METEOROLOGICAL MONOGRAPH SYNOPTIC METEOROLOGY NO. 1 /1976

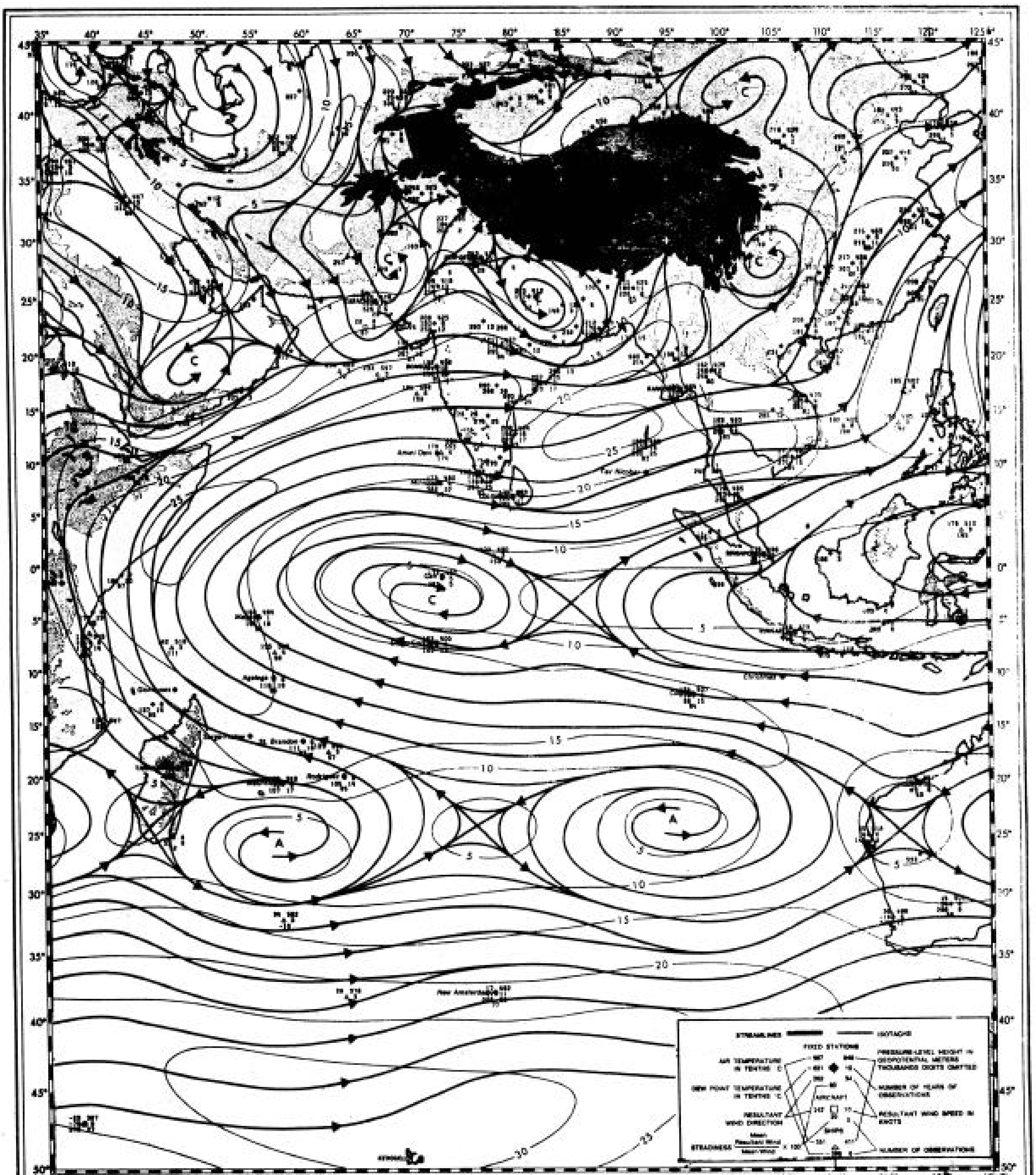
SOUTHWEST MONSOON

Y. P. RAO

DIRECTOR GENERAL OF OBSERVATORIES

INDIA METEOROLOGICAL DEPARTMENT

JUNE 1976



Frontispiece

Reprinted from the Meteorological Atlas of the International Indian Ocean Expedition - Volume 2 - Upper Air.

Mean Wind Flow - July - 850 mb

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P R E F A C E

The Southwest Monsoon has been a subject of intensive study for over four centuries now. In the last hundred years, Meteorologists in India have looked into various aspects of the monsoon problem. Of late, global interest is developing for increased understanding of the monsoon circulation through extensive data collection programmes. My main objective in preparing this monograph has been to present in a single compilation a reasonably exhaustive summary of work done on the subject, particularly in India. The bulk of the writing was accomplished during the two year period 1973-1975.

In deciding to publish the monograph, I was very much encouraged by the comments of Mr. J.S. Sawyer, F.R.S., Director of Research, U.K. Meteorological Office and Dr. R. Ananthakrishnan, F.N.A., former Director of the Indian Institute of Tropical Meteorology, Poona. My long association with Dr. B.N. Desai has contributed much to my appreciating the traditional knowledge on synoptic aspects of the monsoon, which is still the mainstay of weather forecasting. My colleagues, Mr. C.R.V. Raman and Mr. V. Srinivasan, gave a very exhaustive criticism which enabled me to improve the monograph. I would like to express my sincere thanks to these fellow meteorologists for their help and encouragement. Some work on Southwest Monsoon, particularly in languages other than English, would have escaped my attention, which I hope to add in a future edition.

Many diagrams from various books and publications have been reproduced in this monograph. I have pleasure in gratefully acknowledging the kindness of copyright holders in allowing reproduction of diagrams.

Mr. K. Gopalan, Mr. O.D, Sharma and Mr. S. Gurumurthy typed the manuscripts at various stages. Mr. S.S. Bhondve, Mr. P. S. Ray and Mr. N.B. Borgave assisted in drafting diagrams. Mr. V. Sadasivan meticulously assembled the manuscript with care and attention,, Mr. M.P. Rao assumed responsibility for graphic art layout and plate making for printing, Mr. N.A. Salve and Mr. V.M. Anarse skillfully printed the monograph on the Departmental Rotaprint machine. Without their willing and enthusiastic cooperation it would not have been possible to bring out the monograph in such a short time. I am thankful to all of them for their assistance.

New Delhi Y.P. Rao

25 June, 1976.

ACKNOWLEDGEMENT

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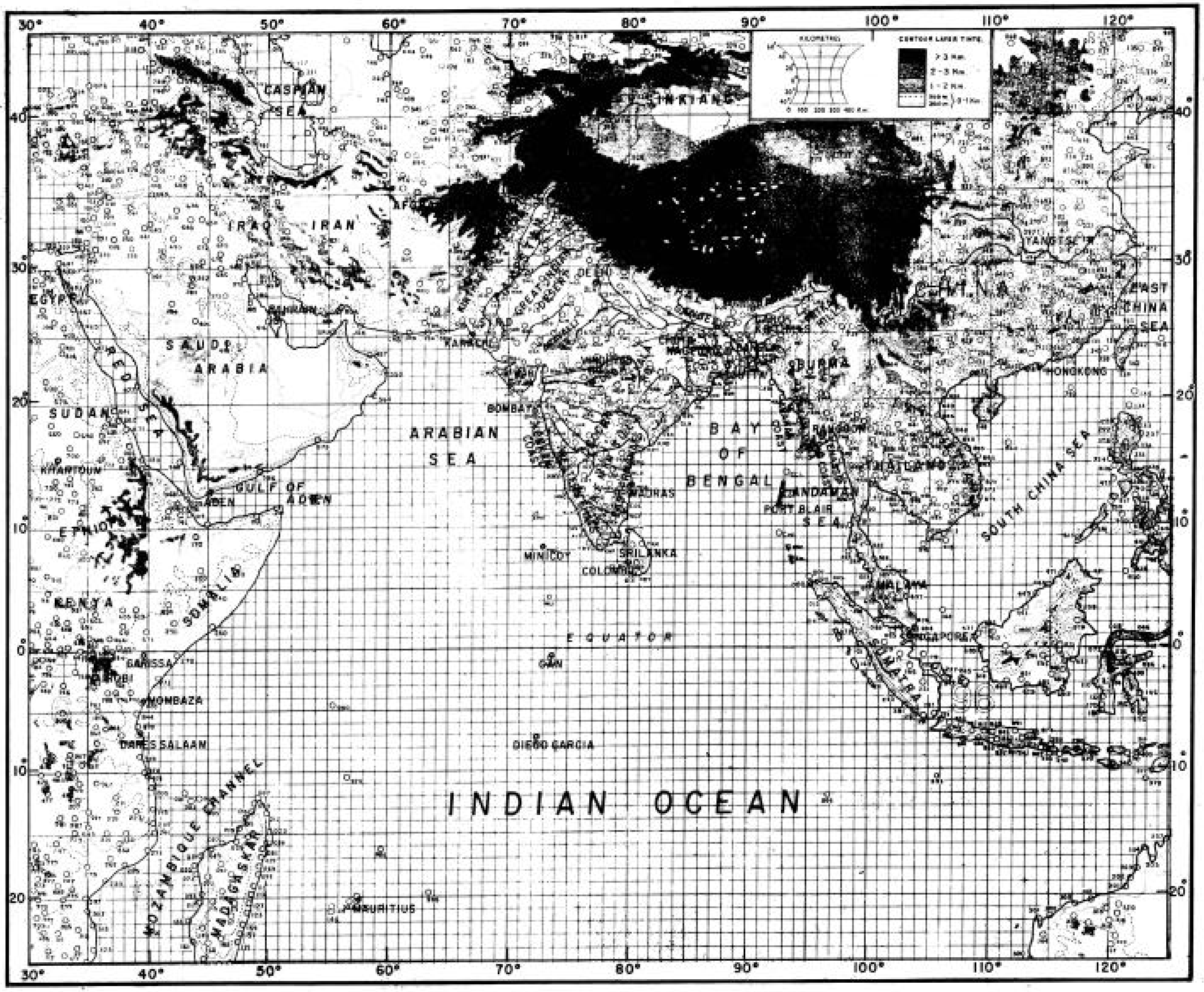
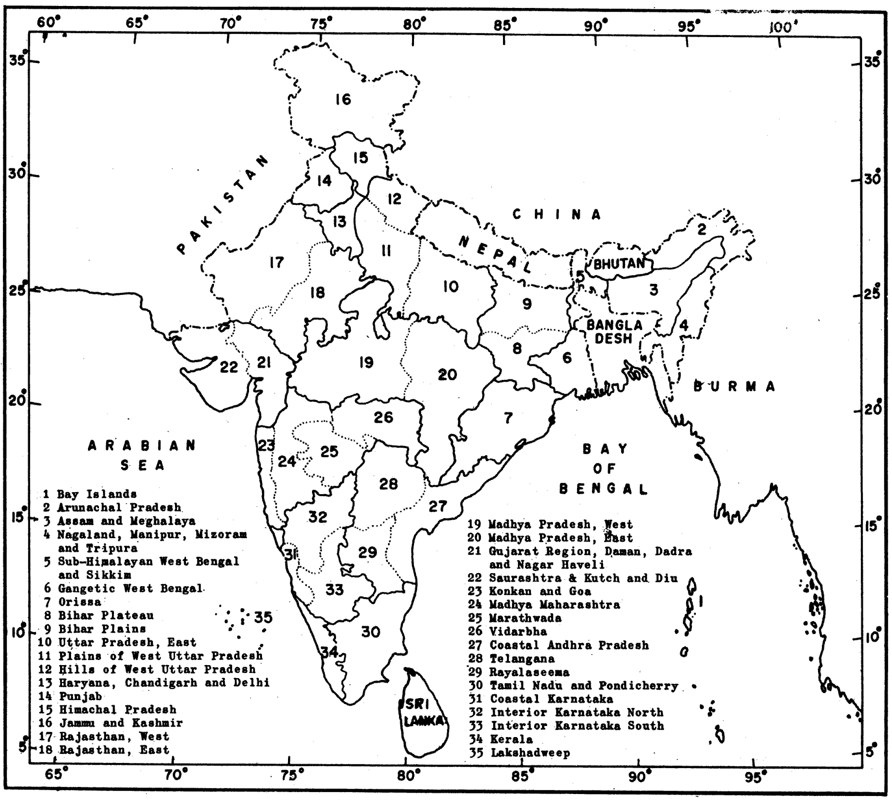


MONSOON ASIA AND THE INDIAN OCEAN



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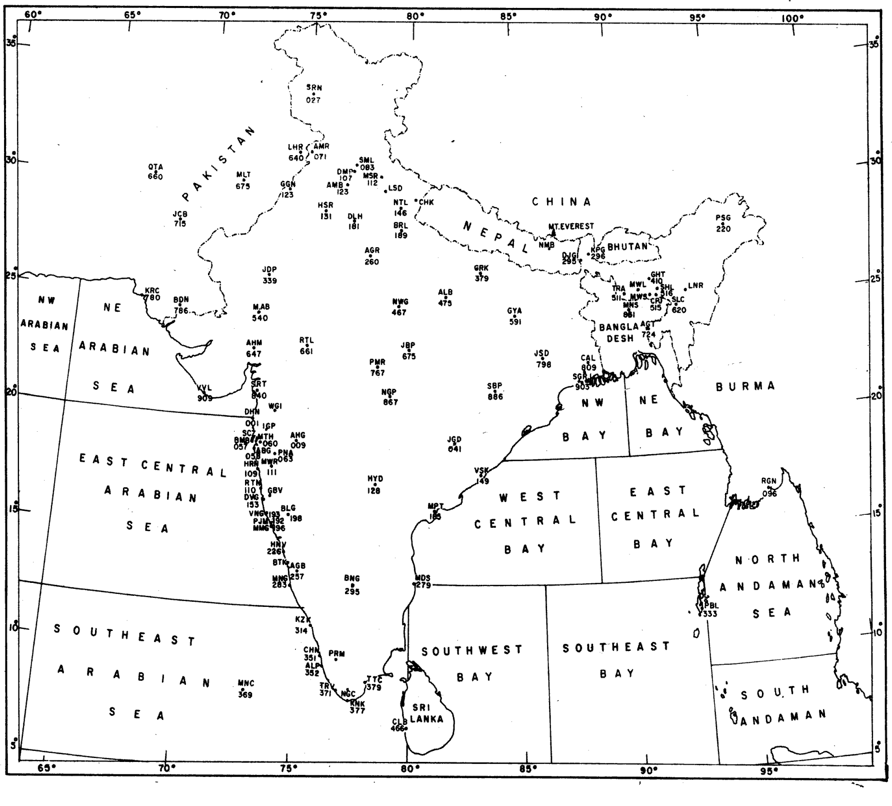

INDIA AND ADJOINING SEA AREAS

CONTENTS

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CHAPTER 1

INTRODUCTION

1.1 The name 'Southwest Monsoon' is used both for the phenomena of rains and southwesterly surface winds and the period during which they occur. Though the Arabic root word 'Mausam' simply means 'Season', in meteorological parlance, monsoon or monsoonal effect has come to mean the seasonal reversal of wind and in that sense the term has been used to describe such changes even in higher latitudes and stratospheric winds. Over most of India and southeast Asia, colossal differences are noticed from preceding or following months or the winter (also called by analogy 'Northeast Monsoon'). The mean rainfall over the Indian plains in the southwest monsoon period of June to September is 925 mm and for the rest of the year only 145 mm. Winds are southwesterly over the Bay and the Arabian Sea in July compared to northeasterly in January. The monsoon air mass is maritime and moist in great depth as against the winter dry continental air. Temperatures in the troposphere decrease to north in winter, while monsoon reverses this gradient. The westerly jet stream of the other seasons is replaced by the easterly jet stream. The meridional circulation of the Hadley Cell is replaced over the Peninsula by southerlies in the lower troposphere and northerlies aloft. Most of the annual rainfall of the densely populated areas of the world, viz. India, southeast Asia, Japan and parts of China is due to the phenomenon of the southwest monsoon and meteorologists should spare no effort to fully unravel its various facets.

1.2 Ramage (l97l) has defined the monsoon area by the following criteria :–

i) the prevailing wind direction shifts by at least 120° between January and July;

ii) the average frequency of prevailing wind directions in January and July exceeds 40 per cent;

iii) the mean resultant winds in at least one of the months exceed $3\,m\,sec^{-1}$;

iv) fewer than one cyclone – anticyclone alternation occurs every two years in either month in a 5° latitude – longitude rectangle.

The area between 35° N and 25° S and 30° W and 170° E satisfies this definition and India and the surrounding seas fall within this area.

1.3 The present review deals with the southwest monsoon over India. Adjoining land areas and the Arabian Sea and the Bay of Bengal are dealt with to the extent they are relevant to understanding the characteristics of the monsoon over India. From the time modern meteorological observations were commenced in India by the British in the eighteenth century, monsoon data have been analysed and there have been contributions regarding their relationship and underlying causes from workers such as Blanford, Eliot, Walker, Simpson and Ramanathan. The advent of radiosonde and radiowind measurements and International Indian Ocean Expedition (IIOE) and the Indo– Soviet Monsoon Experiment (ISMEX) over the Indian Ocean in the recent years have enabled further understanding of the southwest monsoon in all its dimensions. This review presents the existing knowledge on the southwest monsoon from the point of view of an Indian synoptic meteorologist, without tracing the historical development.

