SYNOPTIC DISTURBANCES

IN THE TROPICS

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Summary :

The importance of tropical disturbances in the energetics and the maintenance of the general circulation is emphasized. The important features of the different disturbances existing over different areas are reviewed. Emphasis is put on our understanding of the energetics of these phenomena.

1. Introduction

The lower tropospheric wave disturbances in the tropics have been extensively studied by using a variety of analysis methods since the pioneering works by Riehl (1945, 1954) and Palmer (1952). Riehl studied synoptic-scale disturbances in the tropical easterlies known as "easterly waves" in the Caribbean area. These low level tropical wave disturbances embedded in the trade easterlies, develop in many parts of the tropical regions. These disturbances form initially as perturbations with weak amplitudes and gradually intensify into organized wave-type circulations during their westward propagation. Occasionally, under favorable conditions, they transform into tropical depressions, hurricanes or typhoons. Since the first description of such waves in the easterlies, easterly waves over various parts of the tropical oceans have been studied by many investigators using data from different sources and using a variety of analytic methods.

In the 1960's, following the implementation of an upper air sounding network in the western and central Pacific, tropical disturbances were studied in that region. Carlson (1969) made the first detailed analysis of the African waves propagating westward across West Africa and the tropical Atlantic ocean. A large number of studies focused on these disturbances using the GATE observations.

The existence of depressions moving north westward from the Bay of Bengal accross India has been known to meteorologists for a long time. Heavy rainfall associated with these tropical disturbances is responsible for a significant percentage of the rainfall over Northern and Central India during the summer monsoon season. One the objectives of the Summer Monsoon Experiment which took place in 1979 was to examine the structure and dynamics of these phenomena.

Wave like convective features have been identified over the South Indian ocean but the lack of data has hampered detailed studies of tropical perturbations over that area.

In this paper, the characteristics of tropical disturbances occurring over different regions are reviewed. The main differences between the different systems will be stressed. Their importance in the global circulation of the tropics will also be discussed. This last point has been addressed by very few scientists although it is important to understand how these perturbations fit in the dynamics of the tropical circulation.

2. Global features of the disturbances

As discussed in the introduction, the existence of tropical disturbances over several areas in the tropics has been demonstrated as early as the 1950s and 60s. The first global view of tropical disturbances was obtained after the implementation of operational meteorological satellites. Time-longitude diagrams of cloud cover at different latitudes (more frequently averaged between 10°N and 10°S) revealed the global nature of synoptic scale systems in the tropics. In particular, this kind of analysis showed that this type of meteorological systems can propagate over long distances (several thousand kilometers). For example, it was shown that African waves generated somewhere around 20°E can propagate over the Atlantic ocean and intensify into cyclones during the season.

Although limited to the tropics and taking place at a time when operational assimilation of data by global circulation models was not performed, GATE (GARP Atlantic Tropical Experiment) in 1974 was the first large scale experiment to cover the tropics and to give a detailed description of meteorological fields in the tropical belt during the summer season. Fig. 1 gives some evidence of the existence of synoptic tropical disturbances in the 3-5 days spectral range. Krishnamurti (1978) carried out the analysis of the upper troposphere (200hPa) horizontal motion fields on a daily basis using simple objective analysis successive correction method for 100 days of summer 1974. The wind vectors were separated into divergent and rotational parts. The intensity of east-west circulation was defined according to the following formula :

$$I_E\ (x,t) = \frac{-1}{(y_2 - y_1)} \int_{y_1}^{y_2} \frac{\partial \chi}{\partial y} dy$$

with χ the velocity potential, and y_1 and y_2 the latitude boundaries of a tropical domain within which the east-west circulations are well defined. In this case, y_1 and y_2 were selected to be 10°S and 20°N.

The wavenumber-frequency spectra of the east-west circulation index was performed. Fig. 1 shows contours of power plotted for wavenumber against period (in days). Besides the existence of 10-to-20 day time scale waves with dominant spatial scales around zonal wave number 4 through 6, there is a dominant spatial scale around zonal wave numbers 7 and 8 corresponding to a westward propagating mode with a typical periodicity of 3 to 5 days. These phenomena correspond to the synoptic-scale tropical disturbances that carry with them vertical circulations.

The First GARP Global Experiment (FGGE) provided the meteorological community with a complete data set covering the entire atmosphere over a one year period (November 1978 - October 1979). In particular, the so-called Level III-b data set produced by the European Center for Medium Range Weather Forecast (ECMWF) has been used to study the global aspects of synoptic tropical disturbances (Nitta et al., 1985).

Fig. 2 shows the time-longitude section of the meridional component v of the horizontal wind at 850hPa during the period from 15 July to 31 August 1979 at 9.375°N where the ITCZ is generally located. Deviations of v from the time mean are plotted. It can be noticed that there exist short period variations propagating westward with periods of 3-7 days in general and that disturbance activities are localized in space and their characteristic features such as periods and phase speeds are largely different over different regions in the tropical belt.

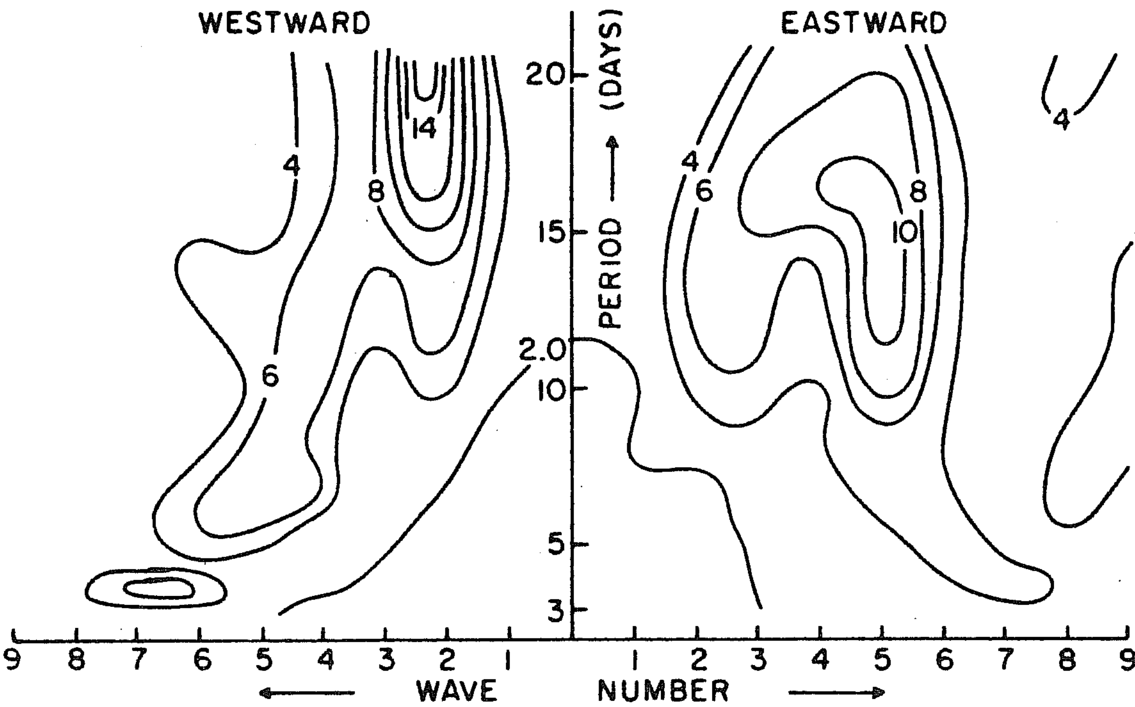


Fig. 1 Power spectral analysis displayed in wavenumber-frequency (1/period) diagram, for the I_B of east-west circulations.

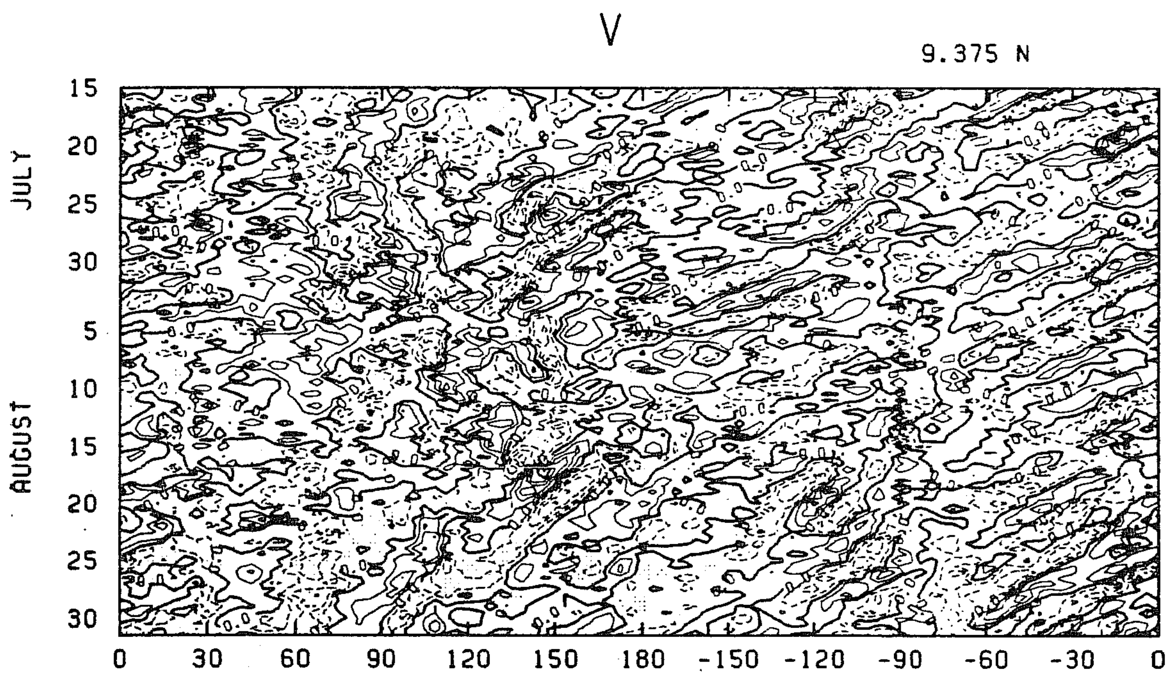


Fig. 2 Time-longitude section of the north-south wind component deviating from the time mean at 9.375°N. The contour interval is 2.5 m s^{-1}. Northerlies are shaded.

