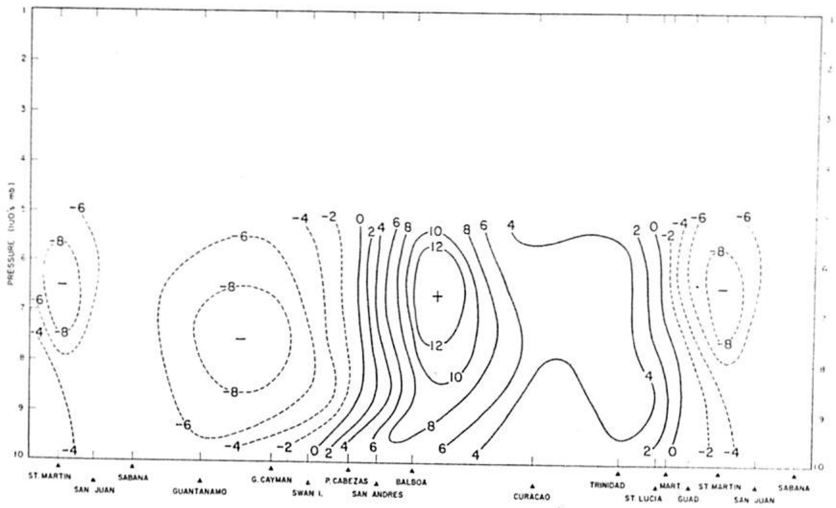


Figure 16. - Distribution around the lateral boundary of the anomalies of Lq. Data for December 1956. Units: tenth of cal.gm.$^{-1}$

Table 6. - Mean energy content of Caribbean troposphere. Units: cal.gm.$^{-1}$

	December 1956					January 1957				
P(100's mb.)	$\overline{gz}$	$c_p\overline{T}$	$L\overline{q}$	$\overline{gz}+c_p\overline{T}$	$\overline{gz}+c_p\overline{T}+L\overline{q}$	$\overline{gz}$	$c_p\overline{T}$	$L\overline{q}$	$\overline{gz}+c_p\overline{T}$	$\overline{gz}+c_p\overline{T}+L\overline{q}$
1-2	33.7	49.2		82.9	82.9	33.7	48.8		82.5	82.5
2-3	25.7	54.6		80.3	80.3	25.7	54.5		80.2	80.2
3-4	20.1	59.0		79.1	79.1	20.1	58.9		79.0	79.0
4-5	15.7	62.3	0.6	78.0	78.6	15.7	62.3	0.5	78.0	78.5
5-6	12.0	64.8	1.2	76.8	78.0	12.0	64.8	1.0	76.8	77.8
6-7	8.8	66.6	2.3	75.4	77.7	8.9	66.6	1.8	75.5	77.3
7-8	6.1	68.0	3.9	74.1	78.0	6.1	68.0	3.2	74.1	77.3
8-9	3.6	69.4	6.0	73.0	79.0	3.6	69.2	5.4	72.8	78.2
9-10	1.4	70.6	8.0	72.0	80.0	1.4	70.5	7.5	71.9	79.4

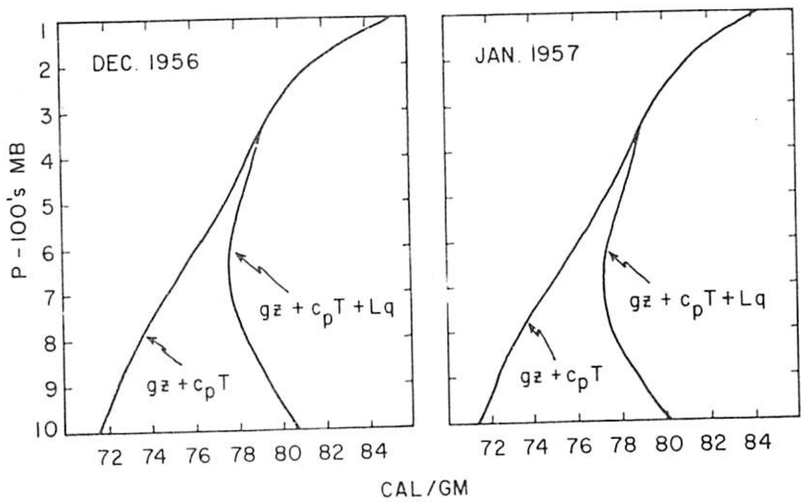


Figure 17. - Vertical profile of the energy content of the Caribbean atmosphere.

Table 7. - Lateral heat flux. Unit: 10^{18} cal.day^{-1}. Positive sign stands for export, negative for import.

			Flux of $(gz + c_pT)$			Flux of Lq			Flux of $(gz + c_pT + Lq)$		
P(100's mb.)	$\bar{c}_n$ (m.sec.$^{-1}$)	Mass Flux (M_h) (10^{16} gm.day^{-1})	By mean motion	By eddy motion	Net	By mean motion	By eddy motion	Net	By mean motion	By eddy motion	Net
					a. DECEMBER 1956						
1-2	-0.5	-26.74	-22.17	-0.04	-22.21				-22.17	-0.04	-22.21
2-3	-0.8	-42.79	-34.36	0.42	-33.94				-34.36	0.42	-33.94
3-4	0.1	5.35	4.23	0.19	4.42				4.23	0.19	4.42
4-5	0	0	0	0.11	0.11	0	0.08	0.08	0	0.19	0.19
5-6	-0.1	-5.35	-4.11	0.09	-4.02	-0.06	0.23	0.17	-4.17	0.32	-3.85
6-7	0.2	10.70	8.07	0.10	8.17	0.25	0.37	0.62	8.32	0.47	8.79
7-8	0.5	26.74	19.81	0.24	20.05	1.04	0.47	1.51	20.85	0.71	21.56
8-9	0.4	21.39	15.61	0.26	15.87	1.28	0.49	1.77	16.89	0.75	17.64
9-10	0.2	10.70	7.70	0.10	7.80	0.86	1.01	1.87	8.56	1.11	9.65
1-5			-52.30	0.68	-51.62	0	0.08	0.08	-52.30	0.76	-51.54
5-10			47.08	0.79	47.87	3.37	2.57	5.94	50.45	3.36	53.81
1-10			-5.22	1.47	-3.75	3.37	2.65	6.02	-1.85	4.12	2.27
					b. JANUARY 1957						
1-2	-0.2	-10.70	-8.83	0.01	-8.82				-8.83	0.01	-8.82
2-3	-1.3	-69.52	-55.76	0.01	-55.75				-55.76	0.01	-55.75
3-4	0.1	5.35	4.23	0.35	4.58				4.23	0.35	4.58
4-5	0.1	5.35	4.17	0.28	4.45	0.03	0.18	0.21	4.20	0.46	4.66
5-6	-0.1	-5.35	-4.11	0.19	-3.92	-0.05	0.22	0.17	-4.16	0.41	-3.75
6-7	0.2	10.70	8.08	0.14	8.22	0.19	0.32	0.51	8.27	0.46	8.73
7-8	0.3	16.04	11.89	0.29	12.18	0.51	0.34	0.85	12.40	0.63	13.03
8-9	0.5	26.74	19.47	0.38	19.85	1.44	0.36	1.80	20.91	0.74	21.65
9-10	0.4	21.39	15.38	0.30	15.68	1.60	0.79	2.39	16.98	1.09	18.07
1-5			-56.19	0.65	-55.64	0.03	0.18	0.21	-56.16	0.83	-55.33
5-10			50.71	1.30	52.01	3.69	2.03	5.72	54.40	3.33	57.73
1-10			-5.48	1.95	-3.53	3.72	2.21	5.93	-1.76	4.16	2.40

Table 8. - Vertical mass and heat flux by mean motion. Positive sign stands for downward motion.