REFERENCE

Ramage, C. S. 1971 'Monsoon Meteorology', Academic Press, New York, p.6

CHAPTER 2

CLIMATIC PATTERNS

2.1 In this chapter the mean distribution of important climatological elements during the monsoon is described, tracing their evolution from the preceding season and decay at the end.

2.2 <u>Sea–level Pressure Distribution</u>

2.2.1 Development of a low due to increased heating over India starts in March itself, with slightly higher pressures over the Arabian Sea and the Bay of Bengal. By April, the global land lows have begun establishing themselves along about 10^{o} N in north Africa and about the Tropic of Cancer in the Indian region and Burma. Peninsular India south of 20° N has a tapering shape and is narrow south of 15^{o} N. The latter portion comes under considerable maritime influence. This makes the heating over land more prominent to the north of 20^{o} N and hence the axis of the low is at a more northerly latitude over India (Fig. 2.l). There is a weak cell over upper Sind and another in Bihar and east Uttar Pradesh. Over the Peninsula, a trough forms with axis along longitude 78° E. A ridge runs from Arabia into the west Arabian Sea where a clockwise wind circulation is found around 14^{o} N, 60^{o} E. A similar circulation is also present over the Bay of Bengal around 14^{o} N, 90^{o} E. The sub–tropical high in the south Indian Ocean is along 30^{o} S.

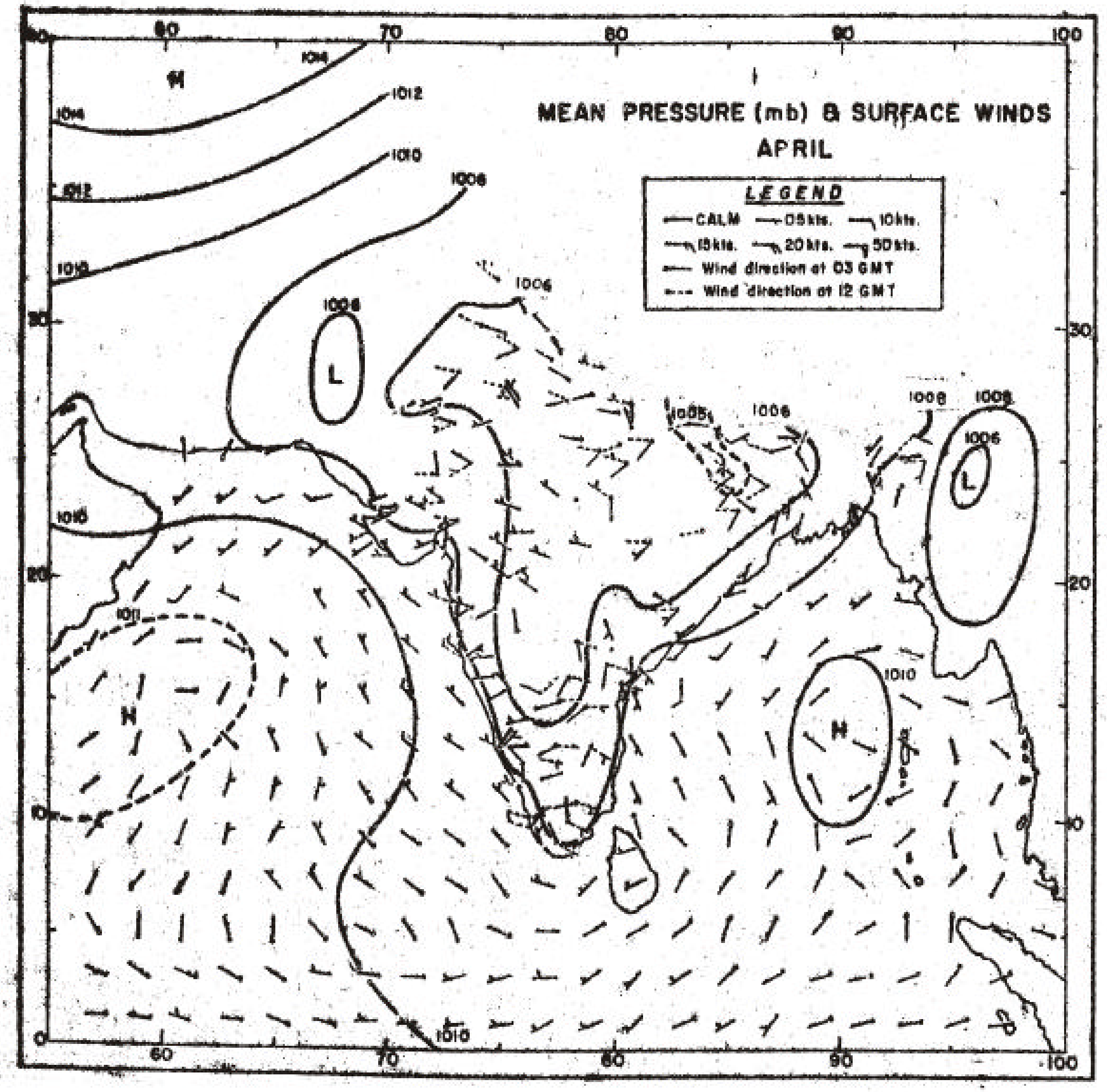
2.2.2 By May, the summer continental low pressure areas completely dominate north Africa and Asia. Its main centre over India is near 30^{o} N, 75^{o} E with an extension as a trough upto Orissa. The heat low is still more marked in June with the main centre over Pakistan. In these months, there are subsidiary centres of low pressure in Africa and other parts of Asia. By May, the trough over the Peninsula has shifted to 79° E along the Madras Coast.

2.2.3 Monsoon activity is maximum in July when the low pressure area ex tending from north Africa to northeast Siberia is most intense. Its main centre is over north Baluchistan and neighbourhood (Fig. 2.2). A trough lies over north India with axis from Sriganganagar to the Head Bay, which is referred to as the 'monsoon trough'. Pressure gradient is strong south of this trough. The Indian Ocean 'High' has strengthened and is centred at about 30° S, 60° E. Pressure continuously decreases over the Indian Ocean northwards of this high pressure belt. Weak ridges are present in the Arabian Sea off the west coast of India and in the Bay off Tennaserim coast and over Burma. The weak trough of the premonsoon months in the eastern Peninsula now lies just off the east coast of the south Peninsular India; it persists through the monsoon months and is more pronounced in September.

2.2.4 In August, the intensity of the Afro–Asian low is decreasing. By September, pressures rise north of 40° N over Asia and the Afro–Asian low is oriented east–west. The pressure gradient south of monsoon trough which is maximum in July and slightly less in June and August, decreases to half in September. The ridge off the west coast is displaced in September to about 65° E.

2.2.5 By October, the trough over northern India shifts to the Bay of Bengal, with the trough line along 13° N and the pressure field is flat over the country (Fig. 2.3). The low pressure belt runs from Africa to the west Pacific between the equator and 20° N with Centres over Africa, the Bay of Bengal 'and the west Pacific. The Asian 'High' is establishing along 50° N and is centred at about 90° E.

2.2.6 To summarize, the chief features of the surface pressure distribution in the monsoon season are the heat low over Pakistan, the monsoon trough thence to Head Bay and the strong pressure gradient to the south.



MEAN PRESSURE (mb) & SURFACE WINDS
APRIL
LEGEND
CALM
05 kts.
10 kts.
15 kts.
20 kts.
50 kts.
Wind direction at 03 GMT
Wind direction at 12 GMT
1014
1012
1010
1008
1006
1011
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70
80
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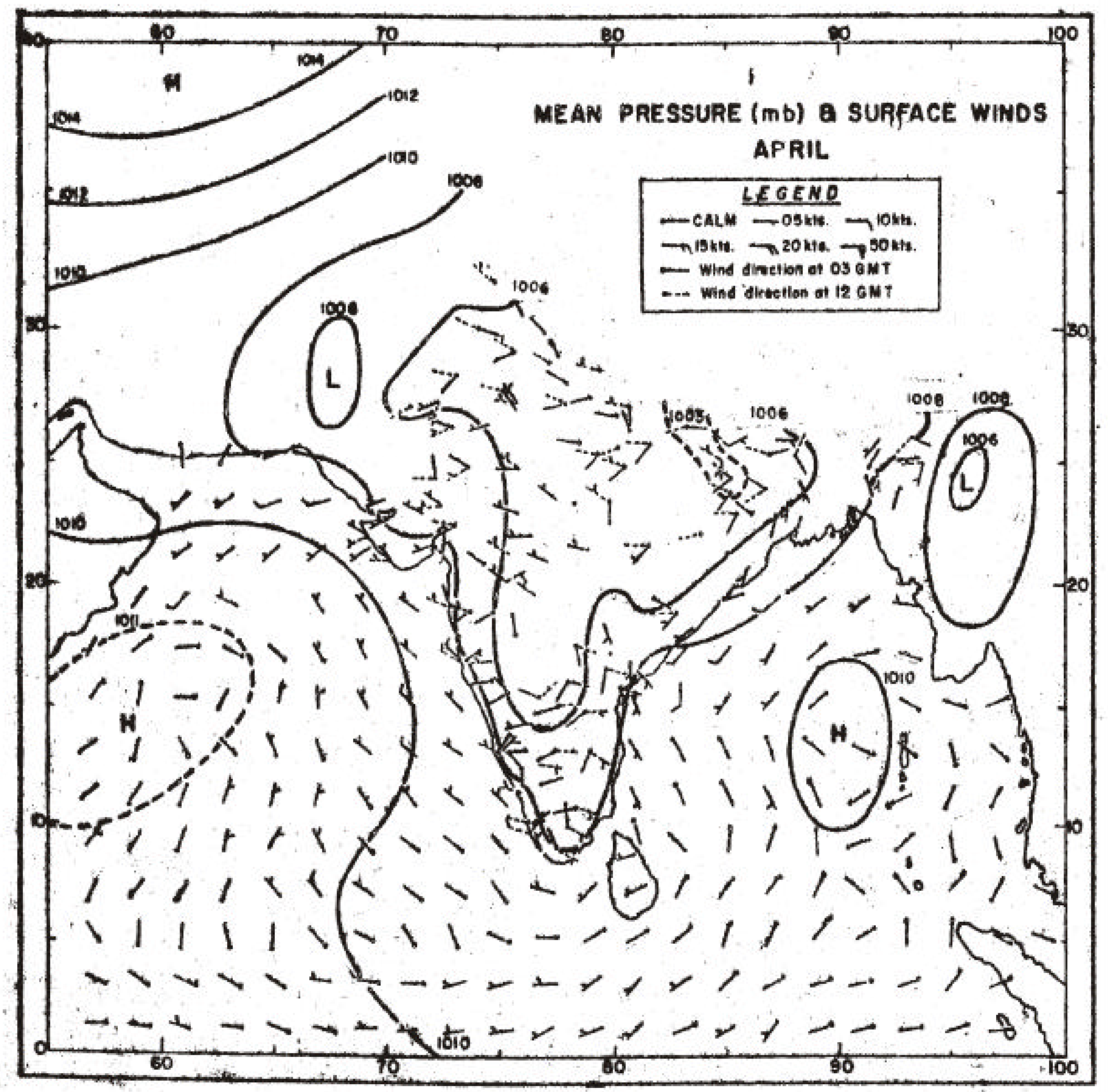
Fig. 2.1


MEAN PRESSURE (mb) & SURFACE WINDS
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LEGEND
CALM
05 kts.
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15 kts.
20 kts.
50 kts.
Wind direction at 03 GMT
Wind direction at 12 GMT
1014
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Fig. 2.1

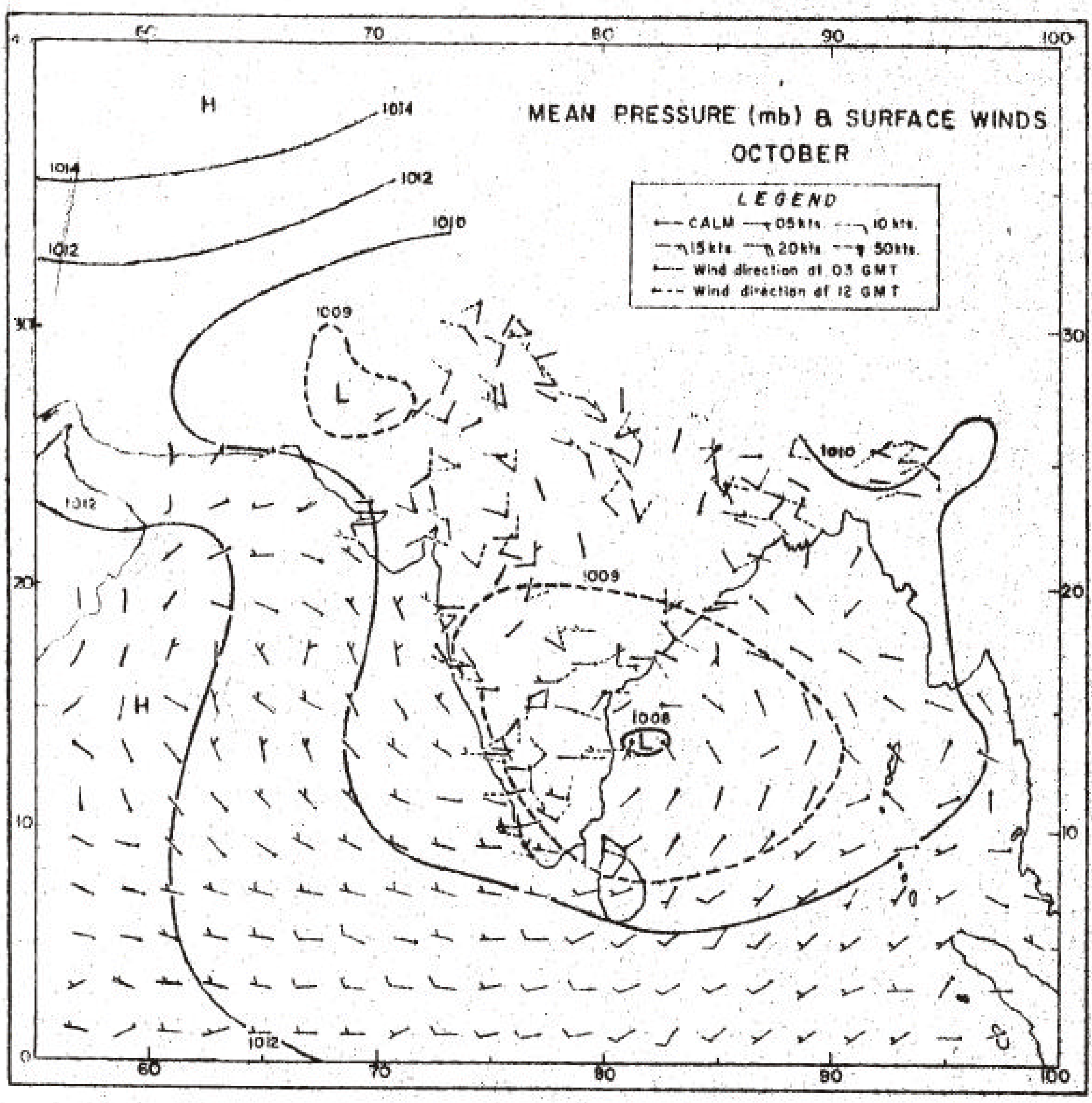
MEAN PRESSURE (mb) & SURFACE WINDS
OCTOBER
LEGEND
CALM
05 kts.
10 kts.
15 kts.
20 kts.
50 kts.
Wind direction at 03 GMT
Wind direction at 12 GMT
1014
1012
1010
1009
1008
H
L
60
70
80
90
100
30
20
10
0

Fig. 2.3

2.3 Surface Winds

2.3.1 In June, winds are from east, to the north of the line running through Lahore, Allahabad and Silchar, Over Gangetic West Bengal winds are mainly from south. Over the rest of India, winds blow from west to southwest, more westerly off the west coast of the Peninsula and practically southwesterly in the Bay. Between 5° N and 15° N and west of 65° E, the speed is 20 kt in the Arabian Sea. Elsewhere the range is 10 to 15 kt.

2.3.2 In July, easterlies prevail over the country to the north of the line through Hissar, Gaya and Silchar (Fig. 2.2). To the south of this line, southerlies occur over West Bengal and southwesterlies to westerlies elsewhere. Strongest winds are in the southwest Arabian Sea. Nearer the west coast of the Peninsula, the direction is westsouthwest to west, except along the Kerala coast where north– westerlies are observed. Mean speeds over land are not more than 10 kt. They are between 10 kt and 20 kt along the west coast. In the Arabian Sea west of 68^{o} E and between 10° N and 20° N, it is over 25 kt. Over most of the Bay and the rest of the Arabian Sea, the speed is about 15 kt. Winds are somewhat weaker between the equator and 5^{o} N. August is similar to July. By September, there is a weakening of pressure gradient and winds weaken, particularly over the sea areas. In September, winds are westnorthwesterly in the Arabian Sea to the east of 65 E.

