# MONSOON RAINFALL AND THE CIRCULATION IN THE AFRO-ASIAN REGIONS

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## ABSTRACT

The July and August rainfalls and the circulation patterns in the African and Asian monsoon regions have been investigated for the period 1891-1974. The interrelationship of monsoon rainfall within this region was examined by correlation between various subregions.

The study of the circulation patterns has shown that there are four major cells in the Afro-Asian region(West Africa 20°W-25°E, East Africa 25°E-65°E, India 65°E-90°E, East Asia 90°E-130°E) with different thermodynamic properties. The relationship between the East African indirect cell and the Indian direct cell was investigated through the study of the Tropical Easterly Jet. It was shown that the East African cell is driven by the TEJ.

The East Asian cell which has its ascending branch in the monsoon trough is examined by correlation with the Southern Oscillation. This work has shown that when the oscillation is strong, the trough fluctuation is in phase with the oscillation but not in phase otherwise. The spatial domain of the center of action appears to contract when the oscillation is weak.

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## 1. INTRODUCTION

The region extending from West Africa (20°W) eastward through India and east Asia (just north of the Philippines) into the adjacent west Pacific (as far as 130°E) experiences a seasonal reversal of wind direction called the 'Monsoon'. In most of this region, the heaviest rainfall is observed during the summer (high sun) period of the respective hemisphere.

In any study of the Monsoon, it is necessary to define the region which one can call the monsoon region. Using January and July surface circulation patterns, Ramage (1971) defined monsoon areas as follows:

" 1. the prevailing wind direction shifts by at least 120° between January and July,

2. the average frequency of prevailing wind directions in January and July exceeds 40%

3. the mean resultant winds in at least one of the months exceeds 3m $sec^{-1}$, and

4. fewer than one cyclone-anticyclone alternation occurs every two years in either month in a 5° latitude-longitude rectangle."

By this definition the monsoon region is limited to the African and Asian region, from 30°N to 25°S and from 20°W to 130°E.

In the present study, the monsoon rainfall in the month of July and August is investigated. Hence, only the region to the north of the equator is considered for the rainfall analysis. The emphasis will be on contemporary relationship rather than on any precursory relationship.

## 2. CLIMATOLOGY OF THE MONSOON REGION

In this section, the general climatology of the monsoon region for the months of July and August is discussed. Figure 1 shows the climatological rainfall for the month of August: regions of heavy rainfall are observed in west Africa, Ethiopia, the west coast of India, central and eastern India, Burma, and the Philippines.

### 2.1 Temperature and Sea-level Pressure

Figures 2, 3, and 4 show the August temperature, July and August sea-level pressures respectively based upon the 10 year period from 1951-1960. Temperature is nearly homogeneous north of the equator with an exception of the hot desert stretching from Sahara to Pakistan. The pressure is generally high in the subtropical Southern hemisphere and low in the Sahara, Arabia, Pakistan (these are heat lows), and the Tong King Gulf. The notable difference between July and August is the development of the East Asian Monsoon trough stretching from Indochina to the Pacific Ocean (northeast of the Philippines). This can clearly be seen in Figure 5, where the difference between the July and August sea-level pressures based upon the 10 year period from 1951-60 is shown. One can also note the pressure drops in Australia, and in the west African rainbelt (near 11°N) and general pressure rise in India.

### 2.2 Rainfall change from July to August

The change of rainfall between July and August is shown in Figure 6. The heavy line delineates the 50 mm isohyet for August with the exception of the west coast of India and Eastern India (Assam), monsoon rainfall generally increases from July to August. The decrease of rainfall in

August in the west coast of India can be explained by the more vigorous evaporation over the Arabian Sea during the month of July. In this region, sea surface temperature reaches its peak in the month of May, (29-30°C). When the southwest monsoon begins in June, increasing wind, which enhances evaporation, and increasing cloudiness both contribute to cooling of the ocean. In addition a pronounced upwelling is observed near Somali Coast. These cooling effects continue until the monsoon terminates in late September.

These effects combine to produce the maximum rate of evaporation in the month of June and July and decreasing rate in August and September (Privett, 1959). The strength of the southwest monsoon, which transports the moisture, is higher in July (see Table 13, 14) than in August. Thus the decreased transport of the moisture toward the west coast of India is the probable cause of the decrease of the monsoon rainfall in west India. The increase of the rainfall in the east coast of India is probably due to more active monsoon depressions during the month of August. Looking further east, a pronounced increase (160 to 200%) occurs in the region of East Asian Monsoon trough which is consistent with the drop of the sea-level pressure from the month of July. West Africa and the regions along 5°S in the Indian Ocean also show substantial increases in rainfall. Thus, one can conclude that the total monsoon rainfall reaches its peak in the month of August.

## 2.3 Transient synoptic systems

In addition to the mean circulation, there are many transient synoptic systems in the monsoon region. Ramage (1971) presented a detailed discussion of this topic in his book on "Monsoon Meteorology". Some of the observations that are important from the previous studies are:

(i) During a study of the recent drought in the Sahelian area of Africa (Tanaka, Weare, Navato, and Newell, 1975), it was observed that the monsoon rainfall in east Africa (east of 25°E) behaves differently from that in west Africa.

(ii) Numerous studies of tropical lower tropospheric disturbances have shown the existence of westward propagating waves with periods of 4 - 6 days (west Africa: Carlson, 1969; Burpee, 1972, 1974, the Atlantic Ocean, the Caribbean: Fett et al., 1974, and the Pacific Ocean: Reed and Recker, 1971).

(iii) In eastern and central India, tropical depressions (Monsoon lows) moving northwest from the Bay of Bengal are observed (Desai, 1951; Raghavan, 1967; and Ramage, 1968).

(iv) Typhoons and tropical storms play an important role in the East Asian Monsoon trough.

These studies have shown that, with the exception of the region 30°E to 65°E, there is evidence for the existence of a mobile lower tropospheric disturbance*. Hence a difference in the rainfall regime between west and east Africa is coincident with a difference of circulation regime in these regions.

*(Aspliden, 1974) has suggested that the lower tropospheric waves reach Africa from India. Burpee's 1972 study of the waves in Africa, however, has found that the African waves originate in the area just west of Khartoum (32°E) on the eastern end of the lower tropospheric easterly jet which is observed near 600 mb.

## 3. RAINFALL VARIABILITY

In order to study the rainfall regime within the monsoon region eight subregions were chosen for the rainfall analysis. These are:

1. WEST AFRICA - (five stations near 11°- 13°N and 14° W - 3°E)
2. EAST AFRICA - (Addis Ababa)
3. WEST INDIA - (four stations on the west coast of peninsula India)
4. BURMA - (three stations on the west coast of Burma)
5. SOUTH CHINA SEA - (Hong Kong and three stations on the Philippines)
6. CENTRAL INDIA - (five stations in central India affected by monsoon depression)
7. EAST INDIA - (four low altitude stations near Assam)
8. TONG KING GULF - (three stations near Tong King Gulf)

These subregions were chosen because each of the regions experiences heavy rainfall in July and August and appears to be governed by similar synoptic control.

Table 1 shows the rainfall statistics for the eight subregions. With exceptions of West India, East India, and Burma, August percipitation is heavier than (or equal to) that of July. The standard deviation indicates that the monsoon rainfall in July and August are most reliable in Burma ($\sigma \sim$ 15 to 20% of mean) and highly variable in the South China Sea ($\sigma \sim$ 30 to 50%).

Tables 2 through 7 show the ten year mean rainfalls for each of the

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*Addis Ababa is chosen because the Ethiopian highland has a distinct maximum of rainfall in the east African region. The comparison of the rainfall of Addis Ababa with those of other stations in Ethiopia with shorter periods of record shows that these rainfalls have a high positive correlation.

available decades. The variability of the long-term precipitation is small in East Africa; apparent secular increase was observed in West India (approximately a 20% increase in 40 years). In the regions of Burma, South China Sea, and East India, the rainfall reached its peak in the decade of 1921-30; there was a decrease to a minimum in the 1940's or 1950's with the recent decades showing some increase.

It is important to note that in none of the above regions is the rainfall declining in the most recent decades. Thus, the numerous droughts that were reported in recent decades in the Afro-Indian regions appeared to affect mostly the region near the fringe of the monsoon rain.

Table 8 shows the ten year autocorrelation (calculated every five years) between the month of July and August. The autocorrelation between July and August rainfall is generally low and highly erratic from decade to decade. In some cases, a positive correlation in one decade is followed by a negative correlation of almost equal strength in the next decade (e.g. Burma 1921-30: $r = +0.58$ followed by 1931-40: $r = -0.47$). Hence one can conclude that, in general, the July and August monsoon rainfalls are not related to each other.

## 4. RAINFALL CORRELATION



The interrelationship of the rainfalls within the monsoon region was investigated by the correlations among the eight subregions defined in the previous section. Table 9 shows the short period rainfall correlation (time period 12 to 36 years). Significant positive correlation exists between Burma and the South China Sea in both July and August. West India shows the positive correlation with West and East Africa in the month of August. These facts show that there is short-term climatic regime which favors synchronization of the rainfall between these distant lands.

The stability of this climatic regime was then investigated by ten year rainfall correlations calculated every five years. Table 10, 11, and 12 summarize the results. Inspection of the table shows that, in general, correlation between the subregions is much stronger and more stable in the month of August. The most stable relationship is between Burma and South China Sea. During the month of August every decade has positive correlation of at least +0.5. In the subsequent section, it is shown that the rainfall in this region is controlled by the East Asian Monsoon trough. The August correlation between West India and West Africa also shows a stable relationship. In all other regions correlations are not stable and some time even reverses their sign. However, many of these regions do show a tendency to be correlated (positive or negative) with West Indian rainfall.

The reasons for the higher spatial and time (stability) correlations of rainfall in the month of August compared to that of July are not clear at present, due to lack of intra-monthly study of the climatology of circulation. The study of the circulation in the northeast Asia region (near Japan) during Bai-u (the season of heavy rainfall in southern Japan during late June and the first half of July) period by Yoshino (1965) shows some insight into this question. Considerable change in the circulation patterns was observed within the month of July. At the beginning of the month, the circulation is a typical Bai-u pattern with the subtropical high located to the southeast of Japan and Bai-u jet(at 500 mb) flows along 35°N over Japan. Some time during the middle of July, Bai-u period is terminated rather abruptly by the sudden movement of the subtropical high toward Japan and the southeast monsoon is established in the Ryuku region. The circulation pattern is now essentially the same as that of August. During the month of August, except for an occasional typhoon, the circulation is very stable in northeast Asia due to stationary subtropical high over southern Japan.

Thus, at least in northeast Asia, the month of July has a much greater intra-monthly variability of the circulation patterns which tends to obscure the spatial and time correlations.

In other regions there appears to be no similar study of the intra-monthly climatology during July and August. In view of the circulation change in northeast Asia during the month of July, it is quite reasonable to suggest that the East Asian Monsoon trough develops during the latter half of July when Bai-u terminates in Japan and the August-type circulation is established in northeast Asia. Thus, the trough becomes conspicuous when the zone of convergence forms between the southwest monsoon in the South China Sea and the southeast monsoon in the western Pacific. As shown in Figures 5 and 6, this is the region of the maximum change in the circulation between July and August. Thus, much greater intra-monthly circulation change during the month of July is the probable reason for the lower spatial and time correlations of the rainfalls in July than in August.

## 5. CIRCULATION OF THE MONSOON REGION

### 5.1 Tropical Easterly Jet

The first comprehensive study of the Tropical Easterly Jet (hereafter referred to as TEJ) over Africa and Asia was done by Koteswaram (1958). He found that the core of the jet exists near 15°N and the current is accelerating east of 75°E and decelerating to the west of 75°E. Thus he suggested that there is a thermally direct cross-stream circulation at the entrance zone (east of 75°E) and indirect circulation at the exit (near 30°E).

Flohn (1964) has made an extensive study of the nature of the jet. This jet is located in the layer 200 - 100 mb from 5°N to 20°N a core of maximum winds ($\approx$ 30 m/sec) at 150 mb. From a mean climatological map of the jet he suggested the existence of an ageostrophic cross circulation. In the entrance region (80° - 150°E) there is a thermally direct cell with uplift to the north of the jet and subsidence to the south. In the exit region (20°W - 70°E) there is an indirect cell with subsidence to the north of the jet and uplift to the south.

In a recent study by Newell, Kidson, Vincent, and Boer (1972), the authors find (Figure 7) that the TEJ weakens rapidly near Ethiopia and adjacent Sudan (between 20°E and 40°E). There is also evidence of small regeneration of the current in west Africa. This suggests that the circulations over east and west Africa are dominated by two distinct cells of circulation.

The cell in east Africa and the adjacent Arabian Sea (25°E - 65°E) is thermally indirect, which converts some of the zonal kinetic energy of the TEJ into available potential energy. West Africa (20°W - 25°E) has a thermally direct cell. With the addition of the direct monsoon cell over India and

and East Asia, there is indication that the summer circulation over the Afro-Asian region is composed of at least a three cell structure.

5.2 Synoptic features

In order to establish the existence and the possible interaction of the proposed three cell structure, the circulation of the Afro-Asian region during August 1970 was investigated. (During this period, all the three cells were stronger than normal as indicated by the rainfall record.)

Figure 8 shows the observed precipitation during August 1970. There is a belt of heavy monsoon rainfall near 12°N in west Africa, the Ethiopian highlands and western India. Virtually no precipitation was observed in the vast region from the Sahara (north of 20°N) to the western border of Pakistan. The monsoon rainfall was heavy to the east of India with the belt of heaviest precipitation along the East Asian Monsoon trough running from Hong Kong through the Philippines to Micronesia.

Figure 9 shows the 850 mb geopotential height and wind field. The flow pattern clearly illustrates the different circulation regimes in east Africa (25°E - 65°E) and west Africa (20°W - 25°E). In west Africa a shallow heat low in the Sahara is overlain by the subtropical anticyclone, centered near Algeria (30°N 0°E). The region near 12°N, where the heaviest monsoon rainfall is observed, has the surface southwest monsoon overlain by a north-east wind from the subtropical high.

On the other hand, east Africa and adjacent Arabia have a northwest flow to the north of Ethiopia and a southwest flow to the west. This results in a strong convergence near Ethiopia (GATE Report #1). The Indian region is predominantly under the westerly monsoon flow. The monsoon trough in East Asia is clearly visible as a convergence zone between the southwest monsoon from the Indian Ocean and southeast monsoon from the Pacific Ocean.

Figure 10 shows the 200 mb geopotential height and winds This level was chosen, because at 150 mb there were several important stations where data was not available. The most conspicuous feature of the map is an immense Tibetan high extending from the Persian Gulf (50°E) to southern Japan (140°E). There is a separate warm anticyclone over the Sahara.

Easterly air parcels to the south of the Tibetan high have a northerly componenet and are accelerating to the east of 65°E. On the other hand, in the region from 65°E to 25°E the wind has a southerly component over Arabia, and east Africa(north of 15°N). Strong divergence is observed over Ethiopia. Easterly air parcels are decelerating over most of this region. In the eastern half (0°-25°E) of west Africa, the easterly flow has a slight northerly component and is accelerating. In the western half (20°W-0°) the air parcels are being decelerated.

Figure 11 shows the temperature at 500 mb where the effect of vertical motion and latent heat liberation is expected to be important. In west Africa, there is relatively cooler air over the Sahara and warmer air over the region of heavy monsoon rainfall (near 12°N). The thermal pattern dramatically reverses itself over east Africa. There is very warm air over the desert of Egypt and the Persian Gulf and relatively cold air over the mountains of Ethiopia. In the Indian region, there is very warm air over northern India and colder air over the Indian Ocean.

Finally, Figure 12 shows the dewpoint depression at 500 mb. Combined with the temperature distribution at 500 bm and observed rainfall, the general characteristics of the circulation are as follows:

(i) In West Africa(20°W-25°E), there is a thermally direct cell with cold dry air sinking over the Sahara (near 25°N) and warm moist air rising near 12°N where heavy monsoon rainfall is observed.

(ii) In East Africa and nearby Arabia (25°E-65°E), there is a strong thermally indirect cell with warm air sinking over the region from northern Egypt to the Persian Gulf and colder moist air rising in Sudan and Ethiopia.

(iii) In the Indian region, there is a well-known thermally direct monsoon cell with rising warm moist air over India.

In addition to the above data, the following observations are important in the region along the East Asian Monsoon trough.

Burma and South China Sea rainfall show a strong and stable positive correlation. (see Table 12)

The heat low which is observed in other regions is absent due to a lack of intense boundary layer heating associated with the desert. (see Figures 16 and 18)

In recent years (especially after mid 1930's) the rainfall near the trough appears to be independent of that over India. (see Tables 10, 11, and 12)

These facts suggest that the thermally direct circulation along the trough can be distinguished from that of India.

5.3 Meridional Cross-section

The study of the circulation and rainfall has shown the existence of four separate cells in the Afro-Asian monsoon region. Though many regional studies of the circulation have been carried out (Flohn,1969, and Dean and La Seur,1974) there appears to be no detailed study of the comparison of the circulation in Africa (2 cell) and Asia during the same period.

In this section, a detailed study of each of the four cells during August 1970 is presented. Figure 13 illustrates the Indian cell. Figures 13a, 13b, and 13c show the meridional cross-sections of the mean zonal wind $\overline{u}$, the mean meridional wind $\overline{v}$, and the departure of the temperature from the meridional average T. This approach was taken to show the distribution of relatively warm and cool air. A similar format is also used in Figures 14, 15, and 16*, which show the meridional cross-sections over East Africa, West Africa and East Asia respectively for the same parameters.

An inspection of these Figures shows that the temperature field and zonal wind field support each other through the thermal wind relation in the higher troposphere. Thus, the mean zonal flow in each of these regions is quasi-geostrophic (except possibly very near the equator).

The circulation and character of each of these cells are summarized in the schematic diagrams shown in Figure 17 (a, b and c) and 18 which show the schematic circulation over India, East Africa, West Africa, and East Asia, respectively**. Regions of relatively moist air (dew point depression less than 8°C) are shaded. Thick heavy lines delineate major features associated with the mean zonal wind. Dashed lines approximate the shape of relatively warm and cold air regions. The arrows represent a simplified meridional and vertical circulation.

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* In Figure 16 longitude is not shown because this is a cross section which is oriented northeastward from Singapore (103°E) toward southern Japan (133°E)

** The tropopause is shown in Figure 18 because of the better sounding network in the East Asian region which enables one to draw a clear cross section up to 50 mb.

Comparing the four major cells in the Afro-Asian monsoon region, there are interesting similarities and differences. In three western regions there is a thermally direct cell in the lowest layer maintained by boundary layer heating in the subtropical desert and by cool moist air to the south. The ascending branch of this shallow cell coincides with the position of the heat low and with the surface position of the I.T.C.Z.

In the mid-troposphere, the character of the cell is different in each of the regions. In the Indian region, the cell is thermally direct and probably increasing the energy of TEJ. The cell in the East Africa region is indirect and is extracting energy from the TEJ. (See Figure 7) Finally, the cell in West Africa is direct below 300 mb and indirect above the 300 mb. These cells in the mid troposphere govern the rainfall in their respective regions.

In the East Asian region, the shallow heat low due to intense boundary layer heating is absent. Thus, a simple thermally direct cell (which coincides with I.T.C.Z.) is observed. It is also important to note that the vertical shear of the wind near the trough axis is small. As suggested by Gray (1968), this condition favors the generation of the tropical cyclones. It is well known by observation that this region is the most productive area for the tropical cyclogenesis. It is also interesting to note that in the two regions where direct mid-tropospheric cells exist, there are observed African Easterly waves and the subtropical cyclone. On the other hand, there is no evidence of a mid-tropospheric disturbance in the indirect cell over East Africa.

## 5.4 Comparison of August 1970 with August 1972

The large scale synoptic features appears to vary considerably from year to year. Some study of this variability was carried out by comparing the previous results with those from the month of August 1972.

Figure 19 shows the 200 mb geopotential height and wind. Compared to August 1970, both the Tibetan and Sahara highs are weaker and displaced to the south. The divergence over Ethiopia is much weaker. The strength of the Tropical Easterly Jet over India is much weaker than that of 1970.

Figure 20 shows the 500 mb temperature. The thermal pattern is much weaker than that of August 1970, which illustrates that all three mid-tropospheric cells were inactive. The intense drought observed in the extensive area of West Africa, Ethiopia and India is associated with the inactivity of these cells.

# 6. THE ROLE OF THE TROPICAL EASTERLY JET

## 6.1 Zonal Kinetic Energy Content

The study of the circulation has suggested that the mid-tropospheric thermally direct West Indian cell and thermally indirect cell in East Africa are related to each other by the Tropical Easterly Jet observed in the 200 - 100 mb layer. It appears that the indirect cell in the East Africa region acts as a sink of the zonal kinetic energy created by the Indian monsoon system. Thus, the kinetic energy content of the atmosphere (especially that of TEJ) at the west coast of India can be regarded as an energy reservoir available to drive the indirect cell located in East Africa. This observation is supported by a study of the angular momentum balance of the Indian summer monsoon (Keshavarmurty, 1968) which has shown that the Indian monsoon region is exporting easterly momentum toward East Africa. These facts suggest that the relationship between the two cells can be investigated by the study of the mean kinetic energy budget. The study of the energy cycle has been carried out for the entire northern hemisphere by numerous authors. The most recent work was done by Peixoto and Oort (1974), and Oort and Peixoto (1974). From the existing data, Kidson (1968) examined the tropical energy cycle and estimates that about $47.2 \times 10^{20}$ erg $sec^{-1}$ of ZAPE* is converted to ZKE** in summer (June - August). More detailed study on the subject has not been carried out because of sparse data.

Since a complete study of the energy cycle could not be carried out due to a lack of an adequate network of stations, only the mean zonal kinetic

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* ZAPE = (mean) Zonal Available Potential Energy

** ZKE = Zonal Kinetic Energy

Both terms are defined by Lorenz (1955).

energy content over India (75°E) and East Africa (32°E) was calculated for July and August of 1965-1974. In each of the regions, meridional cross sections (30°N - 6°N) of the (900 mb to 75 mb) mean zonal wind were used to compute the zonal kinetic energy.

