Sensible and latent heat exchange in low latitude synoptic scale systems

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ABSTRACT

The transfer equations are applied to a set of forty-six days of data collected from an oceanographic research vessel on a station in the low latitude western Atlantic. In this region the lapse rate and shear in the boundary layer of the atmosphere is such that the transfer equations may be applied with a reasonable degree of confidence. In particular, the most accurate results are likely when large transfers occur, the least accurate when small transfers occur. Under these conditions a linear dependence of the eddy exchange coefficient with wind speed and stability were used to compute latent and sensible heat transfer from time averages over one hour of specific humidity, temperature and wind speed at 6 m and at the sea surface. Sea surface temperatures were measured at 10 cm and compared with infrared radiometer measurements.

Individual synoptic scale systems that moved over or close to the point of observation were examined. Over limited regions of these disturbances latent and sensible heat transfers were found to increase by an order of magnitude. Integrated over the whole disturbance the energy flux is found to be double the undisturbed values. By using both streamline analysis and a rainfall amount and occurrence technique, the frequency and size of synoptic systems are determined. This makes possible the construction of summer, winter and annual maps of latent and sensible heat transfer for the tropical Atlantic. Significant differences are found when they are compared with results of earlier workers. The role of synoptic scale disturbances in the atmospheric energy balance is emphasized by these results.

Introduction

Considerations of energy sources, conversions and transports in the earth-atmosphere system soon lead to the conclusion that the tropical oceans play an extremely important role in the heat and energy balance of the atmosphere. The greater part of the incoming short wave solar radiation in either direct or diffuse form is absorbed at the earth's surface. Before this energy can be utilized by the atmosphere, it must be transferred across the earth-atmosphere interface. Therefore, the atmosphere is fuelled mainly from below. As shown by Malkus (1962, p. 93), and others, the most important transfer process at the earth-atmosphere interface is evaporation which provides about half of the atmosphere's fuel in the latent form of water vapor. More than half of this water vapor fuel is supplied to the lower troposphere by the tropical oceans between 30 N and 30 S latitude. If we add to these considerations the observation that the overall radiation balance of the earth-atmosphere system is positive between about 35 N and 35 S latitude, and negative poleward, the validity of the opening statement is justified in general terms.

In the tropical regions of heat input the net transfer of water vapor and sensible heat is from the ocean to the atmosphere. Shear turbulence within the first tens of meters of the trades insures thorough mixing and upward transport of both water vapor and sensible heat. Turbulent eddies continue the upward transport through the subcloud layer to the cloud layer where convective cells in the form of cumulus clouds probably play the dominant role in the vertical transport of energy. Within the undisturbed trades high values of wind steadiness occur through a considerable depth of the lower troposphere. Under these undisturbed conditions, gradients are likely to stabilize and the energy input at the surface will reach an equilibrium value dependent upon the vertical removal from the surface and upward transportation of energy. Riehl, Malkus and others (1951) demonstrated that these processes create a moist convective layer, gradually deepening along the airflow. In the trade wind regions of high evaporation, most of the moisture is retained in vapor form rather than rained back into the oceans. This accumulated water vapor is transported equatorward "at a rate of energy export easily two orders of magnitude greater than the rate of kinetic energy consumption by all the global winds and sea currents combined" (Malkus, 1962).

Within the equatorial trough region much of this water vapor is condensed, releasing its latent heat which is carried to great heights in cumulonimbus clouds. Riehl & Malkus (1958) have shown that relatively few such large clouds concentrated in a limited number of synoptic scale systems (e.g., the vortical and other disturbances common in this region) are able to transport this energy, now in the form of sensible heat and potential energy. But now a distinction should be made between suggestions based upon mean budget conditions and deductions based upon day-to-day synoptic changes. If average trade wind conditions prevail for some period of time, gradients of water vapor and temperature will approach some constant value, and the magnitude of the energy transports in the trades will be governed by this. Within the trade wind-equatorial trough system, synoptic scale disturbances will produce internal convergences of water vapor and sensible heat and a vertical flux of energy. But synoptic scale systems, as was shown by the *Crawford* cruise of 1957 (Garstang, 1958), significantly increase the transport of both latent and sensible heat from the ocean to the atmosphere. Gradients, particularly of temperature, in the air immediately above the sea surface steepen markedly and wind speeds increase during synoptic disturbances. The total flux of water vapor and sensible heat at the surface over the area of a synoptic disturbance is, therefore, likely to be significantly greater than during undisturbed conditions. Thus, the mean picture is constrained by overall planetary budget requirements and departures from it are primarily due to the travelling synoptic systems which form a vital link between the ocean and the atmosphere. An attempt will be made here to evaluate the role that organized synoptic scale disturbances play in determining the distribution of latent and sensible heat transfer over the tropical oceans.

Analytical tools and available data

In restricting our attention to the tropical oceans, we encounter the fortunate situation that the lower decameters of the tropical atmosphere are, by and large, barotropic and close to neutral stratification. Under these circumstances, the most practical equations for the computation of the flux of latent and sensible heat are the bulk aerodynamic equations expressed below:

$$LE = Q_e = \varrho\, L\, C_D(\bar{q}_o - \bar{q}_a)\,\bar{u}_a, \tag{1}$$

$$Q_s = \varrho\, C_P\, C_D(\bar{\theta}_o - \bar{\theta}_a)\,\bar{u}_a, \tag{2}$$

where Q_e and Q_s are, respectively, the eddy vertical transport of latent and sensible heat, and are directly proportional to the differences between the mean specific humidity and mean potential temperature measured at the surface and at some point above the surface, multiplied by the mean wind speed measured at the same point above the surface. If the atmosphere above the sea surface is close to neutral stratification, then the most restrictive assumptions that must be made in order to arrive at the equations are most nearly satisfied.

Forty-six days of observations were made on two separate cruises, each of twenty-three days duration, on a fixed station. Both cruises took place during the months of August and September. The first in 1957 when the ship (*R. V. Crawford*, Woods Hole Oceanographic Institution, Garstang, 1958) was stationed near 11 N 52 W; the second in 1963 when the same ship was stationed near 13 N 55 W (La Seur & Garstang, 1964).

The degree to which conditions in the air above the water departed from neutral stratification was assessed by computing hourly values for the bulk Richardson Number, R_B,

$$R_B = \frac{g}{T}\, z\, \frac{\Gamma_h}{\Gamma_m^2}\, \frac{\bar{\theta}_a - \bar{\theta}_s}{\bar{u}_a^2}, \tag{3}$$

where g is the acceleration of gravity; T is the air temperature in degrees Kelvin at height a; z is the height of the observation (6.0 m) above the sea surface; $\bar{\theta}_a$ the potential temperature at 6.0 m; $\bar{\theta}_s$ the potential temperature at the sea surface; $\bar{u}_a$ the wind speed at 6.0 m; each of the latter quantities averaged over one hour centered on the hour. The coefficients Γ_h and Γ_m represent the gradients of heat and momentum which, in turn, are based upon the original profile coefficient concept for water vapor developed by Montgomery (1940). According to Deacon & Webb (1962) at heights of about 4.0 to 8.0 m and for conditions from neutral to moderate thermal stratification, Γ_h and Γ_m have values around 0.1. Thus the ratio in (3) was taken as equal to 10.

Ninety-six per cent of the observations obtained on the *Crawford* fall in the range

$$-0.325 < R_B < +0.025;$$

sixty per cent fall in the range

$$-0.040 < R_B < +0.020.$$

Conditions are, therefore, relatively close to neutral stratification. For this range of stability, with the actual observed maximum range of air-sea temperature difference (except for three observations in showers) between −2.88C to +0.17C, stability depends upon wind speed rather than upon lapse rate. It clearly follows that *the stronger the wind, the more closely governing conditions are met.*

The saving grace for formulations such as (1) and (2) for the computation of air-sea energy transfers is that, *by and large, the higher the transfer, the more accurately these formulae predict it.* As will be shown in later sections, the largest exchange occurs at times of strong winds. Here the shearing is unlikely to permit maintenance of large lapse rates over the open ocean and, even if these did occur, the strong wind shear keeps the Richardson Number in a reasonable range. When the wind approaches calm and the air-sea temperature difference remains large (unlikely), the formulae are in trouble and give particularly bad results for sensible heat exchange. But all the exchanges are then relatively small and perhaps even negligible for large scale budget and dynamic considerations.