2.4 Surface Temperatures

2.4.1 Mean of maximum and minimum temperatures reduced to sea–level at a lapse rate of 6° C per km will be discussed now. Air temperatures over sea areas have been taken from the Monthly Meteorological Charts of the Indian Ocean (1949), published by the London Meteorological Office.(Fig. 2.4).

2.4.2 Land gets progressively heated after December–January and by April, temperatures are of the order of 33° C to 35° C over Deccan and adjoining inland areas upto about 24° N. Temperatures along the coasts are between 28° C and 30° C. The gradient of temperatures is steep, normal to the west coast, and reaches about 5° C per degree longitude at some places, while it does not exceed 2.5° C per degree longitude towards the east coast. The gradient to the north is also gradual. The spatial range of temperature is about 8° C. The highest temperatures over India and Burma occur at about 20° N.

2.4.3 In July, the southwest monsoon causes extensive cloudiness. Clouding is heavy between 17^{o} N and 24° N in the central regions, west of 77^{o} E and south of 17^{o} N in the Peninsula, and to the east of approximately 85^{o} E in northeast India. Temperatures are even in these regions, being generally between 28° C and 29° C; but on the west coast, temperatures of the order of 26° C to 27° C are experienced which are lower than mean air temperatures over the open Arabian Sea. Similar feature is noticed along the Burma coast also. Over the country, to the east of 77^{o} E and south of 17° N, temperatures are 30° C to 31° C. This large difference of 4° C is partly due to Fohn effect and to lesser cloudiness. The hottest areas in India lie over west Rajasthan with still higher temperatures further west. The spatial range of temperatures over India is about 9° C. Thus the region of highest temperatures shifts from near 20° N in April to 28° N by July.

2.5 Rainfall

2.5.1 The rainfall distribution in the principal rainy season of India, the southwest monsoon period, lasting from June to September, is shown in Fig.2.5. Except in Kashmir and neighbourhood, the extreme south Peninsula and the east coast areas, the annual rain is mainly accounted for by the falls in this season. Orographic influence is dominant in the distribution of rainfall in this season, as the prevailing winds blow almost at right angles against the western ghats and the Khasi–Jaintia hills. There is rapid increase of rainfall to the north of a line running from Ahmednagar to Masulipatnam upto the southern slopes of the Vindhyas.

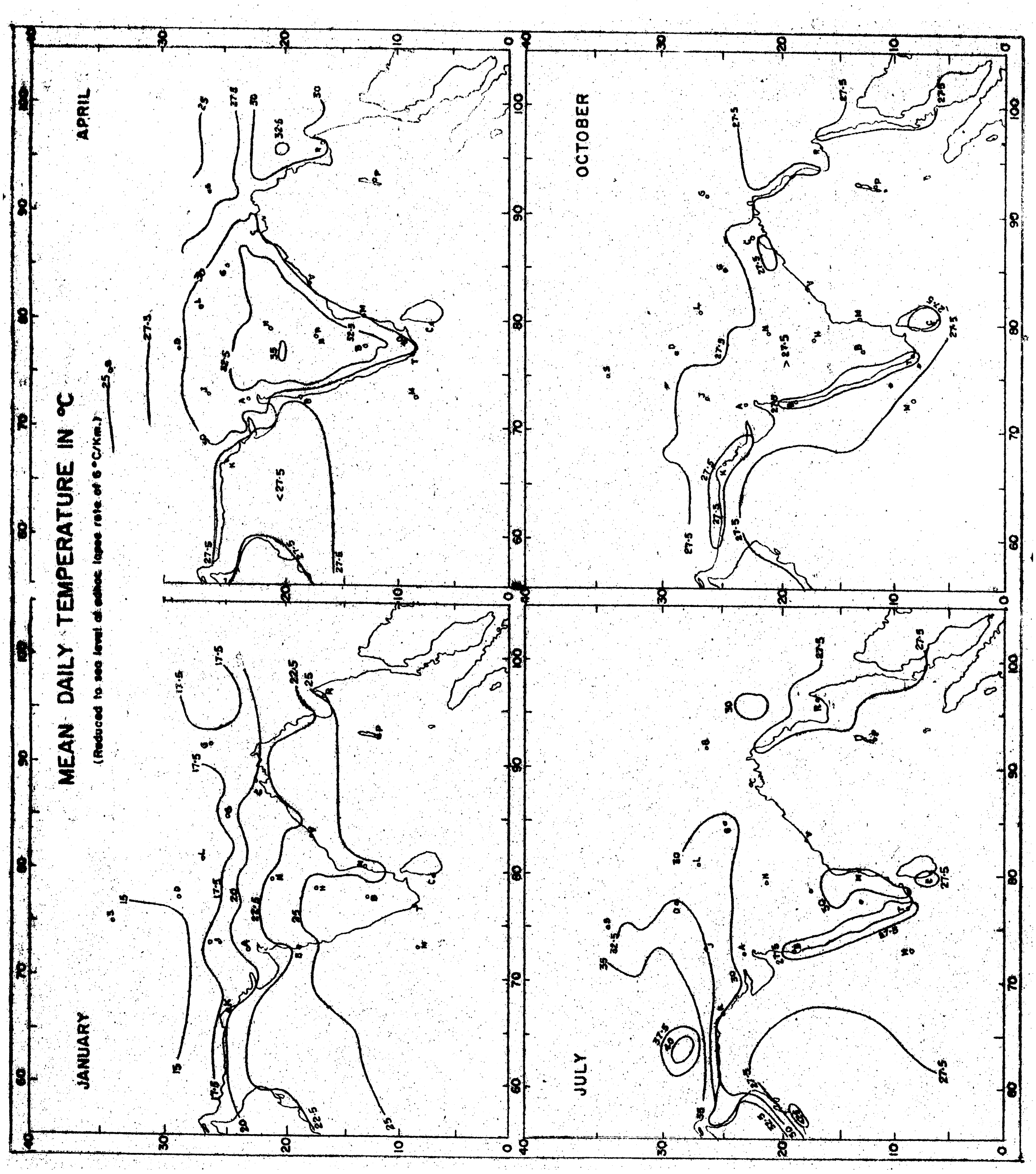
MEAN DAILY TEMPERATURE IN °C
(Reduced to sea level at adiab. lapse rate of 6°C/Km.)
JANUARY
APRIL
JULY
OCTOBER

Fig. 2.4

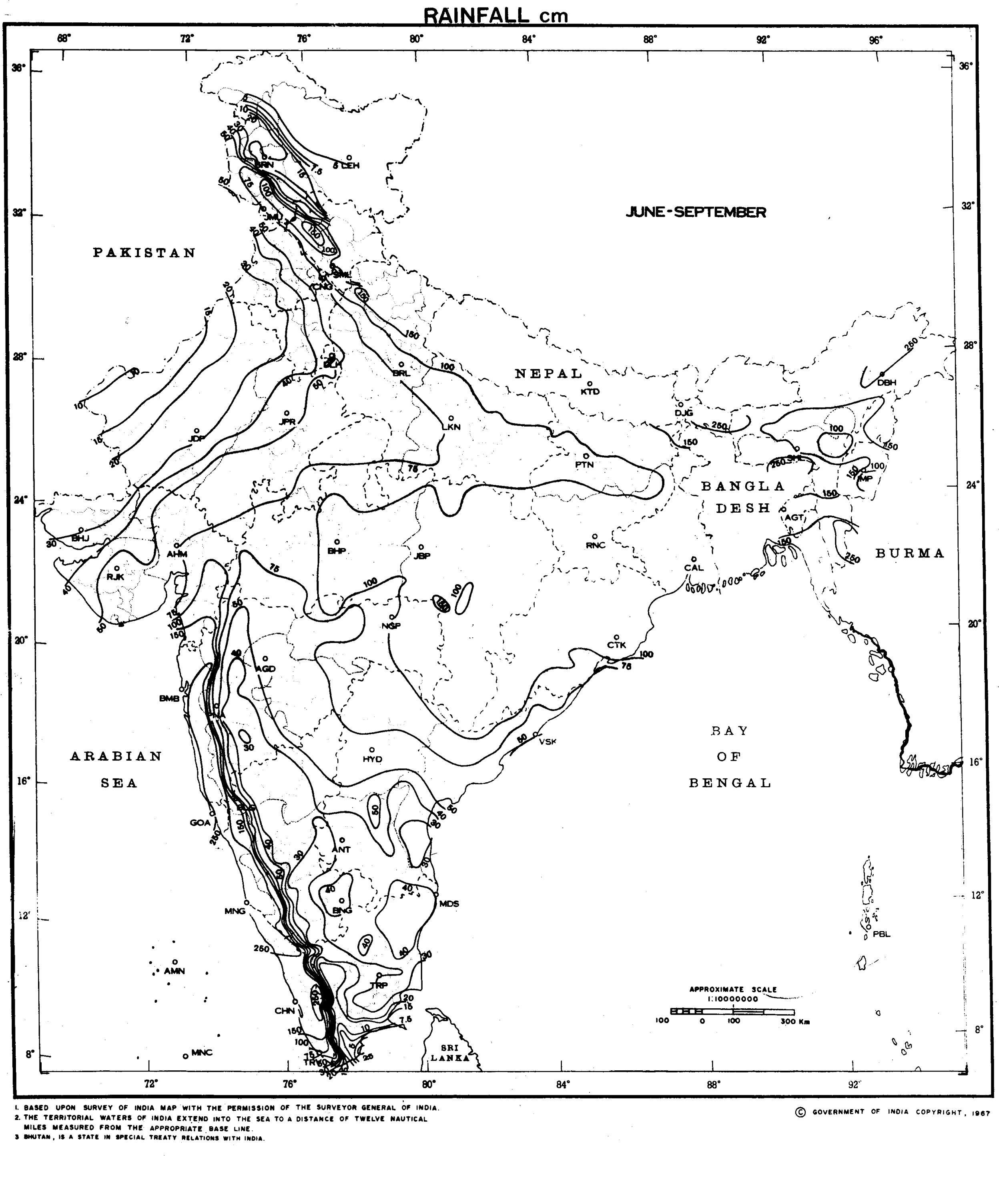
RAINFALL cm
JUNE-SEPTEMBER
PAKISTAN
NEPAL
BANGLA DESH
BURMA
ARABIAN SEA
BAY OF BENGAL
SRI LANKA
APPROXIMATE SCALE 1:10000000
100 0 100 300 Km
1. BASED UPON SURVEY OF INDIA MAP WITH THE PERMISSION OF THE SURVEYOR GENERAL OF INDIA.
2. THE TERRITORIAL WATERS OF INDIA EXTEND INTO THE SEA TO A DISTANCE OF TWELVE NAUTICAL MILES MEASURED FROM THE APPROPRIATE BASE LINE.
3. BHUTAN, IS A STATE IN SPECIAL TREATY RELATIONS WITH INDIA.
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Fig. 2.5

In the north Indian plains, a minimum rainfall belt runs from northwest Rajasthan to the central parts of West Bengal, practically along the axis of the monsoon trough. Rainfall decreases generally from the hills of the western and eastern ghats towards the coast.

2.5.2 Rainfall decreases very rapidly southwards along west coast from 9.5^o N to Kanyakumari. The rainfall at Kanyakumari in this season is about the same as in the Great Indian Desert. To the east of the western ghats between 8^o N and 10^o N, rainfall decreases considerably with a very steep gradient across the eastern slopes. Rainfall is only 2 cm in some places in the coastal strip in extreme south Tamil Nadu. With all the significant amounts of rainfall occurring over the ghats, a saving feature of economic interest is that all the important rivers of south India emerge out of the western ghats to flow east through the plains having rainfall of the order of that in west Rajasthan.

2.5.3 Hills and mountain ranges cause striking variations in rainfall distribution. On the southern slopes of the Khasi–Jaintia hills rainfall is over 800 cm while to the north, in the Brahmaputra valley, it drops to about 120 cm. Cherrapunji's annual rainfall of 1142 cm (at elevation of 1313 m) is obviously due to orographic lifting but its magnitude requires to be quantitatively explained. From the west coast, rainfall increases along the slopes of the western ghats and rapidly decreases on the eastern lee side. No definite information is available about the increase of rainfall with elevation and the height at which the rainfall attains the highest value. In the higher reaches of the western ghats, there are places with seasonal rainfall of 500 cm. Within 80 km on the lee side, rainfall is only 40 cm.

2.5.4 From the coast of West Bengal and the hills of Orissa, rainfall decreases inland. Further westwards, the Chota Nagpur hills, the Maikala Range and the Mahadeo hills cause an increase of rainfall, with lesser amounts in the valleys in between. The Gir hills in Kathiawar have more rainfall than the neighbourhood. Mount Abu in Aravallis has a rainfall of 169 cm while the surrounding plains have only 60 to 80 cm.

2.5.5 Across northern India, a line of rainfall minimum runs from 28.5^o N, 75^o E to 25^o N, 88^o E which is paradoxically close to the monsoon trough. Area to south of this rainfall minimum falls in the track of monsoon depressions which are responsible for much of the rainfall. In tracts further to north, there is probably the influence of the Himalayas in increasing the rainfall. Apart from this, there is also a decrease of rainfall from east to west, from about 120 cm in West Bengal to less than 20 cm in the Great Indian Desert in west Rajasthan.

2.5.6 In the Himalayas, observations are extremely scanty, particularly from higher elevations where there is added difficulty of measuring snowfall. Rainfall measured in river valleys may not be representative of the hill slopes. Between the Great Himalayan Range and the plains, there are the Pir Panjal, the Siwalik and the Mahabharat Ranges. Most of the available observations are from these ranges. Rainfall increases up the slopes of these foot hills, presumably decreases on their northern slopes and increases again on the Himalayan slopes. Annual rainfall at Chaunrikharka (2,700 metres) is 228 cm and at Namche Bazar (3,300 metres) only 94 cm (Dhar and Narayanan, 1965). Both are in Nepal and the distance between the two is hardly 16 km. Therefore, we can tentatively conclude that above some elevation near 3 km, rainfall may decrease with height on the Himalayan Range. In the eastern Himalayas, rainfall is more than in the western portions. In the east, annual rainfall of 400 cm has been recorded but less than 200 cm in the west.

2.5.7 Rainfall in the Andaman and Nicobar Islands in the southwest monsoon season is about 140 to 190 cm, while in Laccadives (Minicoy and Aminidivi Islands) and Maldives in the Arabian Sea, it is only about 100 cm though both the groups are in the same latitude belt. Calicut on the mainland in the west coast, however, gets 235 cm, more than the Bay Islands.

2.5.8 Ramakrishnan and Gopinath Rao (1958) have computed the average amount of water precipitated as rain over each degree square of land area south of 20^{o} N in each month. Their figures for the whole season are given in Table 2.1.

Table 2.1

Average amount of water precipitated as rain on land in each degree square from June to September (in 10^7 cu metres or 10^7 metric tons)

Latitude (° N)	**Longitude (° E)**												
	72–73	**73–74**	**74–75**	**75–76**	**76–77**	**77–78**	**78–79**	**79–80**	**80–81**	**81–82**	**82–83**	**83–84**	**84–85**
20–19	632	2126	500	639	1104	906	1057	1290		1383	1406		701
19–18	819	2530	431	991	805	944	874	874	1562	1410		723	224
18–17		2462	470	481	622	822	716	693	834	705	493	183	
17–16		2571	638	413		638	567	437	697	529	719		
16–15		1043	1507	427		451	478	356	209				
15–14			2019	739	369	369	453	357	34				
14–13			1569	2228	395	467	419	467	125				
13–12			677	2815	613	457	457	541	93				
12–11				1258	1474	265	459	402					
11–10					2278	436	375	169					
10–9					1661	377	194						
9–8					248	321	5						

2.5.9 Figure 2.6 shows the coefficient of variation of rainfall, which is the ratio of the standard deviation of the season's rainfall to the mean amount. This varies from 60 per cent in the western desert to 20 per cent or less in the most rainy areas. In the southeastern tip of the Peninsula where this is not the rainy season, it is 100 per cent. The coefficient of variation decreases with increasing rainfall upto about 100 cm and then does not vary, as pointed out by Rao et al (1972). Standard deviations of rainfall are generally comparable over the whole country but the wide differences in variability are due to the differences in mean rainfall. These fluctuations in rainfall affect economic activities, more so in areas of poor rainfall. Ramdas (1958) has presented a diagram showing years when the monsoon rainfall in each sub–division was twice or half the normal (Fig.2.7)

2.5.10 Mean monthly rainfall amounts are not uniform during this period. Broadly, rains increase with the setting of the monsoon, reach a maximum in July and then decrease. But, Arabian Sea Islands get more rain in June than in July, while Kerala has about the same in the two months, both areas being more rainy in May. In Greater Assam and sub–Himalayan West Bengal, June and July are equally rainy, decreasing thereafter. Some parts of Assam get slightly less rain in July. In Bihar, Uttar Pradesh, Gangetic West Bengal, east Madhya Pradesh and parts of Orissa, July and August have the same amounts. In the Peninsula between 19^{o} N and 16^{o} N and east of 76^{o} E, an increase in September is noticed, apparently due to the effect of depressions and lows forming at lower latitudes. Coefficient of variation of rainfall of individual months is more than for the whole season, as may be expected. For July alone, the coefficient is 30 per cent in the more rainy parts and 100 per cent in the Rajasthan desert.