Tables 13, 14, 15 and 16 show the results. Not only is the year to year variability of TEJ kinetic energy large(standard deviation is about 30% of normal kinetic energy), but the July to August persistence is low. The column 5 and 6 of Table 15 and 16 show the ratio of ZKE over East Africa to that of West India. This shows that on the average about 55% of ZKE is lost between West India and East Africa.

This loss of the ZKE can be interpreted as a part of the mean kinetic energy budget over the region. The equation for the mean kinetic energy budget can be written as follows:

$$\frac{\partial K_M}{\partial t} = C(P_M , K_M) + C(K_E , K_M) - D(K_M) + B(K_M)$$

where $K_M$ = mean kinetic energy

$C(P_M , K_M)$ = conversion between mean available potential energy and mean kinetic energy

$C(K_E , K_M)$ = conversion between mean kinetic energy and eddy kinetic enegy

$D(K_M)$ = dissipation of mean kinetic energy into heat

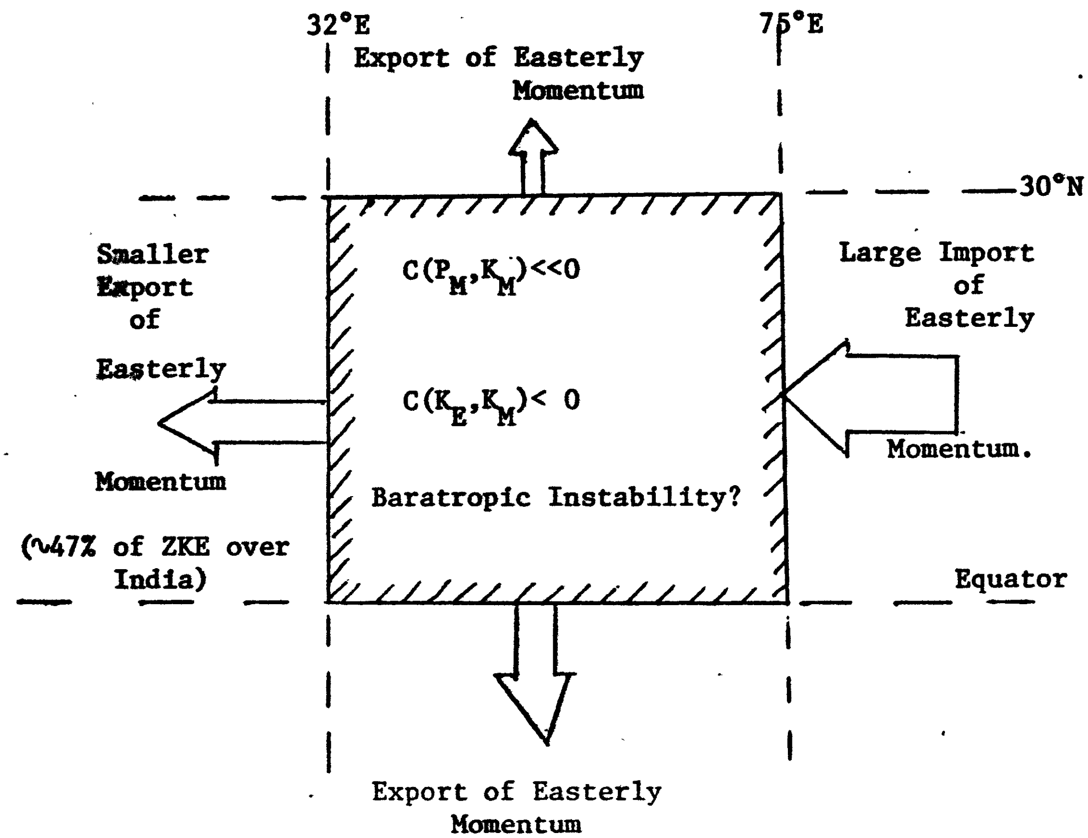
$B(K_M)$ = boundary flux of mean kinetic energy

Thus if one considers a region bounded by latitude walls at 30°N and the equator and by two meridional cross-sections in India (75°E) and East

Africa (32°E), the previous result can be interpreted as a sum of the four possible processes which change the kinetic energy content of the region.

Because an accurate calculation of each of the four processes cannot be carried out due to lack of data over the Indian Ocean, some qualitative idea of the magnitude of the processes will be discussed. The presence of a strong thermally indirect cell centered around 40°E implies that $C(P_M , K_M)$ should be large and negative (mean kinetic energy converted into mean available potential energy). Tantaway (1974) has observed some evidence of the existence of eddies in the TEJ over East Africa. This along with the evidence suggested by Keshavumurti (1968) that there is a convergence of eddy momentum transport near the axis of the TEJ suggests $C(K_E , K_M)$ should be negative (mean kinetic energy is converted into eddy kinetic energy). The determination of the magnitude of this term requires daily sounding data near the TEJ and therefore is deferred to further study. The fact that this term is probably negative implies that there is evidence of instability (probably barotropic) in the TEJ.

The boundary flux term $B(K_M)$, crudely estimated from the circulation pattern, should be negative for the northern and southern boundaries. The magnitude of the flux at these boundaries should be smaller than the flux across the meridional cross-sections (which is very large and positive). Finally, the frictional dissipation term $D(K_M)$ should be small and positive at the level of TEJ. The summary of the discussion is shown on the following page.




Size of arrow is proportional to subjective estimate of magnitude of flux.

6.2 Relationship between mid-tropospheric cell and rainfall.

The zonal kinetic energy content of TEJ calculated in the previous section is compared with the rainfall and the strength of the East African mid-tropospheric indirect cell (defined as 500 mb temperature difference between Helwan and Malakal). The results are shown in Table 17. In the month of August it is clearly evident that the kinetic energy content of the TEJ and the strength of the indirect cell in East Africa is closely related to the rainfall in West India. The rainfall in East Africa has a high correlation with TEJ in East Africa. On the other hand, the rainfall in West India is not related to the kinetic energy content of the TEJ at West India (see Table 13 and 14). The relation of West Indian rainfall with the sea-level pressure gradient (obtained from the pressure difference between Trivandrum and Veraval)however shows that in the month of July, the orographic component of the rainfall appears to be dominant and still plays an important role in August (see Table 17).

In this study, the kinetic energy content of the TEJ at West India is measured just upstream of the West Indian rain area (see Appendix A for a list of the upper air stations). At this point, the following assumptions are considered for the mean kinetic energy budget over the region defined in the previous section.

(i) The boundary flux of the mean kinetic energy at the northern boundary is small.

(ii) The export of the mean kinetic energy at the southern boundary is constant.

(iii) The instability of TEJ is either small or extracts a fixed proportion of the mean kinetic energy.

(iv) The indirect cell in East Africa is driven by the mean zonal kinetic energy.

If the above assumptions are true, the rainfall in East Africa should have a consistent high positive correlation with West Indian rainfall. Observations show (see Figure 10) that assumption (i) is true. The energetic of thermally indirect cell satisfy assumption (iv). This leaves (ii) and (iii) both of which required a dense network of data in the Arabian

Sea for evaluation. Re-examination of Table 10 and 11 shows that in the month of August there are several decades when the rainfall correlation is moderately high (e.g. 1964-1955 +0.80). This suggests that the conditions (ii) and (iii) are fulfilled from time to time which results in a tendency for these rainfalls to be positively correlated with each other.

## 7. EAST ASIAN MONSOON TROUGH

The study of the rainfall correlations has shown that the Burma and South China Sea precipitation have a stable positive correlation. The investigation of the circulation has shown the existence of thermally direct cells with no evidence of a heat low. Hence, the sea-level pressure fluctuations may provide an important key for monitoring the fluctuations of the East Asian Monsoon trough.

Table 18 shows the correlation of sea-level pressure at three stations near the axis of the trough to the South China Sea rainfall. All stations have strong negative correlations with higher correlation in the month of August. Thus, when the trough intensifies, rainfall increases in the South China Sea region. Figures 21 and 22 show the sea-level pressure departures from 1951-60 for August 1958 and 1960 respectively. In 1958, a large area of positive pressure anomalies extended from Ryukyu to northeast India. The rainfall in Burma and South China Sea was 55 and 61 percent of the long-term mean. August 1960 shows a contrasting situation. The monsoon trough was much stronger than normal (pressure was up to 5 millibars below normal) and the rainfall in Burma and South china Sea was 102 and 120 percent of the long term mean respectively. A similar situation in July 1972 caused massive floods in the Philippines(South China Sea precipitation was 272 percent of normal). Looking back at Table 1, we note that the South China Sea region has the highest variability of the monsoon rainfall in July. This fluctuation is due to the high variability of the East Asian Monsoon trough as shown in Table 19b;(compare the first two stations with the other stations) which shows the mean and the standard deviation of sea-level pressures near the axis of the trough.

At this point the relationship between the monthly sea-surface temperature and fluctuation of the East Asian Monsoon trough was investigated. The sea-level pressure at Ishigaki Jima was correlated with the sea-surface temperature in a region 10° latitude by 20° longitude centered at 25°N 130°E for the month of July and August 1951-73. The result was not significant (correlation coefficient was +0.21 for July and -0.01 for August).

7.1 The role of tropical storms

The East Asian Monsoon trough is a favorite breeding ground for the tropical cyclone. Atkinson (1971) gives climatological value (1959-68) of 30.5 tropical storms and typhoons per year in the West Pacific. In the months of June, July, and August these figures are 1.6, 5.0 and 6.8 storms for the respective months.

Since the number of storms is large (especially in the months of July and August), monsoon troughs and typhoons may be intimately related. The study of this relation was carried out by comparing the sea-level pressure at Ishigaki Jima to the frequency of typhoons (including tropical storms) in 5° latitude by 5° longitude boxes in the region defined by 20°N and 35°N latitude and 120° and 140° longitude(see Figure 23).

In order to take into account slow moving tropical storms, the following definition of typhoon frequency was used:

(i) In each of 13 boxes (5° latitude by 5° longitude), if tropical storm exists on a given day at 00Z GMT, it is counted as one typhoon frequency for that box.

ii) If two tropical storms exist at two different boxes on the same day, each box gets one typhoon frequency.

iii) The monthly typhoon frequency for each of the boxes is the sum of the daily typhoon frequencies.

iv) The total monthly typhoon frequency is the sum of the monthly typhoon frequencies of all 13 boxes.

Thus, slow moving storms will increase the typhoon frequency. Sixteen months of July and August of 1956-65, and 1967-72 were used to obtain the total monthly typhoon frequencies. The results are shown in table 20. There are considerable differences in the typhoon frequencies between the months when pressure is high in the trough ($\geq + 0.90$) and when the trough is stronger than normal ($P \leq -0.80$). The application of the student t-test (see appendix B) has shown that this relationship is statistically significant.

Figures 24 and 25 show the monthly typhoon frequencies for each of the 13 boxes. When the trough is strong, not only is the frequency high, but the tropical storm tends to move slowly near the Ryukyu islands. On the other hand, when the trough is weak, the frequency of the storm is low and they have a tendency to avoid the East China Sea by moving west toward Hong Kong or north toward Southern Japan.

### 7.2 The Relationship of the East Asian Monsoon trough to the Southern oscillation.

The southern oscillation was discovered by Walker and Bliss(1932, 1937) and subsequently became recognized as an important mode of the fluctuation of the tropical atmosphere. In general, it consists of the exchange of air between the eastern and western hemispheres.

Figure 26 illustrates the Southern Oscillation defined by Walker and Bliss for the 3 months of June-August. The numbers in the figures are the correlation coefficients of the meteorological parameter of each station to the Southern Oscillation Index which is defined later. There are several important features in the figures.

i) The center of action lies over the Pacific and Indian Oceans.

ii) There is a tail of positive pressure correlation stretching from the Pacific Ocean through the monsoon trough to north-east India.

iii) Rainfall near the Philippines has a negative correlation to the Southern Oscillation.

These features give an impression that the fluctuation of the East Asian Monsoon trough is in phase with the Southern Oscillation. This possible relationship was investigated by reviewing the work done by Walker and Bliss, Troup(1965), and Trenberth(1975) and subsequently studying the fluctuations of the Southern Oscillation for the period of 1891-1974.

Walker has obtained the following formula for defining the Southern Oscillation during the months of June-August:

S.O.I.=(Santiago pressure)+(Honolulu pressure)+(India rain)
+(Nile flood)+ .7(Manila pressure)-(Batavia(Djakarta)pressure)
-(Cairo pressure)-(Madras temperature)- .7(Darwin pressure)
- .7(Chile rain)

where the so-called primary center of action was given full weight and the secondary center was given a smaller weighting. Table 21 shows the correlation of ten parameters with the Southern Oscillation Index.

Troup modified Walker's definition and retained only the station pressure to obtain the Index. He also found that the oscillation has declined in the recent decades(1921-50).

Trenberth recently has studied the Southern Oscillation and the fluctuations of Sea-level pressure at Darwin, Apia, Tahiti and Easter Island. He found a phase difference of about one season between Tahiti and Easter Island(the latter leading). On the other hand, Tahiti and Darwin pressures were almost exactly out of phase. Thus using a study by Kidson(1975) as a reference, an index using only Tahiti and Darwin data was suggested by Trenberth:

$$SO_2 = \frac{P(\text{Tahiti}) - 1.2\, P(\text{Darwin})}{1 + |SON|}$$

where: SON = 1.26P(Tahiti) + P(Darwin) is the measure of noise present in the data.

P= seasonal pressure (3 months average) anomalies, in millibars.

In the current study, it is essential to have a simple homogenous index without the missing data from Walker's period to the present. Some of the parameters in Walker's index have significant gaps in the data(e.g. Djakarta 1965-74). Trenberth's index can only be used up to 1935 due to the lack of data from Tahiti.

Thus four stations near the center of action which have a continuous record from 1891 to 1974 were chosen and used to approximate the fluctuation of the Southern Oscillation. These are: Santiago ( +0.84 ), Honolulu ( +0.76 ), Colombo ( -0.72 ), and Darwin ( -0.68 ). The quantities in the parenthesis are Walker's correlations with his Southern Oscillation Index. For each of the stations, seasonal ( June-August ) sea-level pressureswere obtained. Then the climatological mean pressures were subtracted to obtain the departures from the normals. These values were then normalized and inserted in to the following equation.

S.O.I.T.=P(Santiago)+P(Honolulu)-P(Colombo)-P(Darwin)

Finally the values of S.O.I.T. were normalized. In all cases normalization was done to keep mean values equal to zero and the standard deviation is equal to unity.

Table 22 shows the ten year pressure correlations calculated every 5 years between each of the four stations chosen to represent the Southern Oscillation. The correlations obtained by Troup are given in the bottom as a reference.

In general, these correlations reflect the tendency observed in the Southern Oscillation. Closer inspection reveals considerable fluctuations from one decade to the next decade. Based on the numbers of high correlations ( arbitrarily defined as greater than 0.5 ), the oscillation was strong from 1891 to 1935 with the exception of the period 1906-1915. In the recent decades only 1941-50 and 1965-74 showed a moderately strong oscillation. The correlation between Santiago and Honolulu shows an abrupt decrease around 1935. It is interesting to note that rainfall correlations between West India and Burma for the month of July and August also declined during a similar period ( see Table 10, and 11). These facts suggest that the extent of the area of high correlation contracted around 1935 and no longer includes the North Pacific in the center of action.

The correlation of the pressures with the Southern Oscillation Index (defined in this paper) is shown in Table 23. Since the index contains the data from four stations, the correlation between these stations and the index is naturally expected to be higher than the station to station correlation. An inspection of the table indicates that 1931-40 and 1951-60 are the decades with least correlation (based on the numbers of high correlation between the four stations used to define the Index).

At this point, the discussion will return to the question of the relationship between the East Asian monsoon trough and the Southern Oscillation. Columns 5 and 6 in Table 23 show the correlation between Ishigaki Jima and Silchar (located in the northeast India) Pressure to the Southern Osciallation Index (defined in this paper). During Walker's period (before 1920), the correlation was positive (about +0.5 0.7) as shown by Walker (see Figure 26). In the subsequent decades, the correlations have declined substantially and in the recent decades (1956-70) the correlation is negative.

Thus the fluctuation of the East Asian Monsoon trough is in phase with the Southern Oscillation during the decades when the oscillation is strong as defined above (1901-20, 1941-50) and is not correlated when the oscillation is weak. This can be explained by the change in the shape of the Southern Oscillation. When the oscillation weakens, it can be suggested that the spatial domain of the center of action contracts.

Thus around 1920, the East Asian Monsoon trough was no longer part of the center of action in the south Pacific. By 1935, the north Pacific was not included. The domain expanded again during the decade 1941-50 followed by contraction during the 1950's and early 60's. In recent years, there were signs of the expansion of the domain. The prime example was 1972 when simultaneous low pressure anomalies in the South Pacific and the monsoon trough were associated with El Nino in Peru and the devastating flood in the Philippines.

The correlations of the seasonal pressures between Ishigaki Jima and four other stations are shown in Table 24. The highest correlation is observed with Silchar. This shows that the East Asian Monsoon trough is a large-scale feateure. Thus the rainfall in Burma and the South China Sea appears to be under the strong influence of this monsoon trough. An inspection of other correlations generally shows a positive correlation with Darwin until the last 20 years when it reverses its sigh which is consistant with the observed negative correlation between Ishigaki Jima and the S.O.I.T.

Finally, Table 25 shows the correlation of the Southern Oscillation Index, Ishigaki Jima, and the Honolulu pressure with the index by Trenberth. The high correlation between the two indices clearly shows that the index used in this study is a reasonable estimate of the fluctuation of the Southern Oscillation.

Figure 27 shows the fluctuation of the Southern Oscillation Index.

The oscillation was strong from 1891-1920, around 1940, 1948-57, and 1965-74. These are consistent with the result from the study of 10 year correlations.

# 8. SUMMARY



The study of the rainfall and circulation of the monsoon region has shown many interesting characteristics of the monsoon system during the month of July and August.

The monsoon rainfall reaches a peak during the month of August with the exception of west India where the transport of moisture from the Arabian sea declines in August. In none of the eight subregions is rainfall declining in the most recent decades. This suggests that numerous droughts which were reported in the recent decades were occuring near the fringe of the monsoon rain.

The study of the autocorrelation of July and August rainfall illustrated that in general, they are not related. The correlations of rainfall between eight subregions were much stronger and more stable in the month of August ( than in July ). The probable reason for this difference is due to the much greater intra-monthly circulation change during the month of July with the circulation resembling August condition by the end of the month.

The most stable rainfall correlation was observed between Burma and the South China Sea. The observed positive pressure correlation between Ishigaki Jima and Silchar shows that the East Asian Monsoon trough is a large scale feature, which governs rainfalls in both regions.

The examination of the upper air circulation has clearly shown the existence of four distinct cells in the Afro-Asian monsoon region. In the regions of West Africa ( 20°W-25°E ), East Africa ( 25°E-65°E ), and India ( 65°E-90°E ), there is a shallow thermally direct cell due to boundary layer heating in the subtropical desert. These cells have a descending branch in the cool ocean area to the south and an ascending branch in the heat low which coincides with the surface position of I.T.C.Z.

In the mid-troposphere ( above the 700 mb level ), there is a thermally direct cell in West Africa, an indirect cell in East Africa which is driven by the TEJ, and the well known thermally direct cell in India. The cell in East Asia ( 90°E-130°E ) is a Hadley type thermally direct cell with ascending branch over the East Asian Monsoon trough.

A detailed examination of the East Asian Monsoon trough shows that there is a direct relationship between the strength of the monsoon trough and the total frequency of typhoons. This is an indirect result of stronger surface convergence between the southwest and southeast monsoons when the trough is stronger than normal.

Finally, the relationship between the East Asian Monsoon trough and the Southern Oscillation shows that when the oscillation is strong, the fluctuation of the trough is in phase with the oscillation but there is no correlation or even negative correlation when the oscillation is weak. It is suggested that when the oscillation weakens, the spatial domain of the center of action contracts. Thus the shape of the oscillation is similar to that found by Walker when the oscillation is strong. When the oscillation is weak,the fluctuation is probably confined to a smaller region centered in the South Pacific and the Indian Oceans.

Figure 1.

