## 2. CLIMATOLOGY OF THE INDIAN SUMMER MONSOON

### 2.1 General

Attempts have been made to present the climatology of the monsoon from time-to-time. Although the net-work of observatories, both surface and upper air, for the Indian region has been satisfactory, that for the neighbourhood is poor. In particular, the vast Indian Ocean has remained unexplored creating a big data-gap. Therefore, the analysis so far found in the literature is confined to the region barring Indian Ocean. However, the monsoon over India is not an isolated system restricting its activity over the Indian subcontinent only. It is a final product of a chain of events in which the Indian Ocean plays a prominent role. Therefore, in order to assess as to what extent the weather systems over the Indian Ocean (south of the equator) influence the monsoon activity over India, a knowledge of winds and temperature, both north and south of the equator, is essential. Also, such knowledge is considered essential in view of the recent controversy about the origin of monsoon. Whereas one group (Pisharoty, 1965; Rama, 1966) opines that it originates over north of the equator, the other group (Rao, 1964; Desai, 1968; Godbole et al, 1970; Ramanadham et al, 1971) strongly believes in the classical theory of the south Indian Ocean as the seat of origin of the monsoon. Koteswaram (1969) has presented wind and temperature

analyses south of the equator upto 20° S but these are only in a vertical cross-section along 80°E. A knowledge of the spatial distribution of winds and temperature at the surface and also at higher levels is necessary for examining the monsoon climatology. After the International Indian Ocean Expedition ( 1963-64 ), routine meteorological data started flowing in on a systematic basis from a few Island stations in the Indian Ocean and this has, to some extent, helped in reducing the data-gap over the region. An attempt is, therefore, made in the following to present the distribution of winds temperature, and mixing ratio on a comprehensive basis.

## 2.2 Data

Values of temperature, dew point, wind speed and wind direction at the surface and at 850, 700, 500, 300, 200 and 100 mb (dew point upto 500 mb only) have been considered for the stations shown in table 2.2.1 for a period of 12 years (1958-69) both for 00 GMT and 12 GMT for the month of July - a typical monsoon month. The location of each station (latitude-longitude) and the number of years for which data are available for these stations are mentioned in the table.

The source of surface and upper air data was mainly the Monthly Climate Data for the World published by ESSA of United States ( 1968-69 ). This was supplemented by the International Indian Ocean Expedition Data

Table 2.2.1 : Radiosonde stations used as source of data for analyses

| Sr. No. | Stations | Lat. deg. | Long. deg. | Years |
|---|---|---|---|---|
| 1 | Aden (Arabia) | 13 N | 45 E | 12 |
| 2 | Ahmedabad (India) | 13 N | 73 E | 5 |
| 3 | Alice Spring (Australia) | 24 S | 134 E | 12 |
| 4 | Allahabad (India) | 25 N | 82 E | 13 |
| 5 | Alma-Ata (U.S.S.R) | 43 N | 77 E | 10 |
| 6 | Baharin (Arabia) | 26 N | 51 E | 12 |
| 7 | Bangalore (India) | 13 N | 78 E | 5 |
| 8 | Bankok (Thailand) | 14 N | 100 E | 10 |
| 9 | Bombay (India) | 19 N | 73 E | 12 |
| 10 | Broken Hill (Zambia) | 14 N | 28 E | 9 |
| 11 | Calcutta (India) | 23 N | 88 E | 14 |
| 12 | Carnarvon (Australia) | 23 S | 114 E | 8 |
| 13 | Cocas Island (Australia) | 12 S | 97 E | 8 |
| 14 | Darban (South Africa) | 30 S | 31 E | 11 |
| 15 | Darwin (Australia) | 12 S | 131 E | 12 |
| 16 | Diego Garcia (Indian Ocean) | 7 S | 72 E | 8 |
| 17 | Entebbe (Uganda) | 0 N | 32 E | 8 |
| 18 | Forrest (Australia) | 31 S | 128 E | 12 |
| 19 | Gan Island (Indian Ocean) | 1 S | 73 E | 10 |
| 20 | Gauhati (India) | 26 N | 92 E | 11 |
| 21 | Gildford (Australia) | 32 S | 116 E | 11 |
| 22 | Giles (Australia) | 25 S | 128 E | 12 |
| 23 | Guryev ( U S S R) | 47 N | 51 E | 9 |
| 24 | Jodhpur (India) | 26 N | 73 E | 10 |
| 25 | Kalgoorlie (Australia) | 31 S | 131 E | 12 |
| 26 | Khartoum (Sudan) | 16 N | 33 E | 12 |
| 27 | Lourenco-Marques(S.Africa) | 26 S | 33 E | 12 |
| 28 | Madras (India) | 13 N | 80 E | 15 |
| 29 | Minicoy (India) | 8 N | 73 E | 3 |
| 30 | Nagpur (India) | 21 N | 79 E | 13 |
| 31 | New Delhi (India) | 29 N | 77 E | 16 |
| 32 | Nicosia (Cyprus) | 35 N | 33 E | 8 |
| 33 | Odessa ( U S S R ) | 46 N | 31 E | 10 |
| 34 | Port Hedland (Australia) | 20 S | 119 E | 12 |
| 35 | Port Blair (India) | 12 N | 93 E | 12 |
| 36 | Pretoria (South Africa) | 26 N | 28 E | 12 |
| 37 | Saigon (Viet Nam) | 11 N | 107 E | 12 |
| 38 | Singapore (Malaya) | 1 N | 104 E | 10 |
| 39 | Srinagar (India) | 34 N | 75 E | 4 |
| 40 | Tanavariev (Malagasy) | 19 S | 47 E | 11 |
| 41 | Taskent (U S S R) | 40 N | 69 E | 10 |
| 42 | Tbilisi (U S S R) | 42 N | 45 E | 10 |
| 43 | Trivandrum (India) | 8 N | 77 E | 10 |
| 44 | Vacos (Maritius) | 20 S | 57 E | 10 |
| 45 | Vishakhapatnam (India) | 18 N | 83 E | 6 |

published by IMC/Hawaii University (1963-64) and also by the overseas data published by the Royal Meteorological Society (1958-1969). Sea surface data have been obtained from the Deutshes Hydrographisches Institute (1960). For the Indian Stations, normals available from the publication by I.Met. Deptt. (1970) have been used. For the remaining stations, normals have been prepared and the values so obtained have been made use of in the study.

## 2.3 Analyses

While analysing the zonal (u) and meridional (v) wind components normals of PIBAL winds over India (I.Met. Deptt., 1970) and also the analyses presented by Frost et al (1963), and Thompson (1965) have been used as guidelines. In doing temperature analyses, the results obtained by the earlier workers, namely, Ramakrishnan et al (1960), Muench (1962), Ananthakrishnan et al (1963), Rao et al (1968) and Rex (1969) have been kept in view to ensure consistency. For examining the moisture field the usual practice is to analyse dew point temperature (Rex, 1969). However, since the dew point temperature is normally reported only upto about 500 mb, its analysis becomes restricted to the lower troposphere from surface to 500 mb. Besides, the dew point temperature, being an indirect measure of the water vapour content, always necessitates transformation into a more direct form