Westward propagating disturbances are clearly seen in the region from West Africa to Central Atlantic (0°-60°W). The period is about 3-4 days and this kind of disturbance corresponds to the African wave. This disturbance appears to be originating in Central Africa near 30°E, has a maximum amplitude near the west coast of Africa (≃ 15°W), weakens as it propagates westward in the Atlantic ocean and becomes obscure over South America (60°W - 80°W).

In the eastern Pacific region, another westward disturbance with periods of 4-6 days appears. Some of these disturbances propagate further west of the date line, but most disturbances appear to be disconnected with those in the western Pacific.

Disturbances with large amplitudes move westward with periods of 5-7 days in the western Pacific region between 120°E and 170°E. The westward phase speed of the disturbances in the western Pacific is slower than those in the central and eastern Pacific. No systematic phase propagation is found in the Indian ocean (50°E-100°E) and long period variations with periods longer than 10 days dominate in this region.

Tropical depressions do exist in the Indian ocean sector but they appear in the Bay of Bengal and cannot show up in the diagram corresponding to a latitude of 9-375°N.

Fig. 3 shows the latitudinal distribution of the mean power spectral densities obtained by averaging time spectral densities of v at individual grid points along the same longitude circle. In the equatorial region between 15°N and 15°S, the disturbances have larger amplitudes over the northern latitudes than over the southern latitudes. There exist two dominant spectral peaks in the northern equatorial region, one at 6-7 days period and the other at 3.5-4 day period. The former variation has the maximum amplitude around 10°N and the latter has maximum amplitude slightly to the south of it. The short period variation certainly corresponds to the African waves whereas the longer period may be a reflection of the wave disturbances appearing in the western Pacific and those in the eastern Pacific.

Fig. 4 and 5 show the global distributions of the integrated power spectra of v for the 3.1-4.4 day and the 5-10 day period ranges, respectively. For the first period, the large amplitude is found in the region between the equator and 25°N from the central Africa to the Central Atlantic ocean. It can be attributed to the dominant spectral peak of about 3.6 day period associated with the passage of the African wave disturbances.

For the 5-10 day period range, the maximum intensities are located from the tropical western Pacific to the east China sea. There exist considerable amounts of disturbance activities from Indochina to North India where the dominant periodicity is 6-7 days. These areas correspond to the monsoon trough region. Large amplitudes are also found in the eastern tropical Pacific.

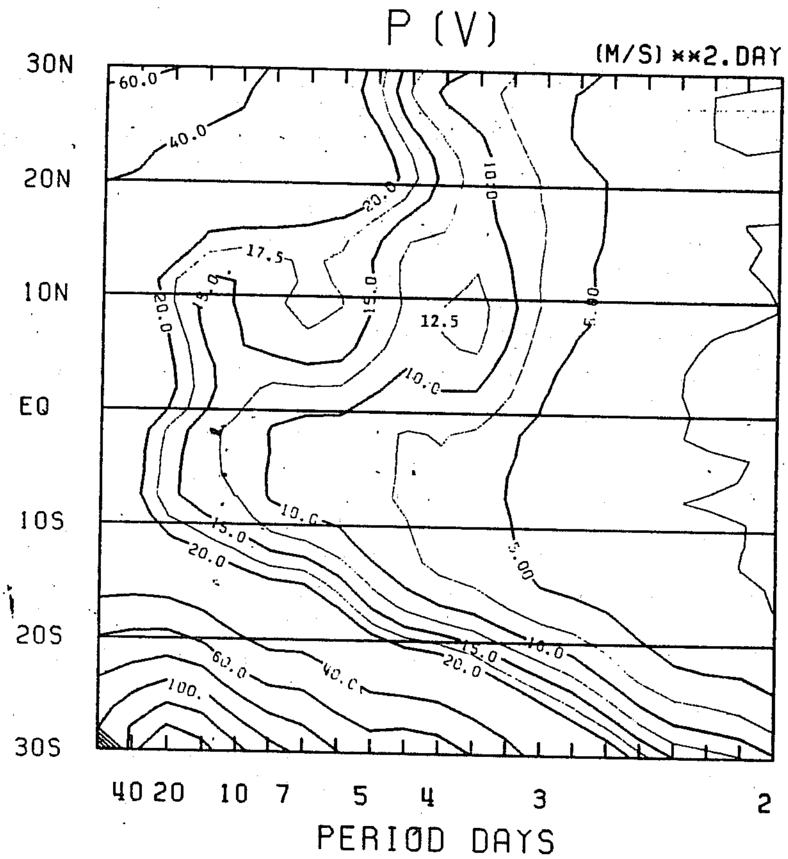


Fig. 3 Zonally averaged power spectra in periods. The contour intervals are 2.5 m^2s^{-2} day for values less than 20 m^2s^{-2} day and are 20 m^2s^{-2} day for values larger than 20 m^2s^{-2} day.

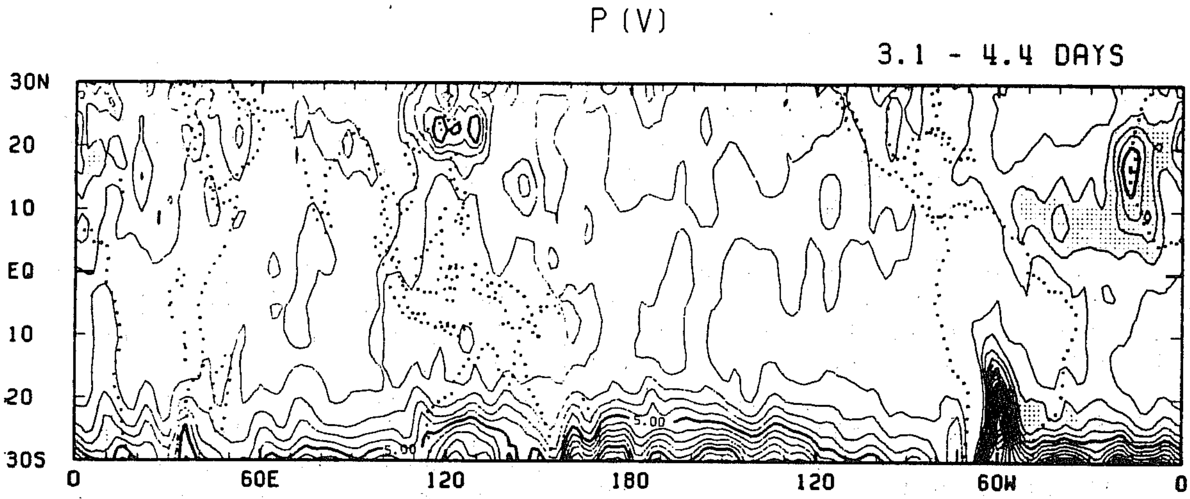


Fig. 4 Global distribution of the v-spectra integrated from 3.1 days to 4.4 days; the contour interval is 1 unit. Regions larger than 2 units are shaded. Units are m^2s^{-2}.

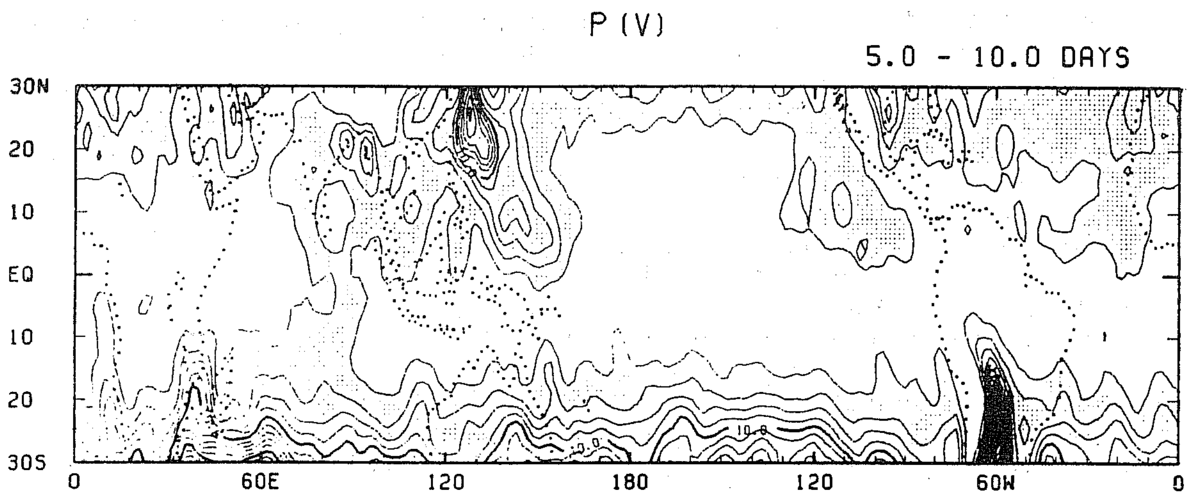


Fig. 5 v-spectra integrated from 5 days to 10 days with contour intervals of 2 units.

Fig. 6 shows the distribution of the integrated OLR (Outgoing Longwave Radiation) - spectra for the 3.1.-10 day period. OLR is obtained from polar orbiting satellites. It is a good indication of convective activity in the tropics.

The area of large variations of the convective activity correspond quite well to those of v as shown in Fig. 4 and 5. Especially, the correspondence is striking in the tropical western Pacific, the Bay of Bengal and the eastern Pacific off the coast of Central America, where the disturbances with periods of 5-10 days dominate. A large amplitude of OLR is also found in the area from the Central Africa to the Atlantic ocean where the African wave disturbances are observed. These areas correspond to tracks of the African wave disturbances. These results indicate that the convective activity is strongly coupled with the lower tropospheric disturbances.