AUGUST 1951-60 PRECIPITATION (MM)

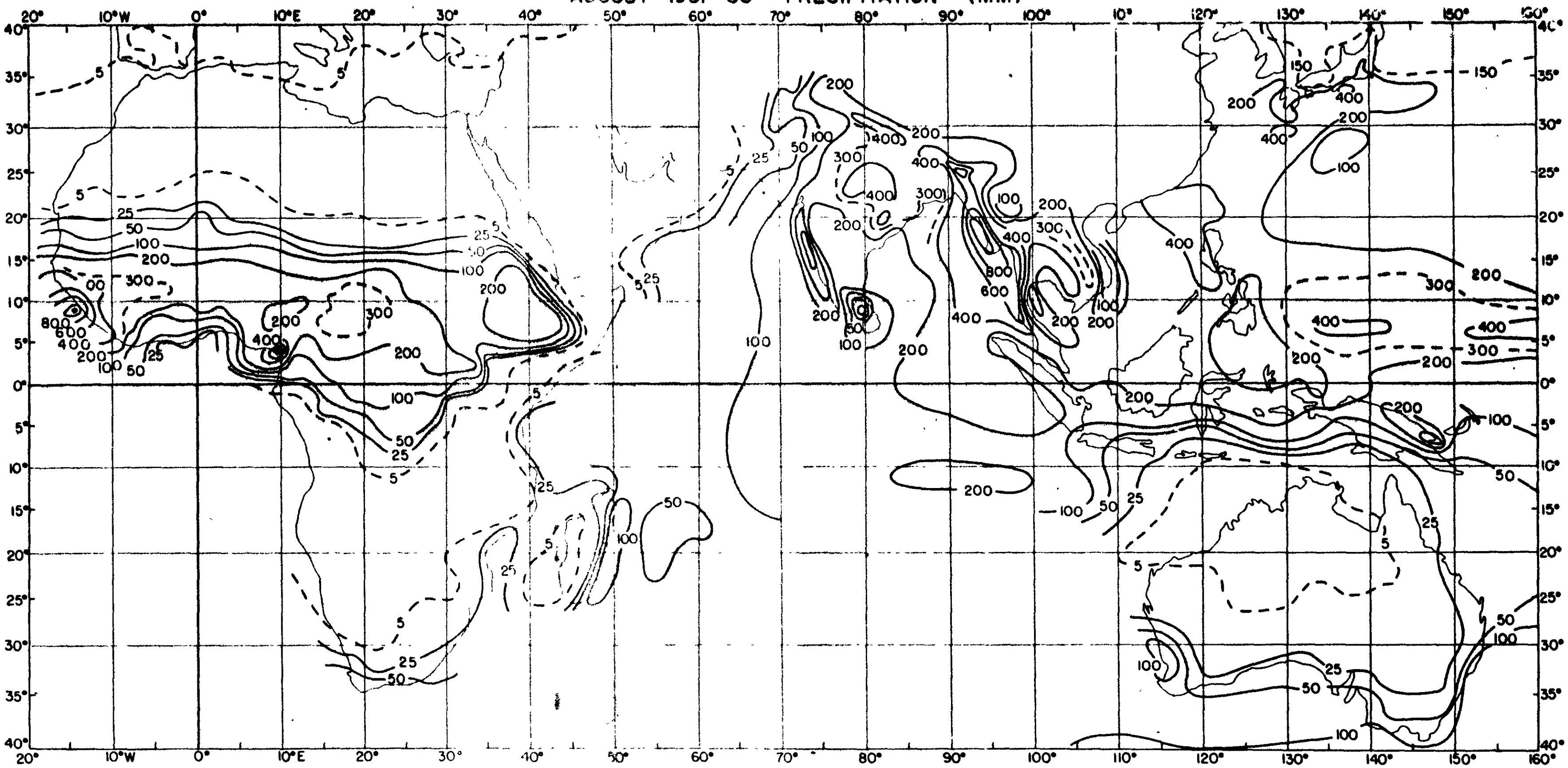


Figure 2.

AUGUST, MEAN TEMPERATURE °C

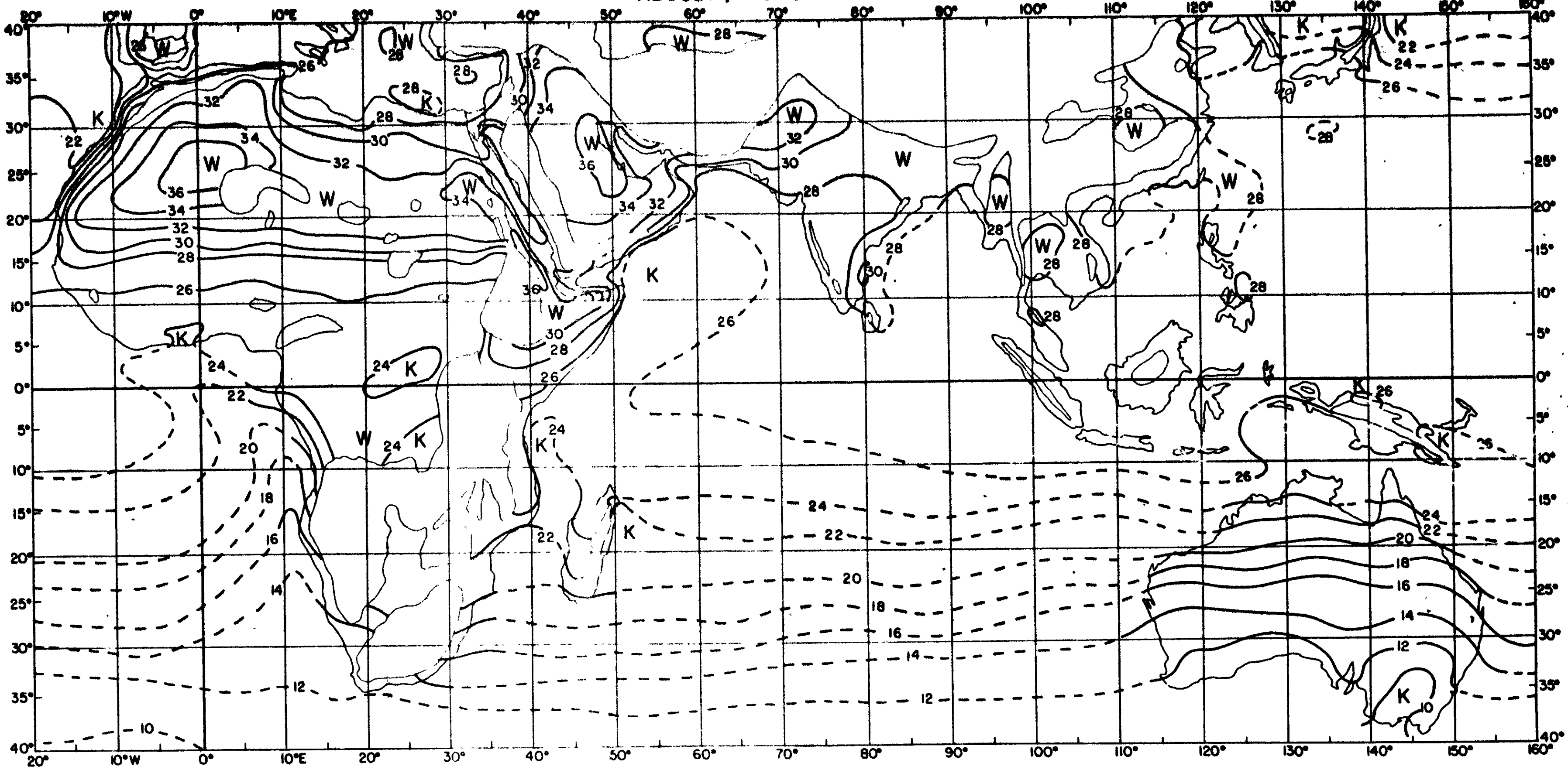


Figure 3.

JULY 1951-60 SEA LEVEL PRESSURE (mb)

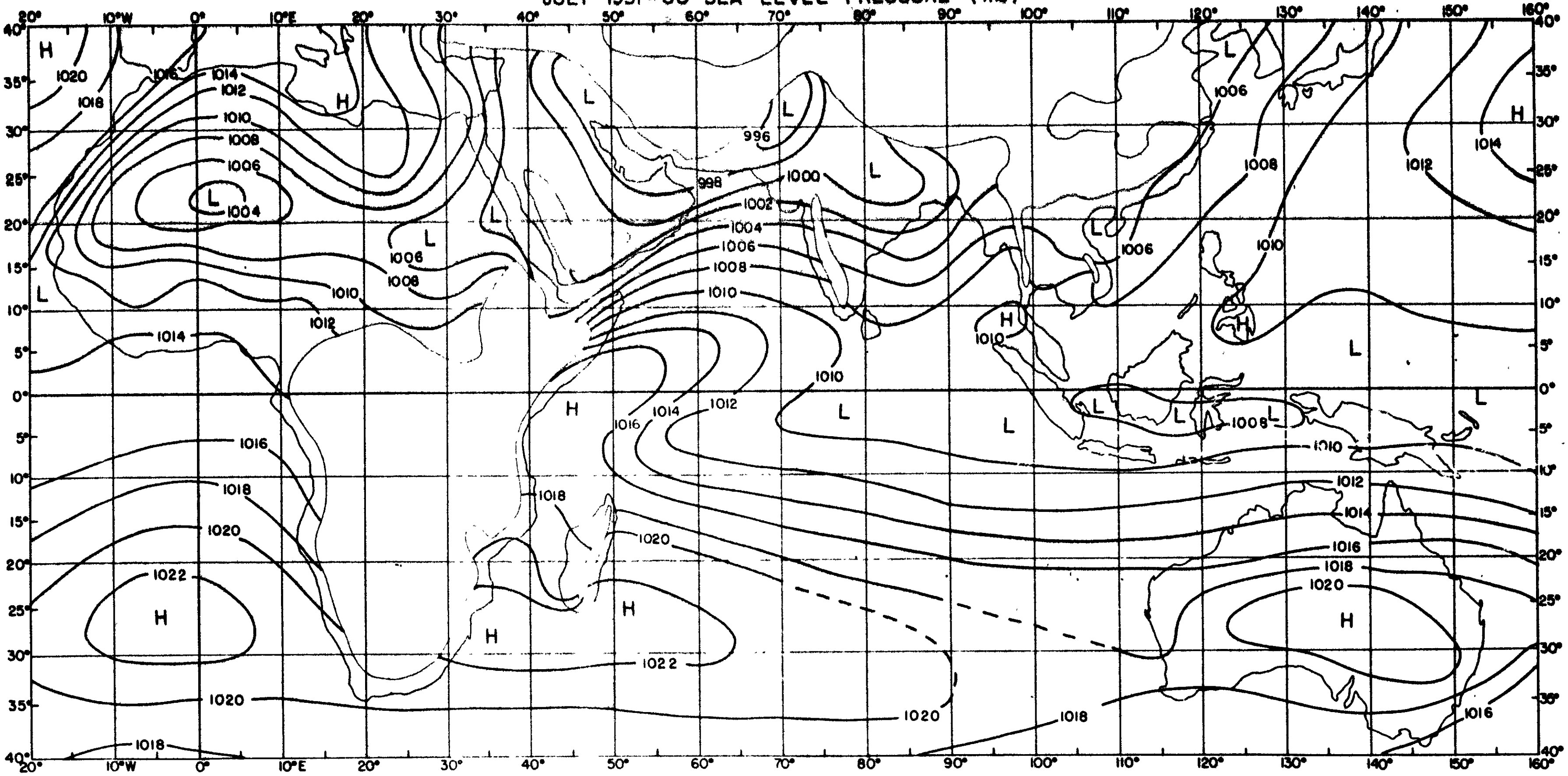


Figure 4.

AUGUST 1951-60 SEA LEVEL PRESSURE (mb)

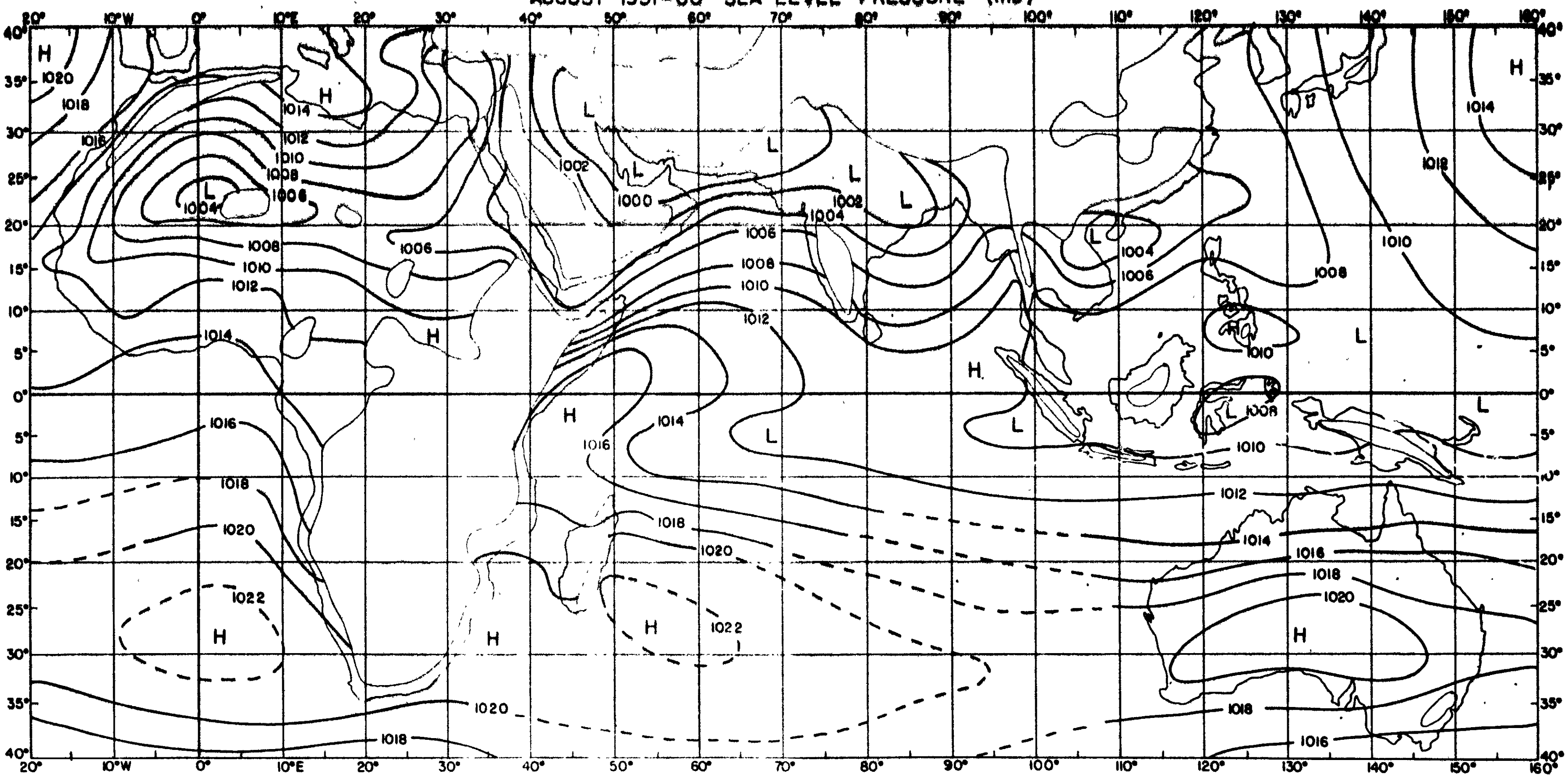


Figure 5.

AUGUST-JULY 1951-1960 (PRESSURE mb)

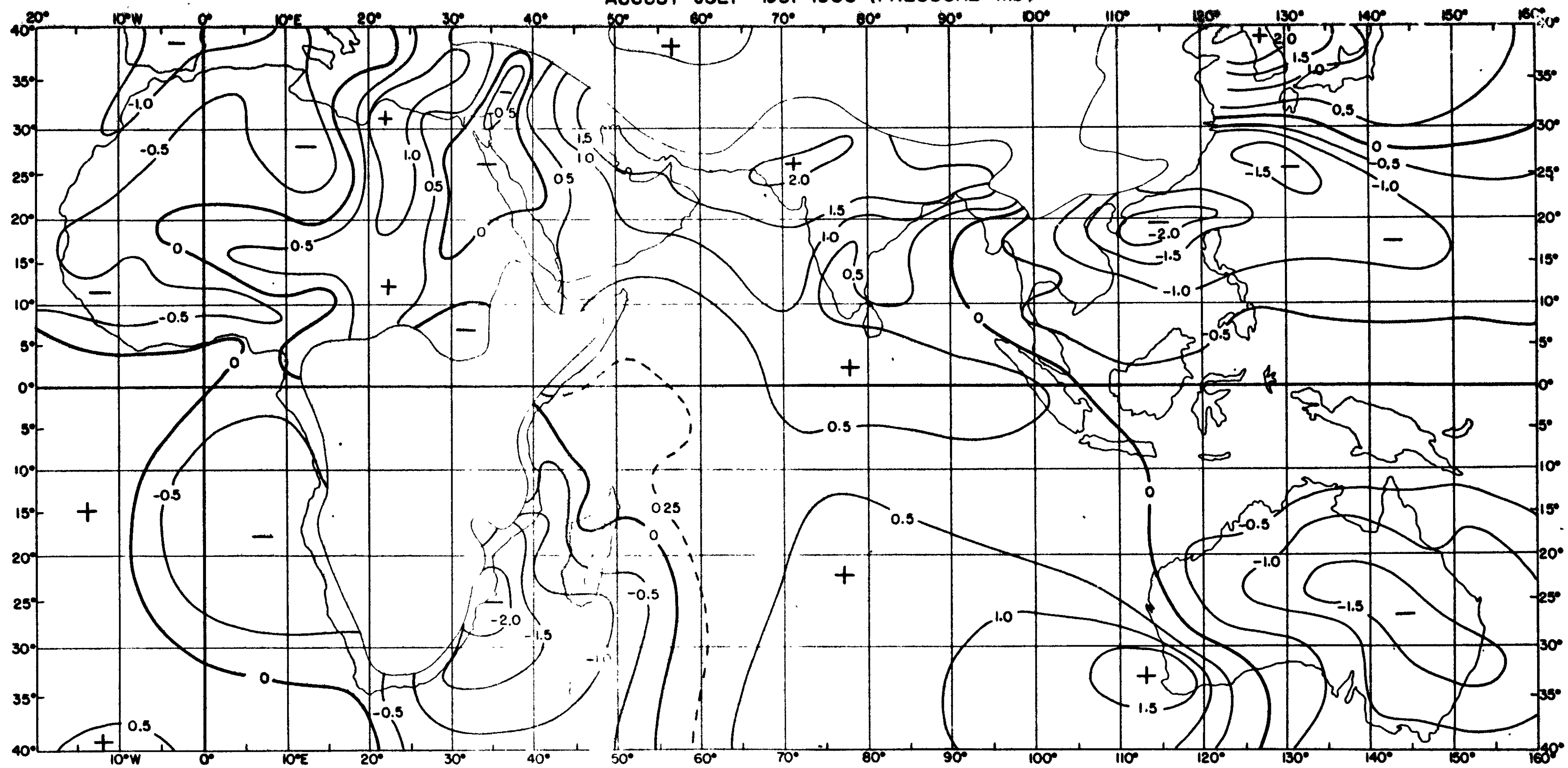


Figure 6.

AUGUST 1951-60 PRECIPITATION (per cent change from July 1951-1960)

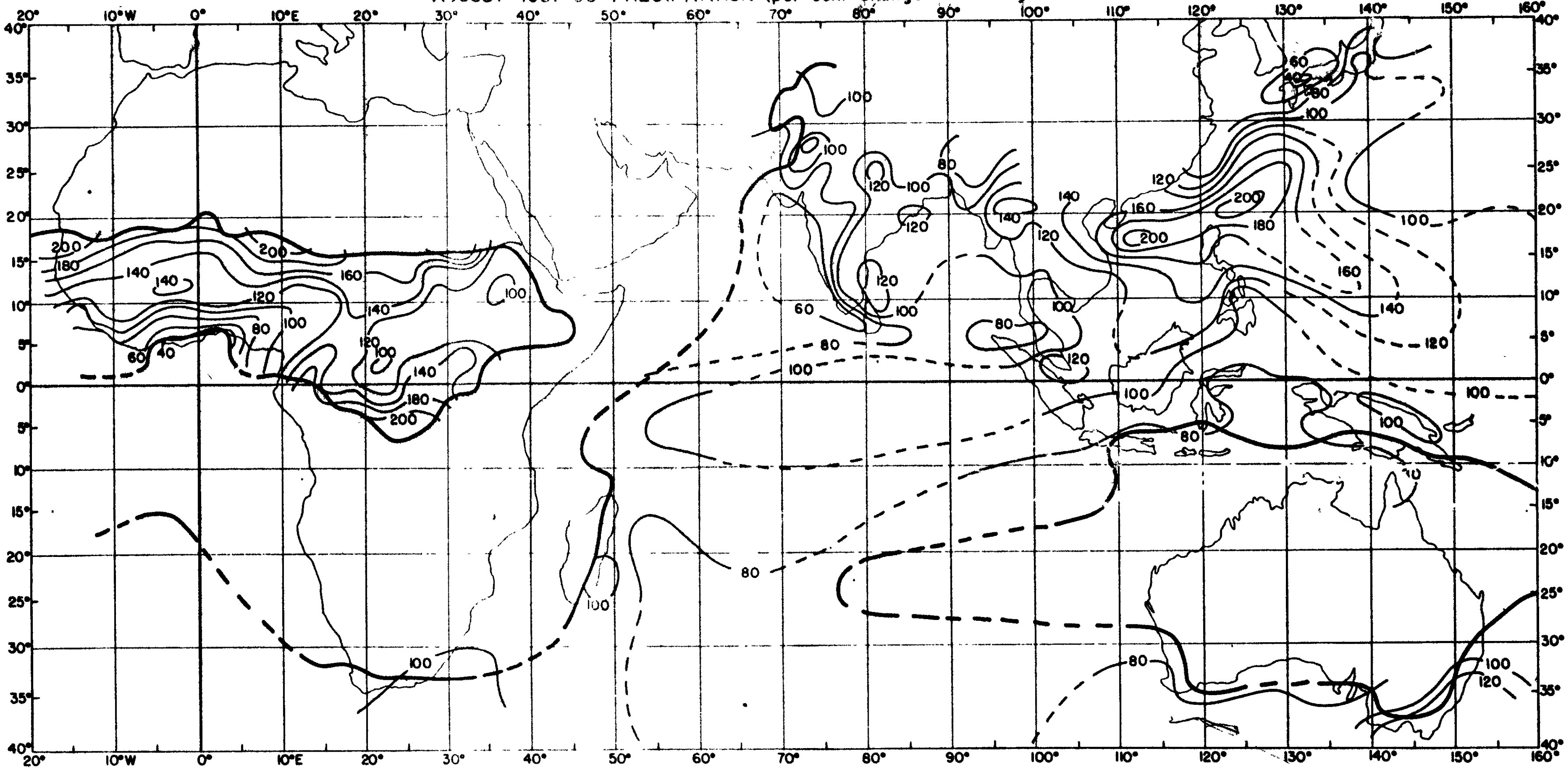
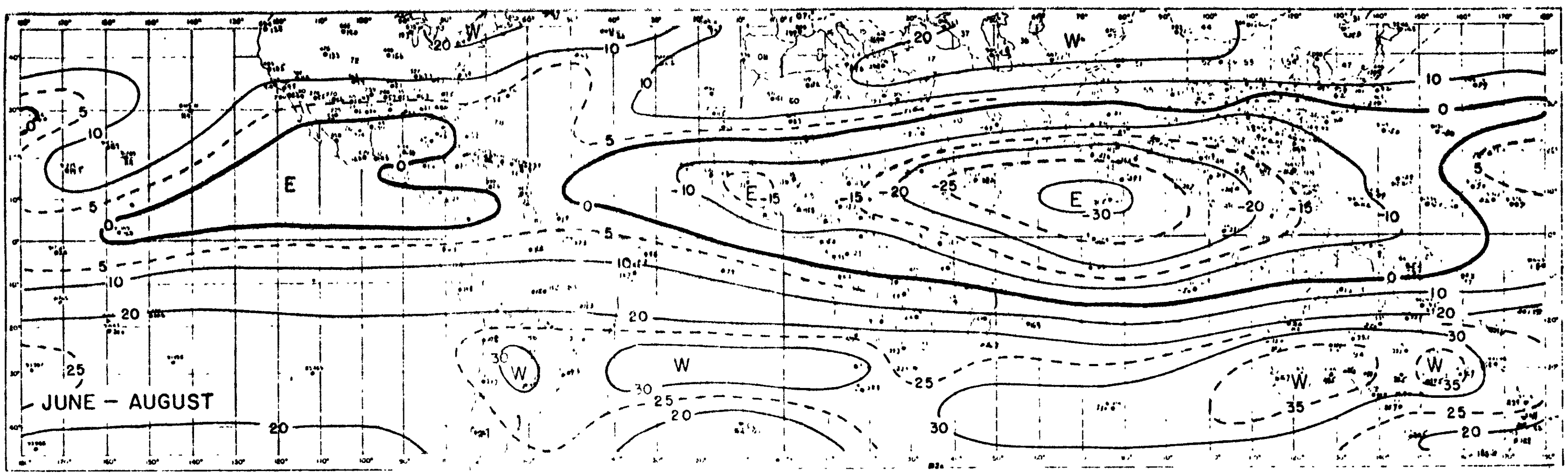


Figure 7.