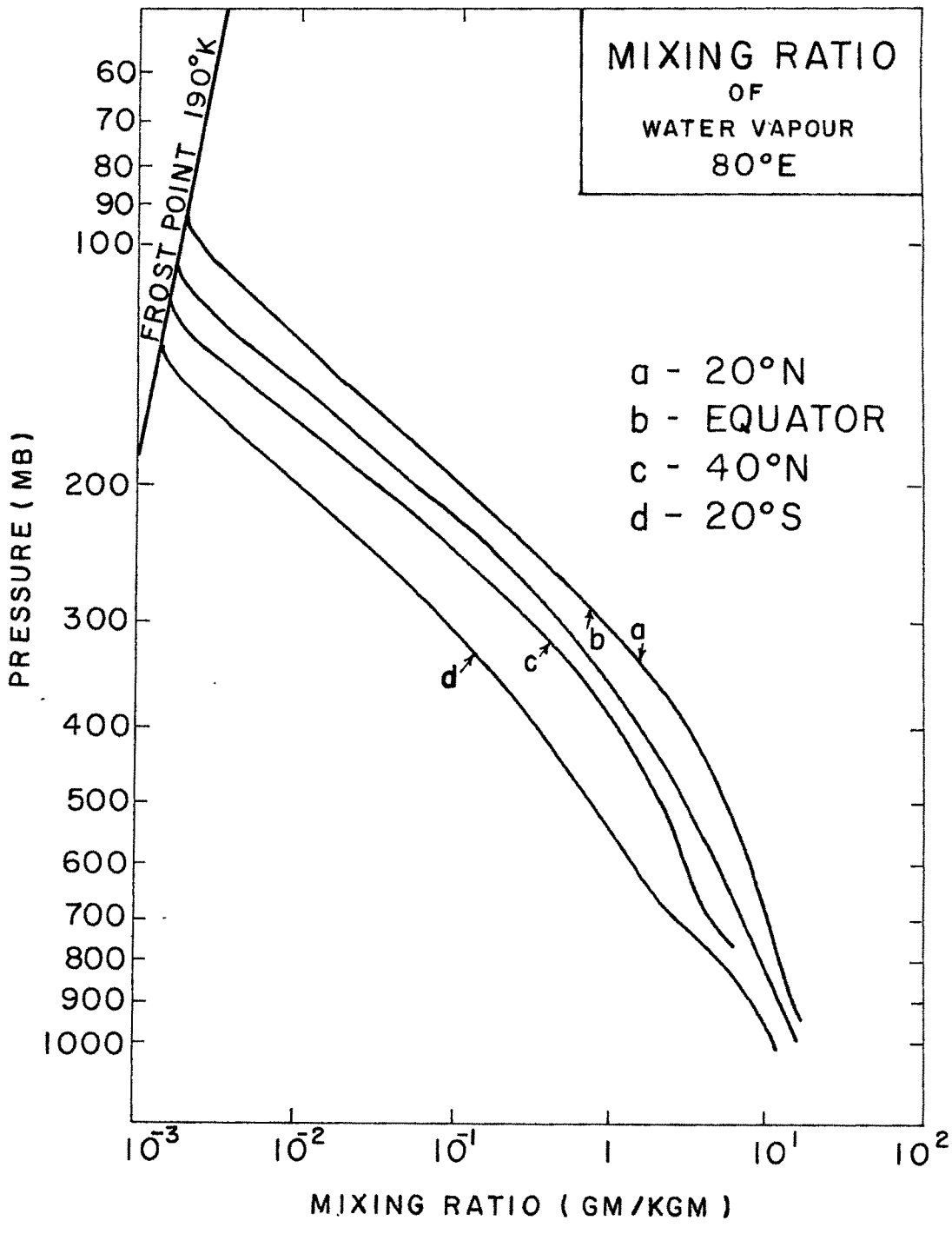



Figure 2.3.1 Distribution of mixing ratio (gm/kgm) with height for latitudes (a) 20°N, (b) equator, (c) 40°N and (d) 20°S along 80°E.

(viz. mixing ratio) for further studies on water vapour content. In the present paper, the mixing ratio of water vapour instead of the dew point has, therefore, been analysed extending to all the levels from surface upto 100 mb. To the best of author's knowledge, this is the first attempt of its kind to present spatial distribution of water vapour upto 100 mb for the region under consideration. The values of mixing ratio above 500 mb were obtained by extrapolation until a value corresponding to a frost point of 190°K is reached and thereafter, for further higher levels, the mixing ratio is assumed to follow a constant frost point of 190°K, irrespective of latitude ( Godbole et al, 1970 ). The extrapolation is done by following the trend of the variation of mixing ratio with height from surface upto 500 mb. As an example, the distribution of mixing ratio with height for different latitudes along 80° E is shown in Figure 2.3.1. The level at which the mixing ratio begins to increase with height (corresponding to a frost point of 190° K) marks the level of tropopause. At 20°N, the mixing ratio remains maximum throughout the troposphere. Also seen from the figure,is the fact that the tropopause height is higher for uniformly higher values of mixing ratio in the vertical column. For example, 100-mb level is already within the stratosphere for lower values of mixing ratio noticed as at latitudes 20°S, equator, and 40°N but it is still within the troposphere for 20°N,

where the values of the mixing ratio are relatively high.

## 2.4 Results and Discussion

The results of the analyses are presented for the region from 20°S to 40° N and from 50° E to 100°E in order to have conformity with the region considered for the numerical simulation of the monsoon.

### 2.4.1 Surface features

The distribution of zonal wind is shown in Figure 2.4.1 A. A belt of westerlies (non-hatched ) is present between equator and about 25°N. The winds are stronger over the sea than over the land. Further,they are stronger over the Arabian Sea than over the Bay of Bengal. The western parts of both the Arabian sea and the Bay of Bengal are subjected to convergence (Zonal) whereas the eastern parts are subjected to divergence (Zonal). The easterlies (hatched) south of equator are due to the subtropical anticyclone over the Indian Ocean. The easterlies north of 25°N are due to two factors. The weak easterlies just south of the Himalayas are essentially due to the monsoon trough whereas those over Afganistan are due to the monsoon low which overlies Arabia,Afganistan and West Pakistan (see figure 1.1.2 A).

The meridional component of wind at the surface is shown in Figure 2.4.1 B. Southerlies (non-hatched )