	December 1956							January 1957					
			Heat Flux (10^{18} cal. day^{-1})						Heat Flux (10^{18} cal. day^{-1})				
p (100's mb.)	Mass flux (M_z) (10^{16} gm.day^{-1})	w (10^2m.day^{-1})	gz	c_pT	c_vT	$gz + c_pT$	Mass flux (M_z) (10^{16} gm.day^{-1})	w (10^2m.day^{-1})	gz	c_pT	c_vT	$gz + c_pT$
1	0	0	0	0	0	0	0	0	0	0	0	0
2	26.74	3.8	7.80	13.96	9.92	21.77	10.70	1.5	3.12	5.55	3.94	8.68
3	69.53	7.2	15.78	39.56	28.09	55.35	80.22	8.3	18.21	45.56	32.33	63.77
4	64.18	5.3	11.36	39.02	27.73	50.38	74.87	6.2	13.25	45.52	32.34	58.77
5	64.18	4.5	8.86	40.82	29.01	49.67	69.52	4.8	9.59	44.21	31.42	53.81
6	69.53	4.2	7.23	45.68	32.40	52.91	74.87	4.5	7.79	49.26	34.96	57.05
7	58.83	3.1	4.35	39.59	28.12	43.95	64.17	3.4	4.75	43.25	30.74	48.00
8	32.09	1.5	1.54	22.08	15.66	23.62	48.13	2.3	2.31	33.02	23.44	35.33
9	10.70	0.5	0.26	7.50	5.33	7.76	21.39	0.9	0.51	14.95	10.61	15.46
10	0	0	0	0	0	0	0	0	0	0	0	0

Values near the surface agree well with those obtained by Riehl et al. [41].

The net horizontal mass flux ($\overline{M}_h$) (table 7) across the lateral boundary is given by

$$\overline{M}_h = \overline{c}_n \, S \, \frac{\Delta p}{g} , \tag{22}$$

where S is the perimeter around the horizontal area. Since a pattern of convergence at high levels and divergence at low levels is present, net downward mass flow takes place at all levels. The downward flow (M_z) was calculated with the continuity relationship by the use of the horizontal flow in each 100-mb. layer. Table 8 presents the results. The mean vertical velocity over the area ($\hat{w}$) was obtained from the relation

$$\hat{w} = \frac{\hat{M}_z}{\rho A} \tag{23}$$

Downward motion was particularly large in the upper troposphere with velocities greater than 700 m.day^{-1} in December and 800 m.day^{-1} in January. In the lower layers the values of $\hat{w}$ compare well with those measured by Riehl et al. [41], in the trades of the North Pacific Ocean. Concerning the high troposphere, it may be noted that Vuorela [52] computed descent of several hundred meters per day, occasionally over 500 m.day^{-1}, in the average around the globe between latitudes 15° and 30°N. for 5-day periods. Therefore, the magnitudes above are not unrealistic. In general, the pattern of mean motion obtained for the Caribbean winter atmosphere supports the picture of the ageostrophic mean circulation for winter by Palmen et al [29] which contains a shallow meridional return flow from the Tropics poleward centered near 200 mb.

iii. Mean Energy Content of the Atmosphere: The contribution of kinetic energy to the total energy is generally found to be negligible and may be ignored in this part of the analysis. A separate balance for the kinetic energy is discussed later.

The space variation in sensible energy, $gz + c_pT$, around the Caribbean Sea (fig. 15) shows higher values in the south and west in the low levels, and in the south and east in the high troposphere. A similar variation is present in the field of Lq (fig. 16), which shows maximum values in the southwestern section of the ellipse. This at once suggests accumulation of latent and sensible energy by the trades during travel across the Caribbean. The latent heat was neglected above the 400-mb. level on account of the low values of moisture content there plus the difficulties in measuring humidities at low temperatures. The data for January (not illustrated) show variations and magnitudes similar to those for December.

Table 6 (fig. 17) presents a summary of the mean total energy content of the Caribbean atmosphere. The sensible energy, $gz + c_pT$, increases steadily with height, while the latent heat energy decreases. As a result the profile of total energy, $gz + c_pT + Lq$, has maxima in the low and high troposphere and a minimum in the layer 700 - 500 mb. This vertical distribution of total energy is characteristic of the tropical atmosphere everywhere (Palmén et al. [29]; Riehl and Malkus [38]).

iv. Results of Flux Calculations: The results of the evaluation of energy flux (equations (20) and (21)) are presented in table 7-a,b. A large export of sensible energy by the mean motion takes place in the lower atmosphere and a large import aloft, so that the net for the entire troposphere is a small difference between two large terms. This, however, can be accepted as reliable in view of the data used here. The standing eddy term is positive (divergence) at all layers except the top one, and small compared to the term due to the mean motion, except in the middle troposphere where $\bar{c}_n$ reverses sign. Yet, because the sign of the eddy term is the same at all heights the integral for the whole atmosphere is a significant percentage of the mean term. Total flux resulted in a net convergence of 3.75 units in December and 3.53 units in January; hence the order of magnitude is the same as that of the bottom row in table 2.

The flux of latent heat was positive at nearly all levels, both for the contribution of mean motion and standing eddy terms. Summation for the troposphere showed an export of 6.02 units in December and 5.93 units in January, of which about 40 percent was contributed by the standing eddy term. Thus the export of latent heat exceeds the import of $gz + c_pT$, and the area acts as a net exporter of heat to other portions of the globe.

b. ATMOSPHERIC RADIATION: The net radiation of the troposphere is determined by an exchange of long-wave radiation with the surface of the earth and space, and by absorption of direct short-wave insolation from the sun. The exchange in the infrared part of the spectrum is such that emission to the earth's surface and space is greater than the absorption of incoming radiation from the surface. Absorption of short-wave insolation falls far short of offsetting the losses due to infrared radiation. Consequently, the atmosphere is a radiational cold source everywhere, including the Tropics.

Moisture, both in the form of water vapor and clouds, is the most important atmospheric constituent in the radiational exchange, both for the infrared and ultraviolet portions of the spectrum. Detailed discussions of the radiational properties of water vapor and of other atmospheric constituents are found in textbooks and other publications (Simpson [44], Elsasser [9], London [21], etc.), so a restatement need not be offered here. The net loss of infrared radiation by the atmosphere results from the vertical distribution of temperature and water vapor, both of which decrease upward. Methods for evaluating infrared cooling by graphical means have been described, especially the Elsasser diagram. Knowledge of temperature and moisture distribution with height, and of the cloud cover, is required for computation.

Absorption of short-wave radiation in the troposphere is accomplished by water vapor, clouds, and dust particles; the largest part of the absorption is due to water vapor. An emperical formula for computing absorption of short-wave radiation by water vapor in clear skies has been offered by Mügge and Möller [26]. Different forms of this formula have been presented by Charney [7], Riehl et al. [41], and London [21]. For a more exact evaluation of the absorption the effect of the cloud cover must also be considered.

For purposes of this study it was found advisable to adopt the recent radiation computations of London [21]. Regional cloud data for the Caribbean during the period of study were not readily available. Calculations of infrared radiation, using the Elsasser diagram, and of absorption of short-wave radiation, using the Mügge and Möller formula, were made for clear skies. Results agreed well with those of London for the latitude belt 10°-20°N. in the same season. London obtained net infrared cooling of 0.247 $ly.min.^{-1}$ in the troposphere including clouds, and short-wave absorption of 0.075 $ly.min.^{-1}$, hence a net cooling of 0.172 $ly.min.^{-1}$ or approximately, 1.1°C. per day. After integration over the Caribbean the daily radiational heat loss is 5.48 units; this amount was used in both months.

c. ATMOSPHERIC HEAT BUDGET: In the discussion below the numbers for the kinetic energy terms have been taken from the computations described in the next section. As already mentioned precipitation heating must be considered as residual.

Table 9 summarizes the results of the evaluation of equation (18). The net sources of heat in December 1956 include 3.75 units due to import of $gz + c_pT$, 0.75 units received from the ocean by conduction, and 0.08 units provided by frictional dissipation of kinetic energy. The sinks are 5.48 units

Table 9. - Atmospheric heat balance. Units: 10^{18} cal.day^{-1}.

	December 1956		January 1957	
	Sources	Sinks	Sources	Sinks
Convergence of $gz + c_pT$	3.75		3.53	
Divergence of kinetic energy		0.09		0.05
Frictional dissipation of kinetic energy	0.08		0.11	
Flux of sensible heat from ocean (Q_s)	0.75		0.76	
Radiational cooling		5.48		5.48
Net	4.58	5.57	4.40	5.53
Precipitation heating (residual)	0.99		1.13	
Precipitation (cm. month^{-1})	2.4		2.7	

by radiational cooling and 0.09 units due to export of kinetic energy. This leaves a deficit of 0.99 units, which should be balanced by precipitation heating. The balanced calculation for January 1957 indicates a precipitation heating of 1.13 units. Corresponding precipitation rates are 2.4 cm./month in December and 2.8 cm./month in January.

An independent estimate of the precipitation heating is obtained from the moisture balance, equation (19) (table 10). Moisture continuity indicates precipitation heating of 1.62 units in December and 1.88 units in January, or precipitation rates of 3.8 and 4.5 cm./month respectively. Comparing tables 9 and 10 this outcome must be regarded as satisfactory and reasonable in view of the approximations and uncertainties inherent in all such calculations; more is said about this imbalance below.