150 mb $\bar{u}$ climatology. Adopted from Newell et al. (1972).

Figure 8.

AUGUST 1970 PRECIPITATION (MM)

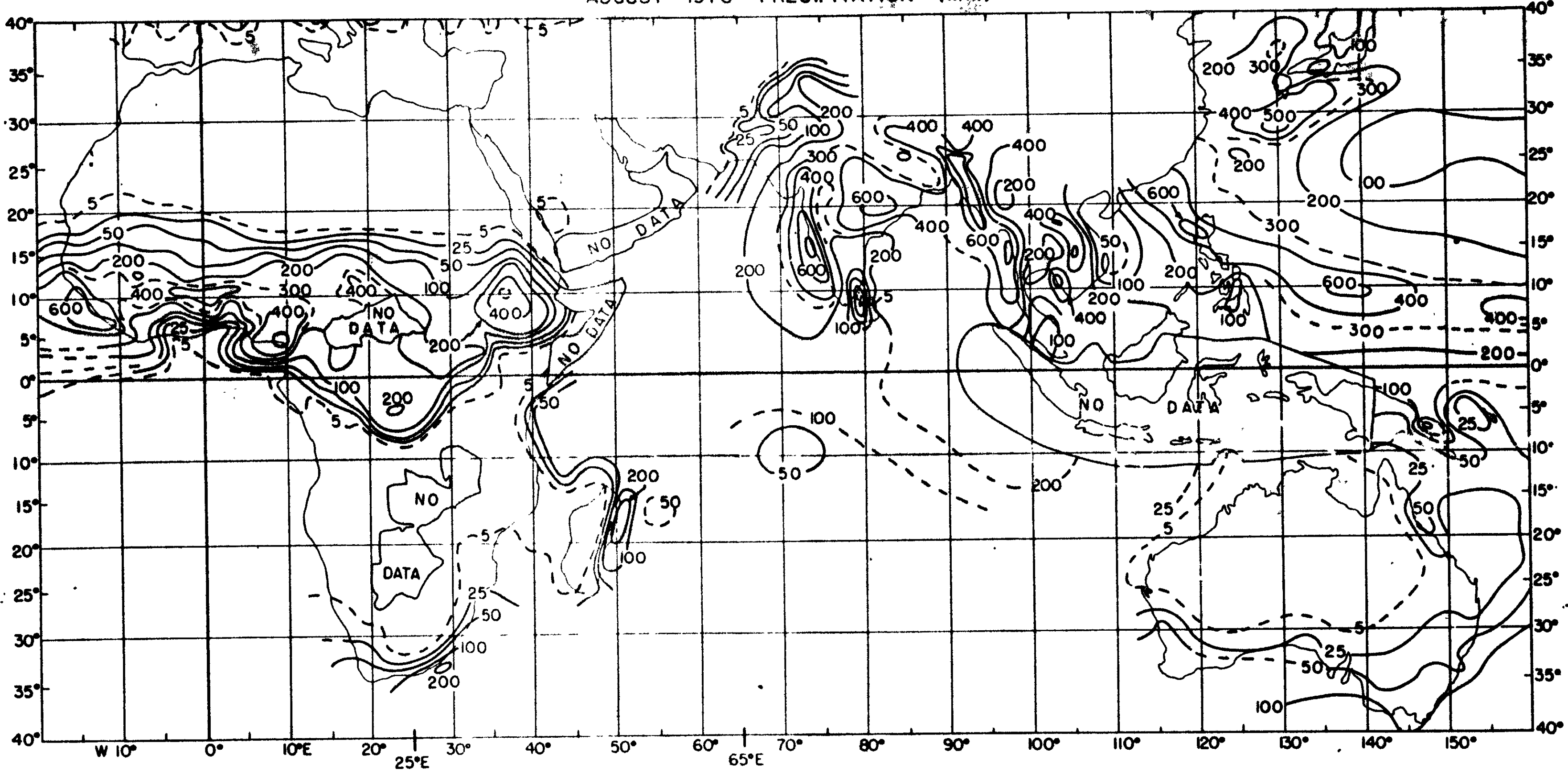
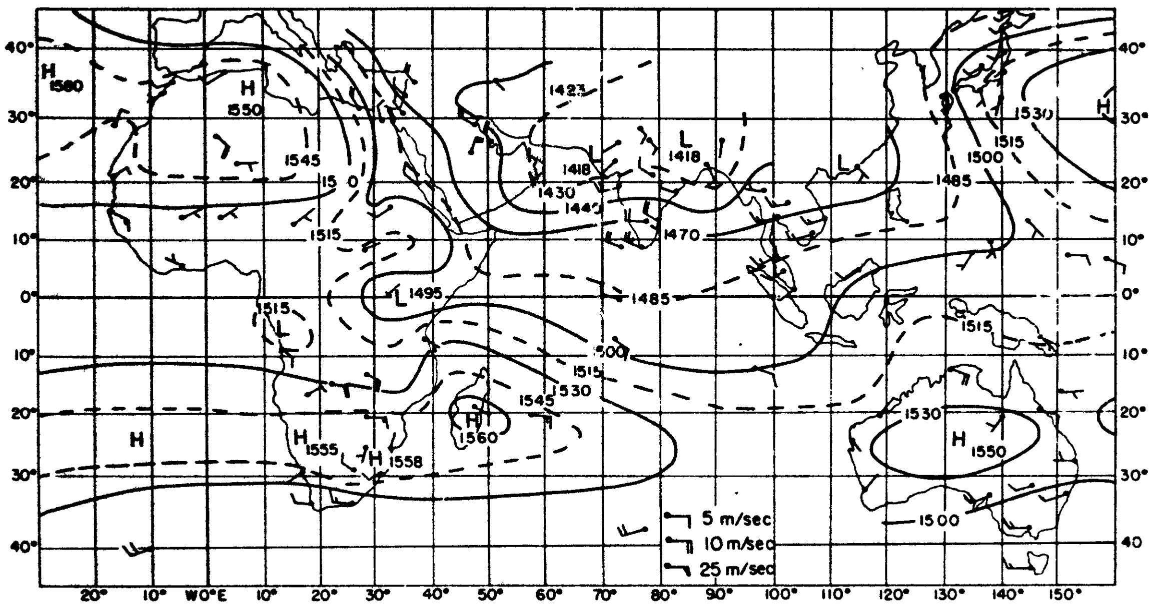


Figure 9.

Figure 10.

AUGUST 1970, 850 MB HEIGHT (GPM) AND WIND




AUGUST 1970, 200 MB HEIGHT (GPM) AND WIND

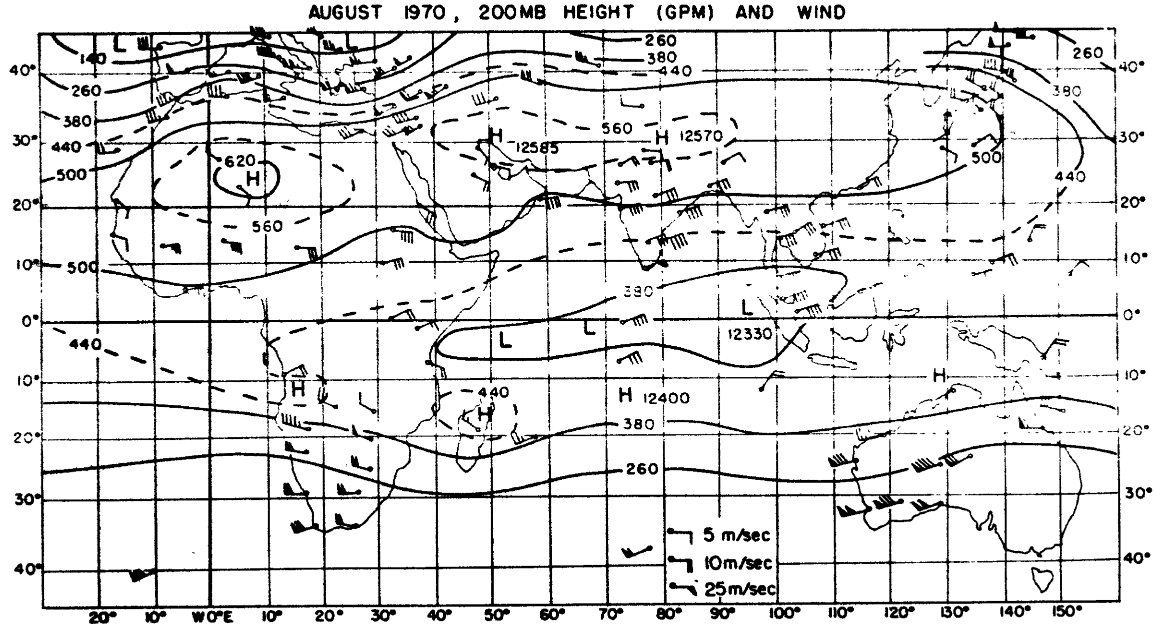
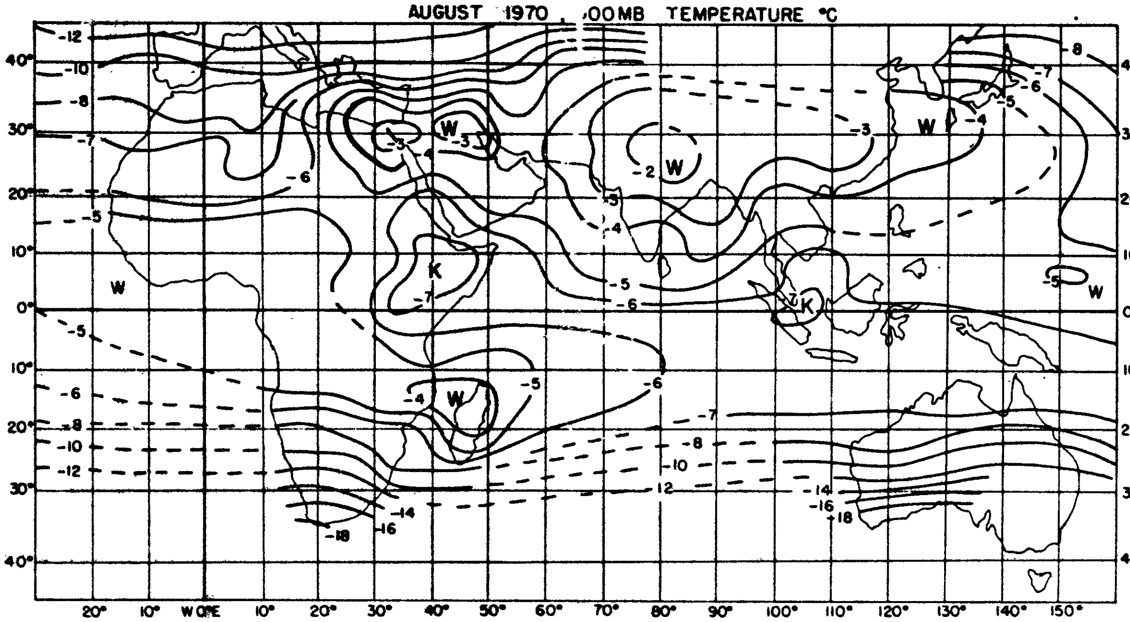


Figure 11.

Figure 12.

AUGUST 1970 , 500MB TEMPERATURE °C



AUGUST 1970 , 500MB DEWPOINT DEPRESSION °C

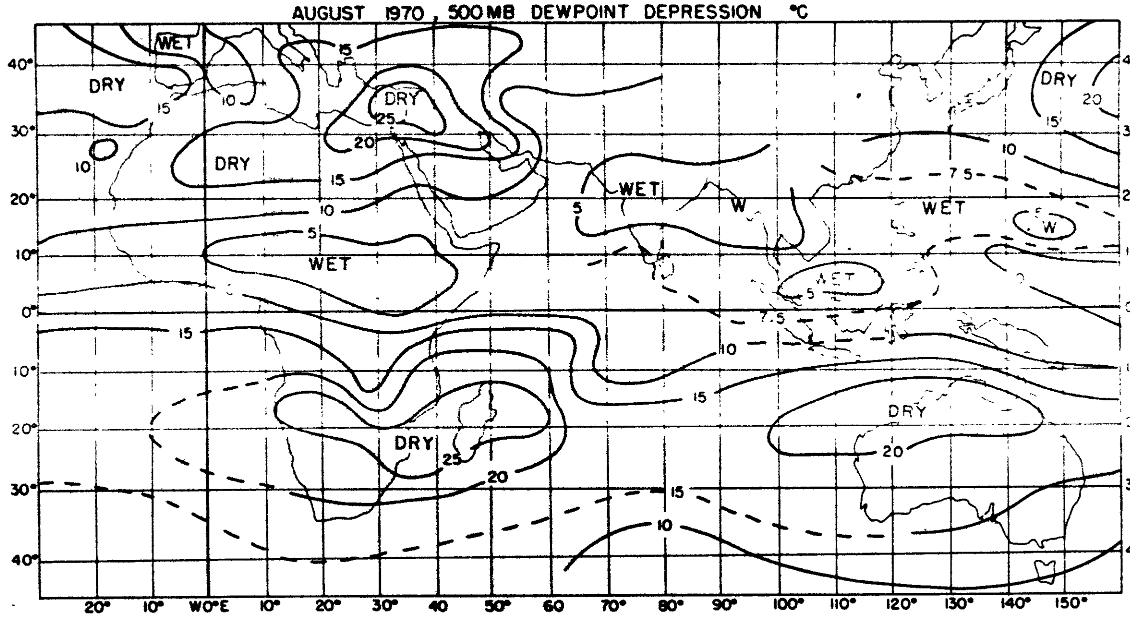


Figure 13.

MERIDIONAL CROSS-SECTION, AUGUST 1970 — INDIA

70°-80°E

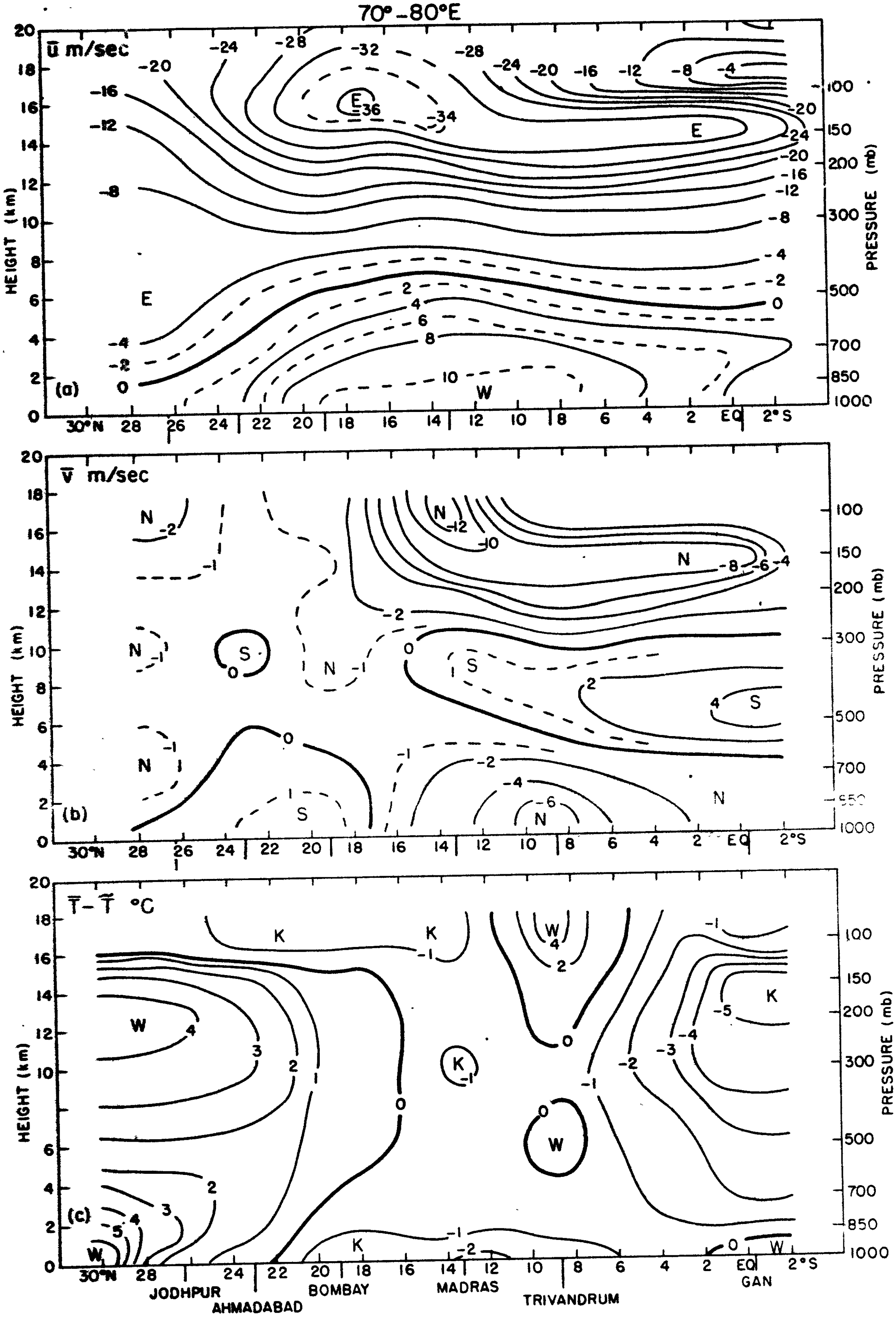


Figure 14.

MERIDIONAL CROSS-SECTION, AUGUST 1970 — EAST AFRICA
30°–40°E

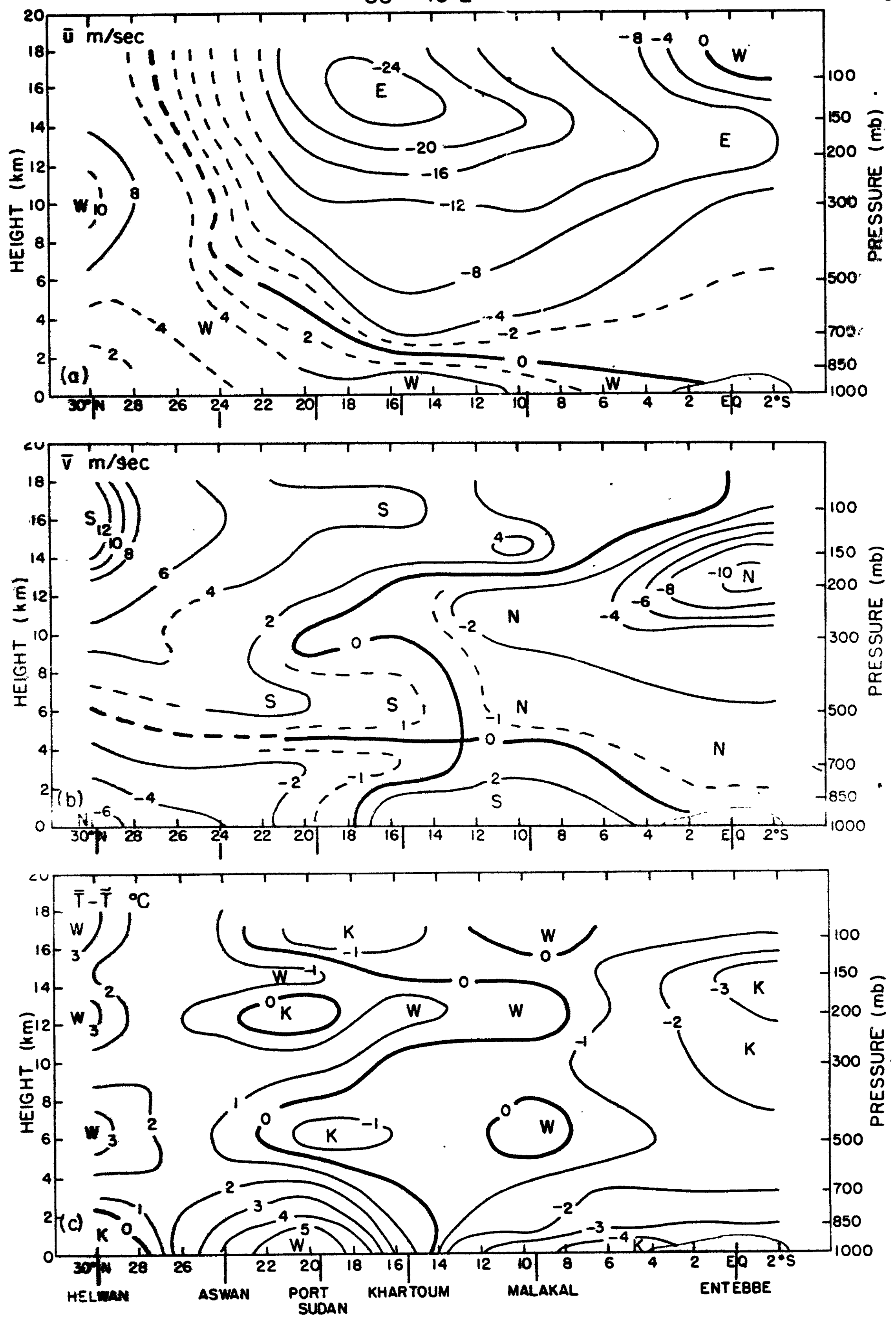


Figure 15.