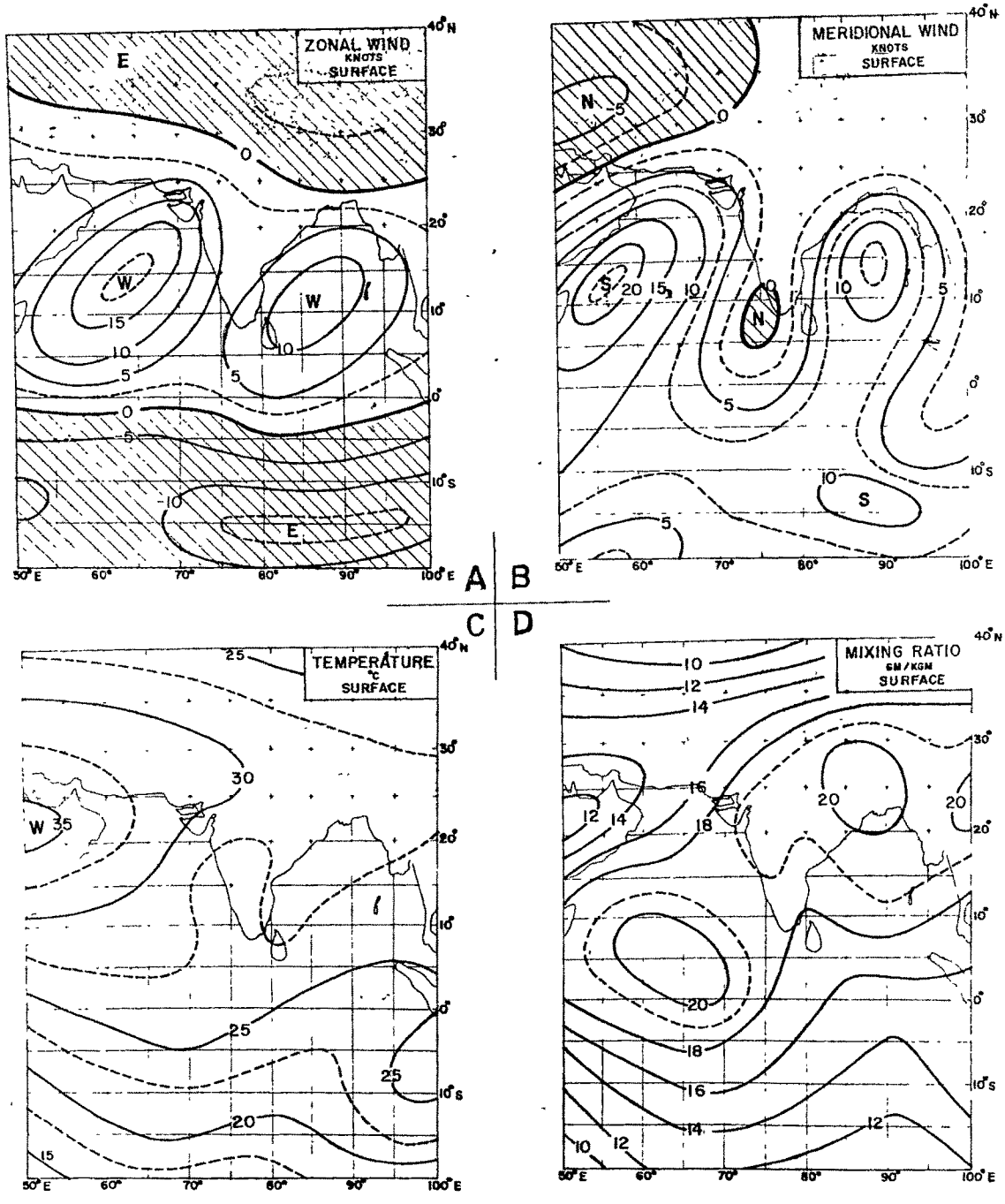



Figure 2.4.1 Spatial distribution of mean (A) zonal wind (kt), (B) meridional wind (kt), (C) temperature ( C), and (D) mixing ratio (gm/kgm) at the surface for July. Regions of easterly (E) and northerly (N) are hatched. Warmest regions are indicated by (W).

prevail over most of the region except over the northwest (i.e. Afganistan) where the northerlies (hatched) prevail. Weak northerlies are also observed over the Kerala coast. The southerly winds are strongest over the south-west Arabian Sea. The structure of zonal and meridional winds when considered together reveal that the axis of the strongest winds is oriented in the southwest-northeast direction off the Somalia Coast. This feature has been reported by Findlater (1969).

The configuration of isotherms at the surface is shown in Figure 2.4.1 C, without reducing the surface temperatures to sea-level as is customarily done (Rao et al, 1968). In this connection it is considered that the surface temperature as observed and not reduced to sea-level is the real temperature at which the earth radiates its energy. Hence it is more realistic to deal with the observed surface temperature rather than with the so-called reduced surface temperature. High temperatures are found over the desert areas in the northwest. The Arabian sea is, in general, warmer than the Bay of Bengal. This is due to less clouding over the Arabian sea than over the Bay of Bengal (see Figure 3.2.1). Over India, the minimum temperature is located over the west peninsular region. This feature is due to deep and extensive clouding observed over the region ( Godbole et al, 1969). The higher temperatures over southeast India and the adjoining Bay region are due to Fohn effect and

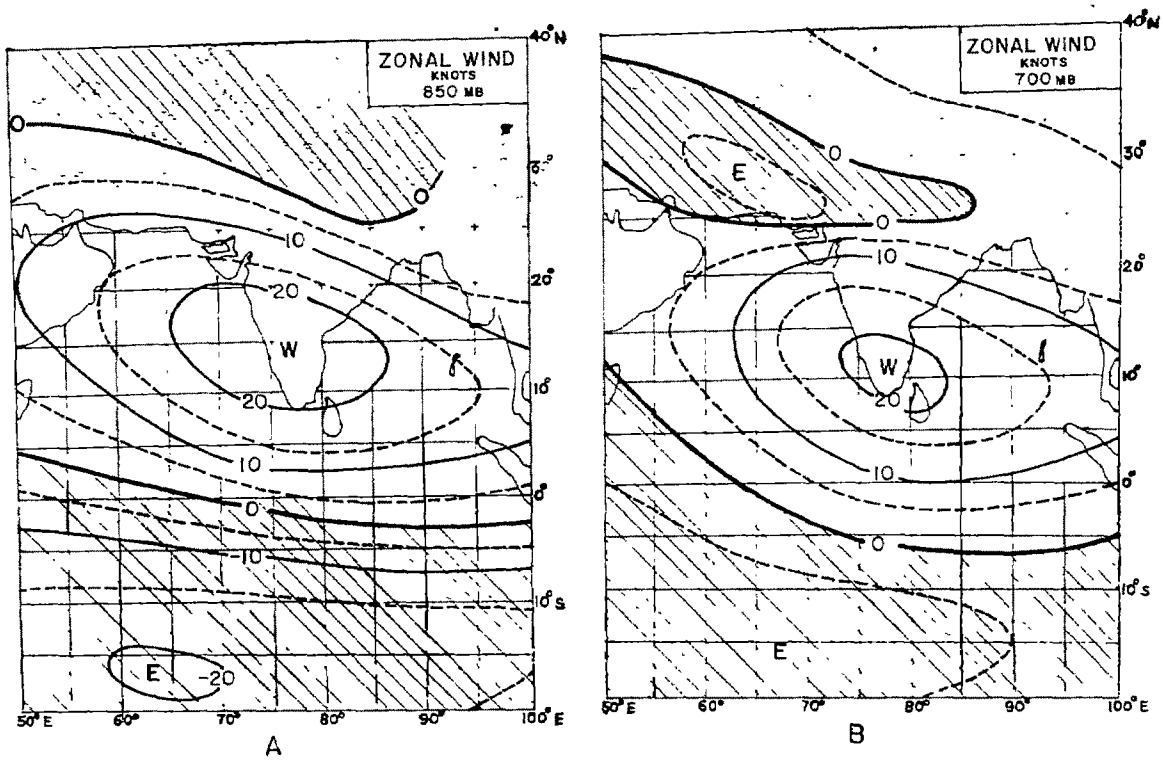



Figure 2.4.2 Spatial distribution of mean zonal wind (kt) at 850, 700, 500, 300, 200, 100 mb for July. Regions of easterly (E) are hatched.