The preceding establishes the fact that if the bulk aerodynamic equations can be used at all, they can probably be best applied to conditions prevailing over the tropical oceans. However, it is also quite clear that the drag coefficient, C_D, is not a constant but is a function of at least height (z), wind speed ($\bar{u}$), and stability (R_B).

From Monin & Obukhov (1954), we may write

$$u_* = \frac{k\bar{u}}{(\ln z/z_0 + \alpha R_B)},$$

$$u_* = \frac{\bar{u}}{(C_a^{-1/2} + \alpha R_B/k)}, \tag{4}$$

where

u^* = friction velocity,
C_a is the adiabatic drag coefficient $C_a = k^2/(\ln z_a/z_0)^2$,
the subscript a = height,
k = von Karman's constant,
z_0 = roughness length,
α = empirical constant as defined by Monin & Obukhov (1954),
R_B = bulk Richardson Number as defined by Eq. (3).

If we assume constant stress and set up Eq. (4) for two different levels, $a = 6.0$ m and $d = 11.0$ m, we can show that

$$C_6 - C_{11}\left(\frac{\bar{u}_{11}}{\bar{u}_6}\right)^2 = aR_B\left[C_6^{3/2} - C_{11}^{3/2}\left(\frac{\bar{u}_{11}}{\bar{u}_6}\right)^2\right], \tag{5}$$

where C_6 and C_{11} are drag coefficients computed from the adiabatic relationship but for conditions where $R_B \neq 0$. When $R_B = 0$, Eq. (5) reduces to the adiabatic case.

Approximately 500 values of R_B were computed from observations made on the *Crawford* in the tropical Atlantic (13 N, 55 W) for wind speeds > 2.6 m sec^{-1}. This restriction meant that not too great a range of stability would be considered. As can be seen from Fig. 1, this gave a range of $-0.30 < R_B < +0.025$, a range which, as shown above, includes 96 per cent of all observations made on the *Crawford*.

Values of C_6 and C_{11} were computed at 6 m and 11 m using a linear dependence of the drag coefficient upon wind speed. Class intervals were then chosen for R_B and the mean value of $C_6 - C_{11}(\bar{u}_{11}/\bar{u}_6)^2$ were obtained for each interval. These mean values were centered on each class interval and plotted on Fig. 1. For the range of stability considered, the fitted line suggests that a linear relationship exists for the quantity

$$a\left[C_6^{3/2} - C_{11}^{3/2}\left(\frac{\bar{u}_{11}}{\bar{u}_6}\right)^2\right].$$

For this fit the constant equals 4.2.

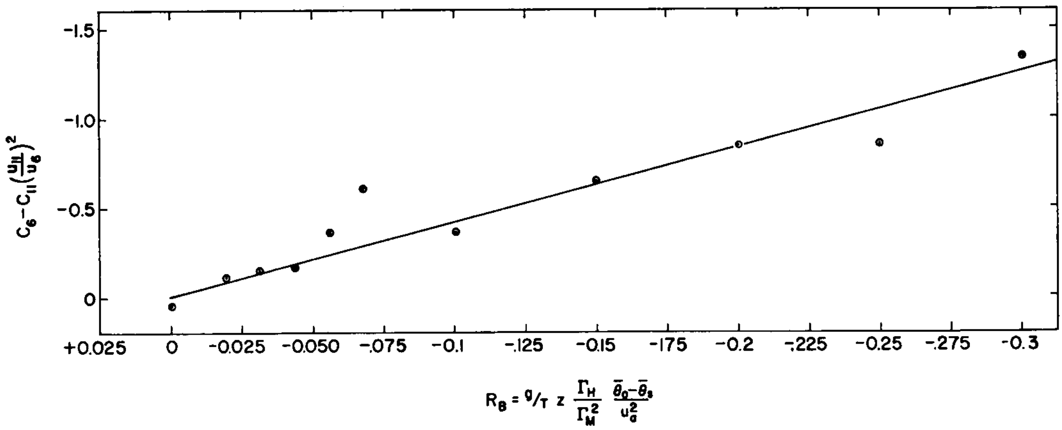

Fig. 1. Departures from the neutral value of the drag coefficient at 6.0 m as a function of stability expressed in terms of the bulk Richardson number.

Taylor (1960) showed that α varies systematically with stability and ranges between 3 and 6 for a stability of $-0.03\ (z+z_0)/L<0$, where $L=$ stability length as defined by Monin & Obukhov (1954). If

$$a\left[C_6^{*3/2} - C_{11}^{*3/2}\left(\frac{\bar{u}_{11}}{\bar{u}_6}\right)^2\right] = 4.2 \quad \text{for} \quad 3<\alpha<6,$$

then the quantity in parenthesis must range between 0.30 and 0.15. These were found to be acceptable values in the sample of 500 observations for the stability range of $-0.30<R_B<+0.025$.

For typical tropical oceanic conditions, therefore, it is considered likely that a near linear relationship exists between the bulk Richardson Number and the departure from the neutral drag coefficient. This linear dependence upon stability is combined with a linear dependence upon wind speed given by Deacon & Webb (1962) to give the following relationship for the drag coefficient at 6.0 m:

$$C_6 = (1.46 + 0.07\ \bar{u}_6 - 4.2\ R_B) \times 10^{-3} \qquad (6)$$

where u has units of m sec^{-1}.

Finally, before using equations (1), (2) and (6) to compute the transfer of latent and sensible heat, an evaluation of the accuracy of the measurements must be made. Wet- and dry-bulb temperatures and wind speeds were measured at 6.0 m above the sea surface at a point 4.7 m on a pulpit ahead of the bow of the vessel. The vessel was either hove to or steaming slow ahead into the wind. Sea surface temperatures were measured 10 cm below the surface, and 4.5 m ahead of the vessel. The temperature sensors were thermistors, and the wind speed was obtained from a recording cup anemometer. Care was taken to avoid radiation effects. Supplementary temperature measurements were made around the ship (out to 140 ft) on three different occasions. These measurements were used to determine any extraneous effects on the pulpit observations. Wind speeds were checked for ship effect in a similar manner to that described by Deacon *et al.* (1956). It is felt that, with these precautions and the corrections later applied, the observations are representative of ambient conditions over the open ocean.[1]

Synoptic variations of latent and sensible heat transport

In order to examine the distributions of latent and sensible heat transfer under varying synoptic scale conditions, each observing period was subjected to careful analysis. The objective was twofold—

1) to delineate synoptic scale disturbances and categorize each system with respect to intensity;
2) to relate the observations made by the ship to each system and delineate the sector intensity that was sampled.

[1] A more complete treatment of ship effect is contained in a doctoral dissertation [Garstang, 1964]. Similarly, details of analysis and computation referred to in subsequent sections of this paper, may be found in the same source.