MERIDIONAL CROSS-SECTION, AUGUST 1970 — WEST AFRICA

5°W–5°E

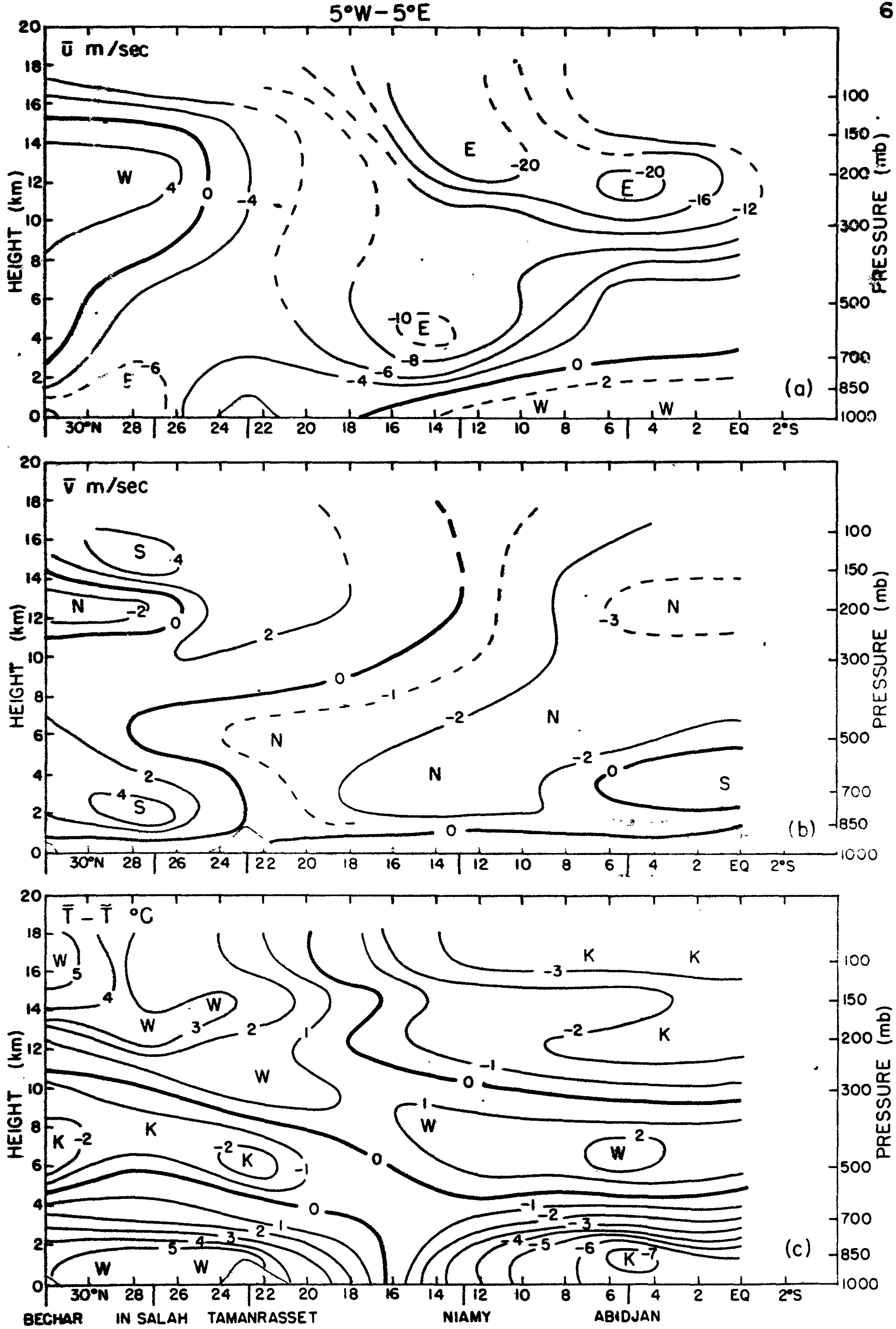


Figure 16.

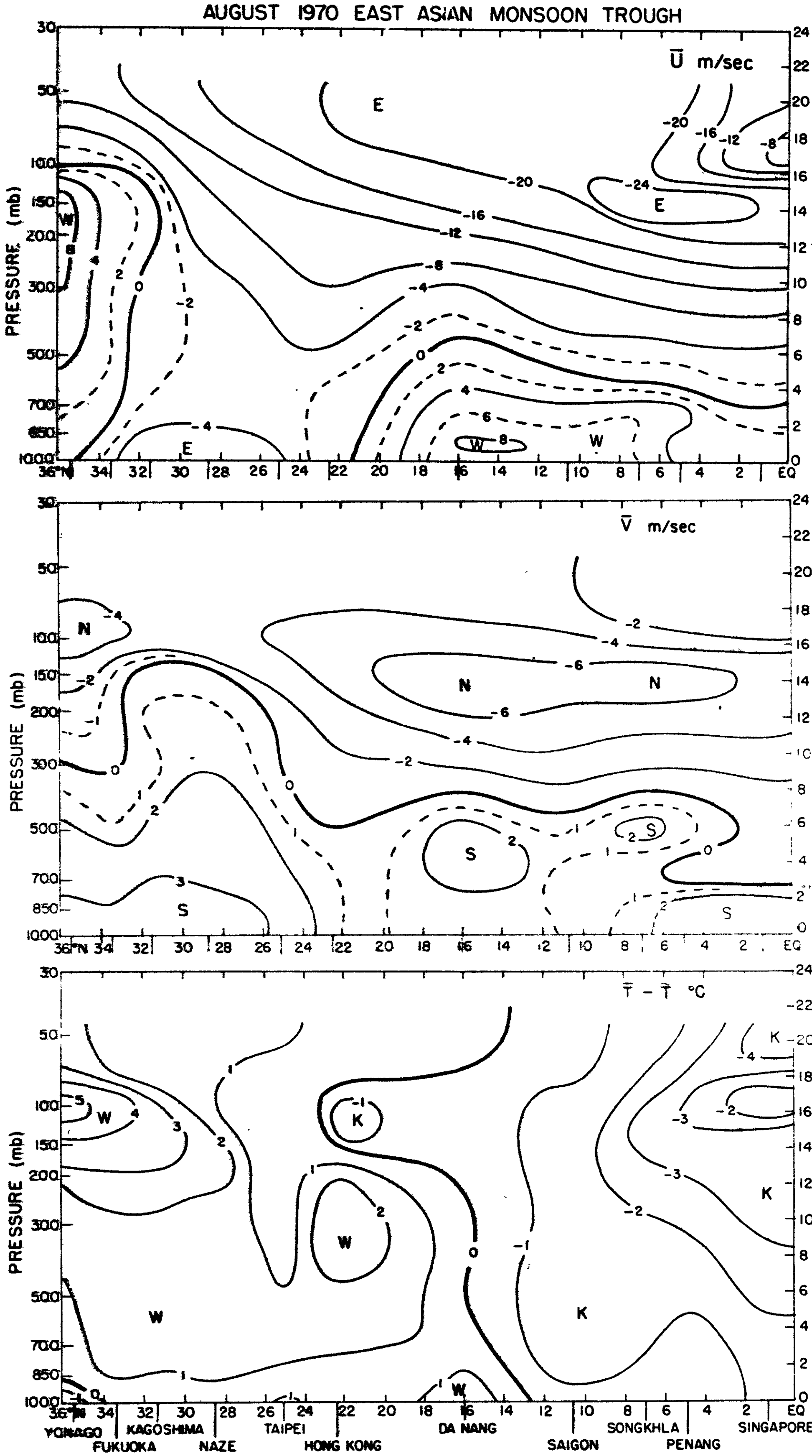

AUGUST 1970 EAST ASIAN MONSOON TROUGH
Ū m/sec
V̄ m/sec
T̄ − T̂ °C
PRESSURE (mb)
HEIGHT (km)
YONAGO
FUKUOKA
KAGOSHIMA
NAZE
TAIPEI
HONG KONG
DA NANG
SAIGON
SONGKHLA
PENANG
SINGAPORE

Figure 17.

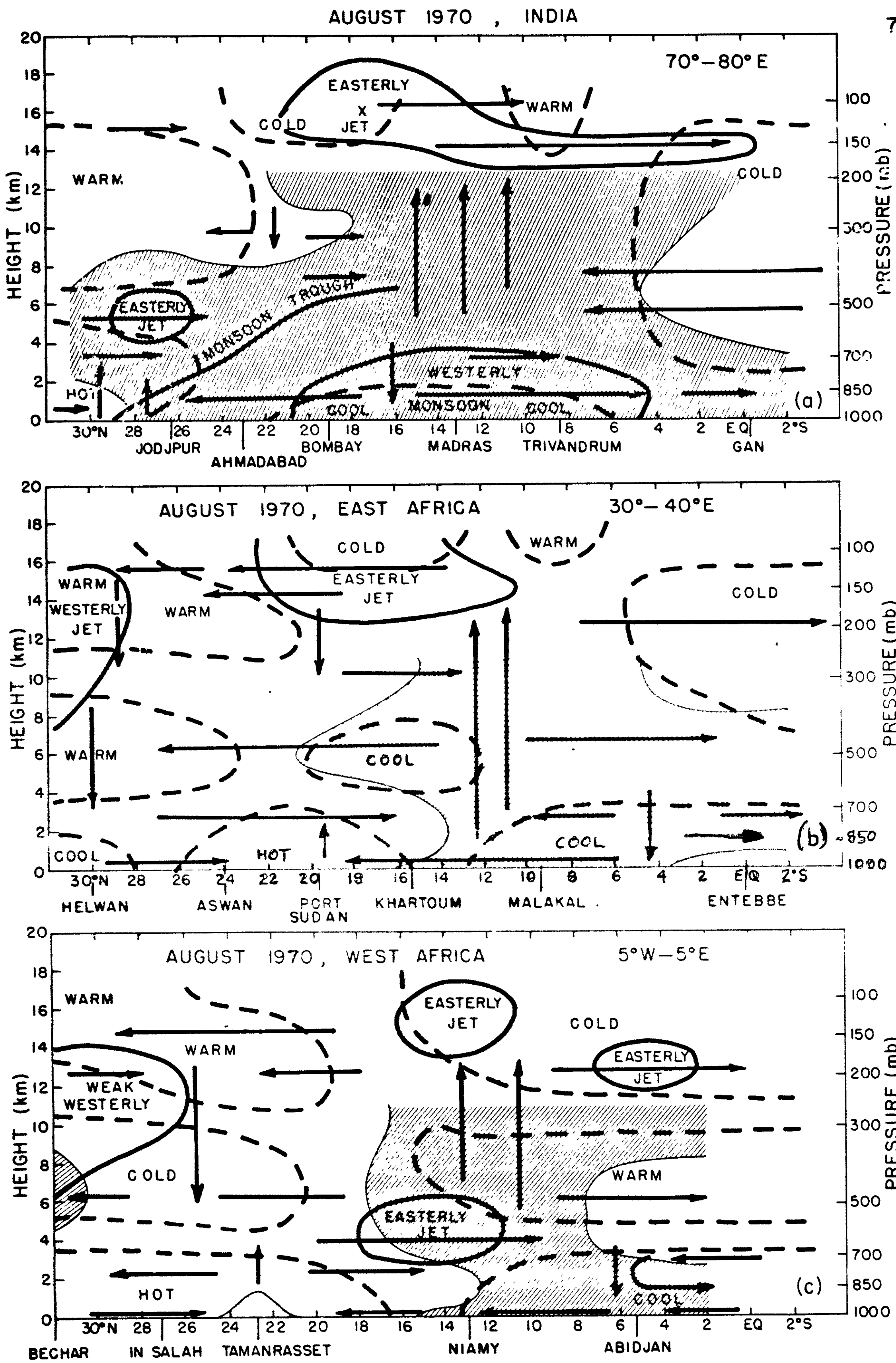

AUGUST 1970 , INDIA
70°–80° E
EASTERLY
JET
WARM
COLD
WARM
COLD
EASTERLY
JET
MONSOON TROUGH
WESTERLY
HOT
COOL
MONSOON
COOL
(a)
HEIGHT (km)
PRESSURE (mb)
30°N 28 26 24 22 20 18 16 14 12 10 8 6 4 2 EQ 2°S
JODJPUR
AHMADABAD
BOMBAY
MADRAS
TRIVANDRUM
GAN
AUGUST 1970, EAST AFRICA
30°–40°E
COLD
WARM
EASTERLY
JET
WARM
WESTERLY
JET
WARM
COLD
WARM
COOL
COOL
HOT
COOL
(b)
HELWAN
ASWAN
PORT SUDAN
KHARTOUM
MALAKAL
ENTEBBE
AUGUST 1970, WEST AFRICA
5°W–5°E
WARM
EASTERLY
JET
COLD
WARM
EASTERLY
JET
WEAK
WESTERLY
COLD
WARM
EASTERLY
JET
HOT
COOL
(c)
BECHAR
IN SALAH
TAMANRASSET
NIAMY
ABIDJAN

Figure 18.

AUGUST 1970 EAST ASIAN MONSOON TROUGH

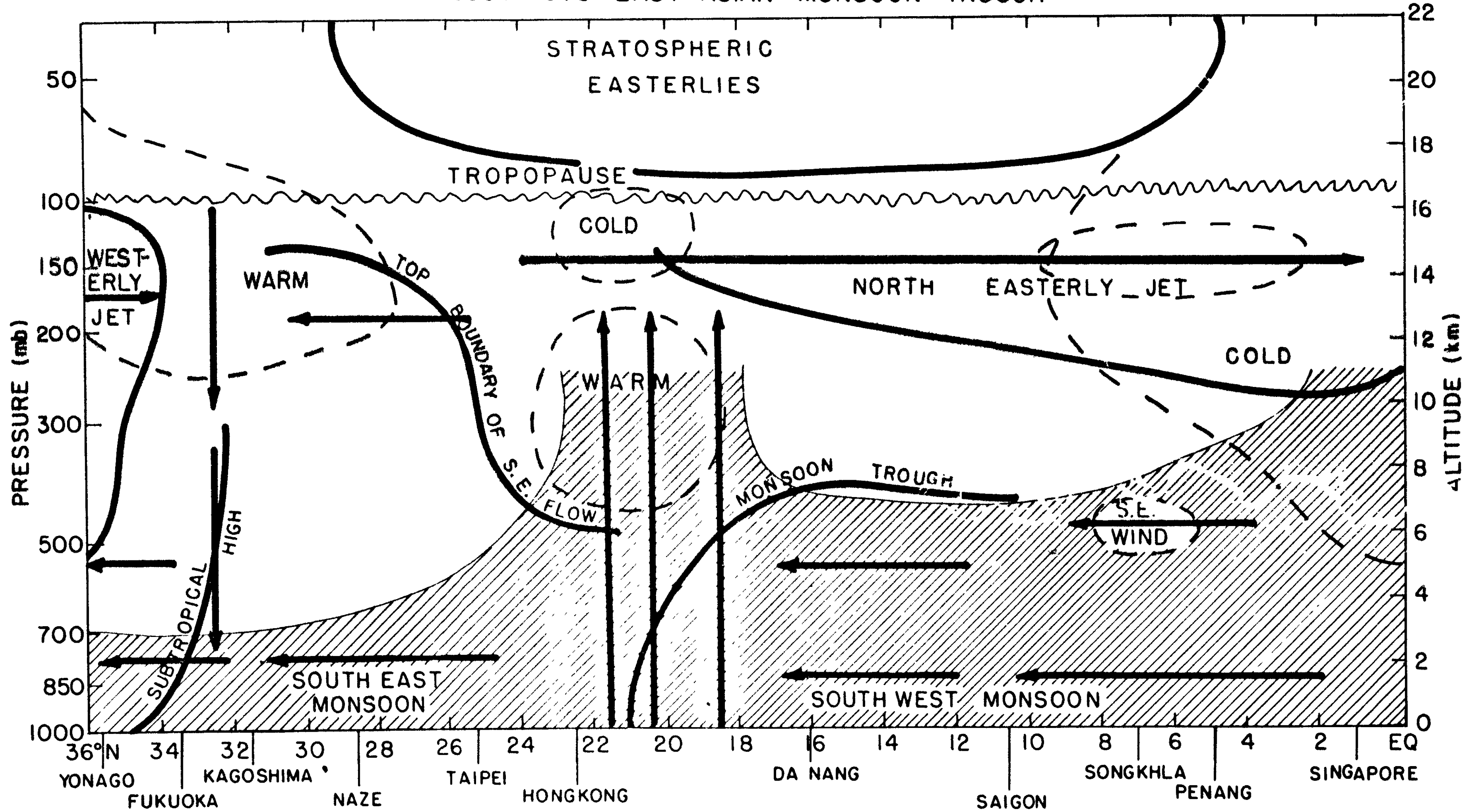
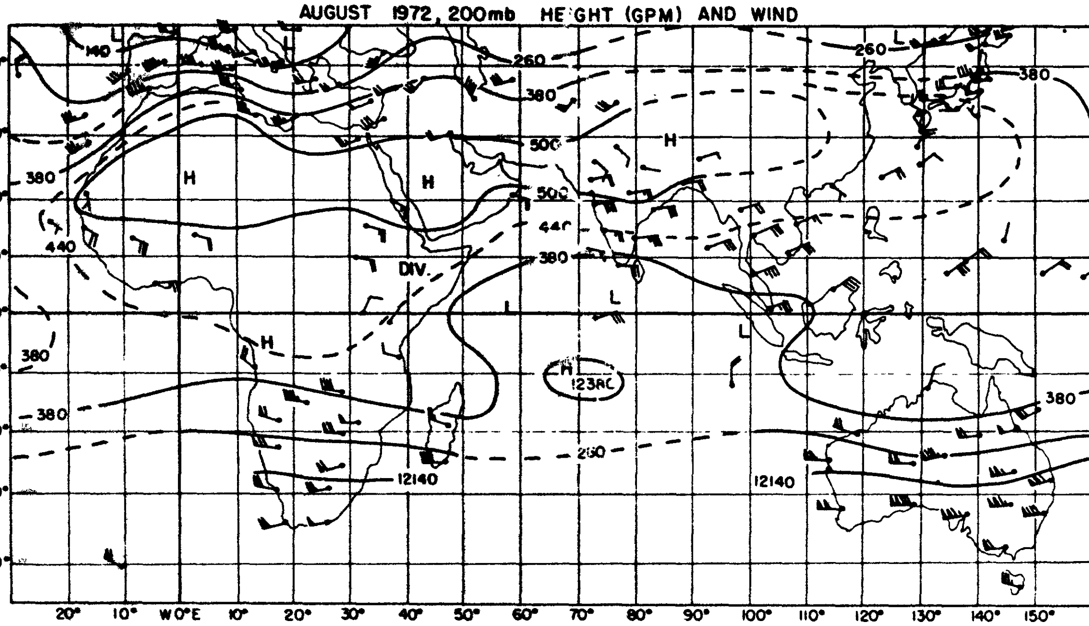


Figure 19.

Figure 20.

AUGUST 1972, 200mb HEIGHT (GPM) AND WIND




AUGUST 1972, 500MB TEMPERATURE °C

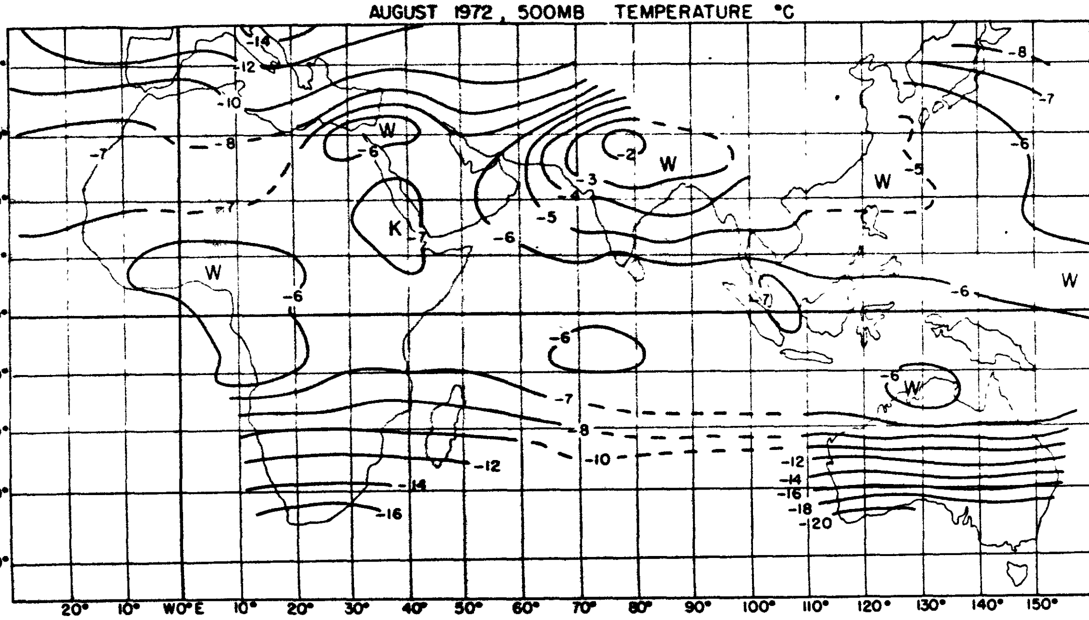


Figure 21.