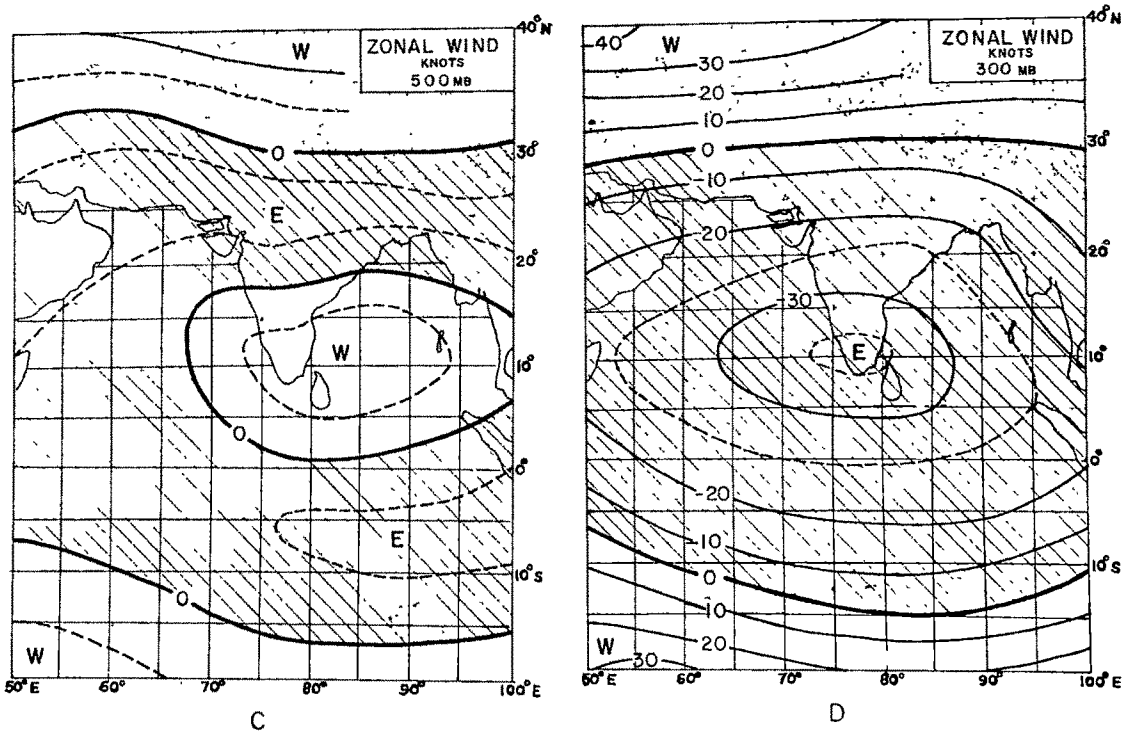
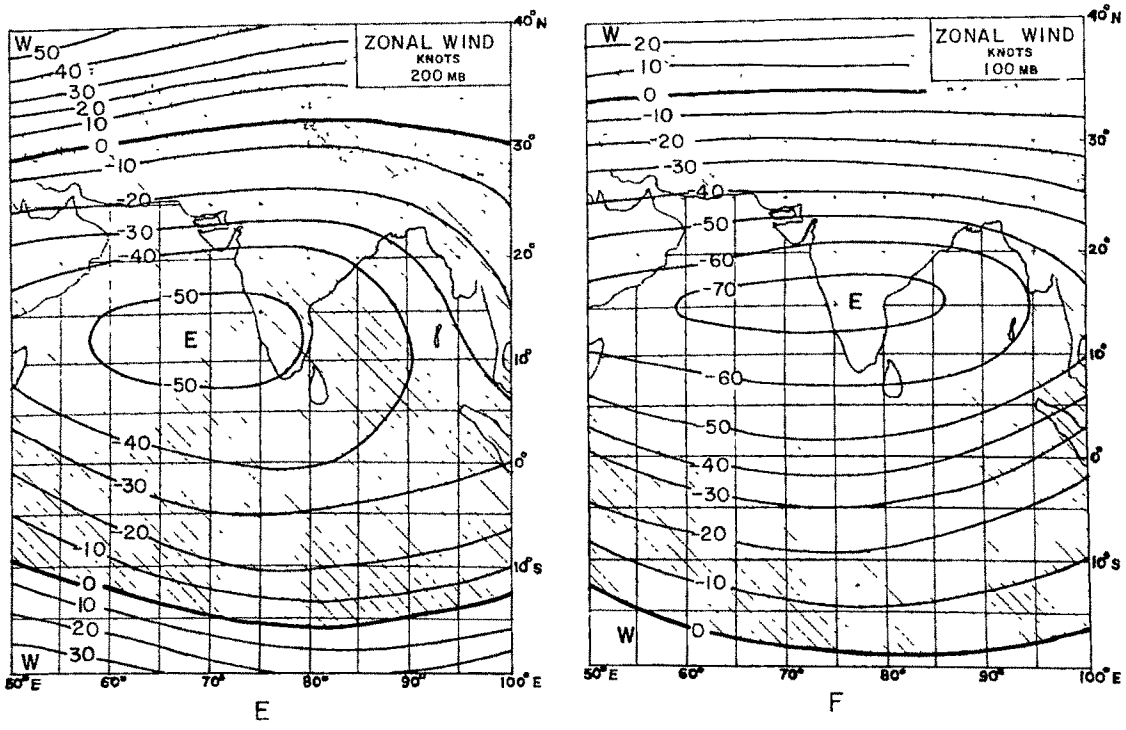



Figure 2.4.2 continued.




E F

Figure 2.4.2 continued.

comparatively less clouding. It may be interesting to note that the analysis presented by Rao et al (1968) for the reduced surface temperature field (to sea level)does not differ widely from what is shown in Figure 2.4.1 C.

In Figure 2.4.1 D, is shown the distribution of mixing ratio at the surface. High values ( > 20 gm/kgm) are found over northeast India and also over Burma. High values are also noticed over the West Indian Ocean. The relatively dry areas are over Arabia and north Afganistan. One interesting point to note here is that the region showing highest values of mixing ratio are seen aligned with the monsoonal sweep over India and Burma.

2.4.2 Upper air features

The distribution of zonal winds at 850, 700, 500, 300, 200 and 100 mb is shown in Figure 2.4.2. Comparing it with Figure 2.4.1 A, it is seen that winds (both easterlies and westerlies ) over land, gain considerably in strength with height from the ground upto 850 mb (from 5 kt at surface to 20 kt at 850 mb, Figure 2.4.2 A). With further increase in height the monsoon westerlies decrease rapidly with height and ultimately reverse in sign above 500 mb. At 500 mb (Figure 2.4.2 C), the wind field is weak. The monsoon trough (not shown in the figure) which maintains