2.5.11 Rainfall amounts vary still more, from day to day. The following convention is in vogue to describe the activity of monsoon by the amount of rainfall:–

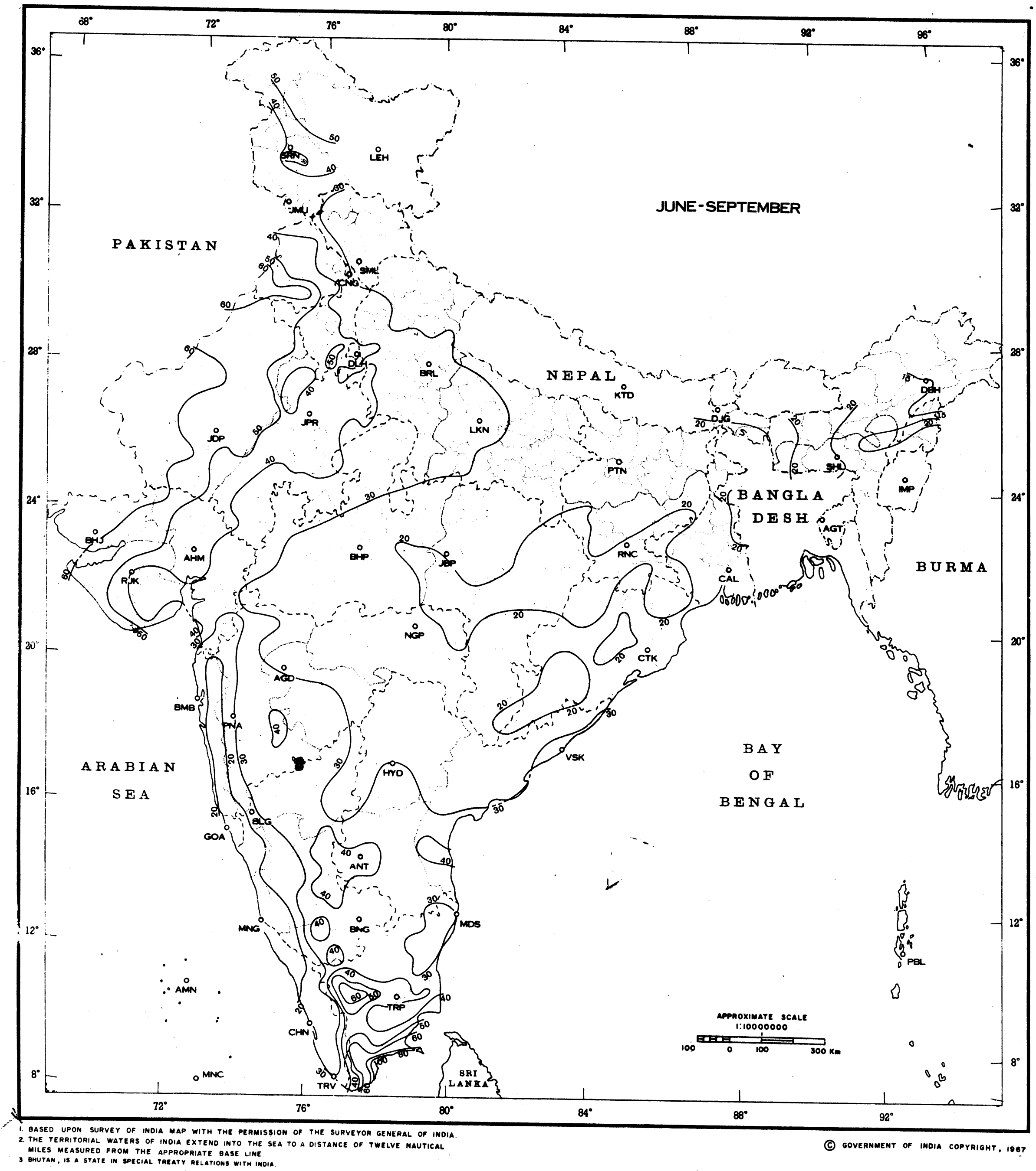


Fig.2.6 Coefficient of variation of rainfall

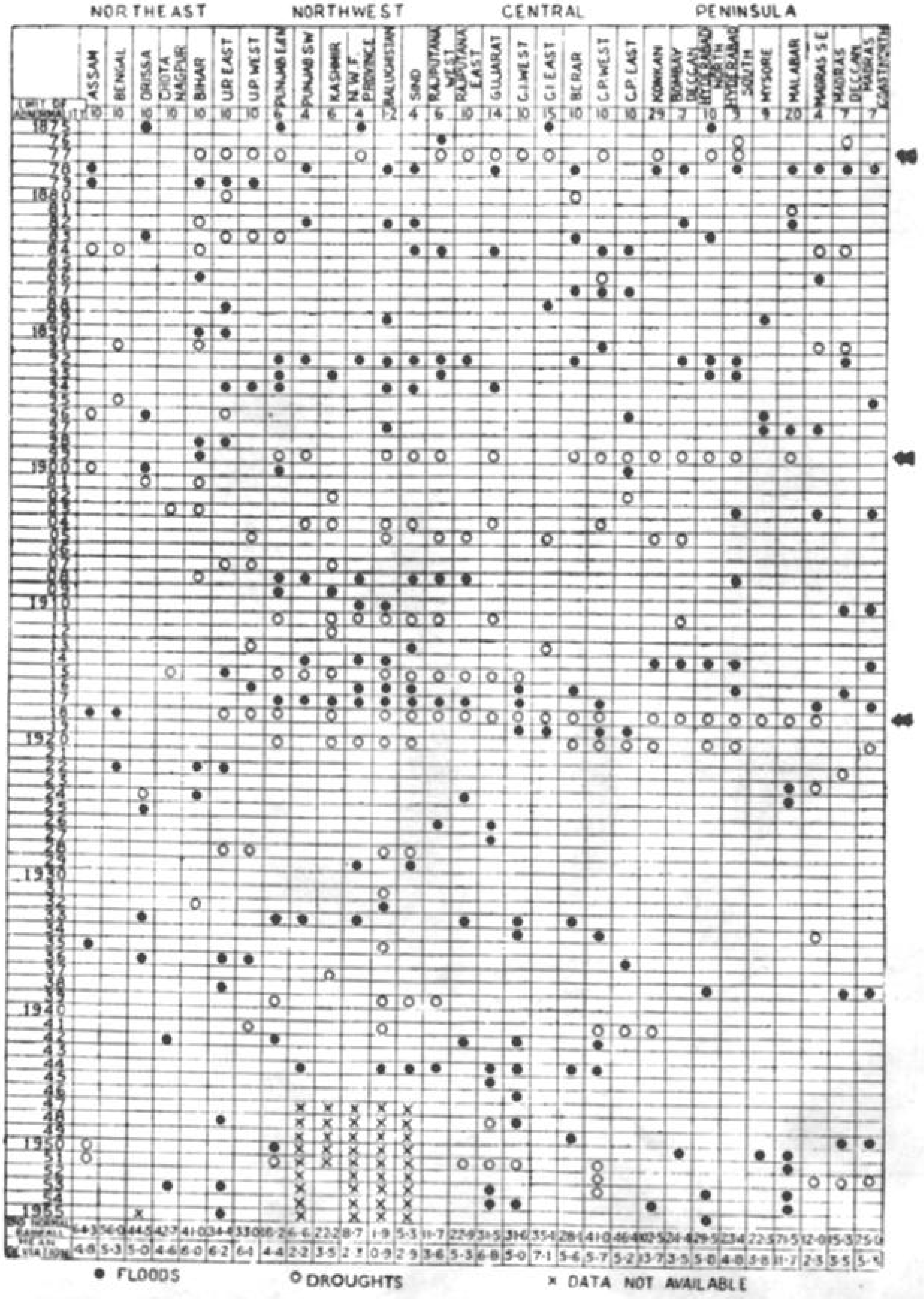


Fig. 2.7 distribution of Monsoon Rainfall (1875-1955).

Description of monsoon	**Rainfall over the area**
Weak monsoon	Rainfall less than half the normal
Normal monsoon	Rainfall half to 1 1/2 times the normal
Active/strong monsoon	Rainfall 1 1/2 to 4 times the normal with at least two stations recording 5 cm if along west coast and 3 cm if elsewhere
Vigorous monsoon	Rainfall more than 4 times the normal, with at least two stations recording a minimum of 8 cm if along the west coast and 5 cm if elsewhere.

Active monsoon conditions decrease sharply after July in Kerala, Coastal Karnataka and Konkan. In west Uttar Pradesh, such periods are more in August. In Madhya Pradesh, July to September are equally active. North Assam and Vidarbha show an increase again in September. In Kerala and Arabian Sea Islands, monsoon is only 'normal' on about half the days.

2.5.12 Heavy rains of the order of 25–30 cm are not infrequent in this period, but they are more probable north of 15° N along west coast and north of 20^o N in the rest of the country, apart from the western ghats. In areas of poor rainfall like Rajasthan, Saurashtra and Kutch, whole season's normal rain may occur on one day. A plain station, Dharampur (Surat District) recorded 100 cm on a single day, (2nd July 194l). Khasi–Jaintia hills are noted for heavy falls of over 75 cm in 24 hours on the windward side. In Assam, heavy rains (>25 cm) are most frequent in June, the frequency dropping to one–third in August, in spite of the 'breaks'. In Uttar Pradesh and Bihar, heavy rains are more frequent in August and September. The high frequency of heavy rains in these areas and in Punjab and Haryana in September, is promoted by interaction with extra–tropical systems. In Madhya Pradesh, all months are equally liable to heavy falls but along the west coast, the frequency drops to a quarter after July.

2.5.13 Berson and Deacon (1965) have shown the occurrence of heavy rainfall during monsoon (June–July) is more frequent at certain epochs of lunar synodic cycle. The lunar effect is much stronger in years of below average sunspot number (Table 2.2 and Fig. 2.8).

Table 2.2

Frequency of heavy spell rain days at Mangalore (India) related to ten divisions of lunar synodic period (Months of June and July in the period 1901–1950)

	Phase in lunar synodic decimals										**Total**	x^2
	1	**2**	**3**	**4**	**5**	**6**	**7**	**8**	**9**	**10**		
Quiet Sun	21	20	30	24	37	29	21	15	11	18	236	27.47
Active Sun	34	28	21	29	27	26	19	21	24	25	254	7.02
All years	55	48	51	63	64	55	40	36	35	43	490	20.00

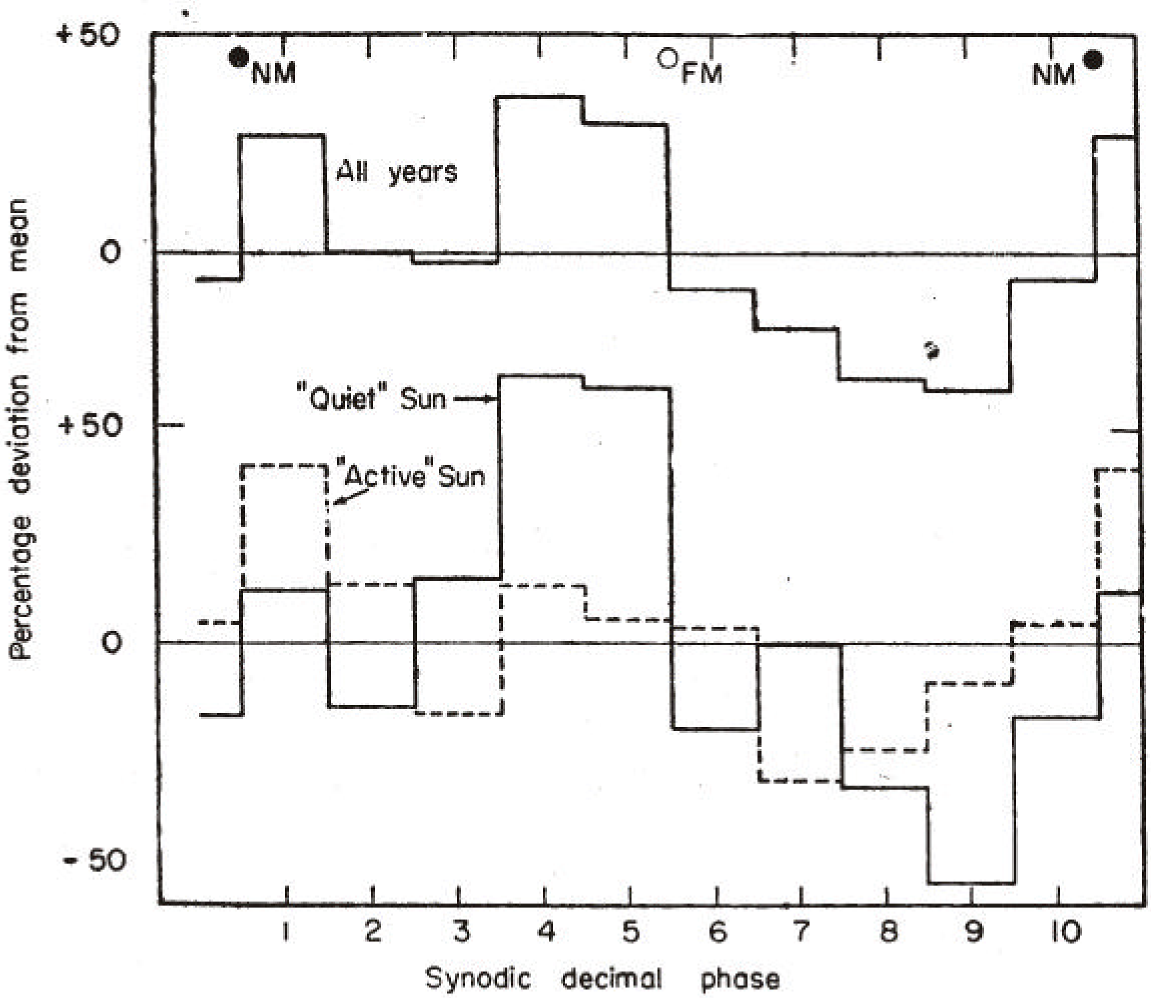


Fig.2.8 - Percentage deviation from mean of the precipitation on days of heavy spell (rainfall exceeding 2 inches) in months June & July at Mangalore, (India) over the period 1901-1950.

2.6 Upper Air Temperatures

Difference between various types of radiosonde instruments, short periods of data and sparse network of stations in southeast Asia make possible only a broad description of the large scale features of the upper air temperature distribution.

By April, a thermal high develops over India at 850 mb with centre near about 22° N, 80° E (Fig. 2.9). Temperature decreases in all directions from this centre. The fall in temperature is about 7° C down to 8° N, while it is 4° C upto 30° N. Further north, meridional temperature gradient is very marked. By 700 mb, the thermal high shifts southwards to 15° N and the temperature gradient is towards north. A ridge appears to run from north Arabian Sea to the Caspian Sea. At 500 mb also the temperature decreases to the north of 12^{o} N and is uniform to the south. Temperature range from 17^{o} N to 30^{o} N is 4° C. At 300 mb, a weak thermal high is over the south Peninsula and the temperature fall from 20° to 30° N is 6° C. Temperature is very flat at 200 mb (within 2° C of 223°A) but a marked increase towards north develops at 150 mb and 100 mb from the central parts of the Peninsula. At 100 mb, the temperature difference is 12° C between 10^{o} and 30° N. Standard deviations of temperature are 2° to 4° C at 700 mb and 500 mb. They are 3° to 5° C from 300 to 100 mb, higher values being in the northern parts of India.

The tropical tropopause occurs in April near 100 mb; to the north of 25° N, the middle tropopause also occurs at 200 mb. The temperature at the tropical tropopause is about 205° A north of 25^{o} N, decreasing to 195° A south of 15^{o} N. The frequency of middle tropopause decreases from January to April over northern India unlike some other areas in the same latitude.

Between 850 and 700 mb the highest lapse rates of over 9° C Km^{-1}, occur from the central parts of the country to Rajasthan. The lapse rate decreases to the south and to the east, becoming 5.5° C Km^{-1} at Trivendrum and Port Blair. From 700 to 500 mb it is over 7° C Km^{-1} over northwest India, while remaining 5.5° C Km^{-1} at Port Blair and Trivendrum. It is between 6° and 7° C Km^{-1} over the whole of India between 500 and 300 mb. Aloft, lapse rates decrease north of 25° N becoming 2° to 4° C in the layer 150 to 100 mb. South of 20° N, high lapse rates of more than 7° C Km^{-1} are between 300 mb and 200 mb, decreasing aloft to 2° to 4° C Km^{-1} between 150 and 100 mb.

2.6.3 In May, the thermal high at 850 mb shifts a little to the northwest and is centred at 23^{o} N, 78^{o} E. Temperature decreases markedly towards east and south, Gauhati and Trivendrum being 8° to 10° C cooler than Nagpur. At 700 mb, the warmest region is still in the central parts but temperature gradient decreases markedly. Due to the higher lapse rate in the continental air mass in the hottest parts, at 500 and 300 mb the lowest temperatures are over the northwestern parts of India though temperature gradients are very slack at these levels. Between 200 and 100 mb, temperatures are nearly uniform over the Peninsula but increase towards north, the gradient increasing with height.