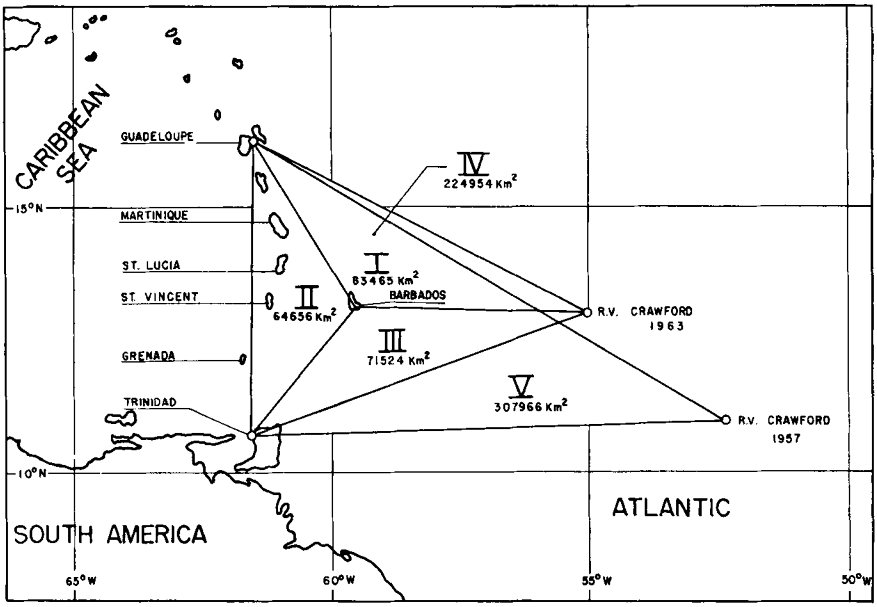

Fig. 2. Location and size of triangles used for the computation of divergence, relative vorticity and vertical motion. I. *Crawford* (1963) /Guadeloupe/ Barbados; II. Guadeloupe /Trinidad/ Barbados; III. Barbados /Trinidad/ *Crawford* (1963); IV. *Crawford* (1963) /Guadeloupe/ Trinidad; V. *Crawford* (1957) /Guadeloupe/ Trinidad.

To achieve this, conventional methods of analysis were supplemented by a number of other techniques. Among these were: divergence, vorticity, and vertical motion computations using a kinematic technique for each of the five triangles shown in Fig. 2; rainfall analysis of six islands in the Lesser Antilles, using a classification based upon occurrence and amount; and a method using wind speed and steadiness as a measure of the occurrence and strength of a synoptic scale disturbance.

Once the systems had been identified, time distributions of latent and sensible heat could be obtained by calculating the hourly values for all forty-six days, and obtaining average values ($\bar{Q}_e$ and $\bar{Q}_s$) for each of the systems. Table 1 shows the initial broadscale breakdown. In the first two rows all undisturbed states can be compared with all disturbed states. The results from both sets of observations show the same trend: *more heat is transferred from the ocean to the atmosphere under disturbed condi-*

Table 1. *Heat transport from the ocean to the atmosphere under varying synoptic conditions*

	$\bar{Q}_e$		$\bar{Q}_s$		$\bar{Q}$		Bowen ratio	
	1957	1963	1957	1963	1957	1963	1957	1963
Undisturbed trades	265.2	371.8	3.1	14.0	268.3	385.8	0.012	0.038
Disturbed modes	309.4	396.2	24.0	28.5	333.4	424.7	0.078	0.071
Weak and moderate trades	254.2	326.0	3.0	14.4	257.2	340.4	0.012	0.044
Moderately disturbed modes	492.8	408.2	35.1	31.2	527.9	439.4	0.071	0.076

Table 2. *Synoptic scale exchange fluctuations in the vicinity of 11 N to 13 N, 52 W to 55 W during the month of August and part of September from the 1957 and 1963 R. V. Crawford cruises*

	$\Delta\bar{T}$ C	$\Delta\bar{q}$ g kg^{-1}	$\bar{u}$ m sec^{-1}	$\bar{C}_D$ 10^{-3}	$\bar{R}_B$	$\bar{Q}_e$ cal cm^{-2} day^{-1}	$\bar{Q}_s$ cal cm^{-2} day^{-1}	$\bar{Q}$ cal cm^{-2} day^{-1}	Mean Bowen ratio
Strong	0.290	5.54	8.12	2.062	−0.008	544.6	12.6	557.2	0.026
trade	0.002	4.47	7.68	1.992	0.001	407.4	3.6	411.1	0.009
regime									
$\bar{u} > 7.7$ m									
sec^{-1}									
$\bar{S} > 98$ %									
Moderate	0.500	5.24	5.71	2.020	−0.037	356.5	13.4	369.9	0.040
trade	−0.080	4.61	5.47	1.827	0.004	272.0	−0.6	271.4	−0.002
regime	0.540	4.78	5.23	2.004	−0.042	295.7	15.4	311.1	0.049
5.01–7.6	0.090	5.36	5.15	1.913	−0.022	314.8	3.9	318.7	0.012
m sec^{-1}	0.200	4.39	5.04	1.933	−0.029	245.4	5.2	250.8	0.046
$\bar{S} > 97$ %									
Weak	0.530	5.65	4.75	2.043	−0.060	312.6	12.5	325.1	0.041
trade	0.680	6.06	4.73	2.047	−0.062	340.0	17.8	357.8	0.048
regime	0.520	5.16	3.96	1.963	−0.054	227.3	5.4	232.8	0.030
< 5.0 m	0.100	5.31	4.47	1.903	−0.031	254.4	2.4	256.8	0.032
sec^{-1}									
$\bar{S} > 97$ %									
Weakly	0.60	5.39	5.43	2.011	−0.044	350.7	18.5	369.2	0.480
disturbed	1.09	4.56	5.02	2.791	−0.361	325.0	32.9	357.9	0.102
mode	0.76	4.41	4.33	2.475	−0.223	255.8	22.6	278.7	0.079
$\bar{u} < 5.5$ m	1.06	5.21	2.86	3.351	−0.418	282.1	40.5	322.7	0.123
sec^{-1}	0.50	5.34	4.00	2.053	−0.268	268.7	12.6	284.0	0.050
$\bar{S} < 75$ %	0.32	4.69	5.48	1.913	−0.016	293.2	9.5	302.8	0.035
Moderately	1.01	4.83	5.79	2.187	−0.078	355.0	32.6	387.6	0.095
disturbed	0.72	4.75	6.99	2.108	−0.038	419.2	27.2	442.5	0.083
mode									
$5.5 < \bar{u} < 7.5$									
m sec^{-1}									
$90\% > \bar{S} > 75\%$									
Strongly	0.74	5.27	7.67	2.109	−0.027	505.1	29.0	534.1	0.057
disturbed	0.71	4.84	9.30	2.213	−0.024	570.1	43.4	612.1	0.081
mode									
$\bar{u} > 7.5$ m									
sec^{-1}									
$95\% > \bar{S} > 90\%$									

tions than during undisturbed conditions. The increase in latent heat transfer is not as large as in the case of sensible heat transfer. The ratio of $\bar{Q}_e$ undisturbed versus $\bar{Q}_e$ disturbed is 1.17 in 1957 and 1.07 in 1963; but for $\bar{Q}_s$ it is 7.74 and 2.04, respectively. On each cruise weakly disturbed modes were fairly frequent and strong trade wind conditions relatively few, so that it is perhaps more meaningful to examine the transfers under weak and moderate trades and moderately disturbed modes. Under these conditions, the ratio of undisturbed to disturbed for $\bar{Q}_e$ increases to 1.94 in 1957, and 1.25 in 1963; and for $\bar{Q}_s$ 11.70 and 2.16, respectively. The relative increase in the Bowen ratio remains essentially the same for both of the above comparisons. While the amount of sensible heat made available to the atmosphere never exceeds 8 per cent of the latent heat transfer, it is significant to note that:

a) this is an average figure throughout the whole disturbed period;
b) for periods up to 6 hours within a disturbed region, the ratio of sensible heat to latent heat may be greater than 0.20 as opposed to undisturbed conditions when, for similar periods, the sensible heat transfer may be in the opposite direction, i.e., from the atmosphere to the ocean;
c) this energy is directly available to contribute

to the instability of the subcloud layer and the development of convective cloud.

Therefore, it is suggested that the role of sensible heat transfer in the formative and developing stages of tropical disturbances is a critical one.