The computed precipitation may be compared with estimates from other sources. Jacobs [17] in particular has presented seasonal precipitation charts over the North Atlantic and Pacific Oceans. His values were determined with several assumptions. First, the annual precipitation totals of Meinardus [24] were reduced by about 25 percent over the Atlantic Ocean to conform with latitudinal precipitation profiles of Wüst, which he considered to be more reliable. Then the annual totals were distributed among the four seasons on the basis of the relative seasonal frequency of precipitation, as given by climatological summaries of ship observations at Greenwich Meridian Noon time. This procedure assumes that the diurnal variation and the intensity of precipitation is the same in every season. Jacobs' charts indicate precipitation of 6 cm./month over the Caribbean in the dry season, and about the same magnitude in the other three seasons. The rainfall obtained in the present analysis is only about one-half of that computed by Jacobs. However, if Jacobs' annual

Table 10. - Atmospheric moisture balance, expressed in the form of latent heat. Units: 10^{18} cal.day^{-1}.

	December 1956		January 1957	
	Sources	Sinks	Sources	Sinks
Divergence of Lq		6.02		5.93
Flux from ocean (Q_e)	7.64		7.81	
Precipitation heating (residual)		1.62		1.88
Precipitation (cm.month^{-1})		3.8		4.4

precipitation is approximately correct, there is every reason to believe that the seasonal variation should be larger than indicated, and that his values for the winter season are too high.

It may be argued, of course, that the uncertainty of all the calculations is such that a precipitation estimate within a factor of two must be regarded as highly satisfactory and/or that computations for one winter are insufficient for comparison with climatic means. Evidently, it will be of interest to repeat the computations for several subsequent dry seasons. On the other hand, it should not be overlooked that the accuracy of the procedures in this paper certainly is not of lower order than that of the statistical route employed by Jacobs and others. A precipitation estimate for an ocean basin from energy balance calculations for atmosphere and ocean has not been made previously to the best knowledge of the author. The line integral technique could not be applied prior to the development of rawin station networks. Hence the technique is regarded as novel and may well furnish the most reliable approach to estimates of precipitation over ocean area that is available.

d. COMMENTS ON THE ATMOSPHERIC BALANCE - EFFECT OF TIME EDDIES: The question was raised earlier as to what extent monthly data, without concern for time variations in a scale less than a month, could satisfy the energy balance relationships. Some uncertainty still exists in that the contribution of the precipitation heating is not a direct measurement and an imbalance is obtained in the evaluation of tables 9 and 10. It is quite possible that this imbalance is due to an over-estimate in the heat convergence or under-estimate of the moisture divergence on account of the time variations. For balance the divergence of latent heat would have to increase by about 11 percent in December, and 13 percent in January, provided the export of sensible heat stayed about the same.

In view of the large values in directional steadiness, contribution of the time eddy term would have to come from time variations in the energy properties or in wind speed. Significant time variations in latent heat are visualized only in the layers near and above the inversion due to up and down pumping of the base of the inversion, or to general upward extension of the moist layer during disturbed periods; it is quite possible that the contribution due to time variations would alter the energy balance appreciably. On the other hand, minor adjustments in the magnitudes of the other two major terms in tables 9 and 10 can provide agreement. For example, if the radiational cooling in table 9 is larger by only 10 percent, the agreement with table 10 becomes nearly perfect, particularly for December. Similarly, a 10 percent decrease in the latent heat transfer from the ocean would bring the precipitation estimate of table 10 to a much better agreement with table 9. Inaccuracies of the order of 10 percent in the estimates of the radiational cooling, oceanic transfer, and/or the atmospheric flux are not unreasonable, but one can see that adjustments of even less than 10 percent, if taken in the proper sense, would eliminate the imbalance between tables 9 and 10. Therefore, it is justifiable to accept those differences as within the limits of accuracy of the data and not due to the effect of time eddies.

It is also possible that perfect balance could be obtained for the two estimates of precipitation and that the contribution by time eddies could still be significant. In that case the test would be to compare the precipitation as computed with real observations. No further comments can be offered on this; it is not possible to provide rainfall figures. The magnitudes obtained appear to be adequate and under those conditions one can state that a satisfactory energy balance has been obtained from monthly mean data only.

e. CONTRIBUTION OF THE CARIBBEAN ATMOSPHERE FOR HEAT BALANCE ELSEWHERE: Table 7a-b indicates a net export of total energy, $gz + c_pT + Lq$, of about 2.3 units in both December and January. This amount represents the contribution of the Caribbean atmosphere for balance of heat in other areas. It comes out of the contribution by the standing eddy terms, and consists essentially of divergence of latent heat in the lower half of the troposphere. This divergence by the eddy terms is a result of the increase in energy content downstream along the trades through pick-up of latent heat from the ocean and is confirmation of the well-known role of the trades as accumulators and exporters of latent heat.

The net divergence of heat from the Caribbean amounts to about 28 percent of the total transfer (Q_a) from the ocean. Riehl et al. [41] calculated a net export of about 20 percent of the heat transfer (Q_a) in the lower 10,000 feet of the North Pacific trades.

5. BALANCE OF KINETIC ENERGY

For horizontal motion, equation (11) becomes, after transformation to an energy equation,

$$\rho \frac{dK}{dt} = - \vec{V} \cdot \nabla p + \rho \vec{V} \cdot \vec{F} , \qquad (24)$$

where K denotes kinetic energy. With differential expansion and use of the continuity equation, we get

$$\frac{\partial}{\partial t} \rho K + \nabla \cdot \rho K \vec{V} = - \nabla \cdot \vec{V} p + p \nabla \cdot \vec{V} + \rho \vec{V} \cdot \vec{F} \qquad (25)$$

Integration over the volume α and use of the divergence theorem, gives

$$\int_\alpha \frac{\partial}{\partial t}\rho K d\alpha + \iint_\sigma \rho K c_n ds dz = -\iint_\sigma c_n p ds dz + \int_\alpha p \nabla \cdot \vec{V} d\alpha + \int_\alpha \rho \vec{V} \cdot \vec{F} d\alpha \tag{26}$$

The first term on the left vanishes in the quasi-steady state conditions considered here. The second integral on the left represents the flux of kinetic energy across the lateral boundary. On the right, the first integral denotes the redistribution of kinetic energy due to boundary pressure work; the second represents production inside the volume; and the third, the kinetic energy sink due to frictional dissipation (Starr [45]).

The evaluation of these terms is discussed below. Under the present observational system, it is customary to work with geopotential heights instead of pressure. If the hydrostatic equation is introduced in the first term on the right of equation (24), the transformations can proceed in the same fashion and equation (26) takes the form

$$\int_\alpha \frac{\partial}{\partial t}\rho K d\alpha + \iint_\sigma \rho K c_n ds dz = \iint z g \rho c_n ds dz - \int_\alpha \rho g z \nabla \cdot \vec{V} d\alpha + \int_\alpha \rho V \cdot \vec{F} d\alpha \tag{27}$$

The first two terms on the right of equation (27) do not have now exactly the same connotation as their corresponding terms in equation 26. For computational purposes, it is convenient to group these two terms together; their sum will be referred to as the net production of kinetic energy.

a. <u>TRANSPORT OF KINETIC ENERGY</u>: Evaluation of the transport term followed the same procedure described for other energy parameters. With the hydrostatic equation and expansion into the contributions of mean and "eddy" terms, we have

$$\iint_\sigma c_n K ds \frac{dp}{g} = \iint_\sigma \bar{c}_n \bar{K} ds \frac{dp}{g} + \iint_\sigma \overline{c_n' K'} ds \frac{dp}{g}. \tag{28}$$

Here the average monthly kinetic energy at each station is defined as $\frac{1}{N}\sum_{i=1}^{31} \frac{1}{2} V_i^2$, rather than as the square of the monthly resultant wind. The rest of the processing for the flux evaluation proceeds as described earlier.

It turns out that the Caribbean - considering monthly data only - acts as an exporter of kinetic energy in both months (table 11). In December this export results mainly from the fact that the air leaves the area with a marked jet-stream-like velocity concentration over the Lesser Antilles. During January this tendency is less pronounced. To get an idea of the magnitude of this export, we can compare the kinetic energy export with the frictional dissipation inside the volume (table 13) and notice that the export in December would be sufficient to balance the frictional loss over an area larger than the Caribbean Sea.

b. <u>PRODUCTION OF KINETIC ENERGY</u>: The boundary term can be processed in the same fashion as the other flux integrals. One can readily see that this term represents a flux of potential energy across the boundary. Then

Table 11. - Lateral flux of kinetic energy. Units: 10^{18} ergs sec.$^{-1}$ positive sign stands for export, negative for import.