Fig. 4, 5 and 6 show that the seemingly scattered wave disturbances and depressions around the global tropics are, in fact, organized on the planetary scale. This has some implications for the general circulation of the tropical atmosphere. Depradine (1980) studied the energetics of large-scale motion using the GATE data sets. Fig. 7 presents the net gain of available potential energy by the long waves (L = 1-3), medium waves (M = 4-5), short waves (S = 6-15) and the zonal flow (Z). The most striking result of this analysis is the net transfer of available potential energy from the short waves to the long waves. This is the largest energy transfer. Short waves correspond to synoptic disturbances and storms. From this result and the previous discussion a clear picture emerges. It is obvious that the placing of individual tropical disturbances favors a planetary scale energy release. This is due to an apparent zonal asymmetry of the tropical regions of storm activity being further north (about 20°N) over the monsoon regions and closest to about 10°N over the oceanic ITCZ. Furthermore, the activity of these disturbances takes place at certain longitudes. This transfer of energy from the storms to the planetary scale is an integral aspect of the tropical general circulation. Only a small proportion of the energy released by the vertical overturnings in the individual storms is used to drive the disturbance itself. This may contradict some results obtained from the study of the energetics of disturbance over a limited domain. It has been noted in different studies that the covariance of the vertical velocity and temperature integrated through the depth of the atmosphere along one wavelength usually shows a net generation of eddy kinetic energy from the eddy available potential energy. What has been misleading in the literature is that in several studies on the limited area energy budgets of a tropical wave, all the energy released from this covariance has been utilized to account for the maintenance of eddy kinetic of the wave itself. Only a small proportion of the energy release by the vertical overturnings in the individual storms is used to drive the disturbance itself. Rather, it appears that most of the energy by virtue of the spatial organization of disturbances on the planetary scale drives the planetary scale tropical motions. In this context, it should be noted that vertical circulations on the planetary scale over the tropics are a statistical result, that is, in

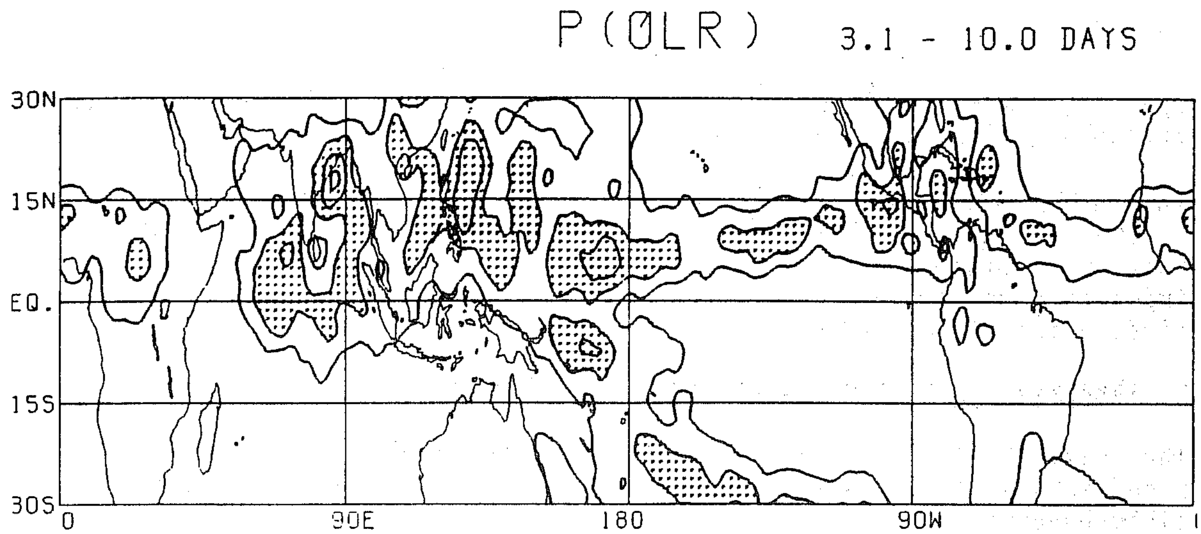


Fig. 6 Distribution of the OLR-spectra integrated from 3.1 days to 10 days. The contour interval is 200 $(W\ m^{-2})^2$. Regions larger than 400 $(W\ m^{-2})^2$ are shaded.

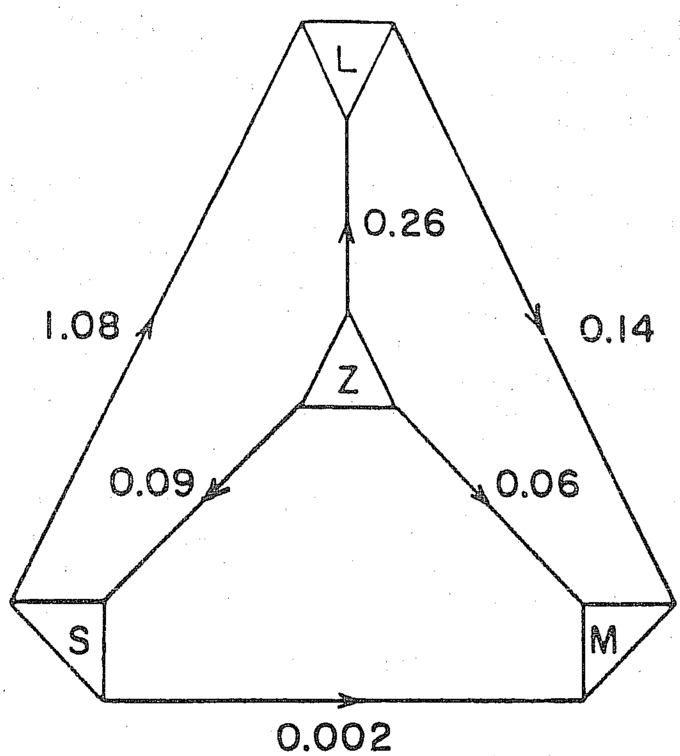


Fig. 7 Net gain of available potential energy by the long waves (L = 1-3), medium waves (M = 4-5), short waves (S = 6-15) and the zonal flow (Z); units: $10^{-6}\ m^2\ s^{-3}$.

reality, parcels ascend in cumulus clouds within the disturbances (Krishnamurti et al., 1983).

This discussion also tells us that results based on the classical Lorenz energy cycle formalism must be looked upon with caution when they are obtained over a limited domain. These results stress the importance of the role played by tropical disturbances in the general circulation of the tropics.

3. Synoptic - scale disturbances in the Caribbean

The summertime tropical Atlantic is populated by major weather-producing systems that are associated with synoptic scale wave disturbances.

Case studies of tropical disturbances have been a very fruitful area of research in the Caribbean region. The easterly wave was extensively studied by Riehl (1945) (summarized in Riehl (1954)). The so-called classical easterly wave is based on his pioneering work.

Fig. 8 and 9 show two time sections at Puerto-Rico on July 11-13, 1944 and Grand Cayman Island during the passage of easterly waves. These two examples represent two of the well-defined examples of wave passage. The following features can be stressed :

- The wave trough tilts eastward with height as shown by the near-vertical solid line. As the wave is moving westward, the upper part arrives later than the lower part. The wind shift shown by the directional change in wind barbs (knots) indicates that the wave extends through the depth of the troposphere.

- The isotherms in this time section (Fig. 9) are shown as a departure of the observed temperature from a tropical standard atmosphere. There is a cold core below the upper trough with higher temperatures to the east at low levels. In this example, the thermal amplitude of the wave at lower levels is somewhat large compared to other such cases. Between 500 and 250 hPa, a reversal in the thermal anomaly structure may be noted : the warm core above the low level cold core can be attributed to the effects of cumulus convection. The thermal amplitude near 300 hPa is about one half that at 900 hPa.

- The moisture field shown by the dashed lines in Fig. 9 indicates that the region east of the wave trough is more moist that the region ahead (i.e. west) of it. The convective weather and large scale upward rising motion tends to occur near the wave axis and to the east of it.

The horizontal structure of these waves has been analyzed by many scientists. Fig. 10 gives the example of the structure of such a wave in the lower troposphere with the trough line indicated by zero meridional wind component.

This model of easterly wave does not encompass all cases. The vertical shear of the large-scale environmental flows varies. Thus, the structure, the scale and the life cycle of disturbances can vary

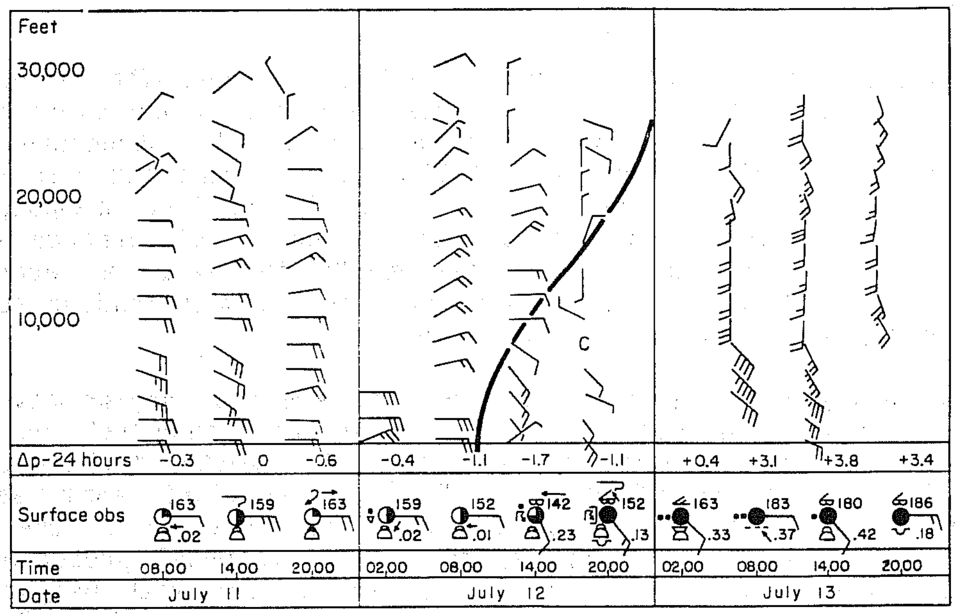


Fig. 8 Vertical time section at San Juan, Puerto Rico. July 11-13, 1944.