2.6.4 In June, the highest temperatures at 850 mb are over extreme north western portions of India and adjoining Pakistan. Temperature decreases markedly over northeast India and the Peninsula. At 700 mb, this decrease in temperature is less. By 500 mb, a weak thermal ridge develops between 25^{o} and 30^{o} N and persists there about upto 100 mb. Temperature decreases south of this ridge at all levels, The level of the tropopause though gradient is less over the south Peninsula south of 15° N is about 110 mb but it is between 95 and 100 mb, north of 25° N.

In May, lapse rates are over 9° C Km^{-1} between 850 and 700 mb over the central parts of the country; it decreases to 5.5° C Km^{-1} at Trivendrum and Port Blair and 6.0° C Km^{-1} at Gauhati. In this layer the lapse rates decrease by the next month.

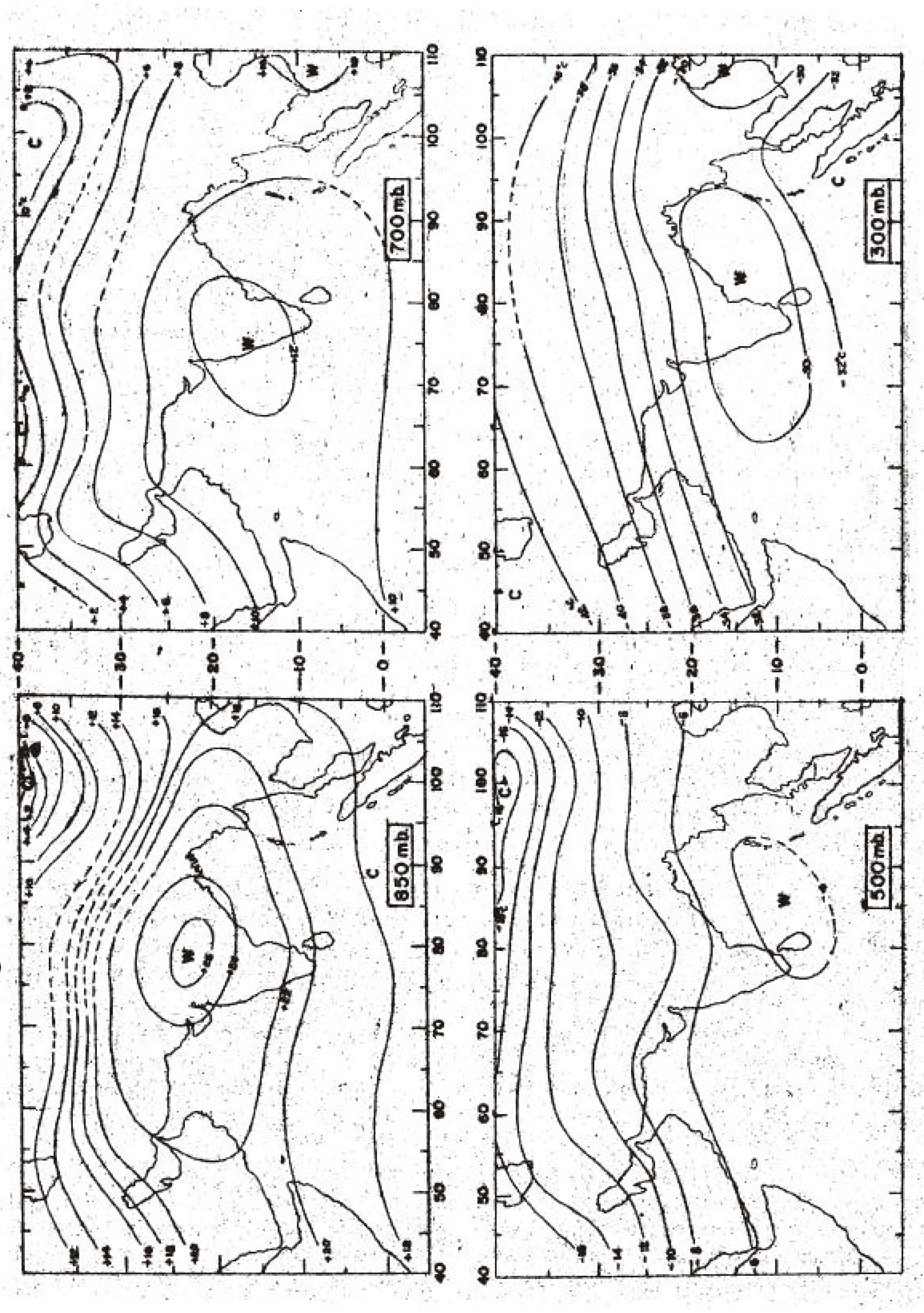


Fig. 2.9 Upper air isotherms (^{O} C) April

2.6.5 In July (Fig. 2.10), at the height of the southwest monsoon, a thermal high lies over Iran, Iraq and central parts of Arabia at 850 mb and a thermal ridge runs from it to Lat. 35^o N to the north of India. Two thermal troughs run – one along the west coast of the Indian Peninsula and the other along the Burma coast – while a thermal ridge is present over the west Bay and neighbourhood. The thermal pattern at 700 mb is nearly similar except that the troughs and ridges are absent to the south of about 20^o N and the thermal gradient over India is less. The thermal high over north America is at a little higher latitude and over west Africa at a lower latitude than over west Asia. At 500 and 300 mb, a thermal ridge runs along 25° – 30^o N with appreciable temperature gradient to the south at 300 mb. At 200 mb, the thermal ridge is along 30° – 35° N with decrease in temperature to the south. Aloft, this thermal ridge disappears. At 150 and 100 mb, the temperature increases from the south to the north, from the Equator to the Pole.

Standard deviation of temperature is about 2° C and 700 and 500 mb, increasing to 2° to 5° C aloft. Standard deviations are higher in the north than in the south.

Tropopause is highest between 25^o and 30^o N, where it is between 100 and 95 mb. The pressure at tropopause increases to the south, being 120 mb at Port Blair and 115 mb at Trivendrum. To the north of 30° N also, pressure at the tropopause increases. The temperature at the tropopause is uniform upto 30^o N, being about 198°A.

Monsoon air mass prevails over most of the country where lapse rates are near saturated adiabatic. Between 850 and 700 mb the lowest values are along the west coast and the highest near Madras which has less rain. They are 5° to 6° C Km^{-1} between 700 and 500 mb and increase with height becoming a maximum at 300–200 mb or 200–150 mb where values of 7° to 8° C Km^{-1} are reached. Aloft, they decrease. Between 150 and 100 mb, higher values are to the north, with 5.7° C Km^{-1} at Delhi and only 2.1° C Km^{-1} at Trivendrum.

2.6.6 The warmest region shifts markedly in the monsoon period, more so in the upper troposphere. The thermal ridge at 200 mb over Tibetan region is an important feature.

2.6.7 Banerjee and Sharma (1967) find from harmonic analysis of monthly upper air temperatures over India, that in the first harmonic, there are two oscillations, one at the surface and the other in the upper troposphere. The first, due to ground heating, is confined to below 700 mb and is in phase with the movement of the Sun. The other is between 600 and 150 mb with maximum amplitude at 350 mb. The maximum occurs from mid–July to mid–August. Its amplitude decreases at 300 mb to one–eighth from 30° N to 10° N. This oscillation is apparently due to a different physical process.

2.7 Upper Winds

2.7.1 The sub–tropical high pressure belt passes over and near India throughout the year. It is masked in the lower troposphere in southwest monsoon season by the effect of continental summer heating. In the middle and upper troposphere, it is prominently seen and defines the wind distribution.

2.7.2 In April (Fig. 2.11 a and b), a trough line runs along 77° E at 900 mb over the Peninsula as at sea level. Weak high cells probably exist over the central Bay and the Arabian Sea. The sub–tropical ridge appears over land near about 18° N at 850 mb and persists at 700 mb. Aloft, it gradually shifts southwards to 8° N at 200 mb. At this level, the anticyclone from the east extends upto Sri Lanka. A trough is also present over northeast India between 900 and 800 mb with axis along 87^o E. Westerlies increase with height over northern India and the adjoining Peninsula. Maximum speeds are about 40 kt near 25° N at 300 mb and 50–60 kt between 25° and 30^o Nat 200 mb, decreasing slowly aloft.

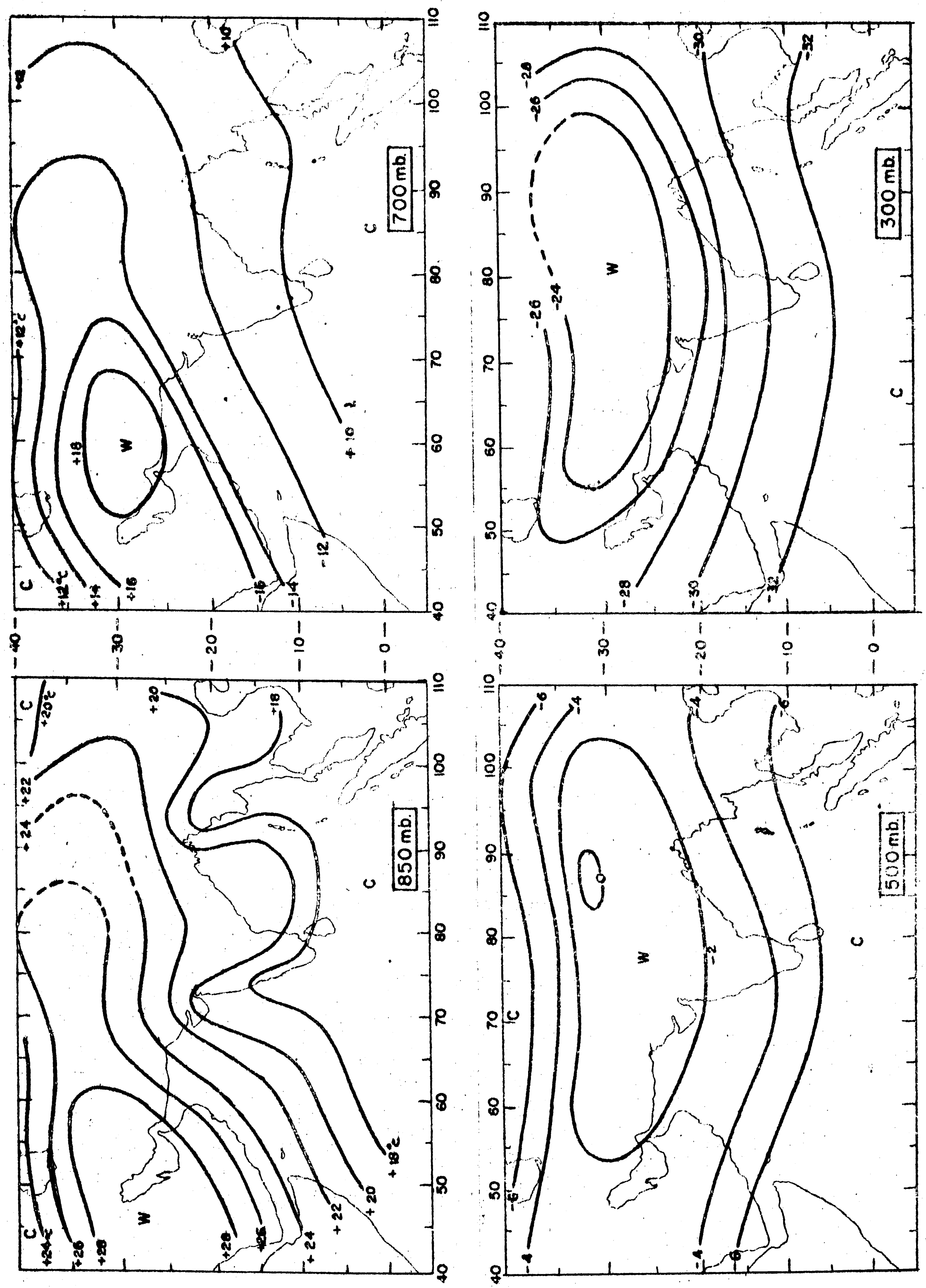


Fig. 2.9 Upper air isotherms (^{0} C) July

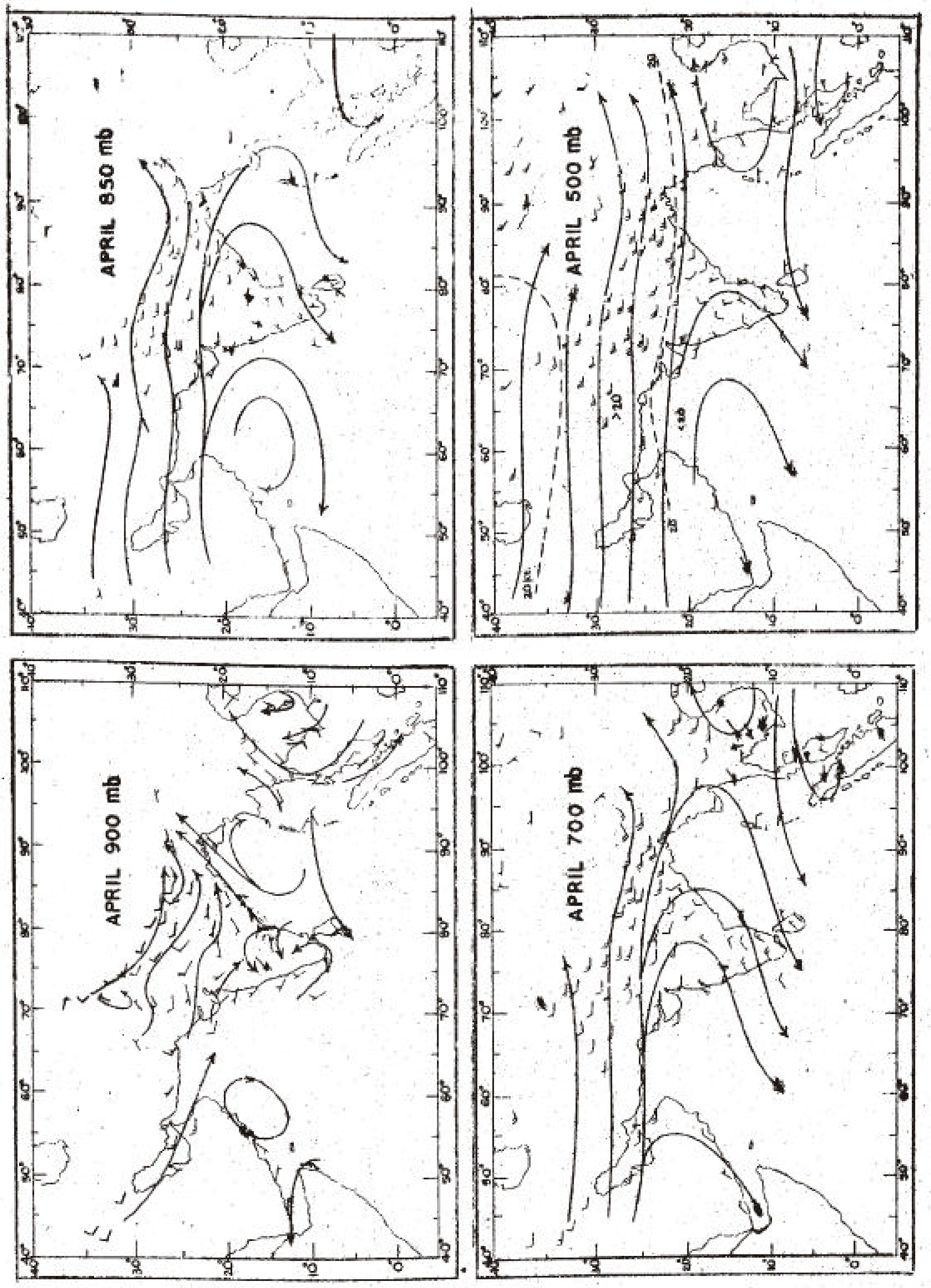


Fig. 2.11(a) Upper Winds

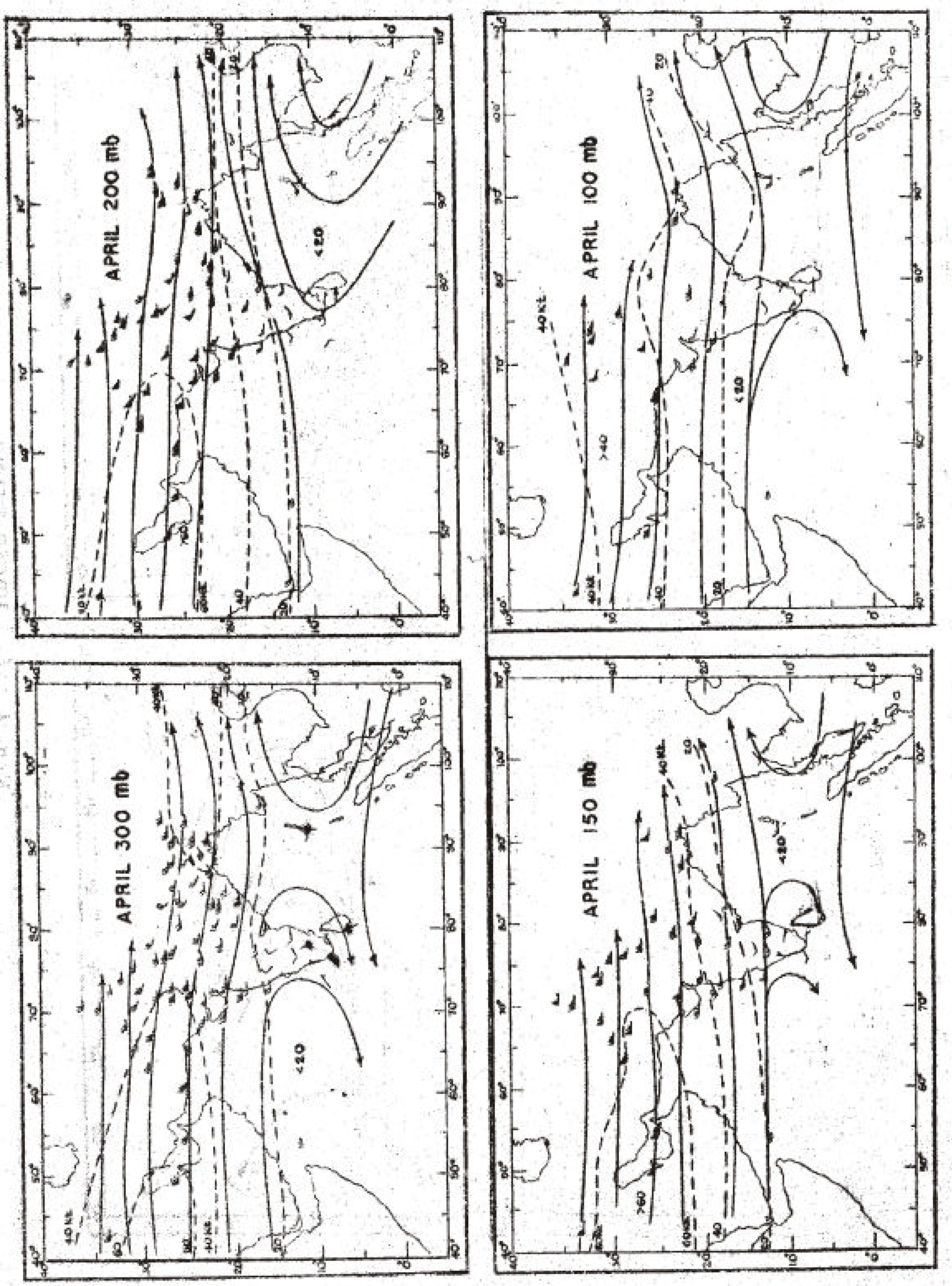


Fig. 2.11(b) Upper winds (missing)