P (100's mb.)	December 1956			January 1957		
	By mean motion	By eddy motion	Net	By mean motion	By eddy motion	Net
1-2	-5.22	14.58	9.36	-1.55	6.75	5.20
2-3	-9.13	35.01	25.88	-12.87	13.19	0.32
3-4	0.52	4.96	5.48	0.44	1.08	1.52
4-5	0	-0.27	-0.27	0.31	0.25	0.56
5-6	-0.18	-0.14	-0.32	-0.30	0.79	0.49
6-7	0.34	0.39	0.73	0.64	1.86	2.50
7-8	1.16	0.12	1.28	1.03	2.14	3.17
8-9	1.33	-0.12	1.21	2.00	1.57	3.57
9-10	0.70	1.62	2.32	1.57	3.67	5.24
1-5	-13.83	54.28	40.45	-13.67	21.27	7.60
5-10	3.35	1.87	5.22	4.94	10.03	14.97
1-10	-10.48	56.15	45.67	-8.73	31.30	22.57

$$\iint_{\sigma} gzc_n ds \, \frac{dp}{g} = \iint_{\sigma} g \, \bar{z} \, \overline{c_n} \, ds \, \frac{dp}{g} + \iint_{\sigma} g \overline{z' c_n'} \, ds \, \frac{dp}{g} \tag{29}$$

The other term requires volume integration and this cannot be accomplished with boundary data alone. Partial evaluation is possible through the following transformation. At first, mean area values of the divergence and geopotential heights are defined:

$$\nabla \cdot \vec{V} = \widehat{\nabla \cdot \vec{V}} + (\nabla \cdot \vec{V})^*$$

$$gz = g\,\hat{z} + gz^* \tag{30}$$

where the circumflex denotes area averaging and the asterisk denotes deviation from the area average. Then

$$\iint_{\alpha} gz \, \nabla \cdot \vec{V} \, dA \, \frac{dp}{g} = A \int g \, \hat{z} \, \widehat{\nabla \cdot \vec{V}} \, \frac{dp}{g} + A \int g \, z^* \, \widehat{(\nabla \cdot \vec{V})^*} \, \frac{dp}{g}. \tag{31}$$

From the definition of divergence,

Table 12. - Production of kinetic energy. Units: 10^{18} ergs sec.$^{-1}$

P (100's mb.)	December 1956 By mean motion	December 1956 By eddy motion	December 1956 Net	January 1957 By mean motion	January 1957 By eddy motion	January 1957 Net
1-2	-0.28	25.57	25.29	-0.33	-4.46	-4.79
2-3	-0.59	61.06	60.47	-1.68	16.87	15.19
3-4	0.17	6.03	6.20	0.06	-2.99	-2.93
4-5	0	-1.05	-1.05	0.13	3.71	3.84
5-6	0.15	3.09	3.24	-0.13	4.56	4.43
6-7	0.26	6.32	6.58	-0.04	7.55	7.51
7-8	1.29	14.78	16.07	0.11	13.78	13.89
8-9	1.03	24.52	25.55	-0.37	24.33	23.96
9-10	0.18	39.78	39.96	-1.99	38.35	36.36
1-5	-0.70	91.61	90.91	-1.82	13.13	11.31
5-10	2.91	88.49	91.40	-2.42	88.57	86.15
1-10	2.21	180.10	182.31	-4.24	101.70	97.46

$$\widehat{\nabla\cdot\vec{V}} = \frac{\bar{c}_n S}{A} \tag{32}$$

and equation(31) is modified to

$$\iint z\nabla\cdot\vec{V}\,dA\,dp = S\int\hat{z}\,\bar{c}_n\,dp + A\int\widehat{z^*(\nabla\cdot\vec{V})^*}\,dp. \tag{33}$$

With these transformations, the production of kinetic energy (P_K) becomes

$$P_K = S\int\bar{c}_n(\hat{z}-\bar{z})\,dp - S\int\overline{c'_n z'}\,dp + A\int\widehat{z^*(\nabla\cdot\vec{V})^*}\,dp. \tag{34}$$

In this equation the first two terms on the right-hand side can be evaluated, $\hat{z}$ being determined from analyzed charts. The third term cannot be computed, as this would require knowledge of the divergence field inside the area, which, even on a monthly basis, cannot be obtained with much hope of accuracy.

Table 12 shows the evaluation of the first two terms. Production takes place in both months from the contribution of the eddy term. The contributions by the mean motion at the boundary and inside the volume cancelled one another.

Table 13. - Kinetic energy balance. Units: 10^{18} ergs sec.$^{-1}$

	December 1956		January 1957	
	Sources	Sinks	Sources	Sinks
Production	182		97	
Divergence		46		23
Frictional dissipation		40		55
Residual		96		19

Thus, the important effect is that depending on $\overline{c_n'z'}$, which may be characterized as a pressure head. Around the boundary, air enters at a higher pressure (or height of an isobaric surface) than it leaves - this is the principal source of kinetic energy.

c. DISSIPATION OF KINETIC ENERGY: The expression for the dissipation of kinetic energy (D_K), the last integral on the right side of equation 27, can be developed as follows. Considering horizontal motion in natural coordinates, the well-known expression for the frictional force takes the form

$$\vec{F} = \frac{1}{\rho}\vec{t}\frac{\partial}{\partial z}\tau_{zt} + \frac{1}{\rho}\vec{n}\frac{\partial}{\partial z}\tau_{zn}, \tag{35}$$

where τ denotes the frictional stress, and $\vec{t}$ and $\vec{n}$ are unit vectors tangent and normal to the velocity vector. Then

$$D_K = \int_\alpha \rho\,\vec{V}\cdot\vec{F}\,d\alpha = \int_\alpha V\frac{\partial}{\partial z}\tau_{zt}\,d\alpha\,. \tag{36}$$

By differential expansion,

$$D_K = \int_\alpha \frac{\partial}{\partial z}V\tau_{zt}\,d\alpha - \int_\alpha \tau_{zt}\frac{\partial V}{\partial z}\,d\alpha. \tag{37}$$

Mean values for the horizontal integrals over the area A are introduced and

$$D_K = A\int_0^\infty \frac{\partial}{\partial z}\widehat{V\tau_{zt}}\,dz - A\int_0^\infty \widehat{\tau_{zt}\frac{\partial V}{\partial z}}\,dz. \tag{38}$$

Integrating from bottom to top of the troposphere, assuming no stress at the tropopause,

$$D_K = -A\,(\widehat{V\tau_{zt}})_0 - A\int_0^\infty \widehat{\tau_{zt}\frac{\partial V}{\partial z}}\,dz, \tag{39}$$

where the subscript zero denotes conditions at the surface of the earth. The first term in equation (39) can be interpreted as the dissipation of kinetic energy due to surface friction, and the second as the internal dissipation within the atmosphere (Palmén [27]).

An estimate of the surface dissipation can be obtained by introducing the expression for the surface stress

$$\tau_o = \rho_o k v_o^2, \qquad (40)$$

where k denotes the drag coefficient. Then

$$D_{K_o} = - k \rho_o \widehat{v_o^3} A. \qquad (41)$$

For the case of a water surface the wind velocity is taken at ship's deck level. The value of k, according to Sverdrup et al. [47], should be about 2.5 x 10^{-3}. More recently Sheppard and Omar [42] obtained values of 1.0 to 1.6 x 10^{-3} in the trade region of the Pacific Ocean. Palmén and Riehl [28] calculated 1.1 to 1.4 x 10^{-3} for the outer zone of tropical hurricanes. In view of these results, a value of 1.3 x 10^{-3} was considered reasonable for this study.