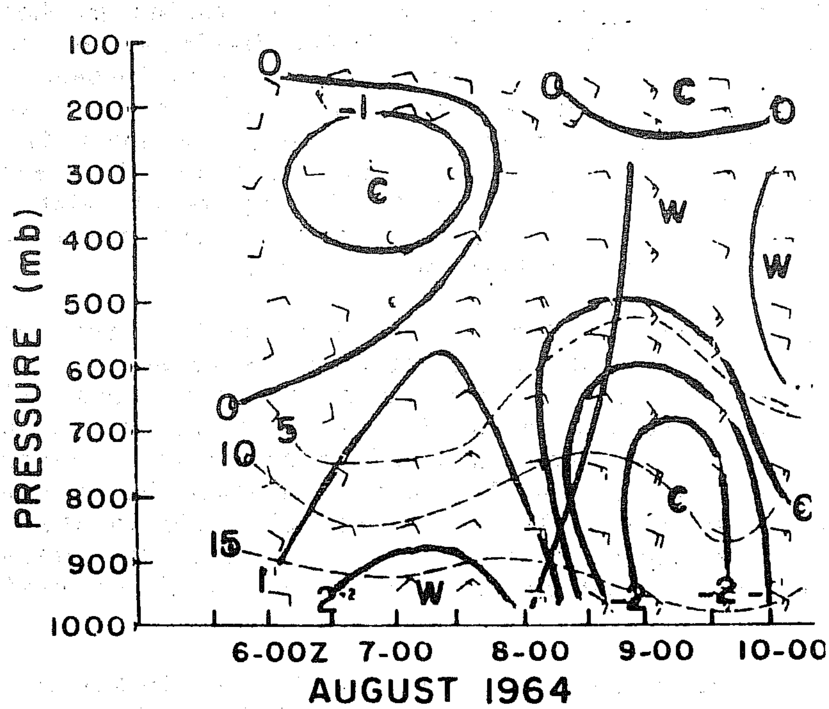


Fig. 9 A vertical time section showing winds (knots), temperature °C (solid lines) and specific humidity (dashed line) g kg^{-1}. The diagram is based on Riehl's study of easterly waves.

considerably. For example, when the easterlies increase with height in the upper troposphere, the Atlantic waves have been known to resemble western Pacific easterly waves. Differencies in the structure have been found, for example, by Yanai (1968). The disturbance which was transforming into a hurricane, was cold cored in the lower troposphere at the beginning. Through an interaction between the disturbance and a pronounced upper-level shear line situated to the west, a secondary warm low was initiated at lower levels near the surface leading to the disappearence of the cold core wave disturbance.

Composite analysis of the GATE data in the Caribbean region revealed the existence of a moderate eastward tilt of the trough axis up to ≃ 700 hPa, with a strong westward tilt above. The westward phase shift was 90° between 700 and 200 hPa (fig. 11). Shapiro (1986) analyzed the three-dimensional structure of a series of easterly waves in the area (during July 1975). In contrast to the classical model presented by Riehl, he found a westward tilt of the trough axis with height. In particular, both the vorticity and the meridional velocity consistently evidenced a near 90° westward phase shift and 200 mb relative to the lower troposphere.

The influence of the basic environment has been studied by Holton (1971) and Shapiro et al. (1988). Holton (1971) developed a simple model to analyze the structure of linear forced wave disturbances in the presence of vertical shear of the mean zonal wind. In the case of a mean wind with westerly shear from the surface up to about 12 km, solutions indicated a surface trough to the west of the heating maximum and a eastward tilt with height in agreement with the classical wave model. With an easterly shear in the basic wind profile, the trough axis tilted westward with height.

Shapiro et al. (1988) concluded that the eastward tilt of the trough axis in the classical picture of the Caribbean easterly wave is not a simple consequence of the vertical profile of the mean zonal wind. Latitude of the disturbance may be an important factor. For example, with a westerly shear in the lower and middle troposphere, a westward tilt above 700mb is found at the latitude of maximum heating when centered at 19°N. An eastward tilt is favored below 400 mb when the heating is more to the south, near 9°N. In that case, the westward phase shift occurs over a narrow layer near the level of maximum heating.

A large number of studies have been devoted to the formation and maintenance of the easterly waves. Because about half of the easterly waves in the Caribbean originates over Africa whereas the other half form over Central Atlantic, the question of the development will be discussed later.

There is a general feeling that the barotropic mechanism may be most important during the initial formative stage of Atlantic waves. Over the western Atlantic, the disturbances loose energy to local zonal flows indicating that the barotropic mechanism is not important in the maintenance of the waves and the role of convection is perhaps more important. In a study of a westward propagating easterly wave (1961

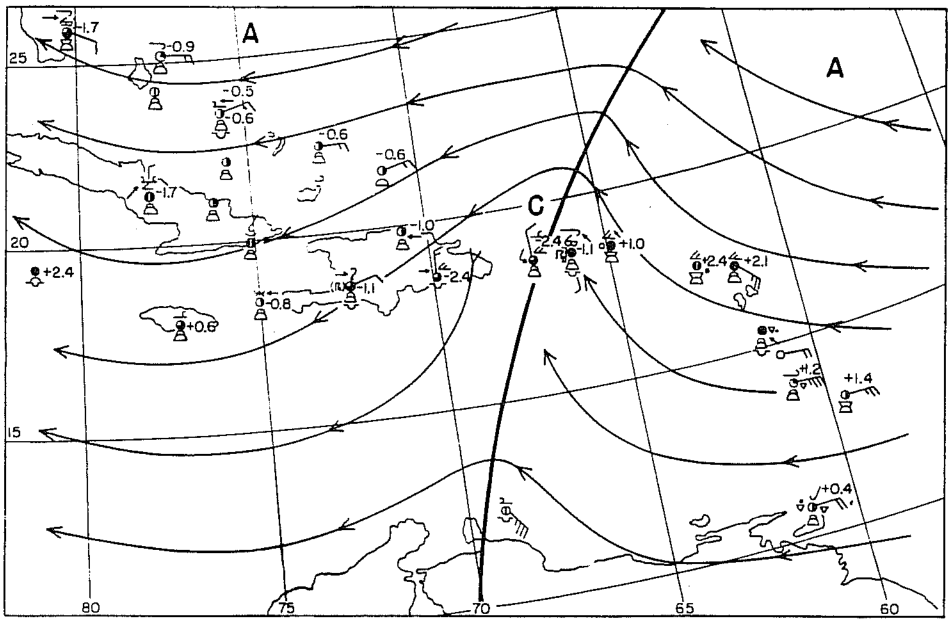


Fig. 10 5,000-foot winds. surface 24-hour pressure changes and weather reports for the Caribbean area, July 12, 1944.

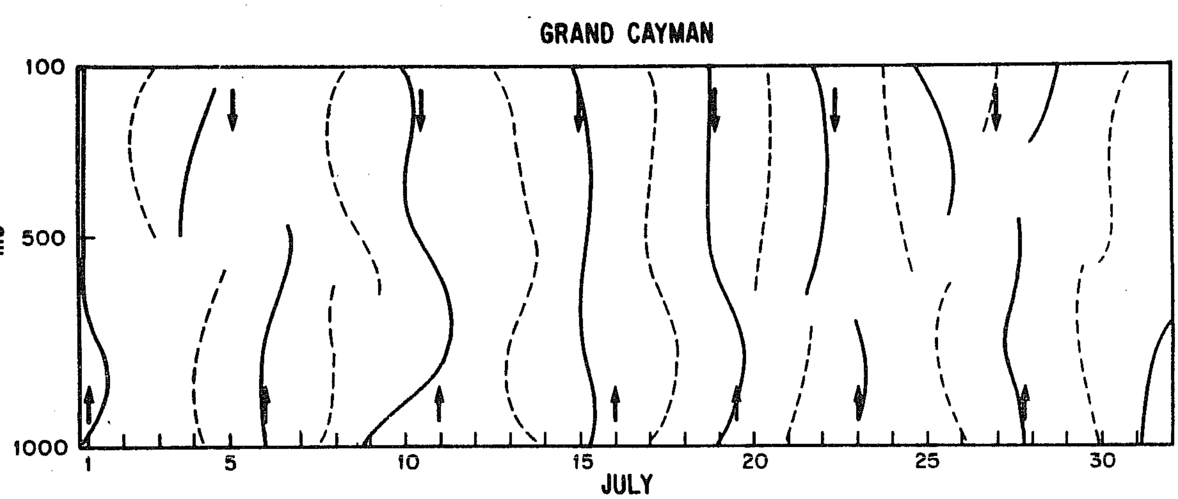


Fig. 11 Troughs (solid lines) and ridges (dashed lines) from meridional winds in 3-5 day band at Grand Cayman in July 1975. Arrows denote troughs from reconstructed dominant combined EOF.

case), Krishnamurti and Kanamitsu (1973) found that the transformation of eddy available potential energy into eddy kinetic energy is the important mechanism for the maintenance of the easterly waves (fig. 12). This mechanism seems to operate in the following manner :

Ascending motion occurs at the wave axis and to its rear. Over this region a warm core is present near the 500 mb surface. Subsidence occurs all around and ahead of the wave axis. The net result of this ascent of relatively warm moist air and descent of relatively colder and dry air is that the generation of eddy kinetic energy occurs from the eddy available potential energy. However, it must be noted that Krishnamurti and Kanatmitsu (1973) results are not obtained from observations but from numerical modelling. Consequently, their results can be questionable.

The westward motion of a non-developing easterly wave is a fairly complex problem. Estimates of the various terms of the vorticity equation are generally used to determine which of these contribute to a positive tendency of vorticity west of the surface trough which will make the disturbance to move westwards. Positive vorticity advection contributes most significantly to this local increase of vorticity ahead of the surface through. This effect is somewhat counteracted by the divergence term of the vorticity equation.

In their study, Krishnamurti and Kanamitsu (1973) looked at the maintenance of the cold low 1° or 2°C which exists within in an easterly wave, especially in the Caribbean where they maintain a thermal structure in near steady state.

Some numerical attempts for that case study with and without convection or radiation temperature gradients showed that the cold core is primarily maintained by ascending motions in the region of the boundary layer of the active portion of the wave. Ascending motions arising due to frictional convergence and heating aloft account for adiabatic cooling. Thus, the cold core is not merely an advective phenomenon that translates from east to west but it is maintained by dynamic and thermodynamic processes.