2.7.3 In May, a trough line runs from Gorakhpur to Kanyakumari at 900 mb A weaker trough also runs from Multan to Badin. Winds all over the country are from some westerly direction. At 850 mb, the only troughs are one over northeast India along 87° E and another roughly along 12° N in the Bay and the Peninsula. The latter trough persists upto 500 mb. The sub–tropical ridge from Arabia extends into the Indian Peninsula at 850 mb and persists upto 500 mb at about the same position. Aloft, this ridge line shifts southwards reaching 14^{o} N at 200 mb, but is again displaced northwards to near 20^{o} N at 100 mb. Aloft of 200 mb, the ridge over the Peninsula is an extension from the east rather than the west. The westerlies over northern India are strongest between 200 and 150 mb, reaching 50 kt at 200 and 150 mb near 30^{o} N. Speed decreases aloft. Easterlies over the Peninsula are not over 20 kt.

2.7.4 In June, by the end of which the monsoon is established over practically the whole of the country, winds are westerlies at 900 mb. The monsoon trough over northern India between westerlies to the south and easterlies to the north is established only towards the end of the month. A trough is present ,over Pakistan and adjoining Iran which persists upto 800 mb. Between 850 mb and 500 mb, a trough line exists over the northeast India between 85^{o} and 90^{o} E. The sub tropical ridge from the west begins extending into Rajasthan at 700 mb and reaches. Gujarat at 500 mb. Weak easterlies appear at 300 mb south of the ridge line along (22 N). The anticyclone from the east extends to 85° E at 300 mb and 70^{o} E by 150 mb. 'The ridge line is at 25^{o} N at 150 mb and near 28^{o} N at 100 mb. Easterlies well to the south of this ridge line strengthen with height and reach 50 kt in some belt over the Peninsula at 100 mb.

2.7.5 In July (Fig. 2.12 a and b) , the monsoon trough runs from Delhi to Calcutta at 900 mb. Westsouthwesterly to westnorthwesterly winds prevail to the south of it and southeasterly winds to north. A weak trough is also present over Pakistan and neighbourhood. Westerly winds over the Peninsula increase with height from ground and reach a maximum between 900 and 800 mb. This level is near 900 mb along the west coast and increases to 800 mb in the eastern Peninsula, Maximum speeds are between 20 and 25 kt. A similar wind maximum near 900 mb has also been found in the Arabian Sea south of 20^{o} N, particularly in the western portions. The monsoon trough shifts south with height and is near about 23° N at 700 mb but becomes diffuse above 500 mb over the Peninsula. The trough over Pakistan is not present at 700 mb; instead the sub–tropical high from the west extends into northwest India. Thus the warm surface low over Pakistan and neighbourhood is replaced by the subtropical high at 700 mb. At 500 mb, this ridge, the easterlies over northern India and a trough from east central Arabian Sea to Orissa are the chief features. Winds at this level are weak. .At 300 mb, apart from the western ridge along 30^{o} N over northwest India and Pakistan, winds are easterly over the whole country. Between 200 mb and 100 mb, another ridge from the east develops to the east of longitude 75^{o} E (with ridge line at 30° N). Easterly winds strengthen with height from 200 mb reaching a maximum at 100 mb. Speeds are between 60 and 80 kt over the Peninsula between 150 and 100 mb. Upper tropospheric easterlies are seen in Gauhati only in mid–monsoon. The strengthening of easterlies with height in Uttar Pradesh and Bihar is less than in central parts and the Peninsula where thermal gradients are stronger.

2.7.6 While the pattern remains the same in August, winds are slightly weaker between 850 and 500 mb. In September, winds are weaker at 900 mb also. The anticyclone from west now extends at 700 mb upto 20° N, 80° E. A cyclonic circulation also covers the north and adjoining central Bay, instead of the trough near 20^{o} N in earlier months. At 500 mb, the ridge has shifted 3° to 4° southwards and middle–latitude westerlies are spreading upto 27^{o} N,

2.7.7 Joseph and Raman (1966) find that the lower tropospheric westerlies over the Peninsula may develop into low level jet stream on many days in July, with a core at about 1.5 km and speeds of 40–60 kt in the core. The jet appears between 8^{o} and 18^{o} N. The wind shear below the core is more than above (Fig. 2.13).

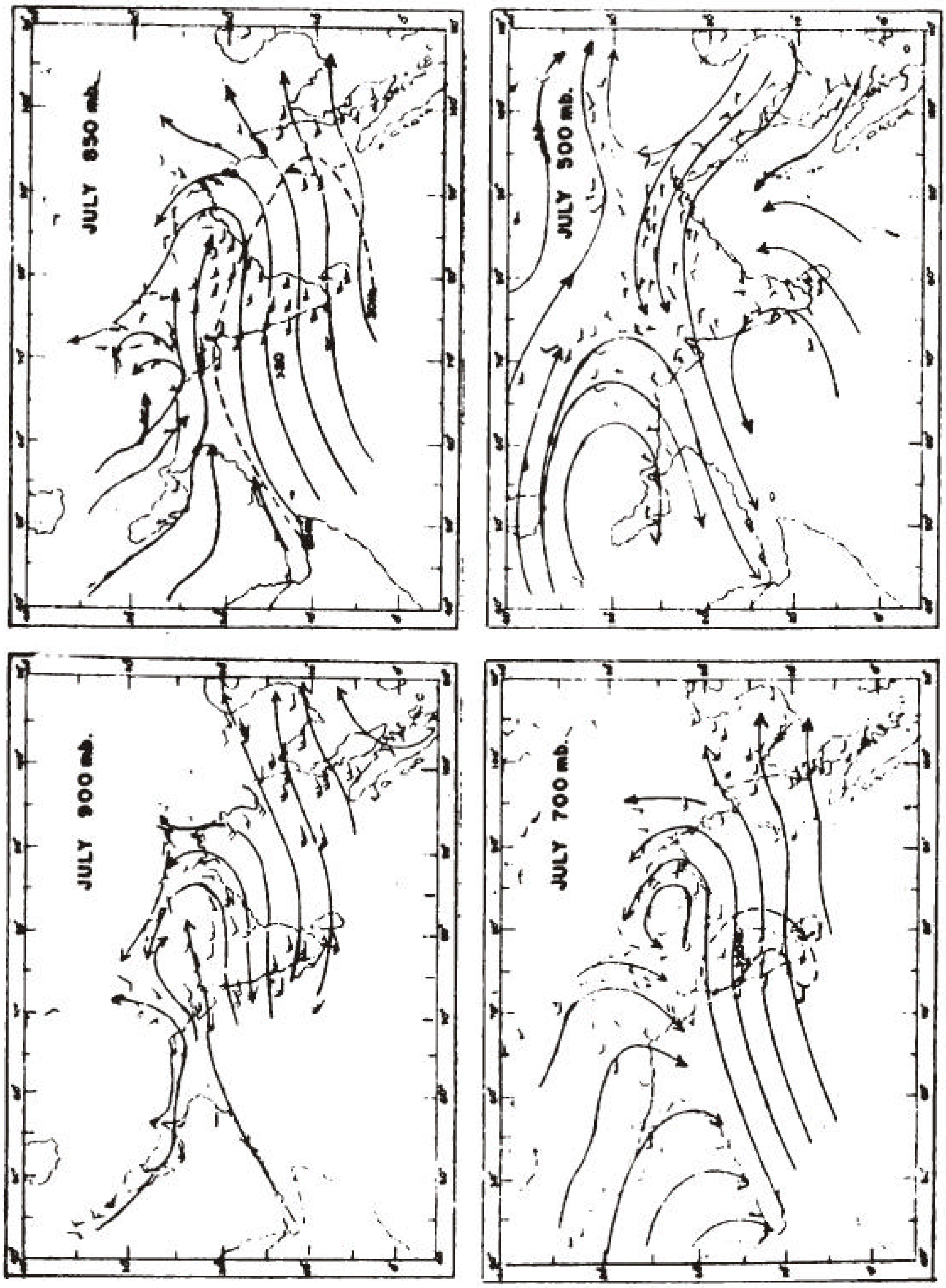


Fig. 2.12 (a) Upper Winds

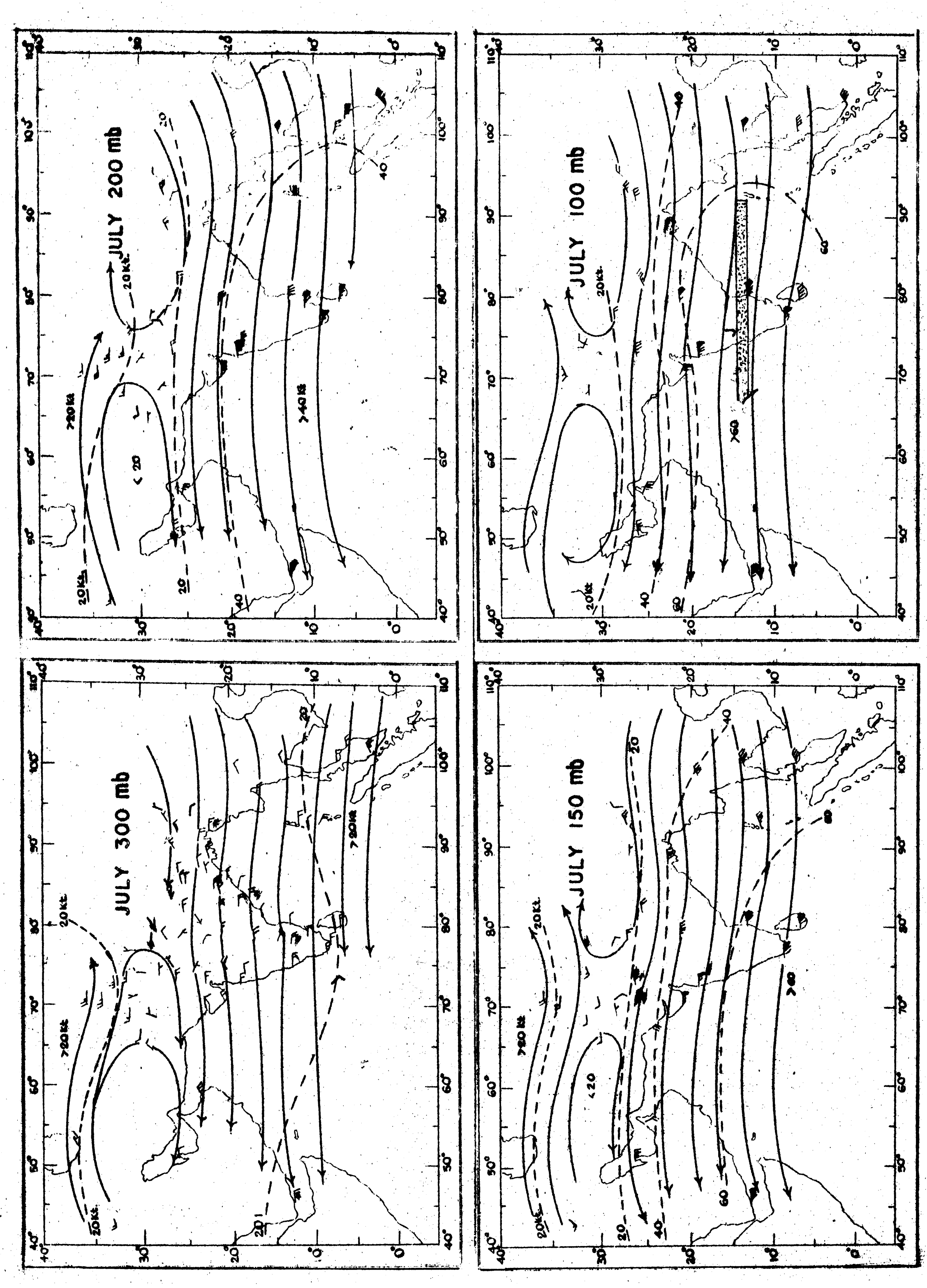


Fig. No.. 2.12(b) Upper Winds

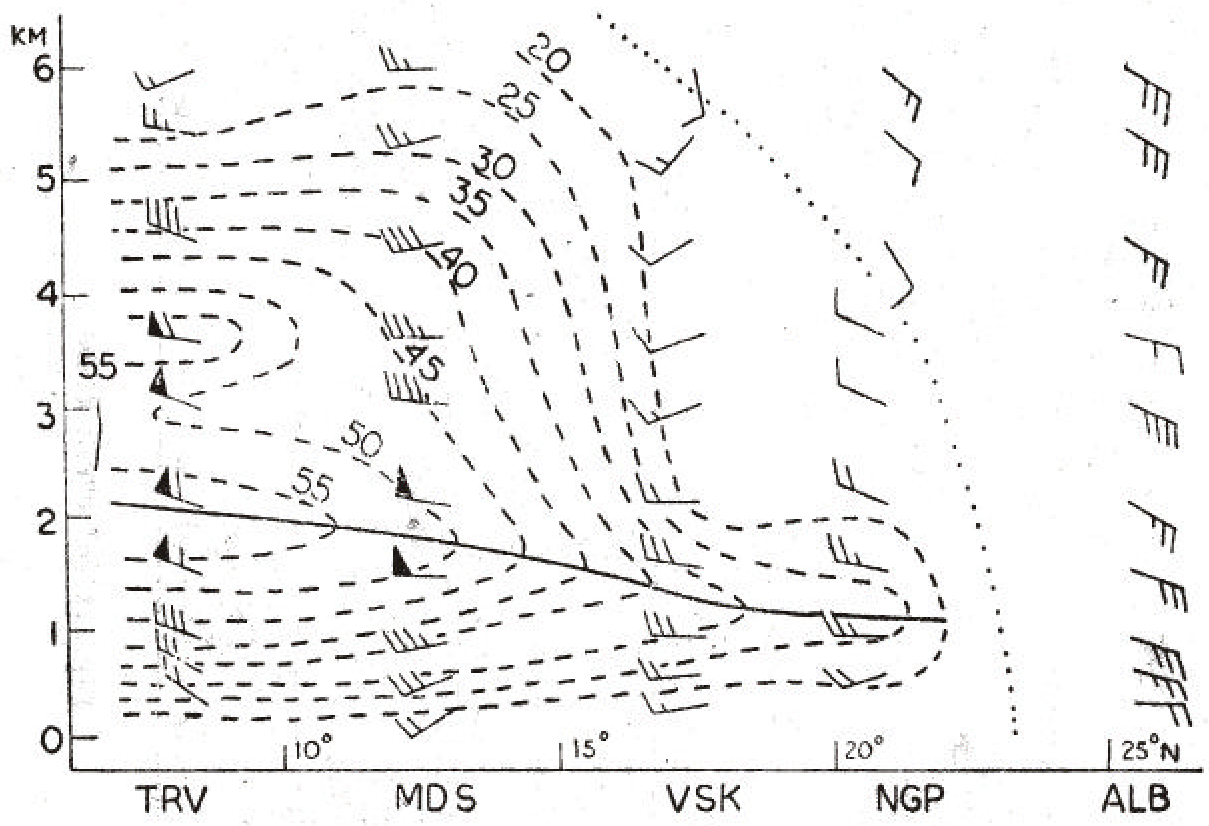


Fig.2.13 - Vertical wind cross section using rawin data only, along 80°E for 13 July 1962, 00 GMT.
(The line of dots shows the approximate line of separation between the easterlies and westerlies. The approximate position of the level of the maximum wind, as obtained from the isotach analysis is marked by the thick line).