Data for computing v_o^3 were taken from the tabulation of ship reports inside the Caribbean Sea during the two months of study mentioned earlier. Each individual wind value was cubed and the mean-cube for the whole sample was determined. The mean air density at the surface was taken as 1.16 x 10^{-3} gm. cm^{-3}. With these data, evaluation of D_{K_o} yields:

<u>Surface Frictional Dissipation</u> (Unit: 10^{18} ergs sec.$^{-1}$)

December	19.9
January	27.4

The second term in equation (39) can not be evaluated from boundary data. The assumption is made here, after Brunt [5] and Palmén [27], that the internal dissipation in the atmosphere is about equal to the dissipation at the ground. With this assumption the dissipation of kinetic energy in the entire volume is approximately double the values listed above.

d. <u>BALANCE OF KINETIC ENERGY</u>: Table 13 summarizes the kinetic energy calculations. The imbalance is rather serious in December, but not in January. It is not considered likely that the imbalance results from the production term. The missing member in equation (34) essentially gives the contributions of cyclones and anticyclones inside the boundary to kinetic energy production. Considering the steady character of the flow in the dry season, cyclones and anticyclones will be few and should carry little weight in the balances. A more important source of error might be the internal dissipation of kinetic energy; but this term would have to be underestimated by a factor of five in December for balance to be achieved and this is unlikely. More probably it is the export that is underestimated. While quantities such as gz + c_pT, and even Lq, are nearly constant from day to day in the dry season, the same is known not to be true for K. As shown initially, the directional stability of the winds is high, but time-cross sections reveal large fluctuations of wind speed, hence also of K - xt with nearly constant wind direction in the high troposphere. Therefore, while one might expect the daily variations of total energy to make a minor contribution to the net export, the same may not be true for kinetic energy. Unfortunately, the time and facilities necessary for computing the energy export from daily winds were not available. This experiment should be made in order to test whether table 13 would essentially balance if the daily transport calculations were given.

Assuming the foregoing to hold, the following picture emerges. Production of kinetic energy takes place in the trades and in the upper return flow by means of pressure heads, i.e., inflow at high and outflow at low pressures in the periphery. This production is used partly to maintain the circulation against friction, and partly the kinetic energy gained at high levels is exported northward and, presumably, becomes part of the subtropical jet stream system.

6. VERTICAL VARIATIONS IN THE HEAT BALANCE AND MECHANISMS FOR VERTICAL TRANSFER

In Section 4 an approximate heat balance was obtained for the entire troposphere. It is necessary, in addition, that balance exist also within each tropospheric layer, achieved by adequate mechanisms for heat transfer from layers of heat surplus to those of deficit. Further analysis of these mechanisms is of interest. For instance, sensible heat from the ocean is introduced into the atmosphere at the surface, while precipitation heat is released mostly in the layers from 900 to 500 mb. Under the conditions existing over the Caribbean during the dry season, upward flow of heat beyond the 500-mb. level is difficult to visualize. The total energy content (fig. 17) has a minimum near 700 mb., so that ascent of air with the properties of the mean atmosphere across the 500-mb. level would not produce heating; rather, cooling would occur in the upper troposphere, unless a particular selective process is in operation.

In a recent study of the heat balance of the equatorial trough zone, Riehl and Malkus [38] encountered this problem when they obtained a significant sink of heat above the 500-mb. level that should be balanced by transfer from below. The writers have suggested that this can only be accomplished by warm, essentially undiluted, updrafts in the interior of thunderstorm cells. Such a mechanism, clearly, will be only a minor contributing factor in the Caribbean dry season, since cumulonimbus activity, although encountered in disturbances, is then at a minimum. Heat balance in the high troposphere must be produced by essentially compression heating.

For this part of the analysis the vertical heat transfer at each level (table 8) was evaluated from the values of vertical mass flow and heat content. Balance was then established at each layer by continuity and the heat surplus or deficit determined.

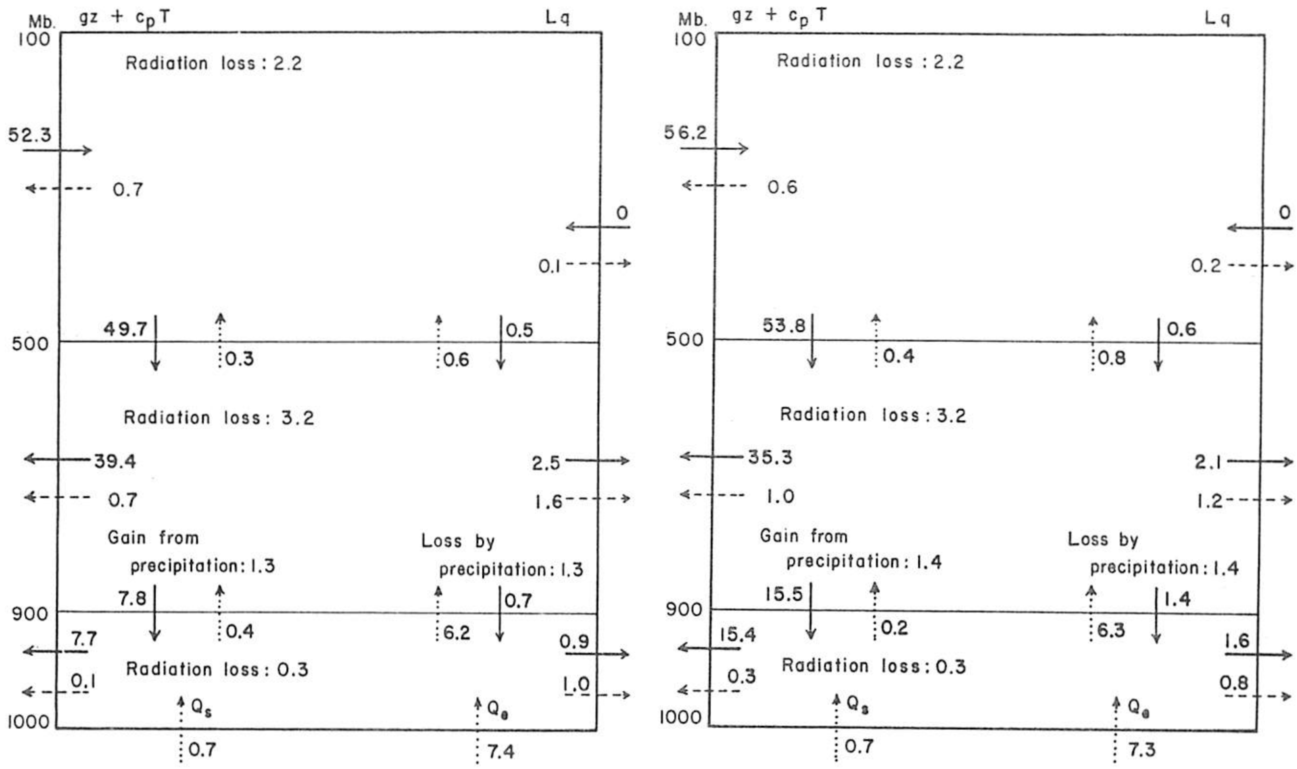
a. <u>VERTICAL DISTRIBUTION OF COLD SOURCES</u>: The results (table 14) reveal that the heat deficit occurs mostly in the layers below the 500-mb. level. Over 80 percent of the total in December is observed below the 700-mb. level and only 11 percent above the 500-mb. level. The total deficit for the whole troposphere, 1.73 units, is balanced by the heating from precipitation, LP, and the sensible heat flux from the ocean, Q_s. The transfer from the ocean enters

Table 14. - Vertical distribution of cold sources. Units: 10^{18} cal. day^{-1}
Positive sign stands for heat divergence (cooling); negative for heat convergence (warming).

P (100's mb.)	December 1956						January 1957					
	Radiational cooling	Heat gain by mean motion	Heat loss by eddy motion	Net cooling	Sensible heat from ocean	Precipitation heating	Radiational cooling	Heat gain by mean motion	Heat loss by eddy motion	Net cooling	Sensible heat from ocean	Precipitation heating
1-2	0.27	-0.40	-0.04				0.27	-0.15	0.01			
2-3	0.48	-0.78	0.42				0.48	-0.67	0.01			
3-4	0.64	-0.74	0.19				0.64	-0.77	0.35			
4-5	0.75	-0.71	0.11				0.75	-0.79	0.28			
5-6	0.81	-0.87	0.09				0.81	-0.87	0.19			
6-7	0.86	-0.89	0.10				0.86	-0.97	0.14			
7-8	0.81	-0.52	0.24				0.81	-0.78	0.29			
8-9	0.54	-0.25	0.26				0.54	-0.40	0.38			
9-10	0.32	-0.06	0.10				0.32	-0.08	0.30			
1-5	2.14	-2.63	0.68	0.19			2.14	-2.38	0.65	0.41		
5-9	3.02	-2.53	0.69	1.18		-0.98	3.02	-3.02	1.00	1.00		-1.19
9-10	0.32	-0.06	0.10	0.36	-0.75		0.32	-0.08	0.30	0.54	-0.76	

the atmosphere at the surface; about one-half supplies the requirement for the surface layer, and the other half is available for transport upward across the 900-mb. level. Above the 900-mb. level the deficit is balanced mostly by the precipitation heating. About 70 percent of this supply is needed in the layer 900-700 mb. where, interestingly enough, most of the trade cumuli are present. The evaluations for January show essentially the same results as in December. About 70 percent of the net cooling occurs in the layers below the 700-mb. level and one-fifth above the 500-mb. level.