AUGUST 1958 SEA-LEVEL PRESSURE (mb) DEPARTURE FROM 1951-60 MEAN (mb)

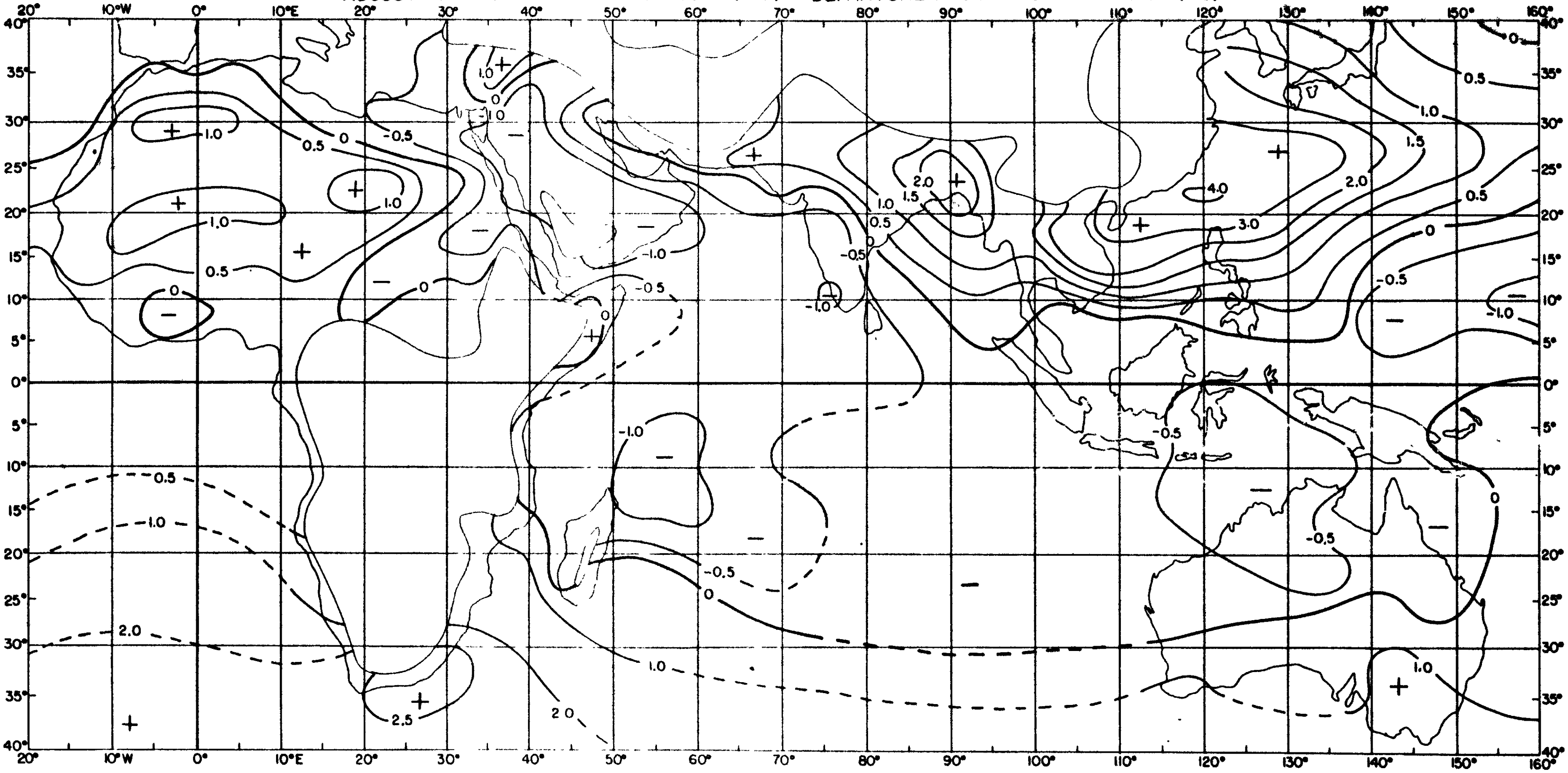


Figure 22.

AUGUST 1960 SEA LEVEL PRESSURE (mb) DEPARTURES FROM 1951-60 MEAN (mb)

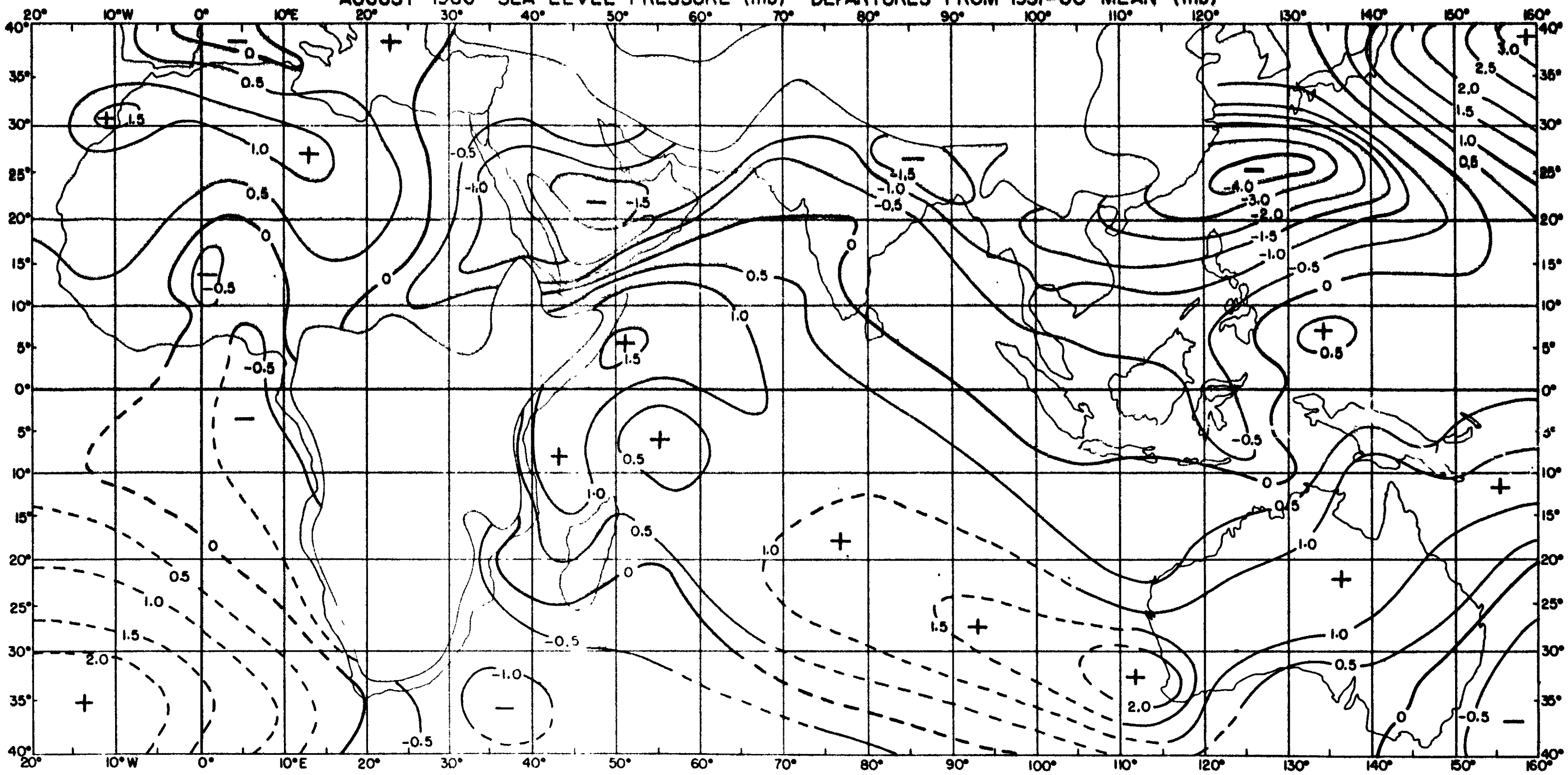


Figure 23.

GRID BOX USED TO OBTAIN TYPHOON FREFRENCY

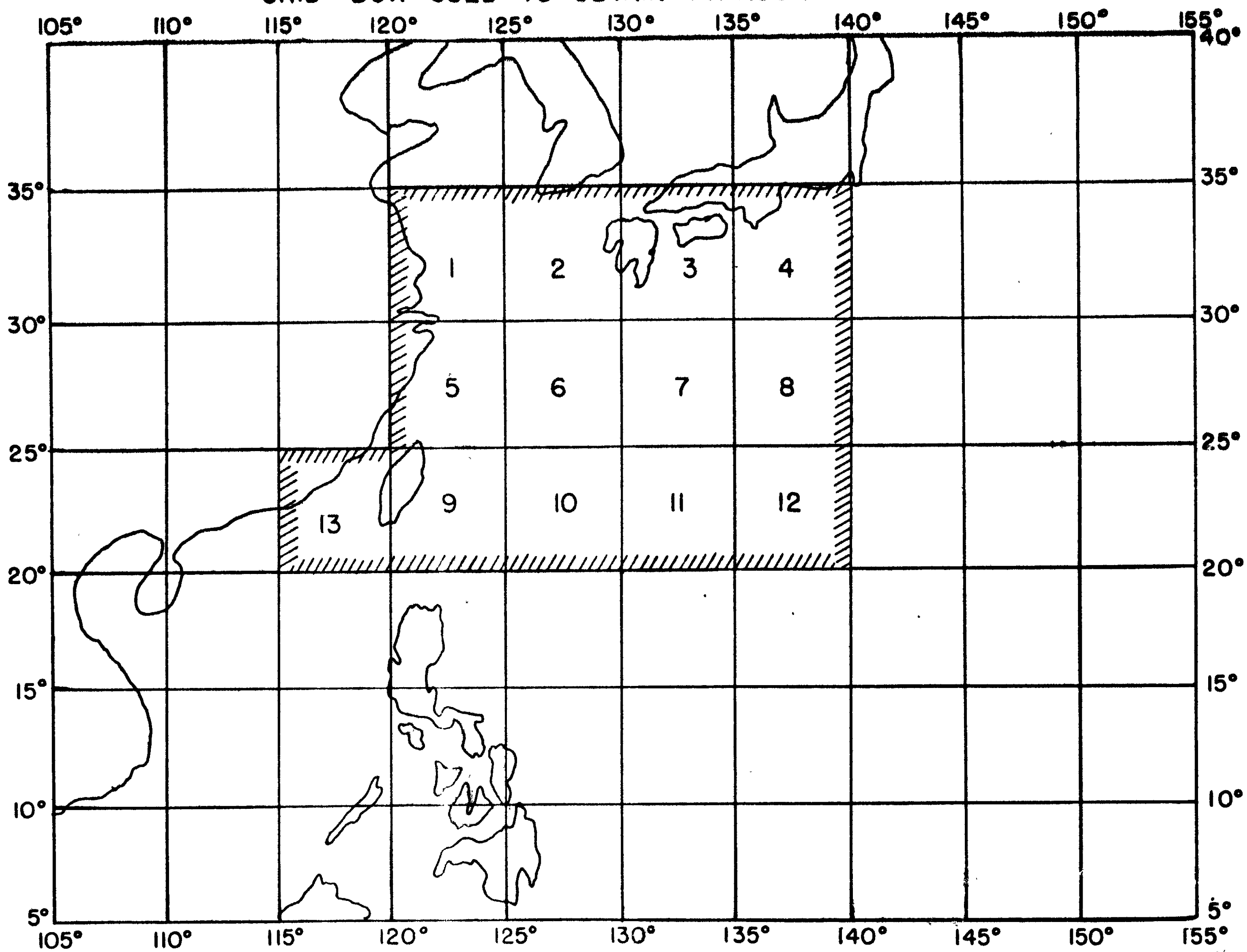


Figure 24.

MONTHLY TYPHOON FREQUENCY (AVERAGE OF 6 MONTHS WHEN P [ISHIGAKI] < -0.8σ)

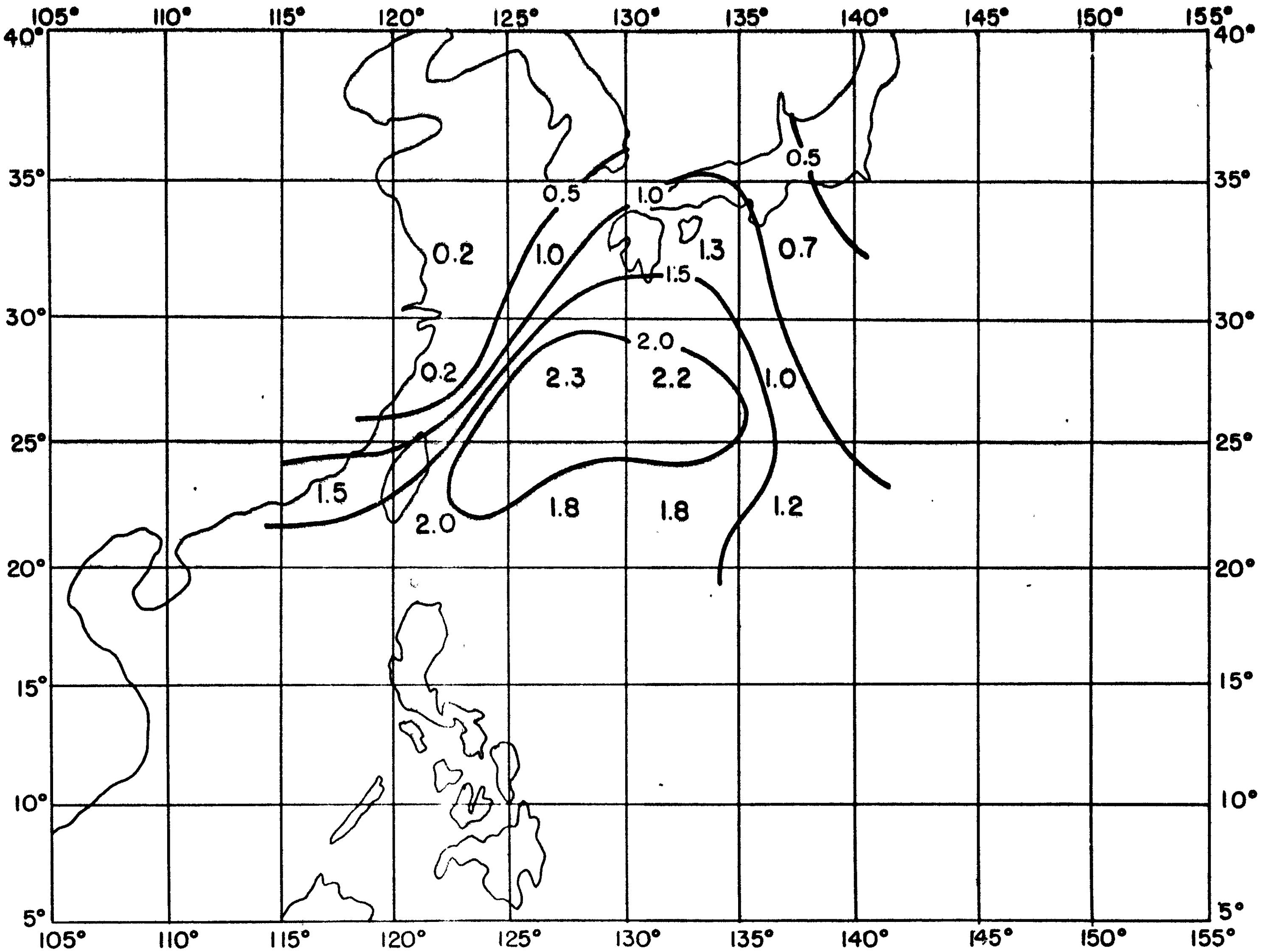


Figure 25.

MONTHLY TYPHOON FREQUENCY (AVERAGE OF 7 MONTHS WHEN P [ISHIGAKI] > +0.9σ)

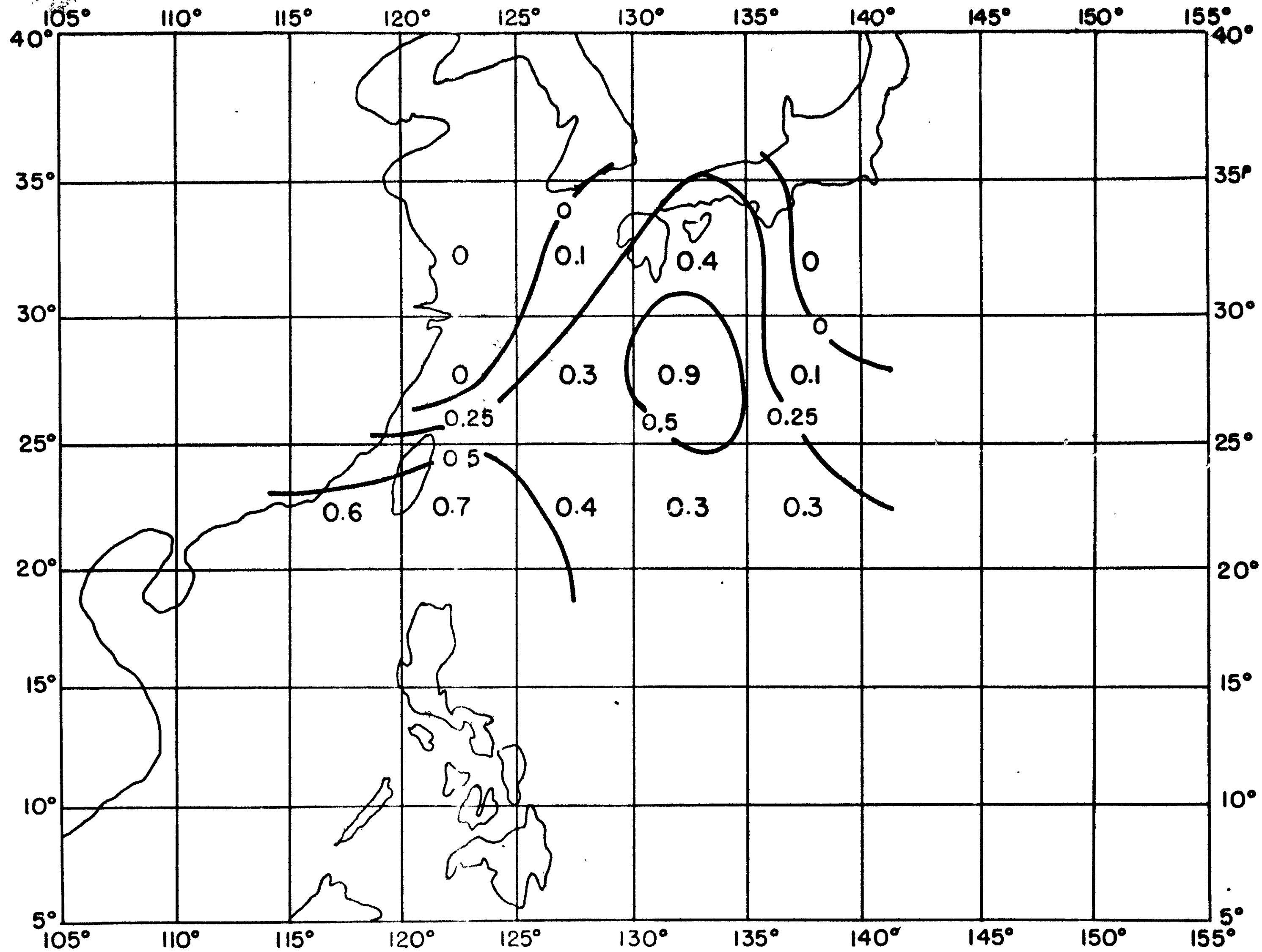
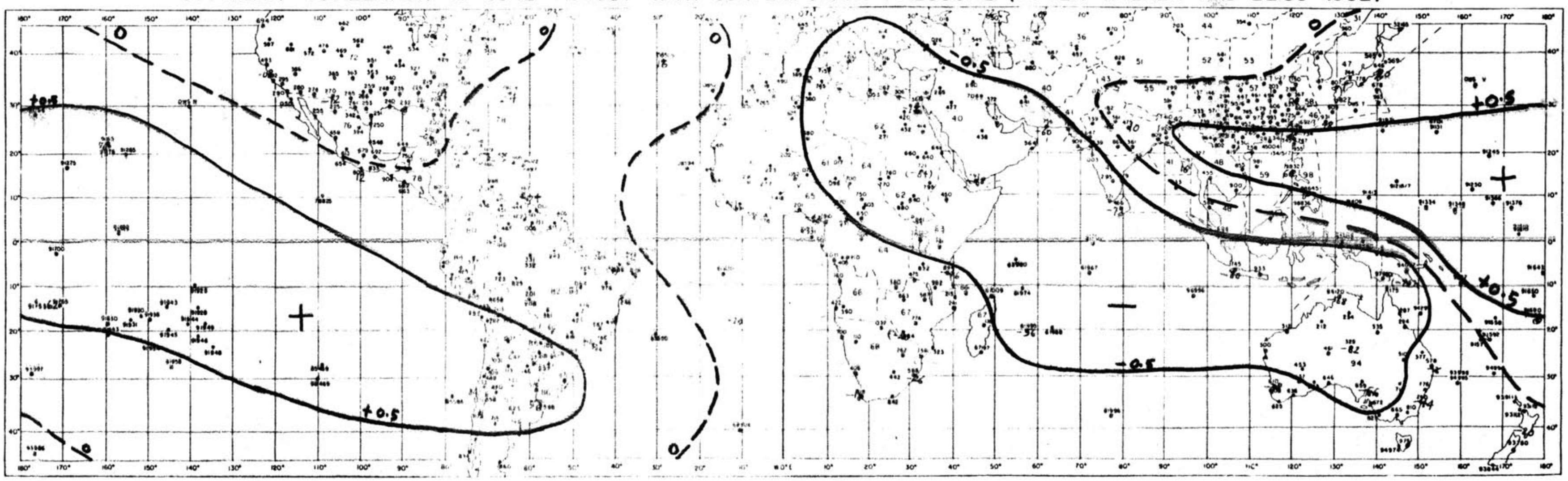


Figure 26.