4. African waves

The existence of African waves was suggested as early as the 1930s. Further evidence came out with the availability of satellite images. Burpee (1972) made a pioneering study on the structure of African waves despite the sparseness of synoptic data over that continent. He also discussed the physical causes of these perturbations. He showed that both the horizontal and vertical shears of the mean flow may be sources of energy for the perturbations due to the existence of a jet stream around 700hPa over west Africa. This jet is a thermal wind due to the temperature gradient between the air above the Sahara desert and the moist and cooler air more south.

GATE which took place during the summer 1974 gave the data set that scientists have been waiting for a long time. During the decade following this international experiment, a certain number of studies

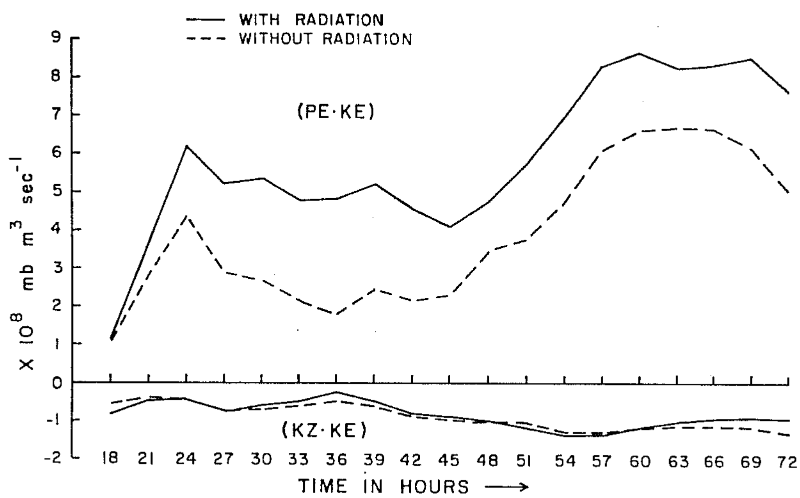


Fig. 12 Barotropic and baroclinic energy exchanges between 18 hours and 72 hours for two experiments, with and without long-wave radiative cooling effects. The energy exchange are evaluated for a latitude belt 12° wide across the active part of the wave and between 100 and 1,000 mb surfaces.

were devoted to describe the structure of Africain waves and to understand their dynamics. Studies have focused on the third phase of the experiment (23 August to 19 September, 1974) because during that period, the perturbations were better organized and more unform in behavior than those observed earlier in GATE and thus were particularly well-suited for compositing. The composite technique has been widely used during GATE with the drawback of smoothing differences between the various cases making up the compositing. The main references are Reed et al. (1977); Thompson et al. (1979), Norquist (1977), Albignat and Reed (1980). The following presentation is largely based on their findings.

A detailed description of the perturbation can be found in Reed et al. (1977). Fig. 13 gives an example of results. The major findings can be summarized as follows :

. The disturbances with an average wavelength of about 2 500 km and a period of 3.5 days are centered in the cyclonic shear zone south of the easterly mid-tropospheric jet. The necessary condition for barotropic instability was met in the region of the jet.

. The wave axis has a pronounced northeast-southwest tilt in the vicinity of 700hPa. This implies a strong northward transport of westerly momentum and a conversion of zonal kinetic energy into eddy kinetic energy.

. In the vicinity of the disturbance path (around 10°N), the systems are essentially vertical below 700hPa and sloped westward above. In the region of strongest baroclinicity to the north of the path they sloped eastward with height in the easterly shear zone below 700hPa suggesting that baroclinic wave growth occurs in this region.

. The disturbances are cold core below 650 hPa, warm core between that level and 250hPa and cold core at still higher levels. Consistent with the thermal wind equation, the vorticity was largest ($3 10^{-5}$ S^{-1}) at the 650hPa level. The largest anticyclonic vorticity is observed in the upper troposphere centered well to the north of the lower level cyclonic center.

. Largest temperature fluctuations occurred at 850hPa in the region of strong temperature contrast north of the disturbance track. The negative correlation between the temperature and the meridional wind component indicated a conversion of zonal to eddy available potential energy. Temperature variations in tropical easterly waves are small.

. The strongest upward motion is found at 700hPa in advance of the trough. Upward motion is correlated with warm temperature anomalies in the baroclinic zone around 850hPa and rather generally in the layer about 300hPa, suggesting the possibility that the waves are maintained, at least in part, by a baroclinic energy conversion.

Norquist et al. (1977) studied the energetics of African wave disturbances using the formalism derived by Lorenz although it does not really apply over a limited domain. The results are presented in

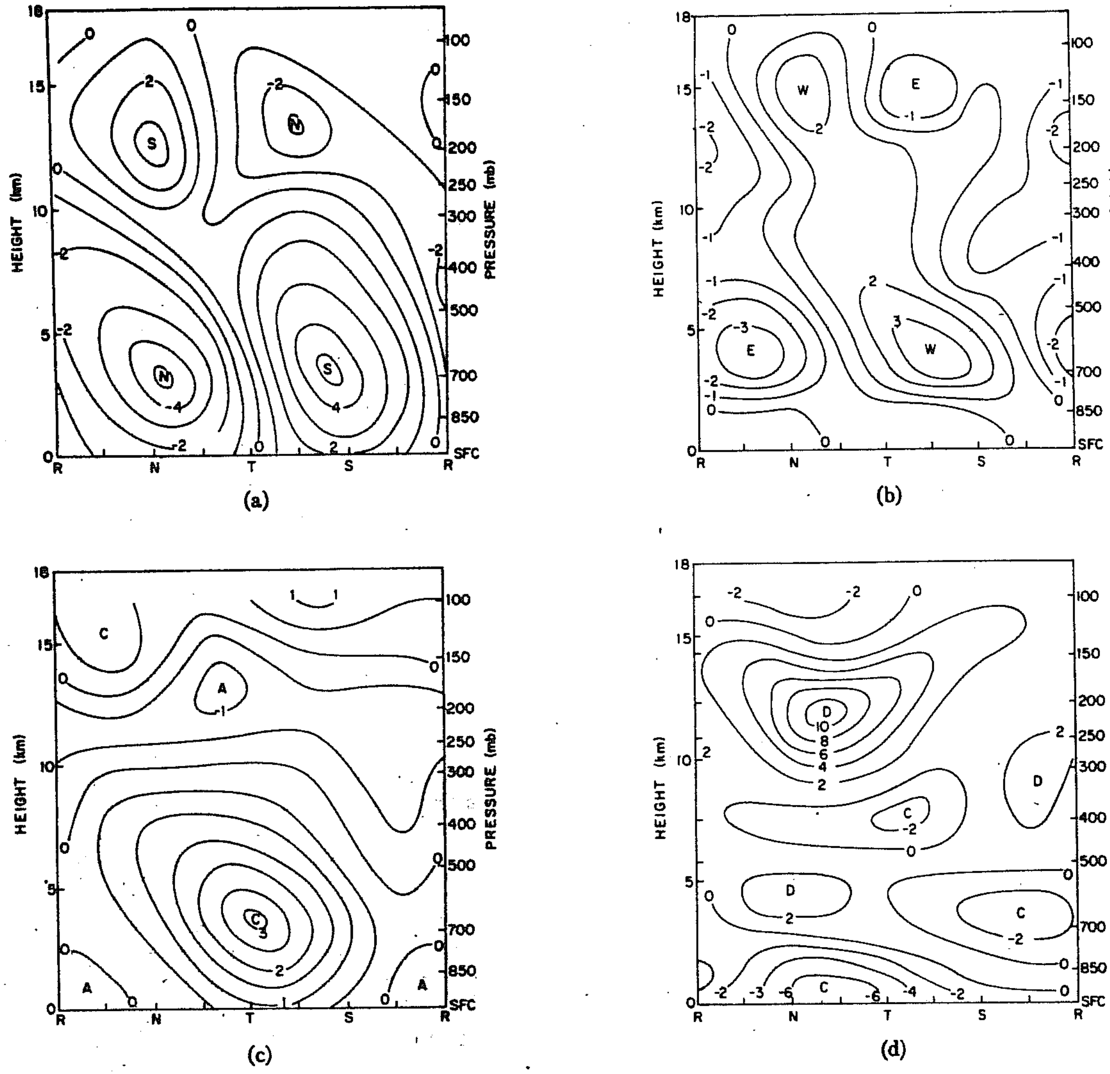


Fig. 13 Vertical cross sections along reference latitude. (a) Meridional wind deviation (m s^{-1}); (b) zonal wind deviation (m s^{-1}); (c) vorticity ($10^{-5}s^{-1}$); (d) divergence ($10^{-6}s^{-1}$). R, N, T, S refer to ridge, north wind, trough, south wind sectors of the wave, respectively.

Fig. 14 for the combined region. These computation indicate that the eddy kinetic energy K_E is about twice the eddy available potential energy A_E and that both are about an order of magnitude smaller than the corresponding mean energies (K_Z, A_Z). The eddy kinetic energy is seen to be maintained almost equally by the barotropic conversion C_K and the baroclinic conversion C_E. It is useful to remind that the first term of C_K is :

$$C_K = - \frac{1}{g} \int_{100}^{P_s} [u'v'] \frac{\partial [u]}{\partial y} dp$$

with a bracket representing a zonal average and a bar a meridional average.

C_E is defined as :

$$- \frac{1}{g} \int_{100}^{P_s} \frac{R}{p} [w'T'] dp$$

The doubling time for the eddy kinetic energy is about 3 days.

The results seem to indicate that the generation of eddy available potential energy by latent heat release is small if the boundary flux is small. However, one must be cautious about this result because the estimation of the vertical velocity and the measurement of temperature are not sufficiently accurate. Furthermore, as said before, the formalism does not really apply to a limited domain.

The results for the subregions indicate some rather striking differences in the energetics of the two regions. For example, the baroclinic conversion C_E is much larger over the land than over the ocean and conversely the barotropic conversion C_K being substantially larger over the ocean.