There are two very favorable observations that stand out in the results above: one, that the requirements in the surface layer are satisfied by the transfer from the ocean alone (since the 900-mb. level lies near the average base of the cumuli, no local release of heat of condensation can be considered below the 900-mb. level); and two, that the bulk of the requirement for precipitation heating falls in the cloud layer, where, presumably, most of it is released. This can be interpreted as evidence that the methods of analysis employed here are generally sound, and the numerical values obtained are generally satisfactory.



A
December 1956

B
January 1957

Figure 18. - Heat balance for atmospheric volume above Caribbean Sea. Units: 10^{18} cal.day^{-1}. Fluxes of sensible heat ($gz + c_pT$) shown in the left margin, of latent heat (Lq) on the right. Lateral and vertical transfers by mean motion indicated in solid arrows; lateral transfer by standing eddy terms in dashed; vertical transfers by diffusion and/or convection in dotted arrows.

b. <u>BALANCE OF TOTAL ENERGY</u>: The balance of total energy, $gz + c_pT + Lq$, is presented graphically in figure 18 for three sections: the surface layer (1000-900 mb.), the rain area (900-500 mb.), and the upper troposphere (500-100 mb.). In order to make it more illustrative, the sources and sinks of sensible and latent heat are indicated separately on the left and right sides of the diagram. In the discussion of the tropospheric budget (tables 9 and 10) an imbalance was noted in the estimates of precipitation obtained from the sensible and latent heat balances. For illustration of a combined budget it is convenient to have the numbers exactly equal. This is achieved in figure 18 by a slight upward adjustment (6 percent in December) in the radiational cooling and a downward scaling (4 percent) in the heat transfer from the ocean. As a result, precipitation heating comes out as 1.3 units for December, and 1.4 for January; that is, the average between the two previous estimates. The upward transfers across the 900 and 500-mb. surfaces by processes other than the mean motion have been determined as residuals.

The balance of sensible heat has been discussed in the previous section; it is dominated by the mean motion, contributions by the eddy motions being generally small. The situation is different in the latent heat budget, which shows sizable contributions by the eddy terms. About 82 percent of the transfer of latent heat from the ocean is exported, and the remainder is precipitated back. One-third of the export occurs in the surface layer and two-thirds in the rain area. There is a requirement for flow of latent heat upward into the rain area of 6.2 units in December and 6.3 units in January; the required flow into the upper troposphere is only 0.6 units (8 percent of Q_e) in December, and 0.8 units (11 percent of Q_e) in January. The vertical transports of total energy across the 900 and 500-mb. levels are, respectively, 6.6 and 0.9 units in December and 6.5 and 1.2 units in January.

The precipitation heating constitutes a gain in the sensible heat balance, and a sink in latent heat. When the total energy is considered, the precipitation cancels out and balance is achieved between the net transfer from ocean, Q_a, the lateral flux, and the radiational cooling. Figure 18 shows a net export of 2.4 units of total energy, which, as discussed previously, is the net contribution of the Caribbean atmosphere for balance in other areas (the numerical difference between this and the previous estimate is due to the rounding off of numbers).

c. <u>MECHANISMS FOR VERTICAL HEAT TRANSPORT</u>: The previous evaluations indicate requirements for a sizable amount of heat transport from the surface layer into the rain area, and a comparatively smaller amount from the rain area into the upper troposphere, which must be accomplished by processes other than the mean motion. Two mechanisms may be considered: diffusion by small-scale turbulent eddies, and vertical currents in the convective scale. It has been shown (Riehl [35]) that over the tropical oceans turbulent eddies are effective only in the layer near the surface (the homogeneous layer); their presence decreasing rapidly into the cloud layer. In the cloud layer and above, transfer by convective currents should predominate. In any case, as already mentioned, turbulent eddies can transfer heat upward near the 900-mb. level, but not near the 500-mb. level.

Assuming that transfer into the cloud layer and above is accomplished only by convective updrafts, we can investigate whether the amount of cloudiness generally present over the Caribbean Sea in winter meets the requirements shown in figure 18. An estimate of the amount of clouds needed can be made using the procedures outlined by Riehl [35] and by Riehl and Malkus [38]. The net heat flux at a given level (the 900-mb. level, for example) can be expressed by the relation

$$H = \hat{M}_z \hat{Q} + \widehat{M_z' \delta Q} \qquad (42)$$

where $\hat{Q}$ is the mean heat content at that level, $\hat{M}_z'$ is the net mass in the updrafts (equal also to their compensating downdrafts), and δQ is the difference in heat content between the updrafts and down drafts. The first term represents the transfer by the mean motion, the second the transfer by convection. Near the 900-mb. level we are dealing essentially with the average cumuli, in which the updrafts are located in a narrow current near the center of the clouds, while the compensating downdrafts are spread over a wider area, mostly in the clear spaces between clouds. The "parcel" idea is invoked and the energy content in the updrafts is assumed to be very close to the content, Q_o, which the parcel possessed when it started to rise near the surface. The mean heat content in the downdrafts, is, of course, less than that of the updrafts, and also less than $\hat{Q}$.

For the case of the heat flux across 900 mb. in December, it is required that $M_z' \, \delta Q = 6.6$ units. As a first approximation we can take $\delta Q = Q_o - \hat{Q}$, that is, the difference between the content at the surface and that at the level in question. From the mean surface conditions over the Caribbean Sea for December 1956 (air temperature 79.2°F., dewpoint 72.8°F.), Q_o is calculated as 81.9 cal.gm.$^{-1}$; $\hat{Q}$, from figure 17 is approximately 79.5 cal.gm.$^{-1}$, and the difference δQ is about 2.4 cal.gm^{-1}. The net upward mass in the updrafts M_z' comes out as 2.7×10^{18} gm.day^{-1}, which represents a very large proportion of the mean mass flow at the level. When viewed in this manner the mean vertical mass flux represents a very small difference between two large terms. The area occupied by the updrafts is given by the relation

$$A' = \frac{M_z'}{\rho w'} .$$

If w' is taken as 2 m.p.s. (Riehl [35]), the area A' is 1.6×10^{14} cm.^{2}, which amounts of 0.7 percent of the total area of the Caribbean Sea. Riehl [35] estimated the area of the updrafts in tropical cumuli as 2 percent of the total cloud area. With this relationship, the area above indicates a mean cloudiness of 35 percent, a very realistic value. However, this calculation is quite sensitive to variations in some of the terms, and represents only a rather crude approximation. For example, the energy difference between the updrafts and downdrafts is probably an underestimate, while the vertical motion w' might be too large for the majority of the clouds. Nevertheless the calculation shows that we are dealing with realistic values, and it illustrates the effectiveness of convective updrafts in the redistribution by the atmosphere of the energy received from the ocean.

The same type of calculation can be made for the transfer across the 500-mb. level. There, however, the quantities involved are rather small, and could probably be considered as within the limitations of the data. One difficulty we encounter at the upper level is that there is no good basis for comparison with what might be considered a realistic picture. No information is available in regard to the number and stage of development of clouds that normally penetrate the upper troposphere. As discussed previously, most of the heat transfer, if real, must be accomplished by undiluted updrafts in the interior of thunderstorm cells (Riehl and Malkus [38]). Presumably some contribution is made also by cloud formations in the category of "bulging cumulus" or"cumulus congestus" that penetrate the high troposphere, but do not attain the cumulonimbus stage. Clouds of this type, and also cumulonimbus, are observed over the Caribbean Sea in winter during disturbed periods, but their occurrence should be generally at a minimum. An evaluation for the upper level in December shows that the required area, A', of undiluted updrafts is of the order of one ten-thousandth (10^{-4}) of the total Caribbean area. This is one order of magnitude less than that calculated by Riehl and Malkus [38] for the equatorial trough region. One would normally expect the convective activity over the Caribbean Sea in winter to be significantly less than that over the equatorial region; therefore, the above results are probably realistic. On the other hand, the evaluations presented here are subject to inaccuracies, the extent of which is very difficult to establish. Therefore, exact results should not be expected in all instances. The calculations, although rather crude, suggest that the balance relationships shown in figure 18 give a correct picture of the conditions over the Caribbean atmosphere in winter.