The synoptic scale fluctuations of exchanges are examined in more detail in Table 2. The mean values for each regime or mode are based upon values computed for each hour through each state. Little change in $\Delta\bar{q}$ is noted between the various states. The mean temperature difference, $\Delta\bar{T}$, however, shows a systematic and fairly large change from undisturbed to disturbed conditions. Mean sea-air temperature differences in excess of 1.0C occur during the disturbed modes, while under the strong trade regime $\Delta\bar{T}$ is less than 0.3C. It should be emphasized that this temperature difference is obtained from continuous recordings averaged over a period of one hour. The hourly mean differences are, in turn, averaged over all hours classified within a given regime or mode. Sea surface temperatures were observed to vary little between periods, although slightly lower temperatures were noted during the strong trade wind regime as opposed to the weak regime. The dominant control upon $\Delta\bar{T}$ is produced by variations in air temperatures. Lower air temperatures during the disturbed modes are ascribed to reduced insolation due to cloud cover, direct cooling by rain showers, as well as the descent from aloft of cool air associated with the condensation-precipitation cycle. These processes are clearly not advective in origin, but are associated with the dynamics of the disturbance involved. While these differences between the undisturbed regime and the disturbed mode are noted, Figs. 3 and 4 show that, with the exception of $\Delta\bar{T}$ in the disturbed case, there is little variation within each mode or regime. The greater dependence of $T_s - T_a$ upon wind speed in the disturbed mode is related to the fact that much lower air temperatures are observed at lower wind speeds under these conditions, as opposed to the undisturbed regime. This leads to the reverse effect in $\Delta\bar{q}$ which is higher at low wind speeds for the undisturbed case. The scalar average wind speed is similar within similar intensity categories, i.e., the weak trade regime wind speed is similar to the weakly disturbed mode wind speed. While it follows from the above that, given similar $\Delta\bar{q}$ and $\bar{u}$, similar $\bar{Q}_e$ should be obtained, it is also illustrated in Table 2 that the stability changes. Under the strong trade regime, the stability as indicated by R_B, is on the average close to neutral. Maximum instability is reached in the weakly disturbed mode, returning to weakly unstable in the strongly disturbed mode. Under the weakly disturbed mode wind speeds are low (< 5.5 m sec^{-1}) but sea-air temperature differences are high, leading to maximum instability. When a drag coefficient with a dependence upon stability is used, then these changes in stability are reflected in changes in C_D. The changes in C_D produce an increase in latent heat transfer during disturbed conditions even though the wind speeds and specific humidity differences are similar to those of undisturbed conditions.

Since there is a fairly systematic increase in $\Delta\bar{T}$, there is a corresponding increase in sensible heat transfer during disturbed conditions. Changes in the Bowen ratio show that this increase is greatest with respect to latent heat transfer in cases categorized under the weakly disturbed mode. This suggests that sensible heat transfer may play an important role in both the formative stages of a disturbance and in the peripheral regions of an organized disturbance.

The maximum exchange of total energy, $\bar{Q}$, takes place in the disturbed mode. The largest exchanges take place when conditions are closest to neutral stratification and, in consequence, closest to conditions which are assumed in the development of the exchange equations. Almost all of the relatively large amounts of total energy transferred from the ocean to the atmosphere during undisturbed conditions is in the latent form. It has been clearly established by other workers such as Riehl (1954, p. 56) and confirmed in this region by La Seur & Garstang (1964a), that the greater proportion (> 50 per cent) of tropical precipitation falls in organized synoptic disturbances. The proportion over the ocean may, in fact, be far larger than this figure. Hence, before a significant proportion of the latent heat accumulated in the tropics can be made available to the atmosphere, it must be advected into organized synoptic disturbances. Condensation and precipitation processes will release part of this latent heat directly into the disturbance. This available energy may then be used both to fuel the disturbance itself, as well

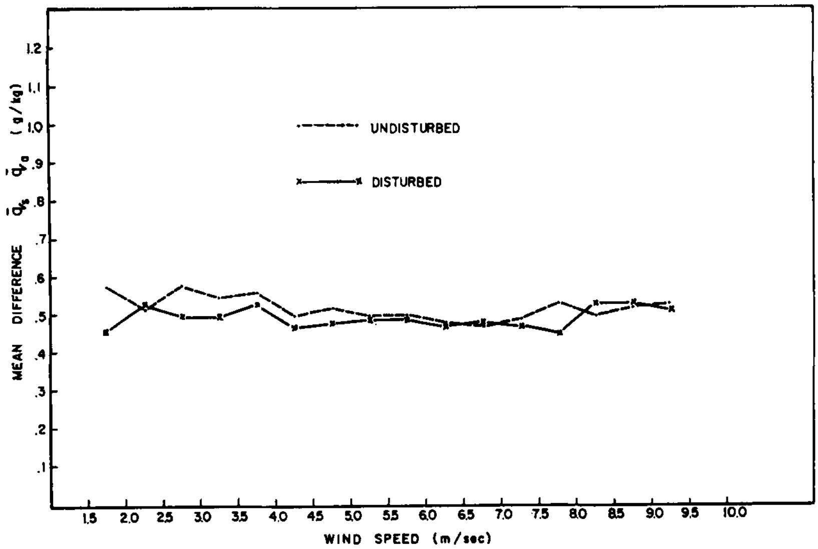

Fig. 3. $\Delta\bar{q}$ as a function of wind speed for all the disturbed modes and for all the undisturbed regimes.

as increase the potential energy in the upper levels of the tropical and equatorial atmosphere. Therefore, the synoptic disturbances not only represent a localized maximum of energy flux, but are also regions of horizontal convergence and vertical transport of the energy supplied to the atmosphere over large regions of the tropics. The extent to which individual synoptic disturbances contribute to the energy budget of the atmosphere is examined below.

Eight identifiable synoptic scale disturbances passed over the ship during the 43 days on station. The ship's hourly position relative to the disturbance could be obtained, and the asso-

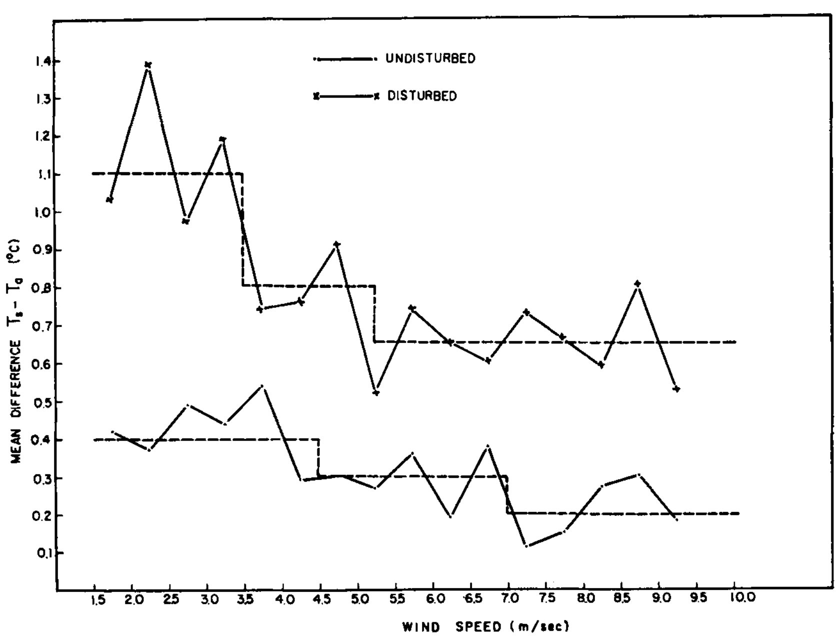

Fig. 4. $\Delta\bar{\bar{T}}$ as a function of wind speed for all the disturbed modes and for all the undisturbed regimes.