2.7.8 In the middle troposphere, winds are very unsteady in July and August but the upper tropospheric zonal easterlies and lower tropospheric westerlies are steady south of 25° N.

2.7.9 To sum up, the chief features of the upper air flow in the season are the monsoon trough sloping southwards with height and the ridges over northern India, Peninsula has lower tropospheric westerlies and upper tropospheric easterlies. The depth of the westerlies increases from the southern tip of the country upto 13° N and decreases further north (Krishna Rao and Ganesan, 1953). The flow from the Arabian Sea across the Peninsula is usually referred to as the 'Arabian Sea monsoon' and that to the north of the trough after some travel over the Bay as the 'Bay monsoon current'. The ridge in the upper troposphere near 30° N and east of 75° E is referred to as the 'Tibetan High'.

2.8 Easterly Jet Stream

2.8.1 Near 100 mb, strong easterlies blow to the south of 25° N in the southwest monsoon period, which concentrate into a core of high winds known as the 'Easterly Jet Stream'. Krishna Rao (1952) first pointed out that Nagercoil shows easterly winds of 50 m sec^{-1} at about 75 mb (19 km) which constitute a part of the easterly jet stream over south India between 7° and 18° N, Koteswaram (1958) studied the easterly jet stream over India in detail. These strong easterly winds near the tropopause are a characteristic feature of Asia and Africa in summer and are not to be found over the Atlantic or the Pacific. In July, winds are easterly 10 to 20 kt at 300 mb south of 27° N and westerly to the north of 32° N with transition in between. As the temperatures decrease to the south from 25° N, easterlies strengthen over the Peninsula to 40 to 50 kt by 200 mb,, At 200 mb, the temperature decrease southwards starts at 30° N itself. At 150 and 100 mb, the temperature gradient south of 20° N is very flat but the southward decrease persists to the north. This northward increase of temperature extends almost upto the Pole. This is a global feature. However, the steep gradient is between 30° and 50 N. The easterlies increase with height upto about 150/100 mb over the Peninsula, at which level they are a maximum. At Madras, the mean wind is 85 /70 kt at 16 km, while at 14 km it is 85°/60 kt and at 18.0 km 85°/50 kt. From the available scanty data, the core of the easterly jet over India appears to be near about the latitude of Madras (13 N) at 100 mb. However, at Trivendrum (08°30' N) the maximum speed is at 14.0 km (150 mb), being 80°/65 kt, higher than that at Madras at that level. Aloft, the speed is less than at Madras. At Gan Island (l° S), the maximum speed of 55 kt is at 150 mb as at Trivendrum and the speed decreases sharply to 25 kt by 100 mb. At Bombay (l9° N), speeds at 14, 16 and 18 km are respectively 55, 70 and 60 kt. The speed at 18 km is higher than that at Madras. Thus while the core of the easterly jet is near 13° N at 100 mb, the highest speeds at 150 mb are further to the south and above 100 mb further to the north.

2.8.2 Even at 100 mb, strong temperature gradient prevails over northern India. Hence easterly winds increase with height even above 100 mb. New Delhi (28 30' N) shows an increase of speed from 10 kt at 14 km to 35 kt at 24 km. The easterlies may not be extending to north of 30° N in the upper troposphere, as the few available winds at Ambala (30° 30' N) show westerlies between 14 and 18 km. The easterlies above 18 km over Delhi may be a part of the global stratospheric easterlies in summer.

2.8.3 The easterly jet stream over the Peninsula is the effect of the southward decrease of temperature in the entire troposphere. This thermal gradient is effective in reversing the moderately strong westerlies of the lower troposphere and building speeds of 80 kt by 100 mb.

2.8.4 The horizontal shear at 100 mb to the south of the jet between Madras and Trivendrum is 0.09 hr^{-1}. To the north, between Madras and Visakhapatnam it may be of the order 0.07 hr^{-1}. The vertical shear between 9 and 16 km varies between 15 and 20 hr^{-1}, the maximum shear being in the layer 5 km below the core. Above the core, between 16 and 18 km the shear is about 22 hr^{-1}.

2.8.5 Temperature and wind data are too scanty to reliably picture the vertical circulation near the jet stream. The same is the case regarding diffluence and confluence in the jet stream. The direction of winds at 100 mb is with in five degrees of 85° at places to the south of 23° N.

2.8.6 In June, Madras, Port Blair, Nagpur and Bombay show increase in the speed of the easterlies from 9 to 16 km. Trivandrum shows maximum speed of 65 kt at 14 km, speed being higher than at Madras at this level. Ambala has fairly strong westerlies between 14 and 18 km. Between 25° and 30° N, westerlies occur at some stations at 14 km and even at 16 km. Easterlies which appear at 18 km at Delhi are weak. The Easterly Jet in June is thus confined to a little more southerly latitude and is also at a higher level in the Peninsula than in July. Speeds are less to the north of 15° N in June but more at some levels to the south.

2.8.7 Conditions in August are similar to July. In September, strong easterlies are confined to south of about 21° N; speeds are less and westerlies appear at levels like 14 km north of 25° N. At Minicoy a maximum speed of 45 kt is at 14 km.

2.9 Meridional Circulation

2.9.1 Mean meridional components are usually weaker than the zonal winds and are not prominently seen in the mean wind charts. In view of their importance for exchange of heat, momentum etc. between different latitudes, meridional circulation over India is separately discussed.

2.9.2 Rao (1962) has studied in detail the meridional circulation associated with southwest monsoon. The mean meridional components for July at rawin stations over India are shown in Fig. 2.14. Except at Trivandrum and Nagpur, at all stations right upto 32° N, there is flow from south in the lower troposphere; aloft, the flow is from north, except in northwest India (Jodhpur, Delhi and Amritsar). At these places, the flow from north in mid-troposphere is capped by flow from south again in upper troposphere. From these, we may picture that to the south of 25° N, there is monsoonal meridional circulation with southerlies below and northerlies aloft. To the north of 25° N latitude, such a monsoon circulation is confined to the lower part of troposphere with the Hadley Cell circulation aloft. The northerlies at surface at Trivandrum are induced partly by the air flow around the southern tip of the western ghats and the presence of a clockwise circulation just south of the equator in lower troposphere at 70° E.

2.9.3 Figure 2.15 presents a vertical cross-section of the mean meridional components for the mean Indian longitude in January and July. In January, over India to the south of about 30° N, northerlies of the direct cell prevail from the surface upto 300 mb and southerlies aloft. Towards the equator these meridional components are stronger, much to the south of the centre (12 N) of the direct cell as found on global scale. Two types of direct circulation have been recognised. Of these, the type associated with the predominant easterly trades has a layer of southerlies at the middle of the general northerlies, the main layer of southerlies being much higher above. Indian data in January also show either weak southerlies or weakening of northerlies at the middle of the layer of northerlies. Along 50°E, upper tropospheric southerlies are absent at Aden and Bahrain where northerlies reach upto the tropopause.

2.9.4 In July (Fig. 2.15) between 12°N and 26°N a simple circulation of southerlies below and northerlies aloft is found over the Indian region, which may be called the 'monsoon cell'. The usual direct cell is very much shrunk in its latitudinal extent and is seen only between 25° and 40° N. Its presence at 30° N is clear from the northerlies in mid-troposphere and southerlies aloft, though weak southerlies of the monsoon cell extend so far northwards (i.e. upto 30° N) in the lowest layers of the troposphere. In this month, Aden and Bahrein exhibit both the lower tropospheric northerlies and upper tropospheric southerlies of the direct cell.

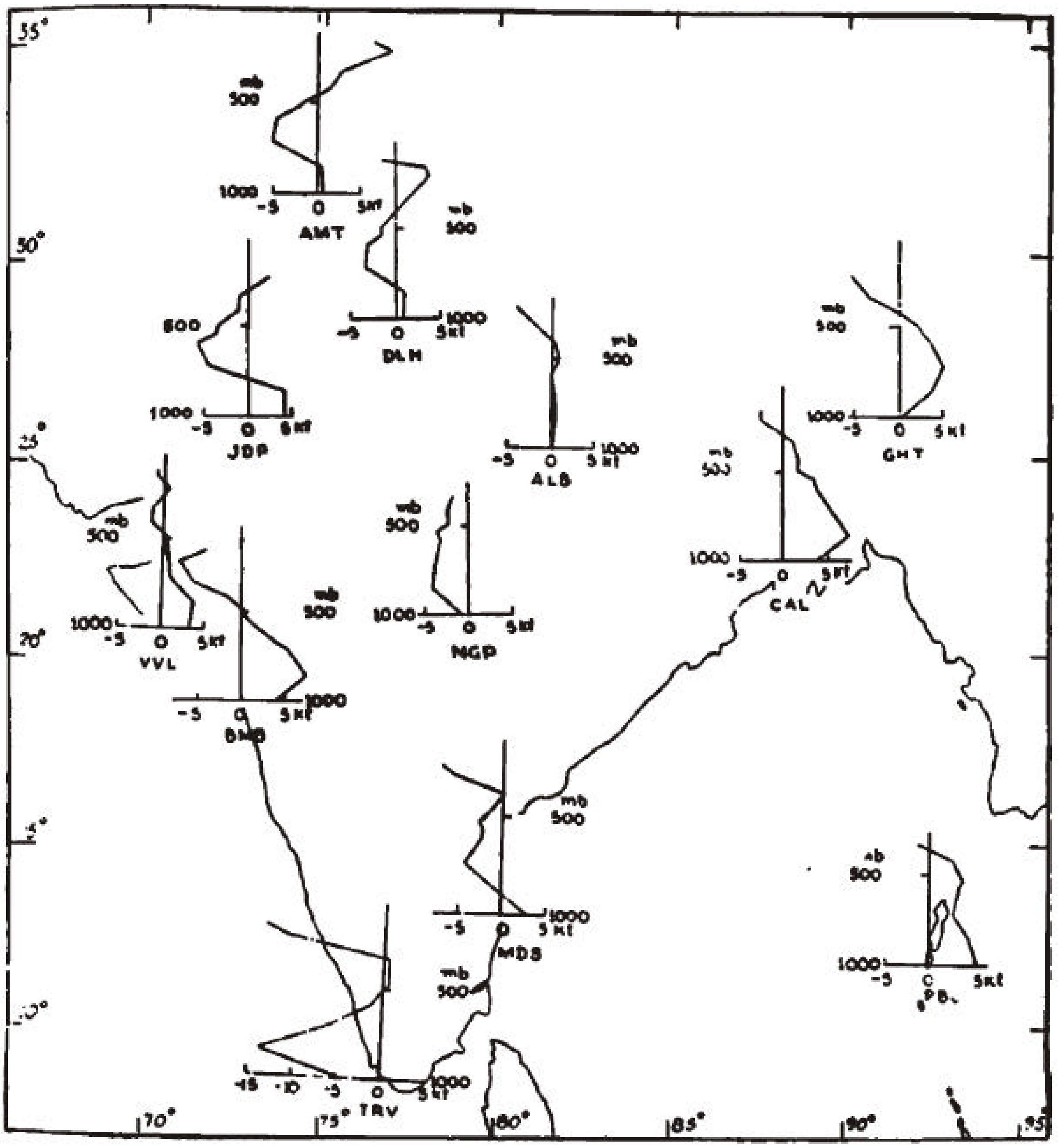
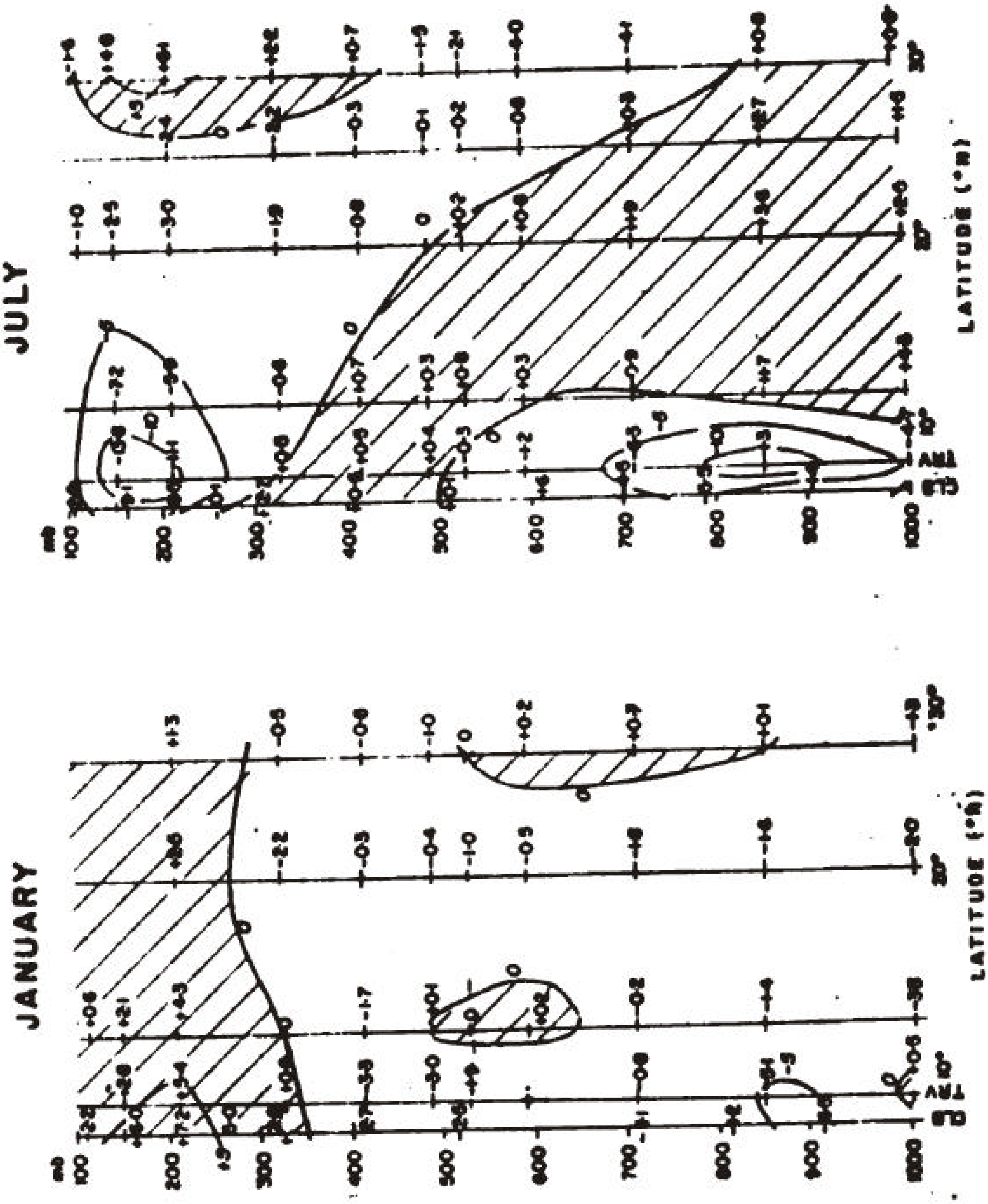


Fig. 2.14 - Mean meridional components for July.
(Component from south is positive)



Fi.g 2.15 Mean Meridional circulation over India

South of 10° N in the Indian area there is a deep layer of northerlies commencing almost from the surface, southerlies above and a marked layer of northerlies aloft. This may not be representative of all longitudes (in the Bay or the Arabian Sea).

2.9.5 Tucker (1959) had constructed the meridional circulation of the northern hemisphere for summer and found between 10^{o} and 20^{o} N, northerlies at surface, southerlies above that, and weak northerlies once again above. This would suggest that the monsoon type of cell, southerlies below and northerlies aloft, was weakly protruding in the middle layers of the atmosphere, into the Hadley Cell further north. In the Indian area the monsoon meridional circulation is below the Hadley Cell north of 25° latitude, while to the south there is no Hadley Cell at all but only the monsoon cell. The equatorial southerlies are stronger in the Indian area and the Hadley Cell is further to the north and smaller in meridional extent. North of 25° N where the Hadley Cell is noticed, it overlies the monsoon cell; the northerlies of the Hadley Cell are of continental origin and warmer.