SOUTHERN OSCILLATION OF JUNE-AUGUST WITH CONTEMPORARY PRESSURE (AFTER WALKER AND BLISS 1932)




SOUTHERN OSCILLATION OF JUNE-AUGUST WITH CONTEMPORARY RAINFALL (AFTER WALKER AND BLISS 1932)

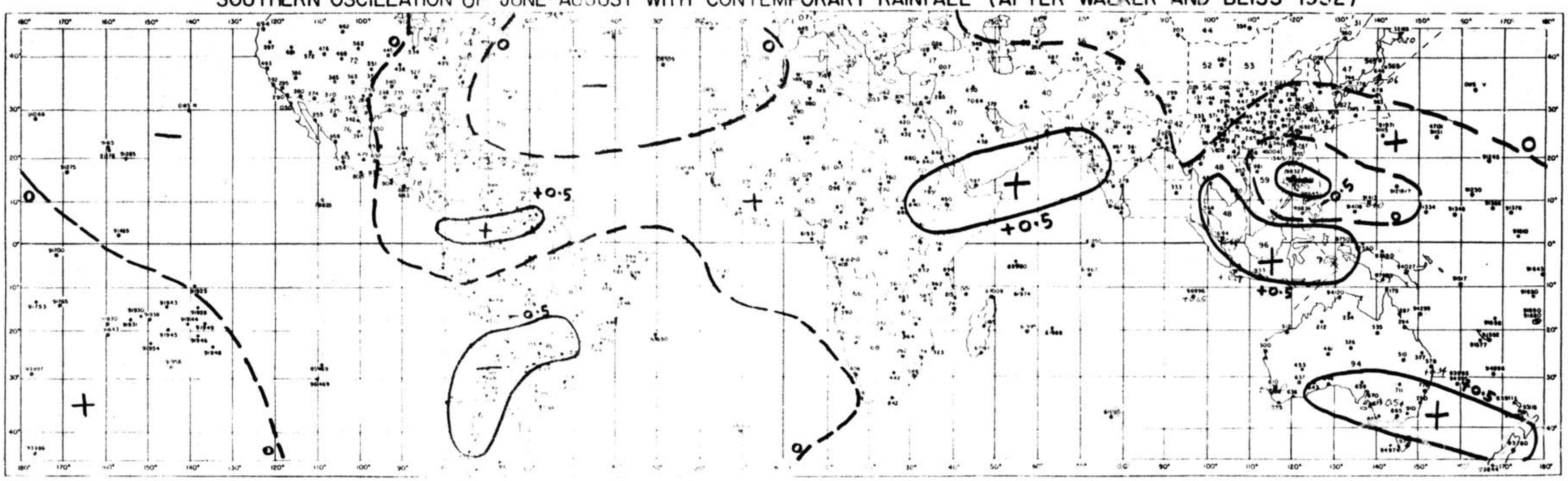


Figure 27.

SOUTHERN OSCILLATION INDEX , JUNE – AUGUST

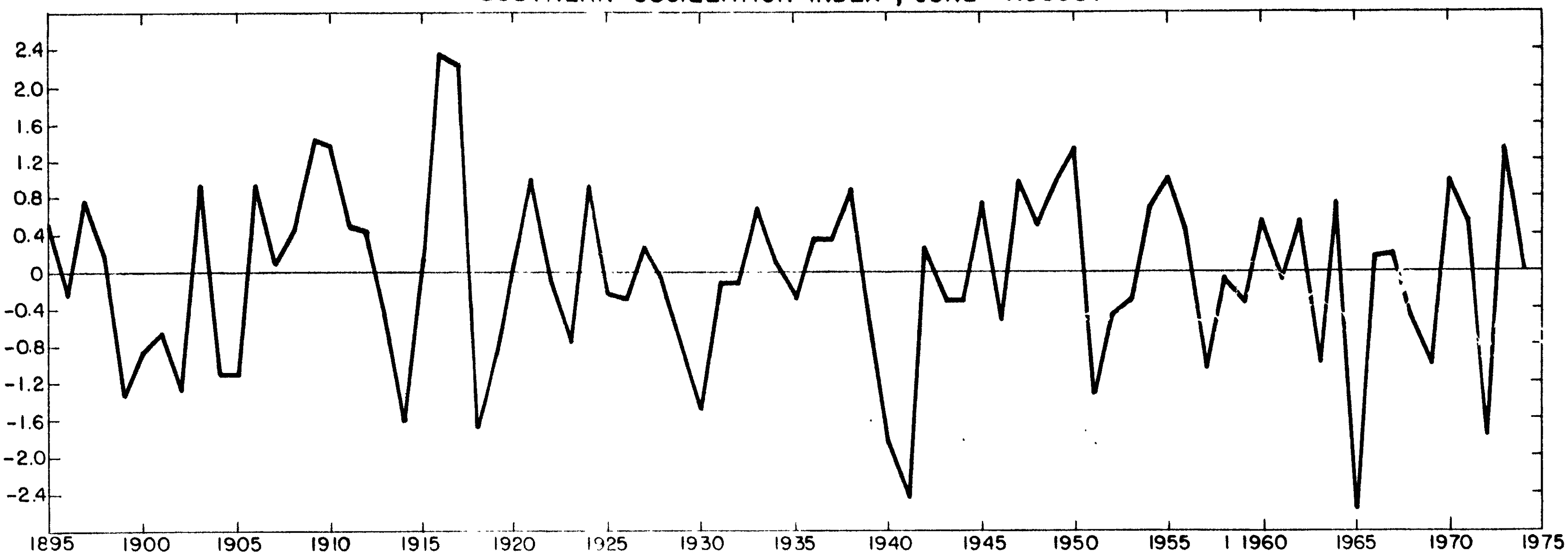


TABLE 1

RAINFALL STATISTICS

A) JULY

| REGION | TIME PERIOD | MEAN PRECIPITATION | STANDARD DEVIATION |
|---|---|---|---|
| | YR | mm | % of MEAN |
| WEST AFRICA | 1941-74(34) | 210.4 | 17.59 |
| EAST AFRICA | 1951-74(24) | 247.0 | 20.56 |
| WEST INDIA | 1941-74(34) | 755.2 | 24.05 |
| BURMA | 1951-74(24) | 827.0 | 15.56 |
| SOUTH CHINA SEA | 1951-74(24) | 291.3 | 50.53 |
| EAST INDIA | 1941-72(32) | 405.0 | 20.14 |
| CENTRAL INDIA | 1941-74(34) | 333.2 | 24.43 |
| TONGKING GULF | 1906-38(33) | 360.4 | 32.11 |

B) AUGUST

| REGION | TIME PERIOD | MEAN PRECIPITATION | STANDARD DEVIATION |
|---|---|---|---|
| WEST AFRICA | 1941-74(34) | 297.4 | 18.96 |
| EAST AFRICA | 1951-74(24) | 288.0 | 31.13 |
| WEST INDIA | 1941-74(34) | 430.6 | 35.99 |
| BURMA | 1951-74(24) | 826.0 | 19.82 |
| S. CHINA SEA | 1951-74(24) | 350.0 | 32.83 |
| EAST INDIA | 1941-73(33) | 356.4 | 20.85 |
| CENTRAL INDIA | 1941-74(34) | 347.1 | 23.09 |
| TONGKING GULF | 1906-38(33) | 370.8 | 40.13 |

TABLE 2

TEN YEAR MEAN PRECIPITATION

WEST AFRICA Unit (mm)

| PERIOD | JULY | AUGUST | TWO MONTH TOTAL |
|---|---|---|---|
| 1941-50 | 196 | 306 | 502 |
| 1951-60 | 219 | 300 | 519 |
| 1961-70 | 211 | 298 | 509 |

TABLE 3

TEN YEAR MEAN PRECIPITATION

EAST AFRICA Unit (mm)

| PERIOD | JULY | AUGUST | TWO MONTH TOTAL |
|---|---|---|---|
| 1902-10 | 268 | 311 | 579 |
| 1911-20 | 281 | 317 | 598 |
| 1921-30 | 287 | 296 | 583 |
| 1931-39 | 269 | 248 | 517 |
| 1941-50 | - | - | - |
| 1951-60 | 239 | 266 | 505 |
| 1961-70 | 254 | 271 | 525 |

TABLE 4

TEN YEAR MEAN PRECIPITATION

WEST INDIA Unit (mm)

| PERIOD | JULY | AUGUST | TWO MONTH TOTAL |
|---|---|---|---|
| 1871-1880 | 640 | 363 | 1003 |
| 1881-1890 | 657 | 352 | 1009 |
| 1891-1900 | 625 | 397 | 1022 |
| 1901-1910 | 698 | 374 | 1072 |
| 1911-1920 | 590 | 366 | 956 |
| 1921-1930 | 695 | 397 | 1092 |
| 1931- 1940 | 691 | 421 | 1112 |
| 1941-1950 | 709 | 396 | 1105 |
| 1951-1960 | 729 | 426 | 1155 |
| 1961-1970 | 819 | 448 | 1267 |
| 100 YR. AV. 1871-1970 | 685 | 394 | 1079 |

## TABLE 5

### TEN YEAR MEAN PRECIPITATION

BURMA Unit (mm)

| PERIOD | JULY | AUGUST | TWO MONTH TOTAL |
|---|---|---|---|
| 1901-10 | 910 | 819 | 1729 |
| 1911-20 | 995 | 853 | 1848 |
| 1921-30 | 1057 | 876 | 1933 |
| 1931-40 | 875 | 730 | 1605 |
| 1941-50 | - | - | - |
| 1951-60 | 765 | 812 | 1577 |
| 1961-70 | 862 | 844 | 1706 |

## TABLE 6

### TEN YEAR MEAN PRECIPITATION

SOUTH CHINA SEA Unit (mm)

| PERIOD | JULY | AUGUST | TWO MONTH TOTAL |
|---|---|---|---|
| 1911-20 | 395 | 397 | 792 |
| 1921-30 | 422 | 390 | 812 |
| 1931-40 | 383 | 365 | 748 |
| 1941-50 | - | - | - |
| 1951-60 | 232 | 372 | 604 |
| 1961-70 | 304 | 327 | 631 |

TABLE 7 

TEN YEAR MEAN PRECIPITATION

EAST INDIA Unit (mm)

| PERIOD | JULY | AUGUST | TWO MONTH TOTAL |
|---|---|---|---|
| 1901-10 | 400 | 354 | 754 |
| 1911-20 | 412 | 370 | 782 |
| 1921-30 | 427 | 392 | 819 |
| 1931-40 | 394 | 341 | 735 |
| 1941-50 | 358 | 364 | 722 |
| 1951-60 | 438 | 323 | 761 |
| 1961-70 | 421 | 378 | 799 |

TABLE 8



TEN YEAR RAINFALL AUTOCORRELATION

( Between July and August precipitation )

| PERIOD | WEST AFRICA | EAST AFRICA | WEST INDIA | EAST INDIA | BURMA | S.CHINA SEA |
|---|---|---|---|---|---|---|
| 1901-10 | - | - | +0.27 | -0.26 | -0.36 | - |
| 1906-15 | - | +0.22 | +0.34 | +0.38 | -0.15 | +0.35 |
| 1911-20 | - | +0.06 | +0.39 | +0.14 | -0.38 | +0.38 |
| 1916-25 | - | +0.08 | +0.36 | +0.02 | +0.11 | +0.10 |
| 1921-30 | - | -0.37 | +0.44 | +0.26 | +0.58 | -0.34 |
| 1926-35 | - | -0.14 | +0.71° | -0.05 | -0.19 | -0.53 |
| 1931-40 | - | +0.04 | +0.36 | -0.45 | -0.47 | -0.21 |
| 1936-45 | - | - | +0.11 | -0.17 | - | - |
| 1941-50 | +0.23 | - | -0.43 | +0.38 | - | - |
| 1946-55 | -0.37 | +0.77° | -0.12 | +0.30 | - | - |
| 1951-60 | -0.54 | +0.46 | +0.27 | +0.53 | -0.01 | -0.53 |
| 1956-65 | -0.53 | +0.23 | +0.04 | -0.17 | -0.06 | -0.18 |
| 1961-70 | +0.01 | +0.25 | -0.24 | -0.25 | 0.00 | -0.06 |
| 1965-74 | +0.57 | +0.60 | -0.05 | - | -0.19 | 0.00 |

° 5% level = 0.65

# TABLE 9

## SHORT PERIOD RAINFALL CORRELATION

### JULY CORRELATION

| LOCATION | PERIOD | CORRELATION |
|---|---|---|
| BURMA-S.CHINA SEA | 1903-22(20 YR ) | +0.65* |
| BURMA-TONGKING GULF | 1921-33(13 YR ) | +0.82* |
| BURMA-S.CHINA SEA | 1928-39(12 YR ) | +0.75° |
| BURMA-TONGKING GULF | 1906-17(12 YR ) | +0.75° |
| WEST INDIA-BURMA | 1921-39(19 YR ) | +0.54 |

### AUGUST CORRELATION

| LOCATION | PERIOD | CORRELATION |
|---|---|---|
| BURMA-S.CHINA SEA | 1905-40(36 YR ) | +0.58* |
| BURMA-S.CHINA SEA | 1951-71(21 YR ) | +0.64° |
| WEST INDIA-WEST AFRICA | 1952-72(21 YR ) | +0.71* |
| WEST INDIA-EAST AFRICA | 1923-39(16 YR ) | +0.74* |
| WEST INDIA-EAST AFRICA | 1946-59(14 YR ) | +0.76* |
| WEST INDIA-BURMA | 1923-40(18 YR ) | +0.54 |

* Significant at $3\sigma$

° Significant at $2.6\sigma$

# TABLE 10

## TEN YEAR RAINFALL CORRELATION

August correlation with West India

| PERIOD | WEST AFRICA | EAST AFRICA | EAST INDIA | BURMA |
|---|---|---|---|---|
| 1901-10 | - | - | -0.45 | +0.54 |
| 1906-15 | - | -0.47 | -0.14 | +0.51 |
| 1911-20 | - | -0.06 | -0.26 | +0.11 |
| 1916-25 | - | +0.10 | -0.48 | +0.24 |
| 1921-30 | - | +0.50 | -0.05 | +0.47 |
| 1926-35 | - | +0.17 | -0.36 | +0.64 |
| 1931-40 | - | - | -0.52 | +0.86* |
| 1936-45 | - | - | -0.43 | +0.26 |
| 1941-50 | +0.30 | - | -0.14 | -0.18 |
| 1946-55 | +0.60 | +0.80* | +0.14 | -0.25 |
| 1951-60 | +0.53 | +0.62 | +0.22 | -0.33 |
| 1956-65 | +0.46 | -0.21 | -0.18 | -0.03 |
| 1961-70 | +0.80* | -0.28 | -0.70° | +0.19 |
| 1965-74 | +0.57 | +0.53 | -0.75° | +0.23 |
| Maximum | +0.87* 1963-72 | +0.80* 1946-55 | -0.81* 1962-71 | +0.86* 1931-40 |

° 5% level = 0.65
* 1% level = 0.78

TABLE 11 

TEN YEAR RAINFALL CORRELATION

July correlation with West India

| PERIOD | WEST AFRICA | EAST AFRICA | EAST INDIA | BURMA |
|---|---|---|---|---|
| 1901-10 | - | - | -0.27 | +0.03 |
| 1906-15 | - | -0.23 | -0.11 | +0.56 |
| 1911-20 | - | +0.50 | -0.40 | +0.61 |
| 1916-25 | - | +0.48 | -0.61 | +0.72° |
| 1921-30 | - | -0.21 | -0.08 | +0.64 |
| 1926-35 | - | -0.39 | +0.60 | -0.29 |
| 1931-40 | - | - | +0.14 | +0.30 |
| 1936-45 | - | - | -0.16 | +0.04 |
| 1941-50 | -0.21 | - | -0.15 | -0.18 |
| 1946-55 | -0.12 | -0.35 | -0.25 | -0.31 |
| 1951-60 | -0.44 | -0.06 | -0.50 | -0.23 |
| 1956-65 | -0.25 | +0.35 | -0.71° | +0.61 |
| 1961-70 | +0.08 | 0.00 | -0.53 | +0.44 |
| 1965-74 | +0.29 | -0.33 | -0.09 | +0.06 |
| | | | | |
| Maximum | -0.45<br>1952-61 | -0.57<br>1964-73 | -0.72°<br>1955-64 | +0.73°<br>1917-26 |

° 5% level = 0.65
* 1% level = 0.78

# TABLE 12

## TEN YEAR RAINFALL CORRELATION

Correlation between Burma and South China Sea

| PERIOD | JULY | AUGUST |
|---|---|---|
| 1901-10 | - | - |
| 1906-15 | +0.45 | +0.72° |
| 1911-20 | +0.48 | +0.65° |
| 1916-25 | +0.58 | +0.73° |
| 1921-30 | -0.09 | +0.58 |
| 1926-35 | +0.71° | +0.55 |
| 1931-40 | +0.62 | +0.67° |
| 1936-45 | - | - |
| 1941-50 | - | - |
| 1946-55 | - | - |
| 1951-60 | +0.26 | +0.73° |
| 1956-65 | +0.58 | +0.76° |
| 1961-70 | +0.52 | +0.62 |
| 1965-74 | 0.00 | +0.51 |
| Maximum | +0.79*<br>1913-22 | +0.86*<br>1953-62 |

° 5% level = 0.65
*1 % level = 0.78

# TABLE 13

## TROPICAL EASTERLY JET

## ZONAL KINETIC ENERGY/UNIT MASS

July India ( 26°N-6°N ) Unit: $m^2/sec^2$

| YEAR | TOTAL(900-75mb) | % of norm. | TEJ(250-75mb) | % of norm. | S.W. MONSOON | % | W. INDIA RAINFALL % of norm. |
|---|---|---|---|---|---|---|---|
| 1965 | 71.4 | 86 | 239.7 | 78 | 41.8 | 106 | 104 |
| 1966 | 71.2 | 85 | 263.4 | 86 | 31.0 | 79 | 111 |
| 1967 | 88.5 | 106 | 321.9 | 105 | 39.6 | 101 | 145 |
| 1968 | 101.5 | 122 | 399.4 | 130 | 50.6 | 139 | 135 |
| 1969 | 97.4 | 117 | 358.0 | 117 | 43.1 | 109 | 97 |
| 1970 | 91.6 | 110 | 350.7 | 114 | 34.8 | 88 | 93 |
| 1971 | 88.1 | 106 | 334.5 | 109 | 36.5 | 93 | 89 |
| 1972 | 77.2 | 93 | 270.2 | 88 | 44.7 | 113 | 83 |
| 1973 | 80.2 | 96 | 281.5 | 92 | 43.2 | 110 | 84 |
| 1974 | 67 | 80 | 251 | 82 | 29.0 | 74 | 160 |
| Av. | 83.4 | | 307.0 | | 39.4 | | |

TABLE 14

TROPICAL EASTERLY JET

ZONAL KINETIC ENERGY/UNIT MASS

August India ( 26°N-6°N ) Unit: $m^2/sec^2$

| YEAR | TOTAL(900-75mb) | % of norm. | TEJ(250-75 ) | % | S.W. MONSOON | % | W. INDIA RAINFALL % of norm. |
|---|---|---|---|---|---|---|---|
| 1965 | 63.6 | 87 | 249.8 | 90 | 11.4 | 44 | 77 |
| 1966 | 78.2 | 101 | 308.8 | 111 | 15.8 | 61 | 41 |
| 1967 | 65.8 | 90 | 227.4 | 82 | 33.5 | 130 | 116 |
| 1968 | 80.9 | 110 | 303.9 | 110 | 32.9 | 128 | 56 |
| 1969 | 65.5 | 90 | 260.1 | 94 | 12.5 | 49 | 80 |
| 1970 | 94.2 | 129 | 359.5 | 130 | 33.6 | 131 | 119 |
| 1971 | 77.7 | 106 | 306.7 | 111 | 21.1 | 82 | 114 |
| 1972 | 58.6 | 80 | 215.5 | 78 | 29.0 | 113 | 67 |
| 1973 | 69.6 | 95 | 260.5 | 94 | 29.0 | 113 | 160 |
| 1974 | 77.3 | 106 | 279.7 | 101 | 38.2 | 149 | 120 |
| Av. | 73.1 | | 277.2 | | 25.7 | | |
| | 88% of July | | 90% of July | | 65% of July | | |

TABLE 15

TROPICAL EASTERLY JET

ZONAL KINETIC ENERGY/UNIT MASS

July East Africa (26°-6°N) Unit: $m^2/sec^2$

| YEAR | TOTAL(900-75mb) | % of norm. | TEJ(250-75) | % | % tile of India ZKE total | TEJ only |
|---|---|---|---|---|---|---|
| 1965 | 24.4 | 63 | 80.1 | 57 | 34.1 | 33.4 |
| 1966 | 38.7 | 100 | 140.8 | 101 | 54.4 | 53.5 |
| 1967 | 54.3 | 141 | 212.5 | 152 | 61.3 | 66.0 |
| 1968 | 43.0 | 111 | 150.0 | 108 | 42.3 | 37.6 |
| 1969 | 42.5 | 110 | 160.5 | 115 | 43.6 | 44.8 |
| 1970 | 48.6 | 126 | 170.0 | 122 | 53.0 | 48.4 |
| 1971 | 40.7 | 105 | 140.6 | 101 | 46.1 | 42.0 |
| 1972 | 15.0 | 39 | 46.0 | 33 | 19.4 | 17.0 |
| 1973 | 42.2 | 109 | 161.5 | 116 | 52.6 | 57.4 |
| 1974 | 36.7 | 95 | 131.7 | 94 | 54.8 | 52.4 |
| Av. | 38.6 | | 139.4 | | 46.5 | 45.3 |

TABLE 16

TROPICAL EASTERLY JET

ZONAL KINETIC ENERGY/UNIT MASS

August East Africa (26°-6°N) Unit: $m^2/sec^2$

| YEAR | TOTAL(900-75mb) | % Of norm. | TEJ(250-75) | % | % tile of India ZKE total | TEJ only |
|---|---|---|---|---|---|---|
| 1965 | 28.8 | 80 | 84.8 | 68 | 45.3 | 33.9 |
| 1966 | 34.6 | 97 | 107.0 | 86 | 44.2 | 34.6 |
| 1967 | 38.7 | 108 | 133.5 | 107 | 58.8 | 58.7 |
| 1968 | 30.4 | 85 | 97.0 | 78 | 37.6 | 31.9 |
| 1969 | 26.6 | 74 | 112.3 | 90 | 40.6 | 43.1 |
| 1970 | 45.8 | 128 | 196.1 | 157 | 48.6 | 54.5 |
| 1971 | 48.1 | 134 | 161.3 | 129 | 61.9 | 52.6 |
| 1972 | 25.3 | 71 | 85.9 | 69 | 43.2 | 39.9 |
| 1973 | 46.6 | 130 | 161.6 | 129 | 67.0 | 62.0 |
| 1974 | 33.0 | 92 | 110.3 | 88 | 42.7 | 39.4 |
| Av. | 35.8 | | 125.0 | | 49.0 | 45.1 |

## TABLE 17

### CORRELATION BETWEEN CIRCULATION AND RAINFALL

| CORRELATION BETWEEN | PERIOD | | CORRELATION |
|---|---|---|---|
| INDIRECT CELL AND | JULY | 1961-74(14 YR) | +0.45 |
| WEST INDIA RAINFALL | AUGUST | 1960-74(15 YR) | +0.69* |
| TEJ ( AT EAST AFRICA) AND | JULY | 1965-74(10 YR) | +0.34 |
| WEST INDIA RAINFALL | AUGUST | 1965-73( 9 YR) | +0.78° |
| TEJ ( AT EAST AFRICA) AND | JULY | 1965-74(10 YR) | -0.19 |
| EAST AFRICA RAINFALL | AUGUST | 1965-73( 9 YR) | +0.82° |
| SEA-LEVEL PRESS. GRAD. | JULY | 1941-74(34 YR) | +0.65* |
| (AT WEST INDIA) AND WEST INDIA RAINFALL | AUGUST | 1941-74(34 YR) | +0.40° |
| WEST INDIA RAIN. AND | JULY | 1965-74(10 YR) | -0.33 |
| EAST AFRICAN RAINFALL | AUGUST | 1965-74(10 YR) | +0.53 |

* Significant at 1 % level.