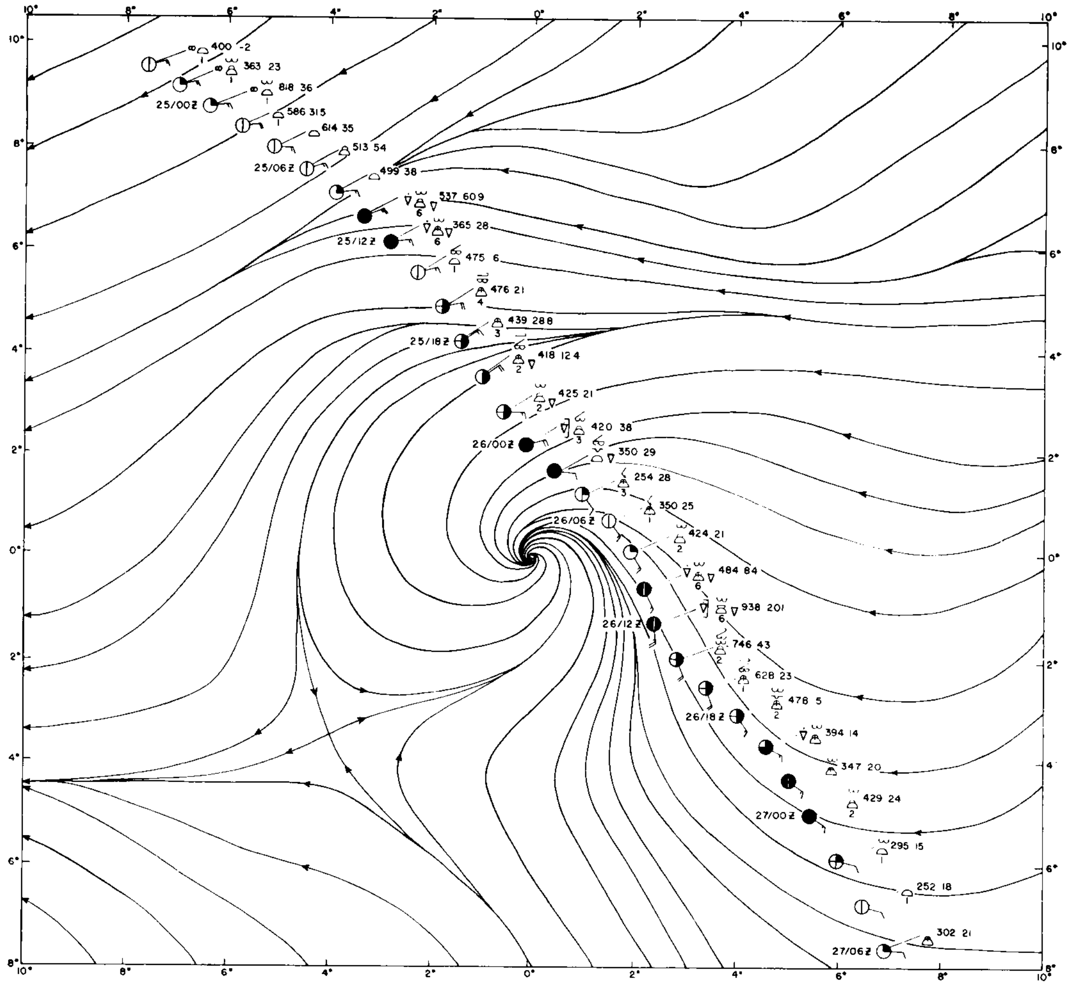

Fig. 5. Streamlines based upon observations composited over the period 00 GCT on 25 August to 06 GCT on 27 August 1963. Observations made on the *Crawford* are plotted every two hours with the corresponding values of latent and sensible heat in cal cm^{-2} day^{-1} on the upper right-hand side of the plotting model.

ciated latent and sensible heat flux identified with each sector of the disturbance. Such a plot (at two hourly intervals) is shown on Fig. 5. No adjustment has been made between the independent streamline analysis (based upon all available data over the time period 0000 GCT 25 August to 0600 GCT 27 August) and the superimposed research vessel observations.

Composite maps were constructed for each disturbance and the following analyses performed:

1) Isogon-isotach analysis to delineate the velocity field.
2) u-v analysis in order to compute surface divergence.
3) Weather and nephanalysis to associate with the surface field of divergence.

Based upon these analyses and backed up by the actual observations through the systems, the temperature and humidity differences contained in Table 3 could be related to each sector of the disturbance. Given the wind speed from the isotach analyses, the time sequences observed on the ship were then extended to spatial distributions covering the entire disturbance. A detailed description of the methods used is

Table 3. *Air-sea property differences based upon Table 2 and Figs. 3 and 4, used to compute energy transfers for the composite disturbances*

	$\Delta\bar{T}$ (C)	$\Delta\bar{q}$ (g kg^{-1})
Strong trade >7.7 m sec^{-1}	0.20	5.0
Moderate trade 5.0–7.7 m sec^{-1}	0.30	5.0
Weak trade <5.0 m sec^{-1}	0.40	5.5
Weakly disturbed A <3.5 m sec^{-1}	1.10	5.0
Weakly disturbed B 3.5–5.25 m sec^{-1}	0.80	5.0
Moderately and strongly disturbed >5.25 m sec^{-1}	0.65	5.0

given by Garstang (1964). The results are summarized in Table 4 for a weak, moderate and strong disturbance. Maximum and minimum values are given over limited areas of 0.5 to 1.0 degree squares.

The large transports are concentrated within the regions usually associated with maximum precipitation. Five centimeters of rain in such a region in one day would represent a release of about 3000 cal cm^{-2} day^{-1}. While the greater fraction of water vapor represented in such an energy conversion must be due to horizontal advection, a significant amount could be accounted for by the transfer processes taking place within such a region. The distribution of sensible heat transport is similar to that of the latent heat transport, with the exception that

Table 4. *Maximum and minimum values of latent and sensible heat transfer obtained for weak, moderate and strong disturbances*

	Maximum Q_e	Minimum Q_e	Maximum Q_s	Minimum Q_s
Weak disturbance $\bar{u}<5$ m sec^{-1}	725	56	38	5
Moderate disturbance $5<\bar{u}<10$ m sec^{-1}	906	132	49	5
Strong disturbance $10<\bar{u}<20$ m sec^{-1}	1446	135	78	7

$\bar{u}$ is a time average over one hour in the region of maximum winds.

strong gradients coincide with the boundary between disturbed and undisturbed states. Since this is the boundary between the regions of maximum and minimum cloudiness and weather, maximum values of Q_s are found within the disturbed region. Maximum values can be further compared with those computed by Petterssen *et al.* (1962) from typical winter cases of cyclones in the western North Atlantic. Maximum values of 1440 cal cm^{-2} day^{-1} for latent heat flux, and 720 cal cm^{-2} day^{-1} for sensible heat flux were obtained. Pyke (1964) computed values for a cyclone in the Gulf of Alaska and obtained values of 350 cal cm^{-} day^{-1} for Q_e and 216 cal cm^{-2} day^{-1} for Q_s. Malkus & Riehl (1959), assuming equations essentially the same as (1) and (2), computed latent and sensible heat fluxes for a moderate hurricane within 90 km of the center of 2420 cal cm^{-2} day^{-1} and 720 cal cm^{-2} day^{-1}. Consistent with the above, they assumed small changes in $q_s - q_a$, a decrease from 5.2 g kg^{-1} in the region 90 to 70 km to 3.5 g kg^{-1} 50 to 30 km from the center. Thus the increase in latent heat from the values observed in the above disturbances and in the trade wind regions is due to an increase in wind speed. They assumed a constant sea-air temperature difference of 2.0C. This they ascribe to adiabatic expansion during horizontal motion toward lower pressure rather than the cooling mechanisms called upon in the above disturbances. This large ΔT together with high wind speeds gives a large sensible heat transfer. In comparison with this value, the highest hourly mean value of Q_s observed on the *Crawford* was 201 cal cm^{-2} day^{-1}. The values utilized by Malkus & Riehl for the moderate hurricane are, therefore, quite consistent with those values computed from the composite storm data. In turn, these are consistent with values of latent heat transfer computed for higher latitudes. The values of sensible heat transfer appear, at the most, to be 40 per cent of the values observed at higher latitudes where pronounced sea-air temperature differences are observed.