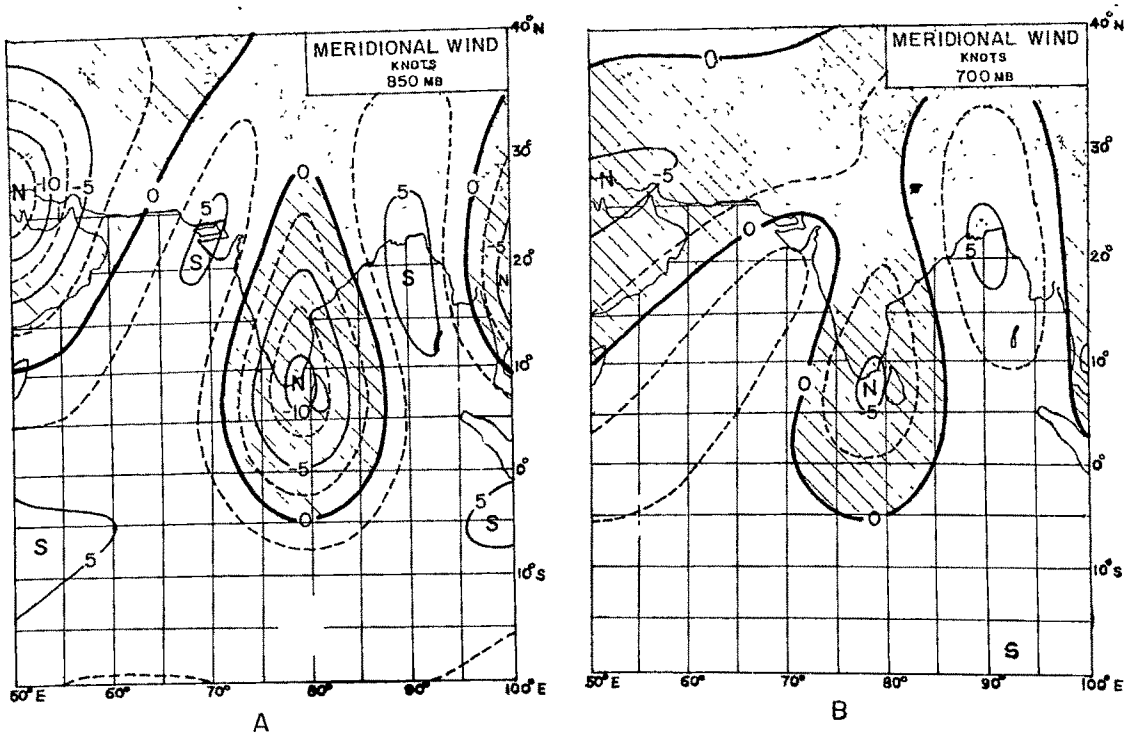
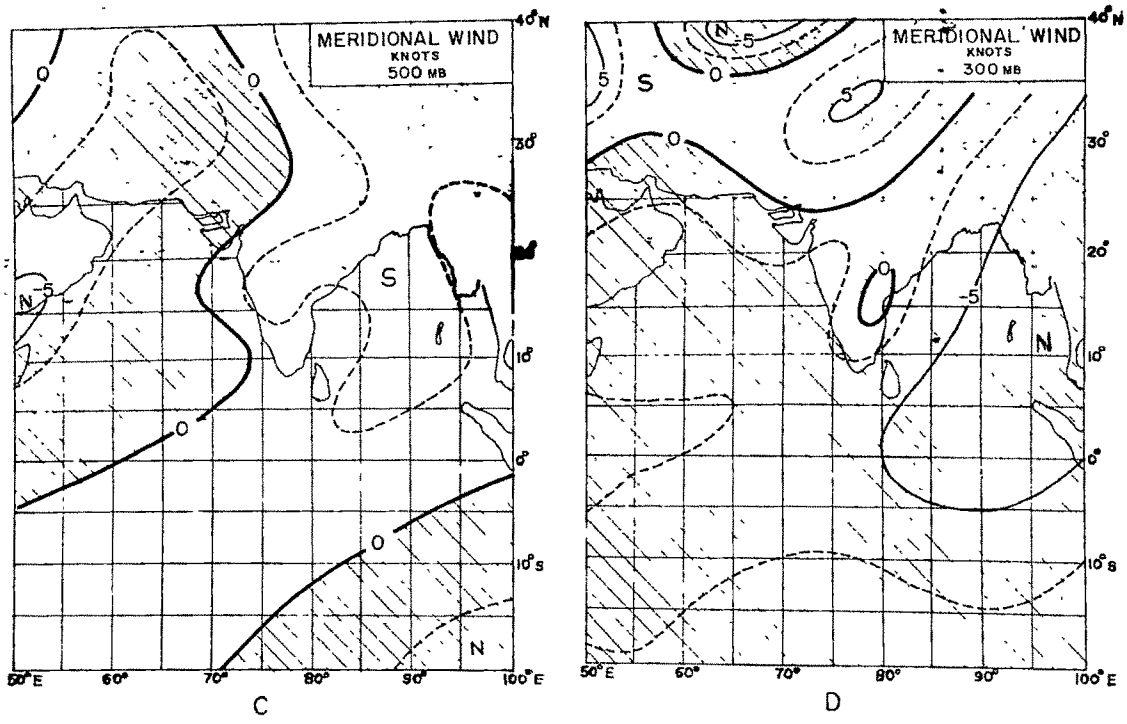



Figure 2.4.3 Spatial distribution of mean meridional wind (kt) at 850, 700, 500, 300, 200, and 100 mb for July. Regions of northerly (N) are hatched.

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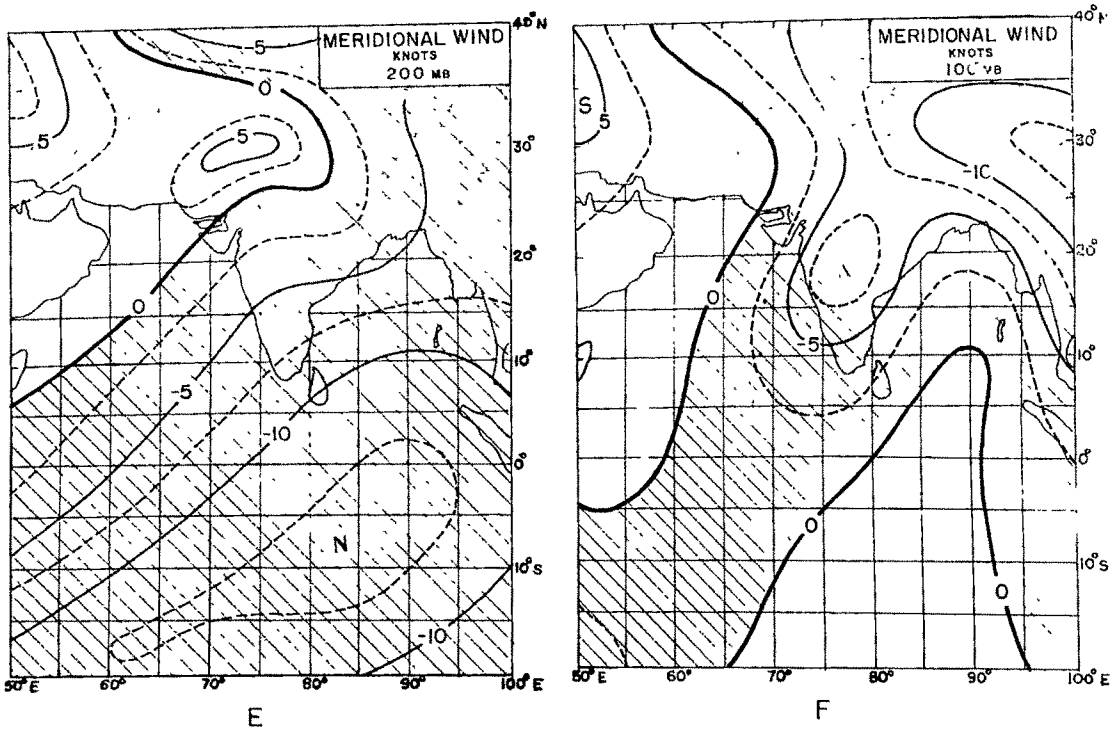



Figure 2.4.3 continued.

easterlies to its north shifts southward with height and becomes diffuse at 500 mb. But the upper level sub-tropical anticyclone which begins to exert its influence from 500 mb upward strengthen the easterlies with height, above 500 mb. Between 200 and 100 mb (Figures 2.4.2 E,F) and south of 25°N, very strong winds with speeds of 70 kt and more develop into what is called the Tropical Easterly Jet (TEJ). The development of TEJ is the effect of the southward decrease of temperature over the region in the troposphere where the thermal gradient is maximum in the northsouth direction. On account of the distribution of isotherms at different levels as seen in Figure 2.4.4 the easterlies strengthen with height. They are capable of reversing the moderately strong westerlies of lower troposphere (surface to 700 mb) and building up speeds of more than 70 kt at 100 mb (Koteswaram, 1958).