° Significant at 5 % level.

TABLE 18 

CORRELATION BETWEEN SEA-LEVEL PRESSURE
AND SOUTH CHINA SEA RAINFALL

(24 YR: 1951-1974)

| SEA-LEVEL PRESSURE | PERIOD ( month ) | CORRELATION |
|---|---|---|
| APARRI | AUGUST | -0.91* |
| ISHIGAKI JIMA | JULY | -0.56° |
| | AUGUST | -0.82* |
| MINAMI DAITO JIMA | JULY | -0.52 |
| | AUGUST | -0.63° |

* Significant at 3 $\sigma$

° Significant at 2.6 $\sigma \simeq 1$ % level

TABLE 19

## SEA-LEVEL PRESSURE STATISTICS

### a) Monthly pressure

| STATION | TIME PERIOD YR | | MEAN PRESSURE mb | STANDARD DEVIATION mb |
|---|---|---|---|---|
| ISHIGAKI JIMA | JULY | 1951-74(24 YR) | 1007.3 | 1.77 |
| | AUGUST | 1951-74(24 YR) | 1006.5 | 1.83 |
| MINAMI DAITO JIMA | JULY | 1951-74(24 YR) | 1009.4 | 2.10 |
| | AUGUST | 1951-74(24 YR) | 1008.0 | 2.38 |
| APARRI | AUGUST | 1951-74(24 YR) | 1007.2 | 1.47 |

### b) Seasonal Pressures (June- August)

| STATION | TIME PERIOD | MEAN PRESSURE mb | STANDARD DEVIATION mb |
|---|---|---|---|
| ISHIGAKI JIMA | 1941-74(34 YR) | 1007.3 | 0.96 |
| SILCHAR | 1941-72(32 YR) | 1002.8 | 1.05 |
| SANTIAGO | 1941-74(34 YR) | 1018.7 | 0.80 |
| DARWIN | 1941-74(34 YR) | 1012.7 | 0.72 |
| COLOMBO | 1941-74(34 YR) | 1009.2 | 0.45 |
| HONOLULU | 1941-74(34 YR) | 1016.7 | 0.66 |

TABLE 20

**TOTAL TYPHOON FREQUENCY**

**Period:** July and August 1956-65 and 1967-72

| Sea-level Pressure at Ishigaki Jima | Number of Month | Average Total Monthly Typhoon Frequency |
|---|---|---|
| $P \geq +0.9\sigma$ | 7 | 4.1 |
| $+0.89\sigma \geq P \geq 0$ | 9 | 9.0 |
| $0.0 > P \geq -0.79\sigma$ | 10 | 9.1 |
| $P \leq -0.8\sigma$ | 6 | 17.2 |
| All Month | 32 | 9.5 |

TABLE 21

CORRELATION OF METEOROLOGICAL PARAMETERS WITH SOUTHERN OSCILLATION

( After Walker and Bliss 1932 )

| Parameters | Correlation |
|---|---|
| SANTIAGO PRESSURE | +0.84 |
| HONOLULU PRESSURE | +0.76 |
| INDIA RAIN | +0.76 |
| NILE FLOOD | +0.72 |
| BATAVIA PRESSURE | -0.80 |
| CAIRO PRESSURE | -0.76 |
| MADRAS TEMPERATURE | -0.72 |
| MANILA PRESSURE | +0.66 |
| DARWIN PRESSURE | -0.68 |
| CHILE RAIN | -0.60 |

TABLE 22

TEN YEAR PRESSURE CORRELATION

| PERIOD | DARWIN v.s. SANTIAGO | DARWIN v.s. COLOMBO | DARWIN v.s. HONOLULU | SANTIAGO v.s. COLOMBO | SANTIAGO v.s. HONOLULU | HONOLULU v.s. COLOMBO |
|---|---|---|---|---|---|---|
| 1891-00 | -0.54 | +0.78 | -0.30 | -0.61 | +0.69 | -0.41 |
| 1896-05 | -0.23 | +0.54 | -0.05 | -0.50 | +0.41 | -0.17 |
| 1901-10 | -0.75 | +0.63 | -0.71 | -0.42 | +0.72 | -0.18 |
| 1906-15 | -0.50 | -0.05 | -0.25 | -0.12 | +0.47 | -0.20 |
| 1911-20 | -0.73 | +0.16 | -0.48 | -0.58 | +0.74 | -0.72 |
| 1916-25 | -0.68 | +0.56 | -0.52 | -0.75 | +0.62 | -0.82 |
| 1921-30 | -0.70 | +0.62 | -0.27 | -0.54 | +0.70 | -0.23 |
| 1926-35 | -0.62 | +0.06 | -0.36 | +0.36 | +0.66 | +0.24 |
| 1931-40 | -0.42 | +0.32 | -0.16 | -0.13 | +0.27 | -0.07 |
| 1936-45 | -0.64 | +0.72 | -0.23 | -0.60 | -0.16 | -0.43 |
| 1941-50 | -0.60 | +0.70 | -0.06 | -0.45 | +0.25 | -0.45 |
| 1946-55 | -0.09 | +0.27 | +0.12 | -0.40 | +0.36 | -0.27 |
| 1951-60 | +0.15 | +0.25 | -0.37 | -0.19 | -0.17 | -0.05 |
| 1956-65 | -0.28 | +0.44 | -0.37 | -0.56 | +0.11 | +0.40 |
| 1961-70 | -0.75 | +0.57 | -0.10 | -0.64 | -0.01 | +0.38 |
| 1965-74 | -0.63 | +0.42 | -0.37 | -0.52 | +0.45 | +0.20 |
| 1882-1920 | -0.57* | +0.46* | -0.66* 1883-1920 | -0.65* 1875-1920 | | |
| 1921-50 | -0.34* | +0.49* | -0.12* | -0.21* | | -0.27* |

5 % level = 0.65
1 % level = 0.78

* After Troup (1965)

TABLE 23



TEN YEAR PRESSURE CORRELATION

Correlation with Southern Oscillation Index (defined in text)

| PERIOD | SANTIAGO | HONOLULU | DARWIN | COLOMBO | ISHIGAKI JIMA | SILCHAR |
|---|---|---|---|---|---|---|
| 1891-00 | +0.89 | +0.77 | -0.76 | -0.84 | - | - |
| 1896-05 | +0.79 | +0.50 | -0.65 | -0.83 | - | +0.32 |
| 1901-10 | +0.88 | +0.80 | -0.93 | -0.68 | +0.59 | +0.81 |
| 1906-15 | +0.75 | +0.72 | -0.74 | -0.40 | +0.65 | +0.66 |
| 1911-20 | +0.92 | +0.89 | -0.76 | -0.71 | +0.47 | +0.21 |
| 1916-25 | +0.86 | +0.87 | -0.81 | -0.90 | +0.42 | +0.20 |
| 1921-30 | +0.94 | +0.73 | -0.78 | -0.73 | -0.11 | -0.00 |
| 1926-35 | +0.79 | +0.83 | -0.71 | -0.11 | +0.25 | +0.49 |
| 1931-40 | +0.64 | +0.60 | -0.76 | -0.49 | +0.27 | +0.18 |
| 1936-45 | +0.67 | +0.49 | -0.89 | -0.87 | +0.31 | +0.09 |
| 1941-50 | +0.71 | +0.54 | -0.82 | -0.91 | +0.50 | +0.30 |
| 1946-55 | +0.74 | +0.56 | -0.51 | -0.75 | +0.20 | -0.37 |
| 1951-60 | +0.37 | +0.60 | -0.66 | -0.62 | -0.16 | -0.28 |
| 1956-65 | +0.73 | +0.40 | -0.83 | -0.62 | -0.56 | -0.08 |
| 1961-70 | +0.89 | +0.19 | -0.89 | -0.72 | -0.58 | -0.40 |
| 1965-74 | +0.91 | +0.55 | -0.82 | -0.60 | -0.07 | - |

5 % Level = 0.65

1 % Level = 0.78

TABLE 24

TEN YEAR PRESSURE CORRELATION

Correlation with Ishigaki Jima

| PERIOD | SILCHAR | HONOLULU | SANTIAGO | DARWIN |
|---|---|---|---|---|
| 1891-00 | - | - | - | - |
| 1896-05 | - | - | - | - |
| 1901-10 | +0.63 | +0.63 | +0.50 | -0.58 |
| 1906-15 | +0.83 | +0.63 | +0.40 | -0.27 |
| 1911-20 | +0.76 | +0.36 | +0.46 | -0.27 |
| 1916-25 | +0.46 | +0.18 | +0.33 | -0.46 |
| 1921-30 | +0.31 | -0.47 | -0.23 | -0.08 |
| 1926-35 | +0.32 | +0.11 | +0.09 | -0.41 |
| 1931-40 | +0.68 | +0.20 | +0.58 | +0.05 |
| 1936-45 | +0.72 | +0.25 | +0.33 | +0.02 |
| 1941-50 | +0.65 | +0.67 | +0.17 | -0.23 |
| 1946-55 | +0.58 | -0.01 | +0.03 | -0.55 |
| 1951-60 | +0.63 | +0.02 | -0.70 | -0.51 |
| 1956-65 | +0.33 | -0.19 | -0.84 | +0.06 |
| 1961-70 | +0.33 | +0.05 | -0.70 | +0.47 |
| 1965-74 | - | +0.49 | -0.07 | +0.35 |

5 % level = 0.65

1 % level = 0.78

TABLE 25

TEN YEAR PRESSURE CORRELATION

Correlation with Southern Oscillation Index ( Trenberth )

| PERIOD | S.O.I.T. | ISHIGAKI JIMA | HONOLULU |
|---|---|---|---|
| 1941-50 | +0.76 | +0.38 | +0.22 |
| 1946-55 | +0.59 | +0.54 | +0.06 |
| 1951-60 | +0.76 | +0.29 | +0.45 |
| 1956-65 | +0.85 | -0.36 | +0.33 |
| 1961-70 | +0.81 | -0.62 | -0.02 |
| 1965-74 | +0.90 | -0.12 | +0.61 |

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# 11. APPENDICES

# LIST OF METEOROLOGICAL STATIONS

## RAINFALL DATA

| WMO # | West Africa | | | |
|---|---|---|---|---|
| 61687 | Tambaconda | 13°46'N | 13°41'W | 46M |
| 61290 | Bamako | 12°38'N | 08°02'W | 332M |
| 65510 | Bobo-Dioulasso | 11°10'N | 04°18'W | 467M |
| 65503 | Ouagadougou | 12°21'N | 01°31'W | 304M |
| 65306 | Kandi | 11°08'N | 02°56'E | 292M |
| | **East Africa** | | | |
| 63450 | Addis Ababa | 09°02'N | 38°45'E | 2408M |
| | **West India** | | | |
| 43057 | Bombay | 18°54'N | 72°49'E | 11M |
| 43197 | Belgaum | 15°51'N | 74°32'E | 753M |
| 43283 | Mangalore | 12°52'N | 74°51'E | 22M |
| 43351 | Cochin | 09°58'N | 76°14'W | 3M |
| | **Burma** | | | |
| 48062 | Akyab | 20°08'N | 92°53'E | 5M |
| 48097 | Rangoon | 16°46'N | 96°10'E | 23M |
| 48110 | Mergui | 12°26'N | 98°36'E | 37M |
| | **South China Sea** | | | |
| 45005 | Hong Kong | 22°18'N | 114°10'E | 33M |
| 98282 | Aparri | 18°22'N | 121°38'E | 4M |
| 98429 | Manila | 14°31'N | 121°00'E | 15M |
| 98637 | Iloilo | 10°42'N | 122°34'E | 14M |

## RAINFALL DATA ( continued. )

### East India

| WMO # | | | | |
|---|---|---|---|---|
| 42411 | Gauhati | 26°11'N | 91°45'E | 55M |
| 42404 | Dhubri | 26°01'N | 89°59'E | 35M |
| 42619 | Silchar | 24°09'N | 92°48'E | 29M |
| 42807 | Calcutta ( Alipore) | 22°32'N | 88°20'E | 6M |

### Central India

| | | | | |
|---|---|---|---|---|
| 42361 | Agra | 27°10'N | 78°02'E | 169M |
| 42475 | Allahabad | 25°27'N | 81°44'E | 98M |
| 42671 | Sagar | 23°51'N | 78°05'E | 551M |
| 42867 | Nagpur | 21°06'N | 79°03'E | 310M |
| 43041 | Jagdalpur | 19°05'N | 82°02'E | 553M |

### Tongking Gulf

| | | | |
|---|---|---|---|
| Peihai | 21°28'N | 109°45'E | 4M |
| Laokay | 22°30'N | 103°57'E | 2M |
| Phu-lien | 20°48'N | 106°37'E | 115M |

### Length of record

| | | |
|---|---|---|
| West Africa | | 1941-74 |
| East Africa | 1902-39 | 1946-74 |
| West India | | 1901-74 |
| Burma | 1901-40 | 1951-74 |
| South China Sea | 1903-40 | 1947-74 |
| East India | | 1901-74 |
| Central India | | 1941-74 |
| Tongking Gulf | 1906-38 | |

# APPENDIX A

## SEA LEVEL PRESSURE DATA ( June-August )

| WMO # | | | | |
|---|---|---|---|---|
| 85577 | Santiago | 33°27'S | 70°42'W | 520M |
| 91182 | Honolulu | 21°19'N | 157°52'W | 12M |
| 94120 | Darwin | 12°28'S | 130°51'E | 27M |
| 43466 | Colombo | 06°54'N | 79°52'E | 7M |
| 47918 | Ishigaki Jima | 24°20'N | 124°10'E | 7M |
| 42619 | Silchar | 24°49'N | 92°48'E | 29M |
| 93938 | Tahiti ( Papeete ) | 17°32'S | 149°34'W | 3M |
| 43371 | Trivandrum | 08°29'N | 76°57'E | 64M |
| 42909 | Veraval | 20°54'N | 70°22'E | 8M |

## Length of record

All stations are 1891-1974 except:

| | |
|---|---|
| Ishigaki Jima | 1897-1974 ( 1950 missing) |
| Silchar | 1893-1972 |
| Tahiti | 1941-1974 (1961 missing) |
| Trivandrum | 1941-1974 |
| Veraval | 1941-1974 |

APPENDIX A



Upper Air Data

1) Indian Cell, 70° -80°E

| | | | |
|---|---|---|---|
| Jodhpur | 26°18'N | 73°01'E | 224M |
| Ahmadabad | 23°04'N | 72°38'E | 55M |
| Bombay | 19°07'N | 72°51'E | 14M |
| Madras | 13°00'N | 80°11'E | 16M |
| Trivandrum | 08°29'N | 76°57'E | 64M |
| Gan | 00°41'S | 73°10'E | 2M |

2) East Africa Cell, 30°-40°E

| | | | |
|---|---|---|---|
| Helwan | 29°52'N | 31°20'E | 141M |
| Aswan | 23°58'N | 32°47'E | 194M |
| Port Sudan | 19°35'N | 37°13'E | 2M |
| Khartoum | 15°36'N | 32°33'E | 380M |
| Malakal | 09°33'N | 31°39'E | 388M |
| Entebbe | 00°03'N | 32°27'E | 1146M |

3) West Africa Cell, 5°W-5°E

| | | | |
|---|---|---|---|
| Bechar | 31°38'N | 02°15'W | 806M |
| In-salah | 27°12'N | 02°28'E | 243M |
| Tamanrasset | 22°47'N | 05°31'E | 1378M |
| Niamy | 13°29'N | 03°56'E | 7M |

4) East Asian Monsoon Trough, 100°-135°E

| | | | |
|---|---|---|---|
| Yonago | 35°26'N | 133°21'E | 8M |
| Fukuoka | 33°35'N | 130°23'E | 14M |
| Kagoshima | 31°38'N | 130°35'E | 283M |
| Naze | 28°23'N | 129°33'E | 295M |
| Taipei | 25°02'N | 121°31'E | 9M |
| Hông kong | 22°19'N | 144°10'E | 66M |
| Da-Nang | 16°02'N | 108°11'E | 7M |
| Saigon | 10°49'N | 106°40'E | 19M |
| Songkhla | 09°11'N | 100°37'E | 10M |
| Penang | 05°18'N | 100°16'E | 4M |
| Singapore | 01°22'N | 103°55'E | 32M |

See the end of the references for the sources of data.

## APPENDIX B

### Student t-test

Application to the problem of the significance of the relation, between typhoon frequency and the strength of East Asian Monsoon trough.

Let P be the normalized sea-level pressure in Ishigaki Jima. Of 32 months there are six months, and seven months during which $P < -0.8\sigma$ and $P > +0.9\sigma$, respectively. The total monthly typhoon frequencies, $x_1$, $x_2$, during these months are:

$x_1$ = 11,30,13,16.and 20 (with $P < -0.8$ )

$x_2$ = 2,4,4,6,3,2, and 8 (with $P > +0.9$ )

Do the months with $P < -0.8$ have significantly higher total monthly typhoon frequency than the months with $P > +0.9$ ?

In this problem:

$N_1 = 6$ $N_2 = 7$ ( numbers of samples )

$\bar{x}_1 = 17.2$ $\bar{x}_2 = 4.1$ ( sample mean frequency )

$S_1^2 = 41.14$ $S_2^2 = 4.12$ ( square of sample standard deviation )

$m = N_1 + N_2 - 2 = 11$ ( number of degrees of freedom )

Thus assuming that the difference of the population means is zero by null hypothesis, one can define.

$$\mu = \frac{\bar{X}_1 - \bar{X}_2}{\left(\frac{\sigma_1^2}{N_1} + \frac{\sigma_2^2}{N_2}\right)^{1/2}}$$

where $\sigma_1, \sigma_2$ are population standard deviations.

Assume $\sigma_1^2 \cong \sigma_2^2$ ( population standard deviations are equal )

then one can define a parameter.

$$\nu = \frac{N_1 S_1^2}{\sigma_1^2} + \frac{N_2 S_2^2}{\sigma_2^2}$$

Student t-test ( continued )

thus t-value can be obtained by

$$t = \frac{\mu}{v} m^{1/2} = \frac{(\overline{X}_1 - \overline{X}_2) \times m^{1/2}}{\left[\left(\frac{N_1 + N_2}{N_1 N_2}\right)\left(N_1 S_1^2 + N_2 S_2^2\right)\right]^{1/2}}$$

Substituting the given values will yield.

t = 4.699

for 11 degrees of freedom, table of t-values show.

5 % level = 2.201   1 % level = 3.106

Thus the relationship in this problem is statistically significant.

## APPENDIX C

### Correlation coeficient

In this study correlation coeficient is obtained by the following methods:

let $X_1$ = first variables of sample size $N_1$.

$X_2$ = second variables of sample size $N_2$.

$N_1 = N_2 = N$ both variables have the same sample size.

$\sum$ = sum of all samples

then; following quantities are calculated.

Variance

$$VAR = \frac{1}{N^2}\left[N \Sigma X^2 - (\Sigma X)^2\right]$$

Standard deviation

$$\sigma = (VAR)^{1/2}$$

Covariance

$$COV = \frac{1}{N^2}\left[N \Sigma X_1 X_2 - \Sigma X_1 \Sigma X_2\right]$$

Correlation coeficient

$$COR = \frac{COV}{\sigma_{X_1}\,\sigma_{X_2}}$$

The significant level of this correlation is obtained by following formula ( as a reference ).

Assume $X_1, X_2$, possess a normal frequency function, then

$$Z = \frac{1}{2}\ln\left(\frac{1 + COR}{1 - COR}\right)$$

the standard deviation of Z is given by $\sigma_Z = (N - 3)^{-1/2}$ where N = number of samples.

if $Z > 2\sigma_Z$ then COR is significant at 5 % level.
if $Z > 2.6\sigma_Z$ then COR is significant at 1 % level.