The integrated transport of latent and sensible heat were computed for all disturbances and mean values for the areas covered by the system are presented in Table 5. These mean values are compared with values which would have been obtained had the whole region been covered by a uniform trade wind, maintaining an average speed within the limits specified, for

Table 5. *Comparison between latent and sensible heat transfer under constant trade wind conditions and synoptic scale disturbances over areas of 10 by 10 and 20 by 20 degrees of latitude*

	Average energy fluxes in the trade wind region		Average energy fluxes in disturbed conditions: Disturbance I averaged over 100° lat²	Disturbances II and III averaged over 400° lat²
	Q_e	Q_s	Q_e	Q_s
Weak trade $\bar{u} < 2.5$ m sec^{-1}	$Q_e < 180$	$Q_s < 5.4$		
Weak to moderate trade $2.5 < \bar{u} < 5.0$	$180 < Q_e < 314$	$5.4 < Q_s < 9.4$		
Weak disturbance			418	15
Moderate trade $5.0 < \bar{u} < 7.7$	$314 < Q_e < 450$	$9.4 < Q_s < 11.3$		
Moderate disturbance < tropical storm			435	17
Strong trade $\bar{u} > 7.7$ m sec^{-1}	$Q_e > 450$	$Q_s > 7.6$		
Strong disturbance < hurricane			480	22

periods of one to four days. Climatic charts for the Atlantic from 15 S to 15 N from August to November (MacDonald, 1938, and Aspliden *et al.*, 1965) show that the average wind speeds are everywhere less than 6.7 m sec^{-1}, and over most of the region less than 5.0 m sec^{-1}. From Table 5 the average value Q_e would be between 180 and 380 cal cm^{-2} day^{-1}, and of Q_s between 5.4 and 10.5 cal cm^{-2} day^{-1}. The occurrence of a moderate synoptic disturbance of dimensions similar to those considered above would greatly alter the distribution of energy input within these latitudes. Only during the months of December through May do average winds in excess of 7.2 m sec^{-1} cover a significant portion of the region 15 N to 15 S. At this time of the year within the moderate to strong trade wind regions, the latent heat flux would approach that associated with moderate to strong disturbances. For the remainder of the year, i.e., June through November, the net input of energy from the surface of the ocean in this region will depend significantly upon the frequency and distribution of such synoptic scale disturbances.

Detailed analysis of the tropical Atlantic for the International Geophysical Year of 1958 has been carried out by Aspliden, Dean & Landers (1965, 1966). The frequency of equatorial vortices for each month in 1958 has been obtained. Preliminary results suggest a frequency of 10 to 15 vortices per month during August, September and October. As shown in Fig. 6, development generally takes place off the West African coast in the vicinity of 5 N to 10 N, with some vortices apparently emerging from the continent itself. As the equatorial trough region migrates equatorward, the number of vortices diminishes and they can be tracked into the Atlantic for limited distances only. During the months of January, and February only a few weak vortices appear in the eastern equatorial Atlantic, none of which migrate into the central Atlantic. Phenomena in the mid- and upper troposphere (such as the tropical easterly jet) appear to play an important role in this seasonal fluctuation in the frequency of these equatorial systems. Figure 6 shows the tracks of eleven equatorial or tropical cyclones. At least nine of these systems followed a track confined to an extremely narrow belt. Mean maps reflect this concentration as an asymptote in the streamline field (Aspliden, Dean & Landers, 1965, 1966).

For the region near the asymptote, a mean number of ten vortices per month can be accepted for the wet season months (July through October), with one to four per month during the dry season (January through May). If, in the mean, the disturbance covered 15 by 15 degrees of latitude and moved at a mean rate of 300 *n mi* within 24 hours, or five degrees of latitude per day, then within a belt 15 degrees

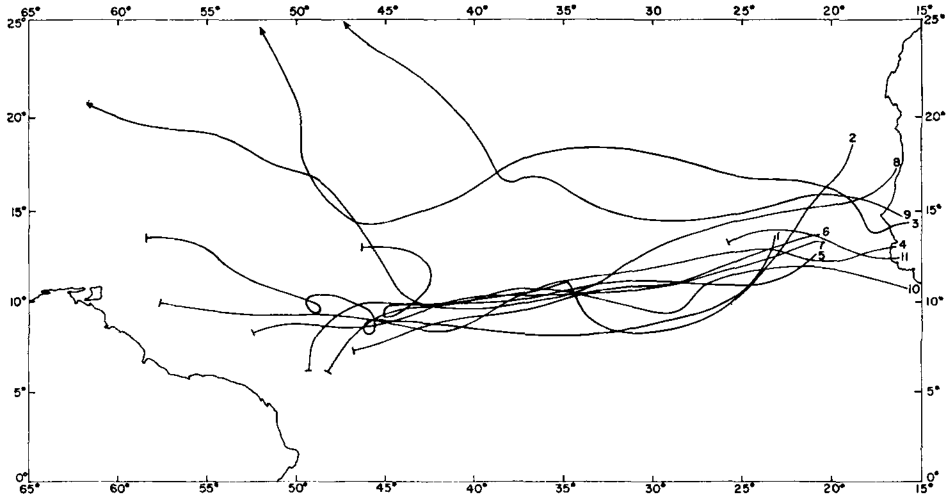

Fig. 6. Tracks of equatorial cyclones for the month of August 1958. The cyclones have been identified *in the surface streamline analysis and tracked until the circulation could no longer be identified.* Tracks of systems originating in July and persisting into August have not been included. (After Aspliden, Dean & Landers, 1965.)

wide across the tropical Atlantic centered on the mean asymptote during any wet season month, the mean sensible and latent heat flux would correspond to the values computed for the composite storms. The choice of a belt 15 degrees wide is suggested not only by the mean size of the disturbances examined, but is also reflected by the charts of wind constancy (e.g., MacDonald, 1938). For the months of July through October a belt 15 degrees wide with wind constancies of 80 per cent or less exist about the equatorial trough zone. A constancy of 80 per cent signifies that during 80 per cent of the time steadiness is better than 90 percent. Therefore, the 80 per cent constancy boundary roughly corresponds to the division between disturbed and undisturbed conditions used above. Outside this belt other synoptic systems would affect the distribution of energy flux, but within the stronger and more persistent trade wind belts such fluctuations would have less effect upon the net transport than in the case in the vicinity of the equatorial trough.

The month of August was chosen as representative of the wet season, and the month of February as representative of the dry season. Computations for the tropical Atlantic have in each case been based upon the mean wind speed values given by Aspliden *et al.* (1965, 1966). On the August map the 15 degree wide region under the influence of disturbances has been centered on the position of the mean asymptote. Latitudinal means were obtained for Q_e and Q_s at one degree intervals from the values computed for the composite storms. These values were applied to 24 of the days of August, the remaining seven days represent undisturbed conditions. The region between ± 7.5 degrees of the asymptote and about 30 N and 20 S was then subdivided on the basis of mean sea surface temperatures (MacDonald, 1938). Based upon these values, the air-sea temperature difference was scaled according to observed values from the 1957 and 1963 data. The difference between the saturated specific humidity and the humidity at 6.0 m was assumed constant at 5.0 g kg^{-1}. Computations (as before) of Q_e and Q_s were then made for five degree squares. This procedure of computing Q_e and Q_s outside the equatorial trough region in August was applied to the whole area in February. Transport during the transition months between wet and dry seasons will depend upon the occurrence and frequency of synoptic disturbances. Thus, an annual mean map of Q_e and Q_s has been constructed from an average of the August and February maps. These six maps are presented in Figs. 7 through 12. The mean value of Q_e for the month of August shown in Fig. 7 shows similarities with the mean annual map of

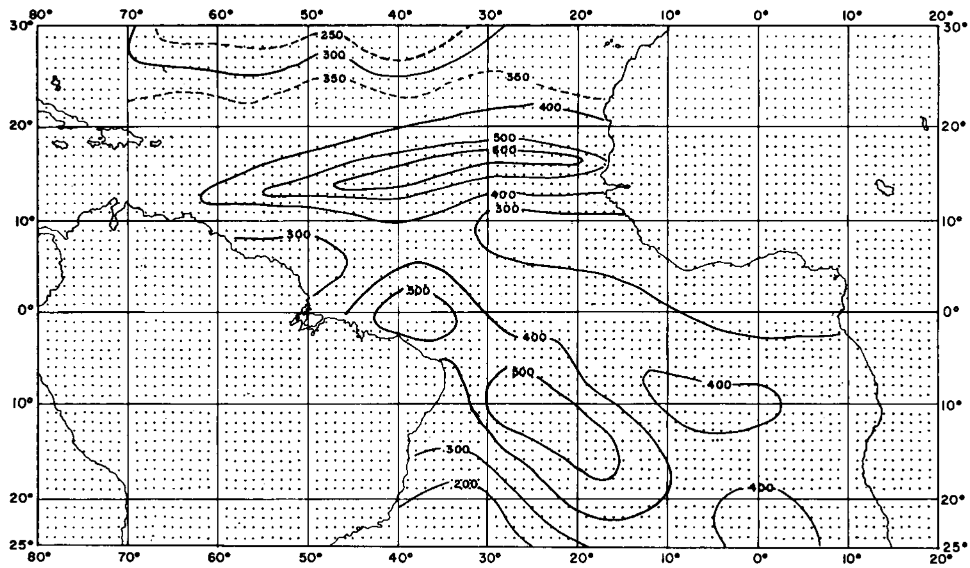

Fig. 7. Mean values of latent heat transport in cal cm^{-2} day^{-1} for the tropical Atlantic for the month of August.