The meridional wind field from 850 to 100 mb, is shown in Figure 2.4.3. Unlike zonal winds, the meridional winds are, in general, weak and alternate in sign frequently at any given level. The speeds are practically one order of magnitude smaller than those of zonal winds especially at higher levels where the general wind flow is predominently zonal in character.

Comparision of Figure 2.4.3 with Figure 2.3.1 B reveals that the strongest southerly flow over the south-west Arabian sea off the Somalia coast at the surface is

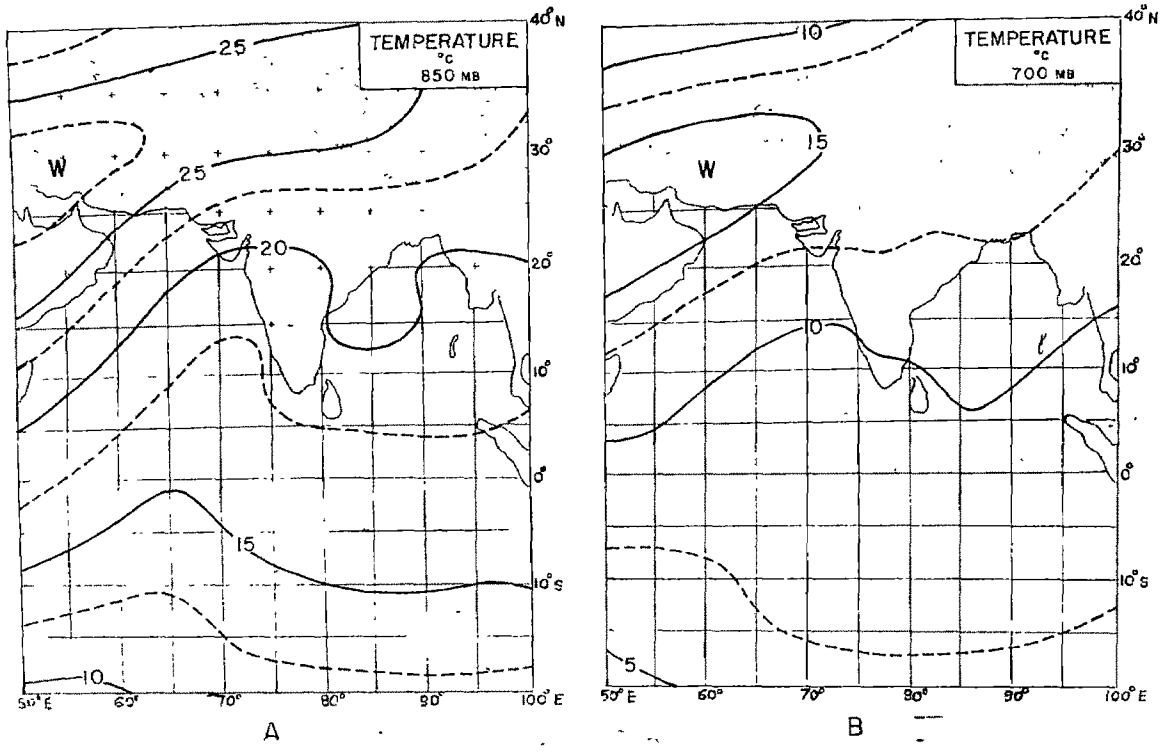
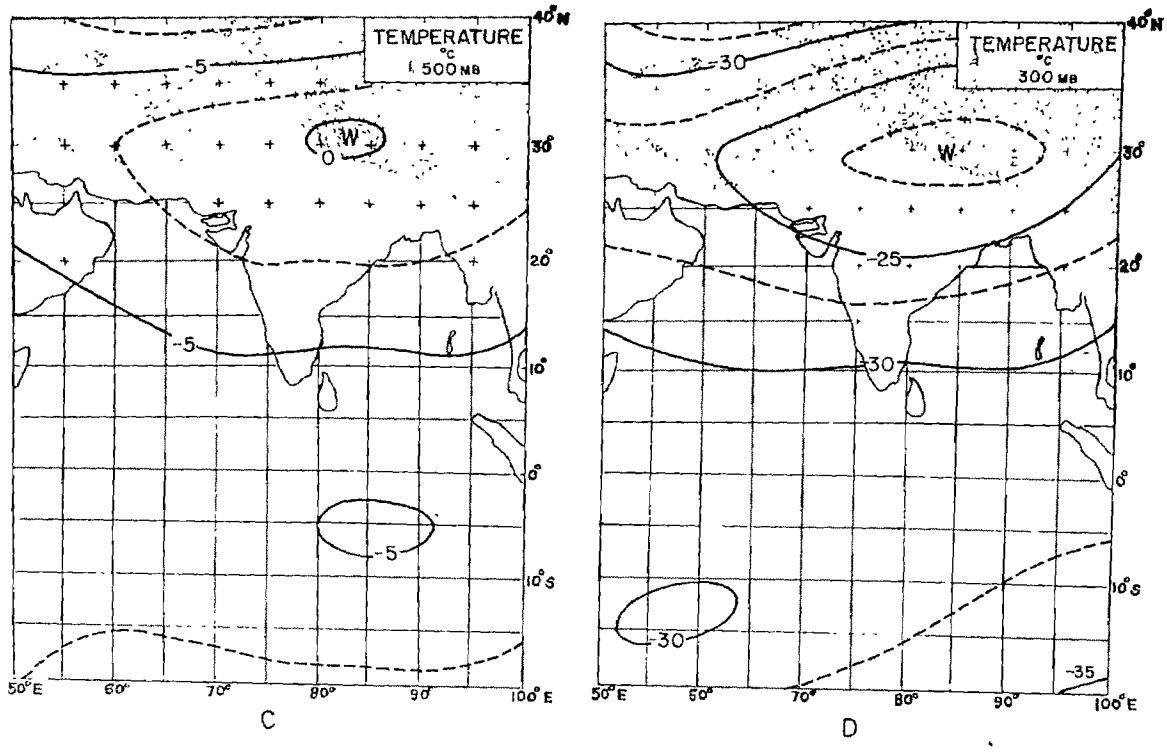



Figure 2.4.4 Spatial distribution of mean temperature (°C) at 850, 700, 500, 300, 300, 200, and 100 mb for July