Although there is some uncertainty in determining C_E, the barotropic conversions based on large samples of wind observations of sufficient accuracy are certainly reliable. The finding that C_K is larger over the ocean than over land can be guessed from the structure of the waves over the two domains. Over the ocean, the perturbations are larger and the wave axis has a more pronounced tilt, contributing to larger eddy momentum flux. The shear of the mean zonal wind is also stronger over the ocean. If frictional dissipation is assumed the same over the two regions, it implies that C_E must be larger over land in order to maintain the eddy kinetic energy balance. However, if frictional dissipation is stronger over the ocean, the argument fails.

If the measured energy conversions are accepted, we can deduce another important difference between the energetics of the two regions. Over land, a substantial generation of eddy available potential energy (latent heat of condensation) is required for

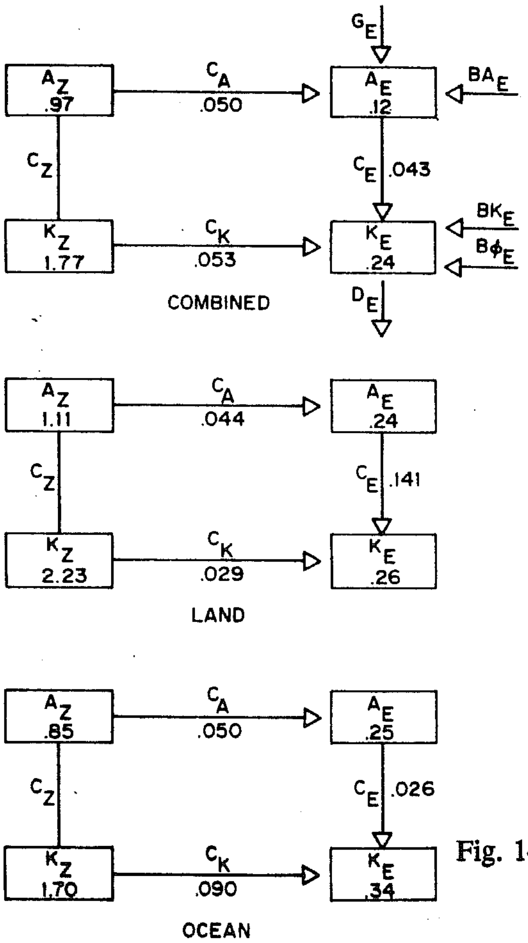


Fig. 14 Partitioned energies (10^6 J m^{-2}) and energy conversion rates (W m^{-2}) for land, ocean and combined regions. Symbols are defined in text.

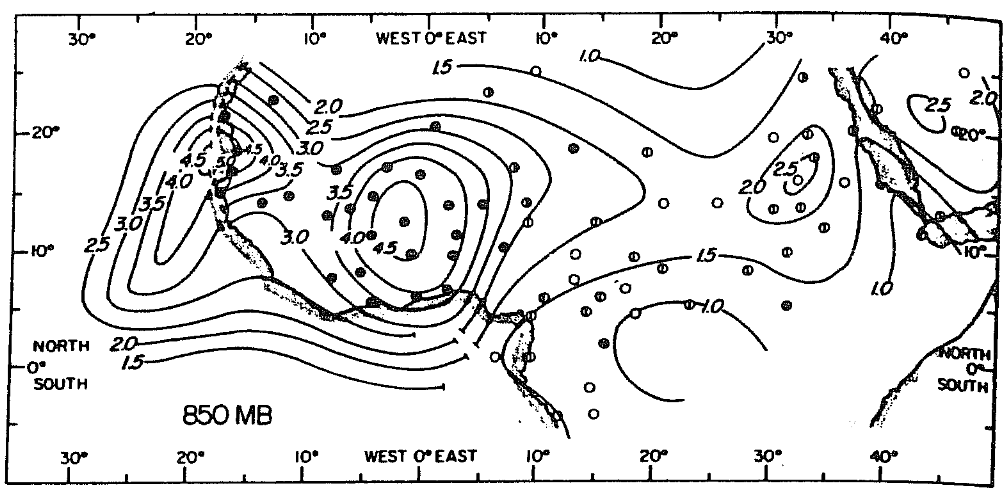


Fig. 15 Amplitude of meridional wind oscillation (m s^{-1}) in the frequency band 0.2-0.4 cpd at 850 mb.

balance. Over the ocean the residual is small and negative, suggesting no significant growth of the disturbances due to latent heat release and an overall decline in energy due to diabatic processes.

In a refined study of the structure and energetics of wave disturbances over the eastern Atlantic, Thompson et al. (1979) found a substantial conversion from eddy kinetic to available potential energy, the generation of eddy available potential energy G_E being even negative. Thus, eddy kinetic energy is transformed into eddy available potential energy and is in turn lost by diabatic heating.

The variations with latitude and height of the leading term in the expression for C_K indicate that the only region of significant conversion is located at the height of the mid-tropospheric jet stream (Norquist et al. 1977). The study of C_E is more complex. The leading term in the conversion of zonal to eddy available potential energy exhibits a positive value near 850hPa at about 15°N. This can be attributed to the correlation of warm temperature anomalies with northerly wind component and thus with southward directed heat flux in the region of relatively strong temperature gradient to the south of the Sahara. Thus it appears that the thermal gradient associated with the mid-tropospheric jet stream provides about the same amount of perturbation energy as the lateral wind shear.

The question of the origin of African waves had been discussed by nearly all scientists who studied these perturbations. Fig. 15 gives the amplitude of the meridional wind oscillation in the period band 2.5-5 days at 850hPa based on GATE data (Albignat and Reed, 1980). This figure indicates that the main growth of the disturbances occurred between 10°E and 0° with the indication of a kind of corridor extending further east. Study of the direction of phase propagation and coherence as given by cross-spectrum analysis does not demonstrate a westward propagation east of 10°E at 850hPa whereas the result is clearer at 700hPa.

Burpee (1972) and Carlson (1969) proposed that the waves originate over Central Africa as a consequence of a joint baroclinic-barotropic instability. In the early stages of the waves, convection is unorganized and latent heat release plays little part in the wave development. As they progress westward, the convection becomes better organized so that over West Africa the condensation heating becomes a dominant factor in their growth and sustenance. Fig. 16 taken from Albignat and Reed (1980) gives the northward flux of zonal momentum ($\overline{u'v'}$) at 700hPa in the frequency band of African waves. The barotropic conversion is proportional to the north-south flux of zonal momentum ($\overline{u'v'}$) and the gradient of the mean zonal wind $\left(\frac{\partial \overline{u}}{\partial y}\right)$

Flux of easterly momentum away from the jet core of westerly momentum toward the core indicates wave growth at the expense of the mean zonal current. It is apparent that the flux reverses sign along almost the jet axis, indicating a down-gradient flux of easterly momentum. The strongest fluxes occur west of 10°E. A minor flux maximum is present over southern Soudan. The mid-tropospheric jet cannot be traced well

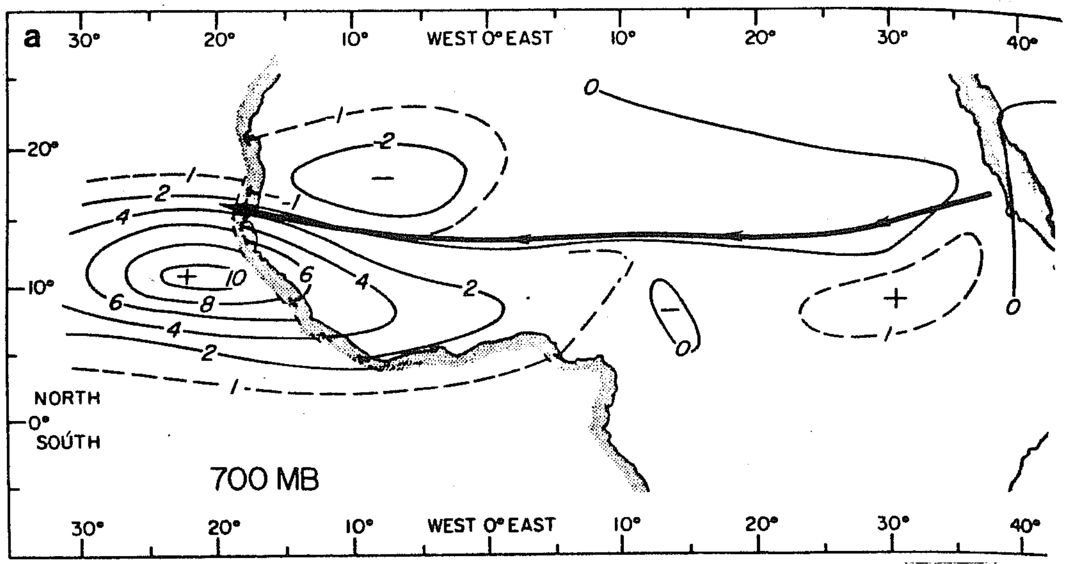


Fig. 16 Northward flux of zonal momentum ($\overline{u'v'}$) at 700 mb in 0.2 - 0.4 cpd frequency band. Units: m^2s^{-2}.