Budyko (1956) only near 30 N. On Fig. 7 a well-defined maximum now appears to the north of the asymptote as a result of synoptic scale systems. A secondary maximum appears in the southwestern tropical Atlantic in association with the maximum in the winter trade of that hemisphere. The overall results indicate a pronounced increase in latent heat transport within the equatorial and southern tropical Atlantic.

The mean sensible heat transport for August shown in Fig. 8 indicates, in general, lower values than reported by Budyko (1956) but higher values than Jacobs (1951). Within the equatorial trough regions values are close to

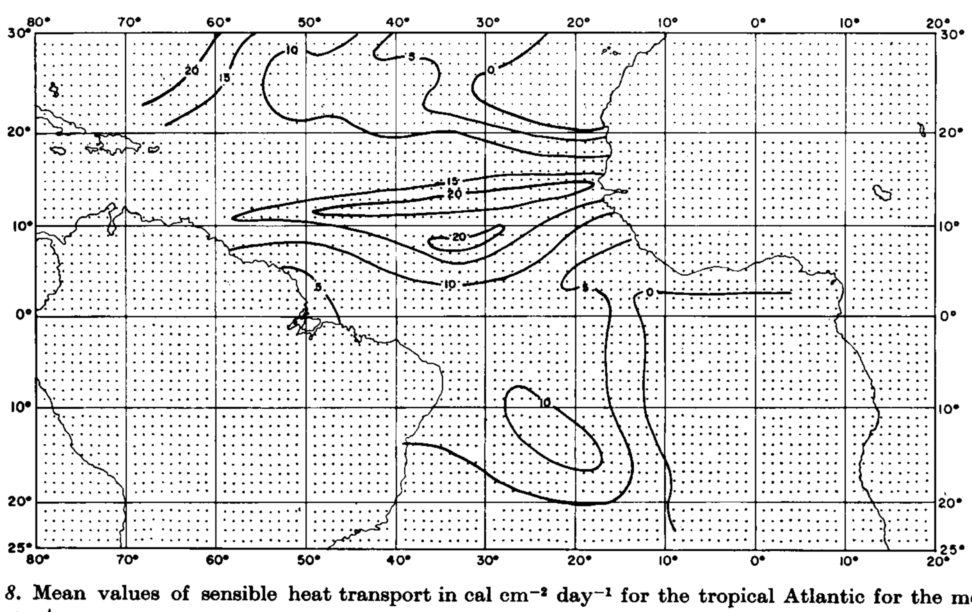

Fig. 8. Mean values of sensible heat transport in cal cm^{-2} day^{-1} for the tropical Atlantic for the month of August.

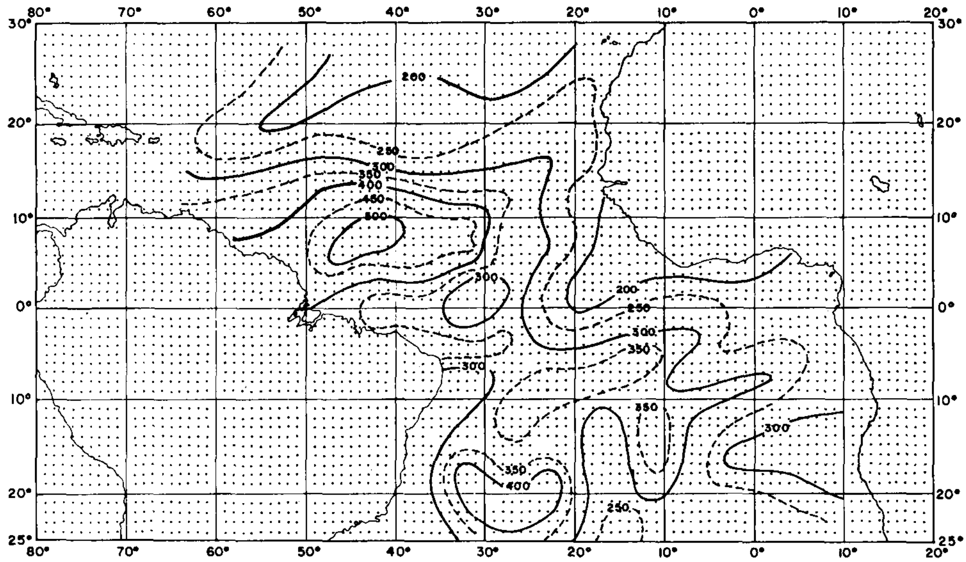

Fig. 9. Mean values of latent heat transport in cal cm^{-2} day^{-1} for the tropical Atlantic for the month of February.

those shown by Budyko. This is due to the added transport associated with synoptic systems. Outside this region the values reported in Fig. 8 are lower than those of Budyko. This is, in large part, due to the exclusion of diurnal effects, a factor not considered by Budyko.

During February the mean transport of latent heat shown in Fig. 9 corresponds in the southern hemisphere to that obtained by Budyko (1956) on an annual basis in the northern tropical Atlantic. A maximum associated with the trade wind maximum of the winter hemisphere appears again.

A marked decrease in the transport of sensible

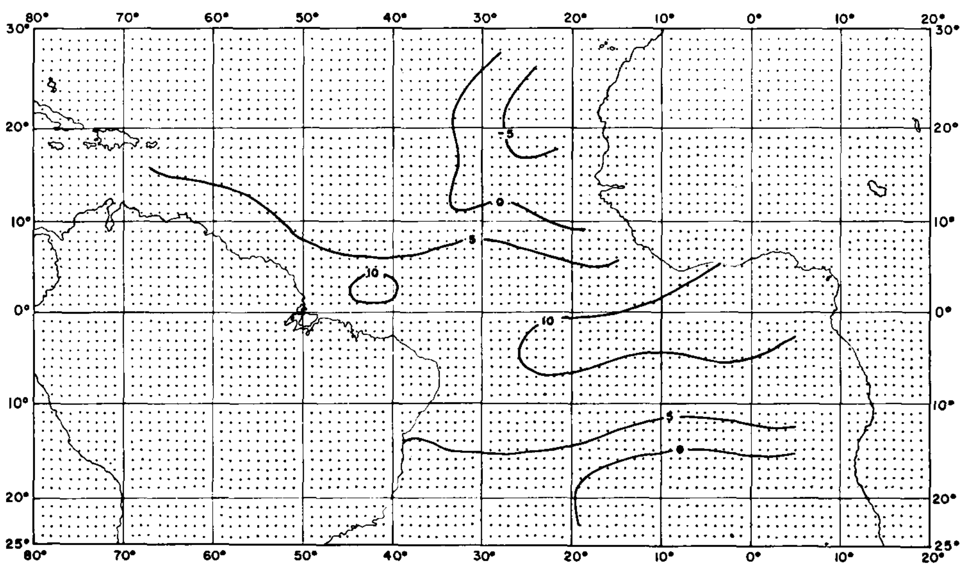

Fig. 10. Mean values of sensible heat transport in cal cm^{-2} day^{-1} for the tropical Atlantic for the month of February.