7. ON THE MAINTENANCE OF THE TROPOSPHERIC CIRCULATION

a. ENERGY CHANGES ALONG THE TRADES: In their study of the northeast trades of the Pacific Ocean, Riehl and Malkus [37] showed, through an evaluation of the energy balance along the trajectory, that the trades are exporters of sensible heat, in addition to latent heat, and under those conditions the wind circulation can be maintained by local energy transformations. There is an increase in potential temperature downstream, which comes from the sensible heat exchange with the sea surface; thus, the air leaving the region has higher energy content than that entering upstream, and heat is exported to the outside. At the same time the heat gained acts to maintain a pressure field characterized by lower pressures at the downstream end of the trajectory. It is well-known that the flow in the lower layers of the trades is non-geostrophic (see fig. 8). The geostrophic wind is at a maximum at the surface, but due to friction the actual wind is at a minimum at the surface and increases to a maximum near the 900-mb. level. The balance of forces along the trajectory in the friction layer is determined essentially by the pressure gradient force acting against the frictional drag. Therefore, the creation and maintenance of the pressure difference along the current become important for driving the circulation. As discussed by Riehl and Malkus, the heat source supplied by the flux from the ocean can support the lower pressures at the downstream end. In the absence of the heat source the changes would be essentially adiabatic, the temperatures at the downstream side would be lower than in the actual case, and maintenance of the pressure field would have to depend on importation of energy from the outside.

It is of interest to investigate the results for the Caribbean Sea to see if the same conclusion is valid. The results in table 7 show export of sensible heat in the trade layers by the standing eddy terms. Since this contribution is due mostly to the difference in properties between the upstream and downstream sides, it indicates higher energy content in the western end of the Caribbean. Actually the contributions by eddy terms show increase in enthalpy and, consequently, internal energy and decrease in potential energy downstream (table 15). However, the above results give the energy changes along the isobaric layers and, it is better for our purpose here to evaluate the energy balance along the trajectory. As shown in figures 4 and 6 the horizontal

Table 15. - Lateral flux of enthalpy, potential and internal energy. Units: 10^{18} cal.day^{-1}. Positive sign stands for divergence, negative for convergence.

	Flux of P. E.			Flux of enthalpy			Flux of I. E.		
P (100's mb.)	By mean motion	By eddy motion	Net	By mean motion	By eddy motion	Net	By mean motion	By eddy motion	Net
				DECEMBER 1956					
1-2	-9.01	-0.05	-9.06	-13.16	0.01	-13.15	-9.31	0.01	-9.30
2-3	-11.00	-0.03	-11.13	-23.36	0.55	-22.81	-16.55	0.39	-16.16
3-4	1.08	-0.01	1.07	3.16	0.20	3.36	2.23	0.14	2.37
4-5	0	0	0	0	0.11	0.11	0	0.08	0.08
5-6	-0.64	-0.01	-0.65	-3.47	0.10	-3.37	-2.46	0.07	-2.39
6-7	0.94	-0.01	0.93	7.13	0.11	7.24	5.04	0.08	5.12
7-8	1.63	-0.03	1.60	18.18	0.27	18.45	12.88	0.19	13.07
8-9	0.77	-0.05	0.72	14.84	0.31	15.15	10.52	0.22	10.74
9-10	0.15	-0.08	0.07	7.55	0.18	7.73	5.35	0.13	5.48
1-5	-18.93	-0.19	-19.12	-33.36	0.87	-32.49	-23.63	0.62	-23.01
5-10	2.85	-0.18	2.67	44.23	0.97	45.20	31.33	0.69	32.02
1-10	-16.08	-0.37	-16.45	10.87	1.84	12.71	7.70	1.31	9.01
				JANUARY 1957					
1-2	-3.61	0.01	-3.60	-5.22	0	-5.22	-3.70	0	-3.70
2-3	-17.87	-0.03	-17.90	-37.89	0.04	-37.85	-26.82	0.03	-26.79
3-4	1.08	0.01	1.09	3.15	0.34	3.49	2.23	0.24	2.47
4-5	0.83	-0.01	0.82	3.33	0.29	3.62	2.36	0.21	2.56
5-6	-0.64	-0.01	-0.65	-3.47	0.20	-3.27	-2.46	0.14	-2.32
6-7	0.95	-0.02	0.93	7.13	0.16	7.29	5.05	0.11	5.16
7-8	0.98	-0.03	0.95	10.91	0.32	11.23	7.72	0.23	7.95
8-9	0.96	-0.05	0.91	18.50	0.43	18.93	13.10	0.30	13.40
9-10	0.30	-0.08	0.22	15.08	0.38	15.46	10.67	0.27	10.94
1-5	-19.57	-0.02	-19.59	-36.63	0.67	-35.96	-25.93	0.48	-25.46
5-10	2.55	-0.19	2.36	48.15	1.49	49.64	34.08	1.05	35.13
1-10	-17.02	-0.21	-17.23	11.52	2.16	13.68	8.15	1.53	9.67

Table 16. - Energy changes along the trajectory of the trades. Units: 10^9 cal. day^{-1}.

	December 1956 (Layer 1000-800 mb.)				January 1957 (Layer 1000-820 mb.)			
	+	-	+	-	+	-	+	-
Increase in heat content	3.9				3.7			
Work by pressure forces		2.9				3.3		
Increase in internal energy			7.2				8.1	
Decrease in potential energy				6.2				7.7
Balance	1.0		1.0		0.4		0.4	

trajectory in the center of the trade current goes essentially from Guadeloupe, on the east side of the Caribbean, to San Andres Island or Balboa, C.Z., on the west side, and the energy changes along the flow are given by the difference in the observations at those stations. The wind speeds (figs. 3 and 5) are such that it would take about two days for the air to move from the east to the west side of the Caribbean Sea. The average sinking motion is given in table 8.

For this analysis, equation (15) can be modified to

$$\rho h - \rho \frac{d}{dt} \left(\frac{p}{\rho} \right) = \rho \frac{d}{dt} \left(c_v T + gz \right), \tag{43}$$

by use of the relation

$$c_p \frac{dT}{dt} = c_v \frac{dT}{dt} + \frac{d}{dt} \left(\frac{p}{\rho} \right),$$

assuming also steady-state conditions and no contributions from kinetic energy and frictional dissipation. As indicated throughout this report, equation (43) defines the energy balance satisfactorily. The heat source, h, can be expressed by the relation

$$h = c_p \frac{T}{\theta} \frac{d\theta}{dt}, \tag{44}$$

where θ denotes potential temperature. Equation (43) is then integrated over a volume of unit thickness, taken from the upstream to the downstream side of the trades, and from the surface to some upper level. Space sections of temperature and potential temperature were constructed from the east to west sides of the Caribbean; the air trajectories were drawn by considering the horizontal and vertical motions, and the terms in equation (43) were measured along the trajectories over intervals of 24 hours. The observations showed an increase in potential temperature downstream of 1.5-2.0°K. in the layers from the surface to around 800 mb. Above the 800-mb. level the downstream gradient in potential temperature decreased, being almost zero near 700 mb. Therefore, the heat source or increase in potential temperature along the trajectory was restricted to the layer from the surface to about 800 mb.

Table 17. - Energy changes along the trajectory of the trades with the assumption of no export of sensible heat. Units: 10^9 cal. day^{-1}.

	December 1956				January 1957			
	(Layer 1000-800 mb.)				(Layer 1000-820 mb.)			
	+	-	+	-	+	-	+	-
Increase in heat content	0				0			
Work by pressure forces		1.8				2.2		
Increase in internal energy			4.4				5.5	
Decrease in potential energy				6.2				7.7
Balance		1.8		1.8		2.2		2.2

In an analysis of this type it would have been better to retain the identity of the trade inversion, located in the mean near the 800-mb. level, as was done in the Pacific study. Then the thermal field in the low levels would be more characteristic of the normal conditions. In spite of this deficiency the integration of equation (43) in the layer from 1000 to 800 mb. gave a satisfactory balance.

The results (table 16) are similar to those obtained in the Pacific trades. There is a net heat source larger than the work done by pressure forces, or an increase in internal energy greater than the loss of potential energy. The gain in heat downstream implies export of sensible heat, and shows the existence of a source of heat that can support the pressure field and the balance of forces in the friction layer. Above the 800-mb. level the effect of the heat flux from the ocean decreases rapidly; the potential temperature lines become nearly horizontal, and a heat gain downstream along the trajectory is no longer obtained. The data for January 1957 give essentially the same picture; the increase in temperature and potential temperature downstream is about the same as in December, but the heat source is limited to a slightly smaller layer due to the larger sinking motion shown by the January data. Table 16 presents the integration of equation (43) over the layer 1000-820 mb., which represents the vertical extent of the heat source. The results otherwise are similar to those of December. The heat source is evidently restricted to the lower layers. In this respect we can notice that the influence of the frictional drag disappears rapidly above the 900-mb. level; the flow above becomes quasi-geostrophic, and the balance of forces must then be influenced by developments in other regions.