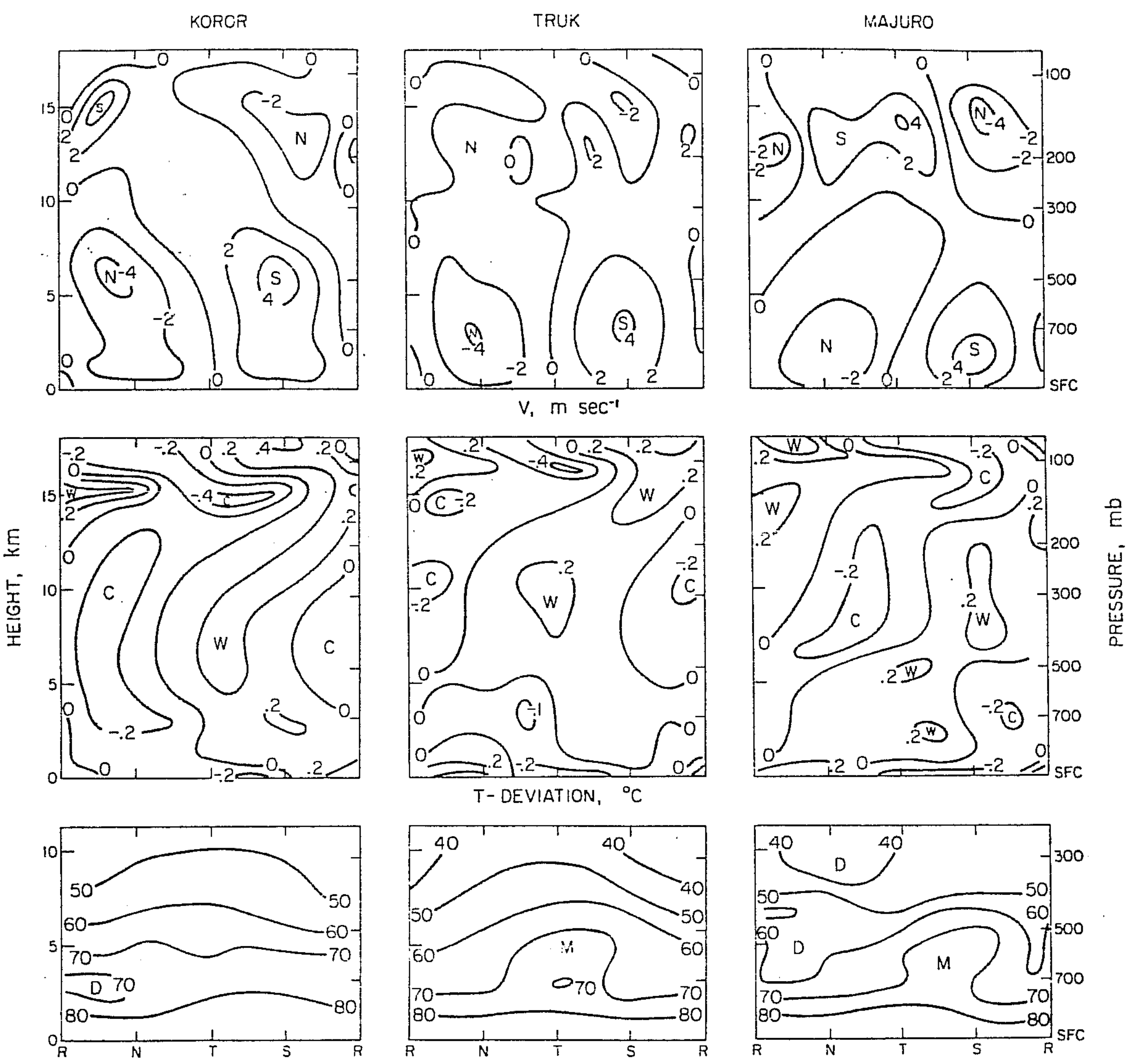


Fig. 17 Composite diagrams of meridional wind component (m sec^{-1}), temperature deviation (°C) and relative humidity for Koror, Truk and Majuro.

over eastern Africa and barotropic energy exchange cannot be called upon for generation of the waves.

The importance for baroclinic conversion can be guessed from the vertical tilt of the system. For the waves to receive their energy through baroclinic conversion, it is necessary for them to lean in the direction opposite to the shear of the basic current i.e. to lean eastward with height at levels beneath the easterly jet stream. North of 15°N a systematic eastward tilt with height is observed indicating that baroclinic conversion contributes to the wave growth near 10°E. East of this longitude the tour is weak.

Several numerical modelling studies have been devoted to the origin of African waves (Rennick, 1976 ; Simmons, 1977 ; Mass, 1979 and Kwon, 1989). Using a 10 level linearized primitive equation model with a parametrization of latent heating and cumulus momentum mixing, a model easterly wave was reproduced by Mass (1979). However, there was some discrepancies especially with respect to the energetics. In particular, a center of downward motion was lying over a center of ascending motion which in conjunction with the temperature pattern and the vertical velocity field implies a conversion from wave kinetic energy to wave potential energy. Thus the wave gets its kinetic energy from the zonal flow via barotropic instability and a part of the kinetic energy is then converted into eddy potential energy which would be eventually destroyed by the heating. This is opposite to observations. Besides, the conversion rate from mean to eddy kinetic energy of the model is one order larger than that of the observation.

The question was reinvestigated by Kwon (1989). In the model it was shown how the growth rate, the propagation speed and the overall composite structure of the observed tropical easterly waves can be well reproduced with a CISK representation of the latent heating. The kinetic energy of the unstable wave is generated via barotropic instability associated with the horizontal shear of the mid-tropospheric jet. However, the vertical shear of the zonal flow also strongly influences the thermal and vertical velocity fields of the wave. From the energetics point of view, the latent heating tends to stabilize the motion at the middle level and to destabilize it below 600hPa, indicating baroclinic instability in the lower troposphere.

5. Synoptic-scale wave disturbances in the Pacific

As said earlier, disturbances exist over the Pacific ocean in its eastern and western sector. They have been studied separately and we will follow the same way in this paper.

Wave disturbances in the western Pacific have been studied in the 1960s by Yanai (1963), Wallace (1971), Nitta (1970) among many others. Reed and Recker (1971) made a detailed analysis of the properties of these systems. The discussion of the tropical disturbances over that area will rely heavily on this last work.

Fig.17 gives the composite diagrams of meridional wind component, temperature deviation and relative humidity at three stations during

July-September 1967 (18 cases). The average wavelength is about 3500-4000 km. They travel westward at an average speed of 7° longitude day^{-1} (i.e. 9m $sec.^{-1}$). Meridional wind fluctuations are strongest near 800hPa (3-4 m $sec.^{-1}$) and 175hPa (2-3 m sec^{-1}), the upper and lower fluctuations being nearly out of phase. Temperature fluctuations are 1°C or less with cold anomalies at lower levels in the region just behind the trough and in the upper troposphere above the trough. There is a region of warm anomaly in between. The vertical motion appears upward everywhere except in the vicinity of the low-level ridge. Convergence predominates at low levels in the waves. The level of non-divergence varies from 500 to 300hPa. The region of strong divergence is centered at 175hPa above the low-level trough zone.

The structure of the waves in the western Pacific undergoes a systematic change as they travel westward. The wave axis which tilts eastward with height in the eastern sector, becomes vertical in the central region and acquires an opposite tilt in the west. This change of wave structure is believed to be caused by the variation with longitude of the vertical shear of the basic current.

Concerning energy exchange, Wallace (1971) and Nitta (1970) found that diabatic heating generates eddy available potential energy which provides the source of eddy kinetic energy. This picture is radically different in the eastern Atlantic as seen above. This opposition in the signs of baroclinic conversion process in the Atlantic and Pacific can be traced to the difference in the vertical velocity fields.

In the GATE area, the strongest upward motions occur at lower levels in the wave trough where negative temperature anomalies prevail. In the western Pacific, the strongest upward motions also are found in the vicinity of the wave trough, but they reached their peak valves at upper levels (500-300 hPa) where the temperature anomalies are positive. Thus the integrated value of $-(p^{-1} [W'T'])$ is different in the eastern Atlantic and in the Pacific ocean.

Following FGGE, easterly waves have been studied over the Eastern Pacific ocean (Tai and Ogura, 1987; Nitta and Takayabu, 1985). This study was the first one in that area that is least explored meteorologically, due to the lack of observations. Low-level waves with a period of 4-6 days are active in the eastern Pacific. Wavelength is of the order of 3000-3500 km. Speed of propagation is 5-7 ms^{-1}.

Maximum wave amplitude is about 4 ms^{-1} at 850hPa. A secondary maximum in the upper-level component, significant in the waves over the GATE area and the western Pacific, is not found. The vertical velocity field shows convection starting from the low level of category 2 and reaching the upper level during the passage of the wave trough.

There is cyclonic vorticity in the trough of the lower level and anticyclonic vorticity for the four to six upper level categories. The temperature anomalies show that waves possess a cold core in the lower level with warm anomalies above it.

The magnitude of horizontal shear in the Pacific ocean is weak. Distribution of absolute vorticity and potential vorticity indicate that both conditions for barotropic instability and internal jet instability is not satisfied in the eastern Pacific. Dynamical causes of wave development remain unclear.

6. Monsoons depressions

Two types of synoptic-scale disturbances exist over the northern Indian ocean region. The first kind is very common and develops several times over the Bay of Bengal during the summer season. This is really what is called a monsoon depression. The second kind of tropical disturbances over that area is the so-called onset vortex which develops once a year, but not every year, at the time of the onset of the summer monsoon.

Several studies have been devoted to the depressions forming over the Bay of Bengal especially when they reach land because they bring a lot of rain. About two of these systems form every month during the months June to October. Prior to the 1970s no real studies on the dynamics and structure have been possible due to sparcity of observations. The first detailed study of monsoon depression with an attempt to model this phenomenom was made by Krishnamurti et al. (1975, 1976) in a serie of two papers. One component of MONEX (Summer 1979) was devoted to the study of monsoon depression especially during its formation stage (Douglas, 1987).

Fig.18 gives a vertical cross-section of the meridional wind. The amplitude is roughly 8m sec^{-1}. Maximum intensity is found near 600hPa. Monsoon depression is, in fact, a vortex whose cyclonic circulation extends from the surface up to 300hPa.

During the westward passage of the monsoon depression, much of the region of the Arabian Sea encounters strong northerly flow of relatively warm air. There is a strong equatorward flux (down the gradient) of sensible heat ($v'T'<0$). This implies a conversion of zonal available potential energy to eddy available potential energy.

Fig.19 gives a vertical meridional cross-section of the zonal wind. The level of strong easterlies north of the depression lies near the 800hPa surface and the intensity is close to 15 m sec^{-1}. The strength of the westerlies south of the storm is between 15 and 20m sec^{-1}. The level of the strongest westerlies descends from 800hPa to the surface during the westward passage of the depression.