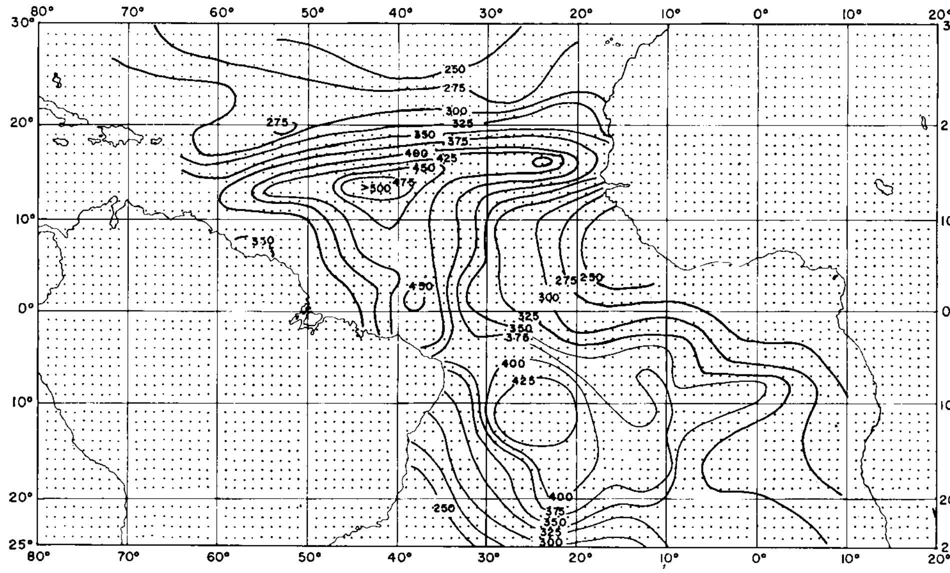

Fig. 11. Mean annual values of latent heat transport in cal cm^{-2} day^{-1} for the tropical Atlantic based upon Figs. 7 and 9.

heat during February is shown in Fig. 10. This is due not only to the inclusion of diurnal effects but also to the fact that fair weather is a mean condition over most of the region during this month.

Figures 11 and 12 show the mean annual distribution of Q_e and Q_s for the tropical Atlantic as calculated in this paper. In the case of latent heat, the most significant features remain the maximum in the equatorial trough region associated with synoptic scale disturbances, and the higher values shown over much of the

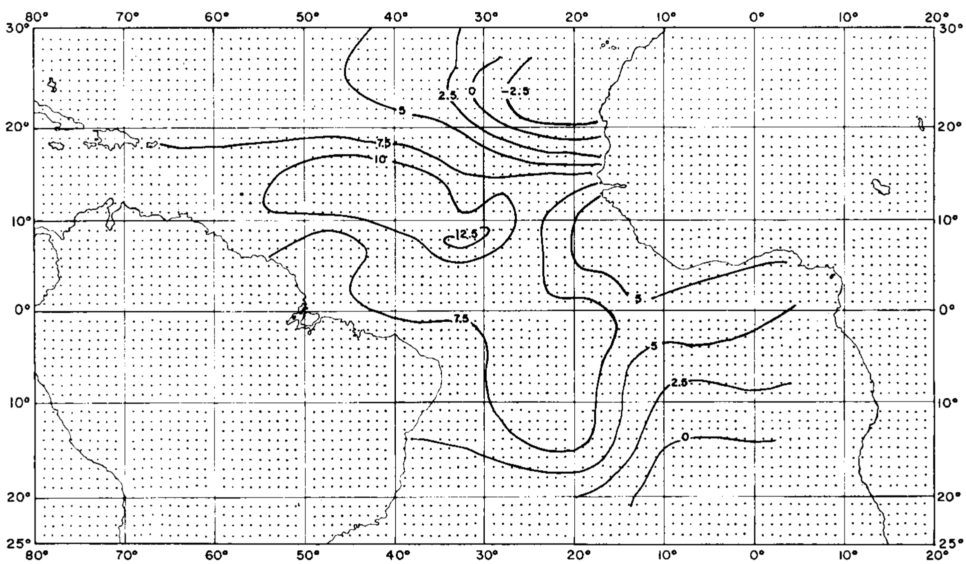

Fig. 12. Mean annual values of sensible heat transport in cal cm^{-2} day^{-1} for the tropical Atlantic based upon Figs. 8 and 10.

southern tropical Atlantic. The integrated effect still produces a maximum within the eastern and central equatorial trough region.

The inclusion of both the synoptic and diurnal effects on the transport of sensible and latent heat produces pronounced variations in both the monthly means and the annual distributions. The inclusion of the drag coefficient as a function of both wind speed and stability results in an increase in the transport of heat, particularly in the regions of high mean wind speeds. Clearly, these effects will apply in all tropical ocean areas. In the central and western Pacific, where the mean annual frequency of synoptic disturbances is higher than in the Atlantic, the effect will be even more pronounced than shown to be the case in the Atlantic.

Conclusions

The strong dependence of latent and sensible heat transport upon synoptic scale disturbances implies that the role of the equatorial trough region in the heat balance of the atmosphere may be even more important than has heretofore been suspected. Not only do the disturbances in this region perform the function of horizontal and vertical advection of vast quantities of energy generated within the trade wind regions, but a large increase in energy transfer occurs within the region of the disturbance. Through the condensation-precipitation cycle, the disturbance then provides a means of releasing this energy. In the mean, the concentration, vertical and horizontal transports of energy in this region prescribe one of the fundamental constraints on the heat balance of the atmosphere.

While it is considered that the magnitudes of the transports derived in the text may be more accurate than those previously published, it is important to note that changes in the coefficients used would not change the direction of the transport, the sense of the synoptic variations or the fact that synoptic systems produce large amounts of latent and sensible heat at the surface. Verification of the absolute magnitude is indeed desirable, but there seems little doubt that a region of maximum transport will occur just poleward of the equatorial trough, and that this region will have maximum values in excess of those appearing on current maps (e.g., Budyko 1956). These conclusions cast some doubt upon the relative magnitude of the various terms in the presently accepted heat budget of the ocean and the atmosphere. In particular, it may suggest a revision of the current estimate of the radiation balance and oceanic heat flux (Budyko 1956). Such a revision would be in agreement with preliminary results that are emerging from satellite cloud climatology that average cloudiness over the tropical oceans is far less than heretofore assumed. However, before such a revision is made, the uncertainties that have entered into the computations made in this paper must be removed.

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ОБМЕН ЯВНОЙ И СКРЫТОЙ ТЕПЛОТОЙ В СИСТЕМАХ СИНОПТИЧЕСКОГО МАСШТАБА В НИЗКИХ ШИРОТАХ

Уравнения переноса применяются к ряду данных за сорок шесть дней, полученных на океанографическом буе на станции, расположенной в низких широтах западной Атлантики. В этой области градиент температуры и профиль ветра в пограничном слое атмосферы таковы, что уравнения переноса можно применять с разумной степенью согласия. В чатности, более точные результаты получаются, когда имеется большой перенос, менее точные, когда имеется маленький перенос. При этих условиях, для подсчета переноса явной и скрытой теплоты над поверхностью моря использовалась линейная зависимость коэффициентов турбулентного обмена от скорости ветра и от устойчивости при осредненных по времени за один час удельной влажности, температуры и скорости ветра на уровне 6 м. Температура поверхности моря измерялась на глубине 10 см и сравнивалась с данными измерений инфракрасной радиации.

Исследовались отдельные синоптические системы, которые двигались над точкой измерения или вблизи нее. Было обраружено, что в ограниченных областях этих возмущений перенос явной и скрытой теплоты возрастал на порядок величины.

Также было найдено, что проинтегрированный по всей области возмущения поток энергии вдвое больше, чем в невозмущенной. Определены временные масштабы и размеры синоптических систем, с использованием как анализа линий тока и количества выпадающих осадков, так и методики типа «да — нет». Это дало возможность составить летние, зимние и годовые карты переноса явной и скрытой теплоты для тропической Атлантики. При сравнении их с результатами предшествующих исследователей найдено заметное различие.

С помощью этих результатов объясняется роль возмущений синоптического масштаба в балансе энергии атмосферы.