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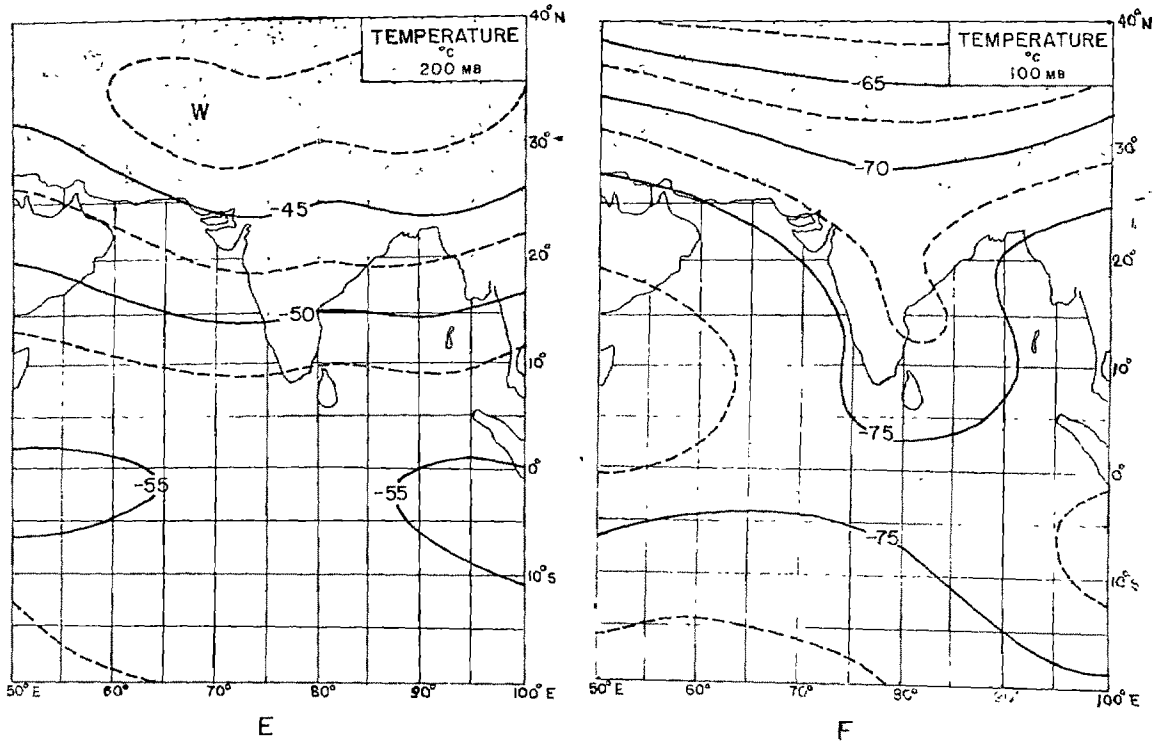



Figure 2.4.4 continued.

confined to a rather shallow layer because over the same region at 850 mb (Figure 2.4.3 A), the winds are very weak. On the other hand, over the other regions, both northerly and southerly winds increase in strength with height, reach a maximum at 850 mb and decrease in strength with further increase in height. Of special interest is the region of northerly winds at 850 mb which extends from 5°S to about 25°N in a rather narrow belt centered along 80°E. The depth of these northerlies is from surface to below 500 mb (Figures 2.4.3 B, C). Northerlies extending to 12° N have also been reported by Rao et al (1968). Since these northerlies are associated with the west wind (Figure 2.4.2), they can not represent the return current of the monsoon. They arise merely due to the pressure configuration over the south peninsular region (see Figure 1.2.1). At 500 mb, the wind field is weak. With further increase in height, the northerlies occupy nearly the whole of the region and also build up in strength. Above 200 mb, they become weak and their extent becomes less. The northerlies in association with easterlies (Figure 2.4.2) at 300 mb and above may be considered to represent the return current of the monsoon (Subbaramayya et al, 1966).

In Figure 2.4.4, is shown the distribution of temperature from 850 upto 100 mb. At 850 mb (Figure 2.4.4 A), a thermal ridge is located between 30°N and 40°N. The peninsular India continues to maintain minimum

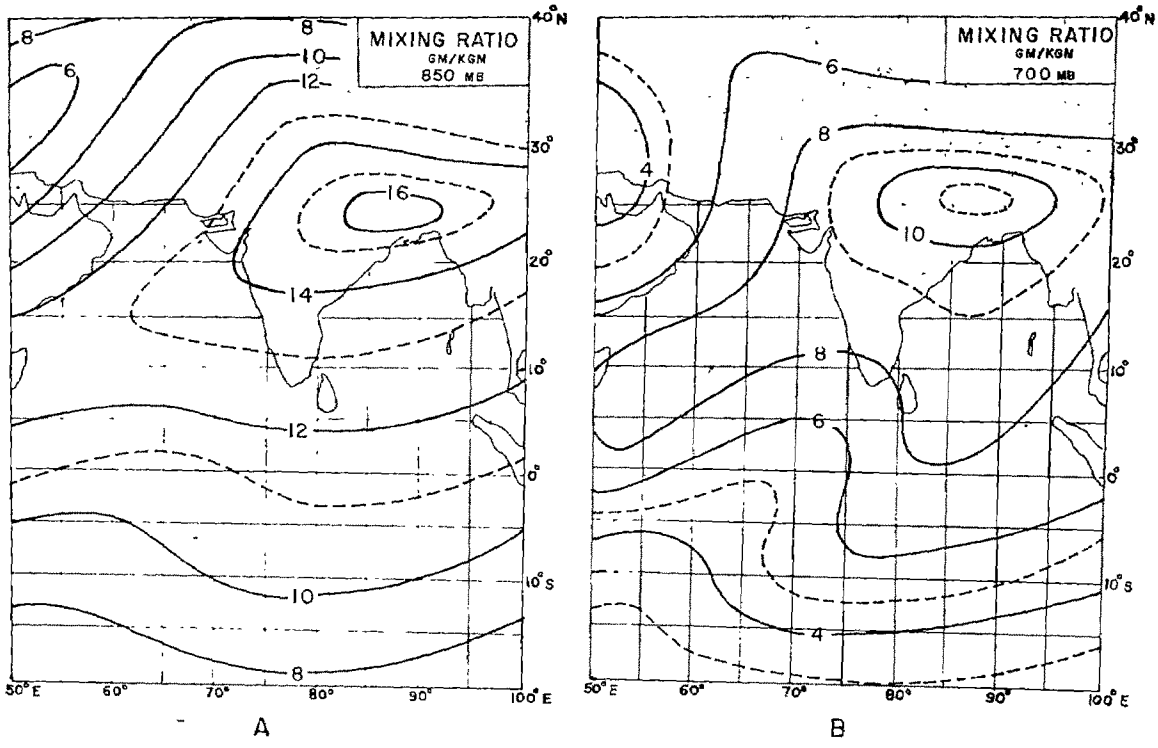
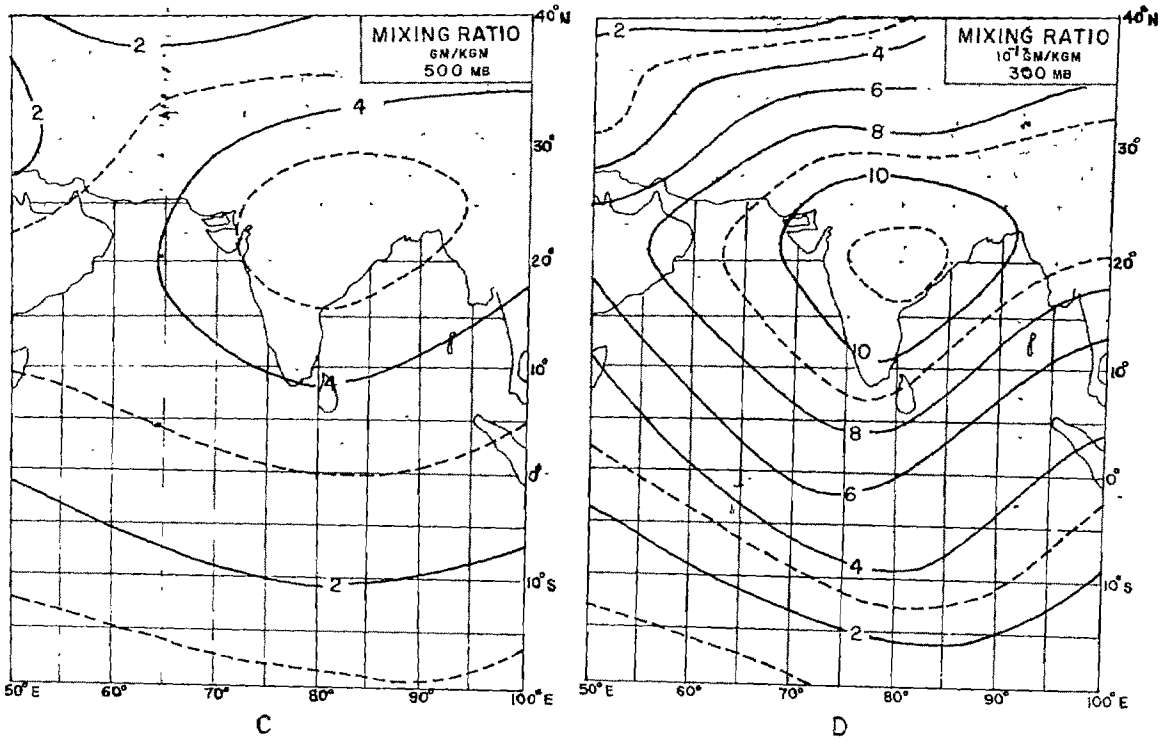