Temperature anomaly is defined as a departure from a latitudinal average at a pressure level. 22°N (latitude of vertical cross-section) passes through the surface center of the monsoon depression (fig.20). Salient feature is the cold core structure. The thermal amplitude of the monsoon depression is largest near 800hPa (between 3 to 6°C). A pronounced warm core ahead of the depression arises primarily from the low-level warm advection of desert air west of the depression. The cold core is very pronounced below 600hPa and there is a reversal of thermal structure above that level. Near 400hPa the amplitude is only

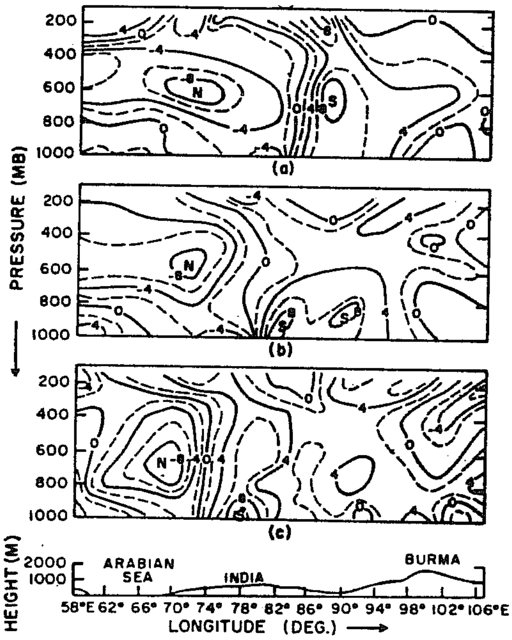


Fig. 18 Vertical cross-section of meridional wind (m/s) along 22°N for 4 (a), 5 (b), 6 (c) August, 1968.

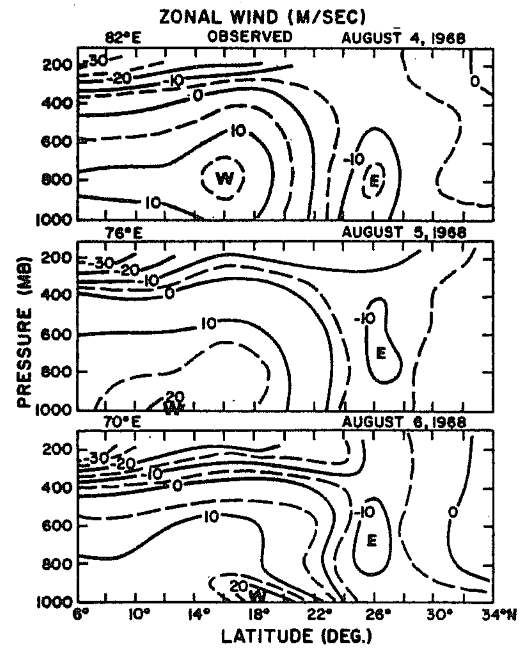


Fig. 19 Vertical cross-section of zonal wind (m/s) along 80°E, 76°E and 70°E for 4, 5, 6 August, 1968, respectively.

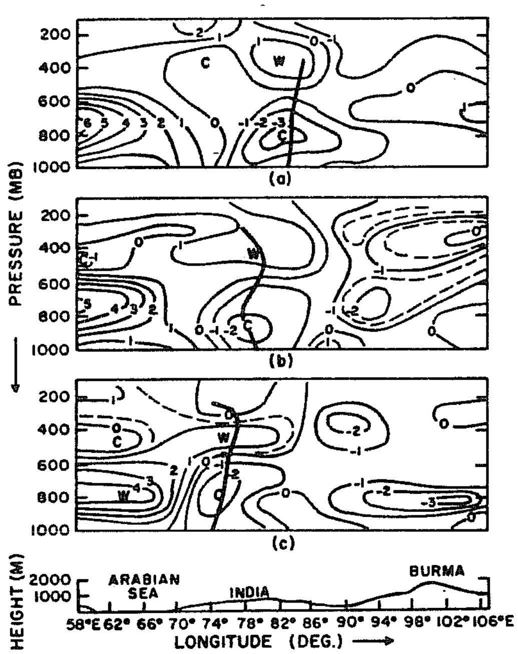
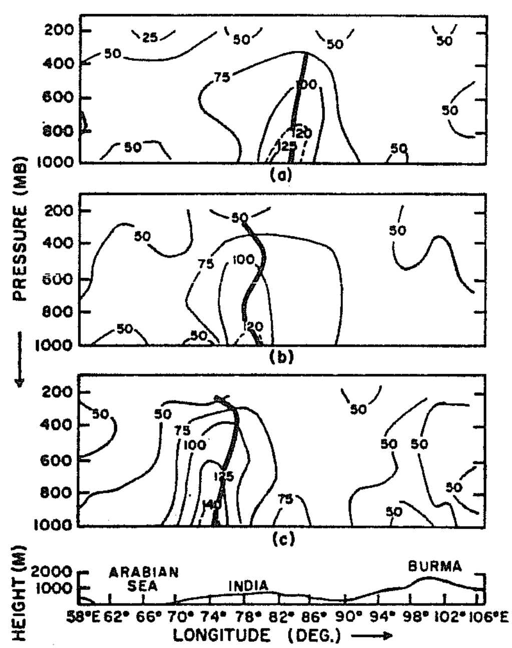


Fig. 20 Vertical cross-section of temperature anomaly (°C) along 22°N for 4 (a), 5 (b), 6 (c) August, 1968.



Vertical cross-section of absolute vorticity (10^{-6} s^{-1}) along 22°N for 4 (a), 5 (b), 6 (c) August 1968.

1 to 1.5°C. The correlation of vertical velocity and temperature (w'T') and of convective heating and temperature (H'T') are respectively important for the conversion of eddy available potential energy to eddy kinetic energy and the generation of eddy available potential energy from convection.

This thermal structure is furthermore different from that of a number of other tropical wave disturbances. The monsoon depression appears to have the most pronounced thermal amplitude among these.

The isopleths of the absolute vorticity show that the disturbance is very vigorous. Largest value is near the surface. It increases as the disturbance approaches the Arabian Sea.

Douglas (1987) studied the evolution of the different terms of the vorticity equation. The divergence term (generation of vorticity by convergence) is positive throughout a deep layer from the surface to 500 hPa. The horizontal advection term is negative in the lower/middle troposphere indicating that lower values of relative vorticity are being advected into the domain from the north. The two other terms (vertical advection of vorticity) and generation by horizontal gradients of vertical motion in the presence of vertical shear are insignificant below 700hPa.

In the middle atmosphere, the divergence and horizontal advection terms are small and the others cancel out. The net tendency is small.

In the upper troposphere, strong divergence is tending to rapidly diminish the vorticity. There is a strong vertical import from below because both the vertical gradient of vorticity and w (upward motion) are large at 300hPa.

Thus, in the rain area cyclonic vorticity is increasing below about 550hPa and is decreasing above this level. The transport of vorticity between the two levels is accomplished by cumulus clouds.

Several studies have been devoted to the understanding of the development of monsoon depressions. Krishnamurti et al. (1977), Saha et al. (1981) looked at predecessors. The former paper attributed the formation of depressions to a very slow westward propagating wave group initiated by a typhoon from the western Pacific. Sadler and Kilousky (1977) found no connections between monsoon depressions and previous storms over the south China sea. Saha et al. (1981) was in agreement with Krishnamurti et al.

Study of the MONEX depression indicates that a preexisting perturbation was present before significant intensification (presence of a middle tropospheric vorticity maximum). The intensification process itself occured with an increase in the thermal anomaly. The very close correspondence of the maximum temperature with the areas of most intense ascent and the growth with time of the thermal anomaly strongly suggests the important role of condensation heating in maintaining and intensifying the depressions. The mean environmental flow about the depressions strongly organizes the distribution of heating and presumably provides a positive contribution to the overall baroclinic energy conversion. Moorthi and Arakawa (1985) suggest that baroclinic instability in the presence of convective heating is a possible explanation for monsoon depression intensification.

Although mainly based on results from modelling, Krishnamurti et al. (1976) showed that the necessary condition for the existence of the combined barotropic-baroclinic instability is satisfied. Eddy kinetic energy increases by this instability.

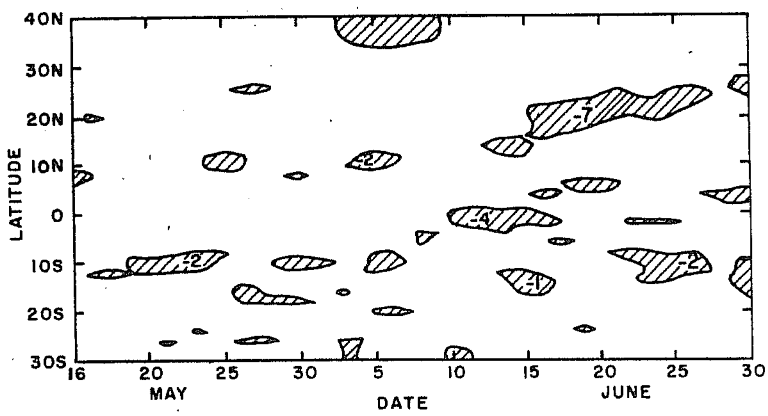


Fig. 21 **A plot of the zonally averaged meridional gradient of absolute vorticity (10^{-2} m^{-1} s^{-1}) at 850mb over the Arabian Sea, as a function of latitude and time.**

The second type of disturbances forming over the Indian region is the onset vortex. It does exist about 80 % of the years. Following MONEX during which a strong vortex formed and observations gathered, studies have been devoted to this phenomenon (Krishnamurti et al., 1981). It is a deep three-dimensional phenomenon, extanding from the surface up to 300hPa. Fig. 21 gives the zonally averaged meridional gradient of absolute vorticity. The necessary condition for barotropic instability is met : strong horizontal shear flow in the low levels provides energy for the maintenance of the vortex in the low levels. The barotropic instability in the lower levels may have been initiated (or triggered) by the gradual descent of the mid-tropospheric circulation which existed prior to the vortex. Douglas (1987) looked at the variation of the absolute vorticity at different pressure level. The result indicates that the onset vortex cannot be the result of barotropic processes since the absolute vorticity is not at all constant.

7. Conclusions

In this paper, the main features of the different synoptic disturbances developing over the tropics have been reviewed. Our knowledge on these phenomena have improved during the last two decades following the implementation of specific experiments in the tropics. However, our understanding on their origin remains poor. This remains one area of research to be explored but it is obvious that we don't have the right data to do this kind of investigation.

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