In the absence of the heat source the balance of equation (43) must be achieved between the potential energy and enthalpy only. In that case the increase in enthalpy downstream would be less than in the actual situation. The balance relationships under this assumption can be examined in table 17. The increase in enthalpy is then only 6.2 in December and 7.7 in January, which implies lower temperatures on the downstream side, and similarly smaller internal energy and work terms than in the actual situation.

When viewed in this manner the increase in heat content along the trades, due to the exchange with the sea surface, makes an important contribution to the balance of forces in lower levels, and the circulation can be sustained by local energy transformations. In the rest of the troposphere, as shown previously, importation of heat is required.

b. ENERGY CHANGES IN THE UPPER EQUATORIAL CURRENT: The energy changes in the upper current are given in the totals for the layer 100-500 mb. Variations in latent heat are negligible aloft; therefore, only the sensible heat transformations need to be considered. The region is an energy sink; importation of heat is required, and it comes about through the mean circulation. The variations in energy content along the current are similar to those of the trades. The contributions by the eddy terms, which depend mostly on the difference in properties between the two sides of the flow, indicate a net increase in sensible heat downstream, due to the increase in enthalpy (table 15). The potential energy decreases, while the kinetic energy increases along the horizontal trajectory.

The energy changes in the upper and lower-level circulations of the Caribbean that have emerged from this study show some interesting, and to some extent paradoxical, relationships. The tropospheric region as a whole requires an outside source of energy to maintain the local circulations and offset radiational losses. The source is actually available in the same volume, in the form of latent heat, but it cannot be completely processed and released locally. An outside agency is required to do the processing and return the energy in a more usable form. So, while the region as a whole exports heat, it is at the same time dependent on atmospheric processes outside the area for its own maintenance. The transformation of latent heat to useable forms of energy is supposedly accomplished in the equatorial region. We should expect to find a net divergence of sensible heat in the upper layers of the equatorial trough, which is, in fact shown in the evaluations by Riehl and Malkus [38].

One interesting aspect of the mean circulation in the Caribbean region described in this report is that the upper equatorial current seems to originate in the same geographical area into which the trades underneath flow. Not much information exists in regard to the synoptic developments that normally take place in that oceanic area, but studies by Alpert [1,2] show that the equatorial trough in the sector between longitudes 80° and 95° W. is located in winter near latitudes 3° - 5° N., just south of the Canal Zone. A possible connection between the trade current, the equatorial trough, and the upper westerlies in the region southwest of Panama can be easily visualized. This subject is not pursued here, but it constitutes an interesting situation that may have a bearing on the evolution of the tropical meridional circulation cell and merits further investigation, if possible by use of special aircraft-reconnaissance missions over the oceanic areas south of Panama.

8. SUMMARY AND SUGGESTIONS FOR FUTURE WORK

The computations of the energy balance of the tropospheric volume, and of the water body of the Caribbean Sea have served as a basis for a discussion of the mean wind circulation over this area in winter and of the energy transformations that take place in the oceans and atmosphere. A measure has been obtained of the contribution of the Caribbean Sea area as an energy source for maintenance of the general circulation. Some interesting results that have emerged from this study are summarized below:

1. The mean circulation over the Caribbean Sea in winter consists of a two-layer atmosphere with the trade current moving west-southwestward in the low levels, and a much stronger upper current centered near the 200-mb. level moving in almost exactly the opposite direction. These upper westerlies apparently originate in the equatorial area west of South America. They show in the monthly mean picture a vertical and lateral concentration of kinetic energy suggestive of jet stream currents observed farther north. Vertical wind profiles show a remarkable concentration of kinetic energy near the 200-mb. level with large values of vertical shear below and above.

2. Computations indicate that the tropospheric flow - except for the friction layer - is quasi-geostrophic. A northward-directed temperature gradient, which shows maximum values near the 400-mb. level, accounts for the decrease in easterlies and increase in westerlies with height. However, the meridional temperature gradient reverses near the 200-mb. level and in the layer above the 200-mb. level temperatures decrease equatorward. This coincides with a rapid decrease in the westerlies between 200 and 100 mb. The maximum in the meridional temperature gradient in the middle troposphere, to the best of our knowledge, has not been properly emphasized before. No attempt was made to inquire into causes; for one thing, the method of analysis used here, i. e., monthly data along a geographical reference system, is not the most effective to bring out clearly the relationships between the wind and thermal fields. Analysis of the data in a reference system relative to the center of the current would be more illustrative.

3. Data on water temperatures below the surface in the Caribbean Sea showed that the annual range in mean monthly temperatures observed at the surface decreased with depth and eventually disappeared at a depth of around 90 meters. The rate of change of heat content in the upper layers was evaluated and was found to be very important in determining the annual variations in the heat flux from the ocean to the atmosphere. This energy flux was measured as a residual in the oceanic heat balance and with turbulent transfer formulae using normal data for a sector of the Caribbean Sea. A fairly good agreement between the two evaluations was noted.

4. A satisfactory balance of energy for the Caribbean troposphere was established with monthly data without concern for the energy flux that could have resulted from time variations of a scale less than a month. However, this should not be interpreted to mean that time variations in the energy parameters in winter are negligible. The balance of sensible heat for the troposphere as a whole shows that the radiation losses inside the volume are compensated by importation of sensible heat through the mechanism of the mean

motion, which consists of convergence aloft and divergence at low levels, by the sensible heat transfer from the ocean, and by the heating due to condensation and precipitation. The moisture balance shows that about one-fifth of the flux from the ocean by evaporation is precipitated back and the rest is exported in the trade layers.

5. The results confirm the prevalent ideas concerning the role of the trades as accumulators and exporters of latent heat. Moisture content and temperature increase downstream along the trade current through pick-up from the ocean surface. A net permanent export of 2.4×10^{18} cal.day^{-1} in the form of latent heat in the trade layers was computed for the Caribbean Sea in December 1956 and January 1957. This energy is available for use in other regions of the globe and represents the contribution of the Caribbean Sea to the maintenance of the general circulation.

6. The distribution of the heat sources and sinks over different layers of the troposphere shows that the heat requirements in the lower troposphere are satisfied by the sensible heat from the ocean and the precipitation heating. The heat losses in the upper troposphere are balanced by the contribution of the mean motion through sinking and compression heating and little transfer is required from the lower levels. This picture differs from that derived by Riehl and Malkus [37] for the equatorial trough area, where heat balance in the layers above the 500-mb. level required a significant amount of vertical heat flow from the low levels.

7. The balance of total energy, $gz + c_pT + Lq$, has been illustrated graphically for three tropospheric layers: the surface, the rain area, and the upper troposphere. The requirements for vertical heat flow were discussed, as well as the mechanisms available for transfer. Rough calculations were used to emphasize the role of convective clouds in the vertical distribution of heat.

8. The analysis of the changes in heat content downstream along the Caribbean trades confirmed the results obtained by Riehl and Malkus [37] in the Pacific Ocean concerning the gain in sensible heat from the oceans by the trades and the role of this heat source in supporting the local circulation through maintenance of the pressure field.

9. The balance of kinetic energy showed that the production inside the volume by pressure forces accounts satisfactorily for the outward flux and the frictional dissipation. The production aloft takes place by a process in which the flow from the equatorial region enters the volume at higher geopotential levels than it leaves farther north. Kinetic energy is exported northeastward into the Atlantic Ocean, and, presumably, it goes into the subtropical jet stream system.

Several lines for further research are suggested by the work presented here. First of all, additional investigations of the wind flow in the upper levels of the Caribbean Sea and other sections of the hemisphere over shorter periods of time are required in order to appreciate fully the nature and role of the equatorial current aloft. A reference system in which the data are analyzed relative to the center of the current should be attempted instead of the geographical averaging used here. The relationship between the equatorial current over the Caribbean and the subtropical jet stream needs further elaboration. Progress should come from the detailed study of individual situations, if possible by supplementing the data from the aerological stations with reconnaissance missions over the oceanic areas.

The energy cycle illustrated by the calculations suggests that studies of time variations in the energy output in the trades and possible reflections in the properties of the wind circulations may be successful in the Caribbean area. This would be more or less in line with the establishment of a tropical index cycle, as suggested by Riehl [36].

Further testing of the methods of analysis employed here is also desirable, although the results of this study were satisfactory, and previous attempts by Riehl [33] and by Gangopadhyaya and Riehl [14] were also successful. Other applications to different types of systems, and over other geographical areas, should be made in order to assess their usefulness and limitations.

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