Figure 2.4.5 Spatial distribution of mean mixing ratio (gm/kgm) at 850, 700, 500, 300, 200, and 100 mb for July.

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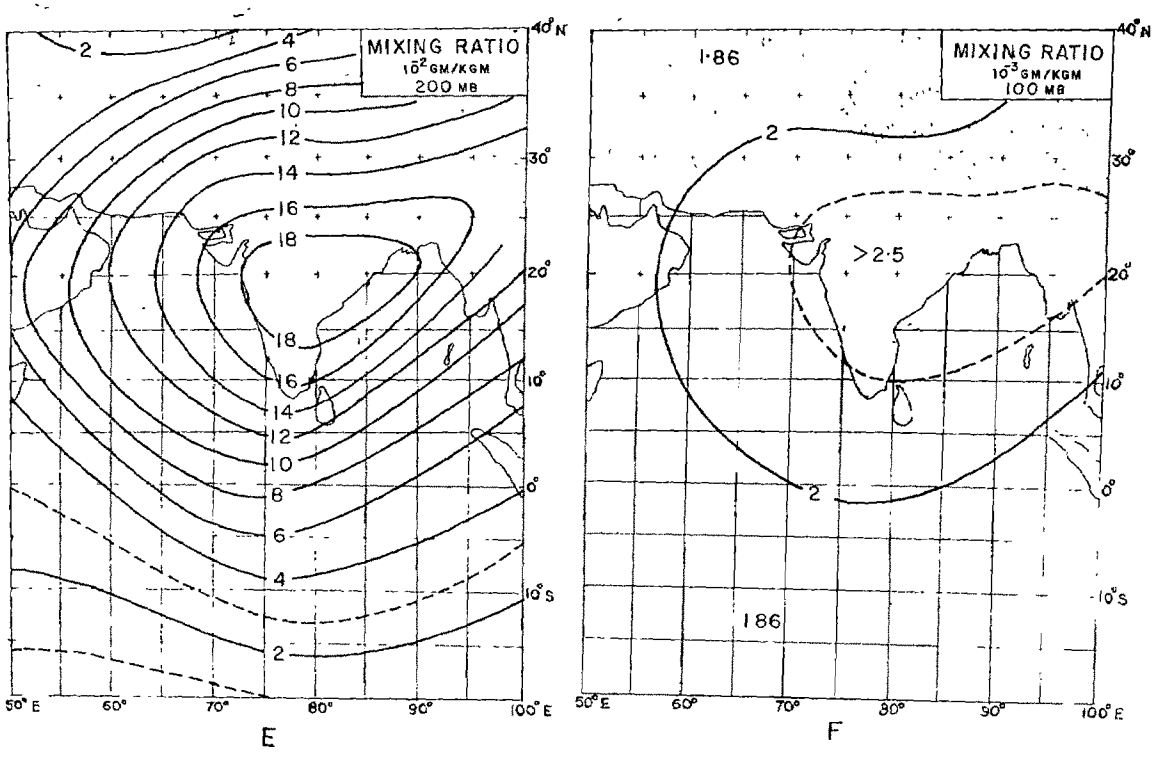


Figure 2.4.5 continued.

temperature as at the surface (Figure 2.4.1 C). At 700 mb (Figure 2.4.4 B), the thermal gradient over India and further south becomes weak. At 500 mb (Figure 2.4.4 C ) a pool of warm air begins to develop over north India and Tibet. The temperature gradient at 500 mb is flat over India and more so over the Indian Ocean. The pool of warm air increases in extent with height upto 200 mb (Figure 2.4.4 E) accompanied by the steepening of temperature gradient to its south. At 100 mb (Figure 2.4.4 F), the pool of warm air disappears and the thermal gradient south of 30° N weakens.

The strong north-South temperature gradient noticed at 200 mb is responsible for the development of TEJ discussed earlier. Also, the pool of warm air which extends from 500 to 200 mb pulls the tropopause upward so that the maximum tropopause height is not encountered over the Equator (unlike the usual feature observed elsewhere) but somewhere over north of the equator as actually observed.

The configuration of mixing ratio of water vapour from 850 to 100 mb is shown in Figure 2.4.5. The patterns at 850 and 700 mb (Figure 2.4.5 A, B ), are similar to the pattern at the surface (Figure 2.4.1 D); the maximum value being located over northeast India with a ridge running across the Arabian Sea in the Southwest-northeast direction. From 500 mb ( Figure 2.4.5 C), the pattern tends to become simple with a maximum centered

over India. From 500 to 100 mb (Figures 2.4.5 C,D,E,F), the distribution of mixing ratio is symmetrical with respect to this maximum. At 100 mb there is no gradient of mixing ratio over the major part of the region. The value of the mixing ratio at this level which is everywhere constant at 1.86 X $10^{-3}$ gm/kgm arises only due to the assumption (Section 2.2, Figure 2.2.1) made in the study, namely, that the mixing ratio at any level in the stratosphere is everywhere constant. The distribution of maxing ratio at 100 mb suggests that the tropopause height should be maximum around 20° N which is what is actually observed.

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## 3. DIABATIC HEAT SOURCES AND SINKS

The earth's surface plays an important part in converting the received solar radiation into sensible heat and in determining how the heat is shared between the atmosphere and the earth. If the earth's surface were smooth and of uniform composition, the global climatic patterns would consist of zonal bands showing no variation with longitude. However, large contrasts in climatic patterns are observed in regions along the same latitude also. The distribution of heat sources and sinks which are responsible for these climatic patterns, depends largely upon the nature of the earth's surface.

The earth's surface absorbs solar radiation in the short wave range, and in turn, emits radiation in the infra red ( long wave range). In tropics the absorp--tion of solar radiation is more than the emission in the infra red. In the polar zones, the opposite is true. The tropical belt is, thus, a heat source and the two poles are heat sinks. The circulation of sea and air should, therefore, operate such that heat is carried poleward. In addition, the troposphere which is, on the whole, at a lower temperature than the earth's surface is a heat sink at all latitudes and so heat should flow from earth to atmosphere. This is accomplished through long wave radiation and also by latent and sensi-