2.10 <u>Some Derived Parameters</u>

2.10.1 Awade and Asnani (1973) have fitted a polynomial to the mean zonal winds in July between 850 and 150 mb between the latitudes 5^{o} and 25^{o} N along the meridian 77.5^{o} E, as given below :–

$$u = (a_0 + a_1 f + a_2 f^2) + (b_0 + b_1 f + b_2 f^2)\log_{10} p + (c_0 + c_1 f + c_2 f^2)(\log_{10} p)^2$$

The value of coefficients obtained by them are

$$a_0 = -673.20, \quad a_1 = -56.94, \quad a_2 = 2.93,$$
$$b_0 = 426.43, \quad b_1 = 39.76, \quad b_2 = -2.93,$$
$$c_0 = -65.41, \quad c_1 = -6.73, \quad c_2 = 0.34,$$

The flow in this region is mainly zonal and values computed by the above formulae agree well with observations. Vertical wind shear $\partial u/\partial p$ as calculated from the polynomial are given in the table below :–

Table 2.3
Vertical wind shear
in kts per 100 mb

Lat.	**(mb)**								
(°N)	**850**	**800**	**700**	**600**	**500**	**400**	**300**	**200**	**150**
5	2.3	2.7	3.7	5.2	7.5	11.2	18.2	34.3	52.3
10	2.2	2.6	3.7	5.3	7.7	11.7	19.2	36.4	55.6
15	2.1	2.5	3.5	4.9	7.2	10.8	17.7	33.4	51.0
20	1.9	2.2	3.0	4.1	5.8	8.6	13.7	25.5	38.6
25	1.6	1.8	2.2	2.7	3.6	4.9	7.3	12.6	18.2

The large magnitude of the vertical wind shear below the easterly jet maximum near about 10^{o} N is well brought out by this representation. The decrease in vertical shear to the north of 10^{o} N is quite marked; at 150 mb, near 25^{o} N, the shear is a third of that at 10° N.

$\partial u / \partial p$ shown in Fig. 2.16 brings out that in whole troposphere upto 150 mb, zonal winds are maximum at 10 ^{o} N. To the south, relative vorticity due to wind shear is anticyclonic upto 300 mb and cyclonic aloft. In the northern latitudes it is cyclonic upto 500 mb and anticyclonic at 200 and 150 mb. Vorticity in the vertical as shown by the mean flow is well marked. Ascending particles will experience increasing cyclonic vorticity south of 10^{o} N and decreasing cyclonic vorticity to north. Awade and Asnani also find that the absolute vorticity on an isentropic surface is positive and conclude that the normal flow is inertially stable as per the criterion given by Kuo (1956).

2.10.2 Anjaneyulu's (1969) computation of divergence and vorticity from mean winds at 850 and 200 mb is shown in Figs. 2.17 and 2.18. At 200 mb, anticyclonic vorticity prevails over the whole region, while at 850 mb cyclonic vorticity is noticed to the east of 75^{o} E in and around the monsoon trough and the west Bay. Weak convergence prevails at 850 mb over most of the country except northwest and in a belt from north Bay across Orissa. The latter is intriguing, as this is the area of genesis of monsoon depressions. At 200 mb, there is convergence in the northwestern portion of the sub–continent and divergence elsewhere. Krishnamurthy (l97l) finds mean divergence in 1967 monsoon at 200 mb over the whole of northern, India, implying ascending motion, in contrast to Anjaneyulu's computation for the northwestern parts of the sub–continent. Well defined mean convergence is present in Krishnamurthy's computation over the Arabian Sea and further to the west, as the easterly jet decelerates westward.

Miller and Keshavamurthy's (1968) computation for the layer upto 900 mb in July, shows convergence in the areas around the heat low and in another belt from Bengal and Orissa to the central parts of the Peninsula and divergence elsewhere.

2.10.3 Das (1962) computed the mean vertical motion over the Indian sub continent for July using the mean flow pattern on adiabatic assumption. He finds pronounced ascent over northeast India and subsidence over the northwestern parts of the sub–continent. He took into account ascent along the slope of the ground which was even of the order of 5–10 cb (12 hr)$^{-1}$. The vertical velocity decreases with height, rather fast upto 800 mb, and to zero at 100 mb as assumed. This vertical distribution is different from the synoptic pattern of vertical velocity increasing upto the level of non–divergence. Vertical motion computed from non linear equations for mean motion cannot perhaps be the mean of the daily vertical motions. Computation of vertical motion from conservation of vorticity and entropy should take into account latent heat release, as ascent of air is along moist adiabat, Mean vertical velocity at 900 mb as found by him is shown in Fig. 2.19. While Das found subsidence at 900 mb, Ramage (1966) found from mean winds, net ascent below 700 mb over the eastern portions of Pakistan associated with the heat low and descent above.These differences bring out the necessity of more exhaustive studies on the mean vorticity, divergence and vertical velocities of the flow from homogeneous data.

2.10.4 Rangarajan and Mokashi (1966) find the standard deviation of zonal winds (S_x) generally more than that of meridional components (S_y) at all levels in the monsoon. At 1.5 and 3.0 km the correlation is significantly negative. Stronger westerly winds are associated with more southerly component. At Trivendrum, in the easterly jet, S_x is nearly twice as large as S_y.

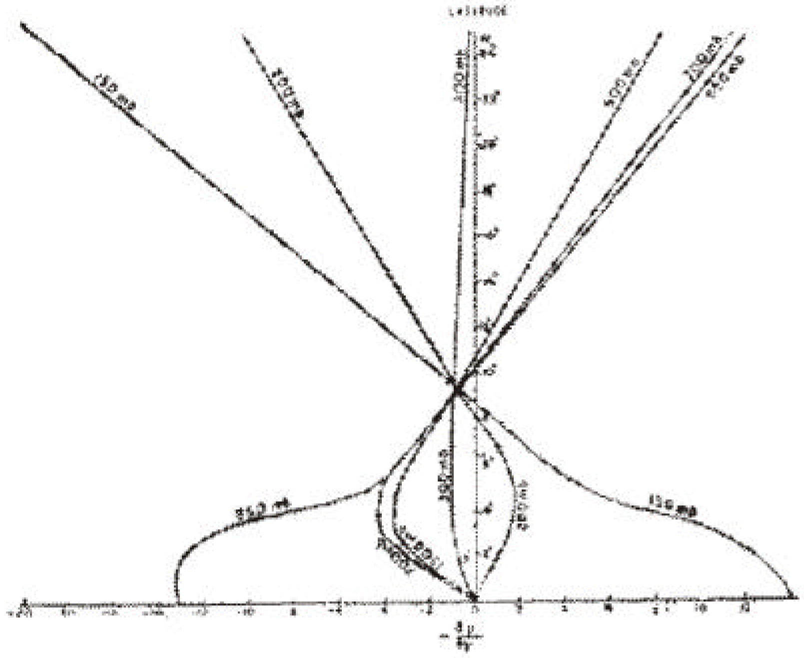
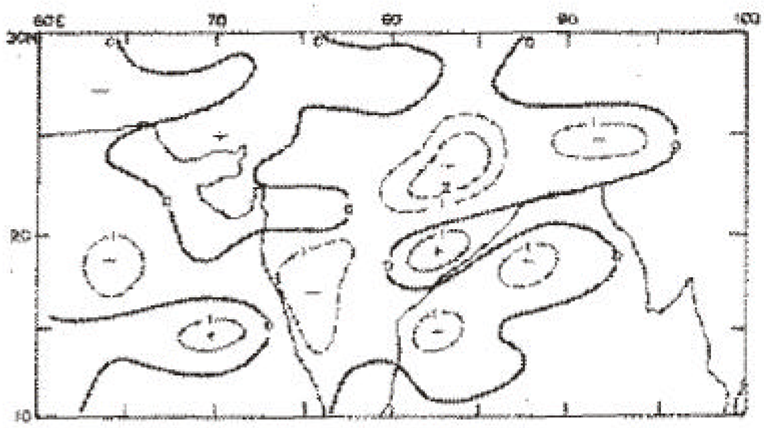
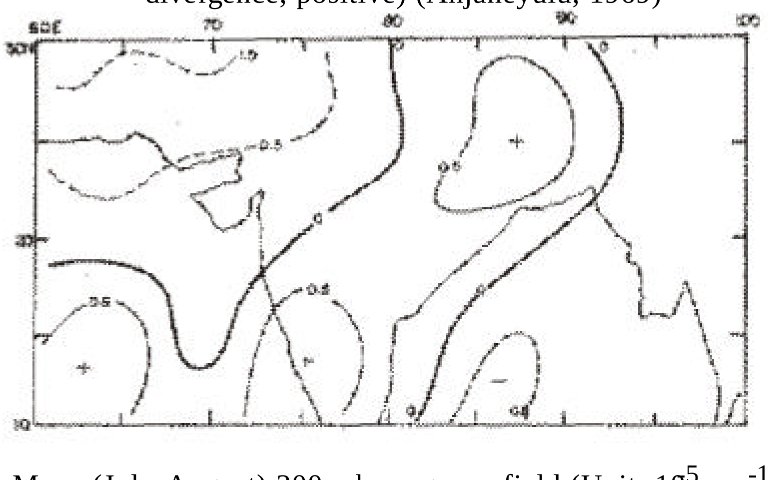


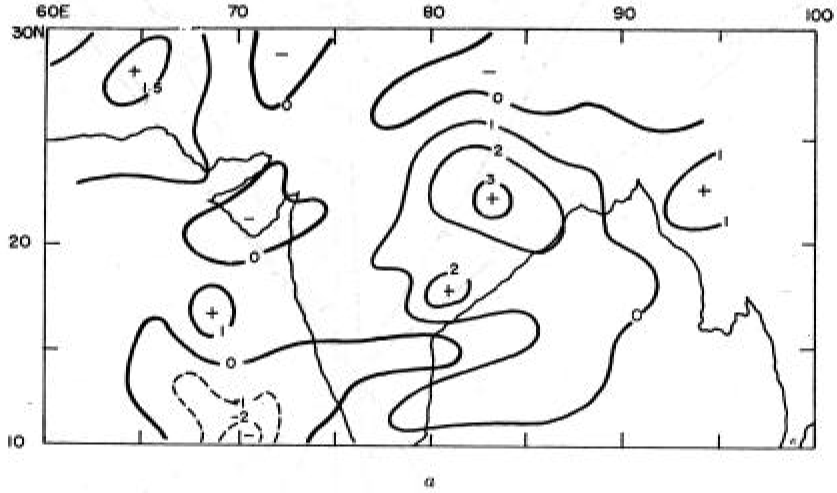
Fig. 2.16 - Values of $\partial u/\partial y$ on isobaric surgace (Unit 10^{-5} sec^{-1})



(a) Mean (July-August) 850 mb vergence field (Unit 10^{-5} sec^{-1})
divergence, positive) (Anjaneyulu, 1969)



(b) Mean (July-August) 200 mb vergence field (Unit: 10^{-5} sec^{-1}
divergence, positive) (Anjaneyulu, 1969)
Fig.2.17



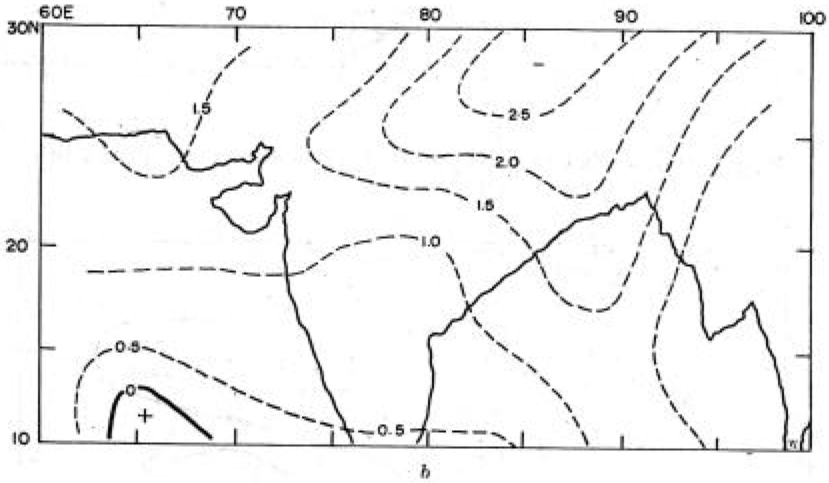


Fig. 2.18 (a) 850 mb relative vorticity fields (Unit 10 $^{-5}$ sec $^{-1}$, cyclonic positive), (Anjaneyulu, 1969)

(b) 200 mb relative vorticity fields (Unit 10 $^{-5}$ sec $^{-1}$, cyclonic positive), (Anjaneyulu, 1969)

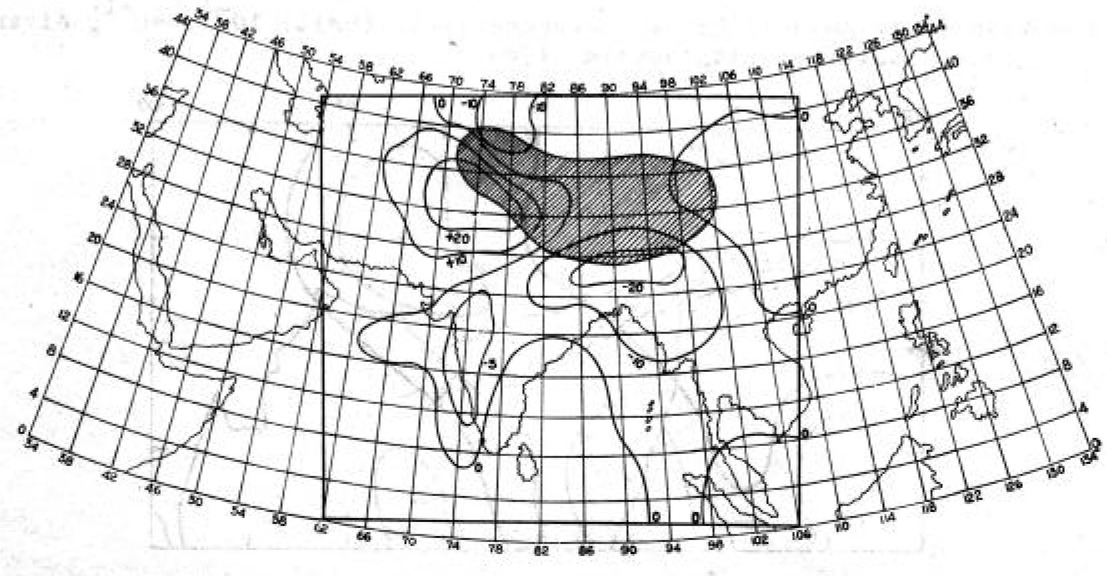


Fig. 2.19Mean vertical velocity in cb (12 hr) $^{-1}$ 900 mb (Das, 1962)

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CHAPTER 3

ONSET AND WITHDRAWAL OF THE MONSOON

3.1 Perceptible changes on account of the monsoon, are (i) winds from about southwest, (ii) decrease in temperature from the heat of April and May and (iii) increase in rainfall. The first is mainly of interest to mariners, while other two have more general impact. Dates of onset can be fixed by the changes in any of these features for any year or from climatological means of these elements for a sufficiently long period. Development of these three features or more recondite aspects like air mass, meridional circulation or easterly jet stream are not simultaneous. Westerlies set in the Arabian Sea in May but the rains only the next month. The India Meteorological Department has fixed the dates of onset and withdrawal with reference to the rather sharp increase and decrease respectively shown by the five–day means of rainfall and the changes in the circulation. The appearance of monsoon synoptic features far ahead of the normal date is referred to as "temporary advance", if it is not maintained. The monsoon rains are some times not easy to distinguish from premonsoon thundershowers. Consequently, the dates of onset and withdrawal of monsoon cannot be fixed uniquely.

3.2 During the course of October there is once again a change from the circulation pattern of the southwest monsoon over the Indian area but the rains increase over south India. This change in circulation should be rightly regarded as the end of the southwest monsoon and it is not appropriate to identify the much later decrease of rainfall in south India as the withdrawal of the southwest monsoon. The displacement of the monsoon air by continental air mass and development of anti–cyclonic flow would determine the dates of withdrawal over north and central India.

3.3 In the interior of the Peninsula, onset of the monsoon may not be seen immediately as a striking increase of rain. Nevertheless, the synoptic practice of fixing the date of onset of monsoon in any year in different areas has a physical basis. The development of the circulation pattern and invasion of air mass are basic criteria and these create conditions favourable for occurrence of more frequent and increased amounts of rain. Whatever may be the uncertainty in forecasting monsoon advance, actual onset can be generally fixed within acceptable limits.

3.4 Figure 3.1 shows the normal dates of the onset and withdrawal of the monsoon as determined from rainfall increase and decrease and other synoptic features. Figure 3.2 shows the five–day rainfall figures at some stations to illustrate increase of rainfall at the onset. Both the dates of onset and withdrawal vary from year to year and may occur even outside the period of June to September, usually regarded as the southwest monsoon period. According to Ramdas, Jagannathan and Gopal Rao (1954), the standard deviation of the dates of establishment of monsoon along the west coast south of 20° N is 6 to 7 days. The earliest date was 17–22 days before, while the most delayed arrival was 10–13 days after the normal date. Their criteria, being slightly different from synoptic practice, resulted in earlier dates being fixed for the establishment of monsoon. Bhullar (1952) gives the standard deviation for the advance into Delhi as 7–8 days from the mean date of 2nd July which he determined from the increase of rains at a number of raingauges around. To illustrate the variability of the advance of monsoon, a histogram giving the dates of onset over Bombay from 1879 to 1975 is given in Fig. 3.3. The extreme dates of onset of monsoon for some parts of the country are given in Table 3.1.

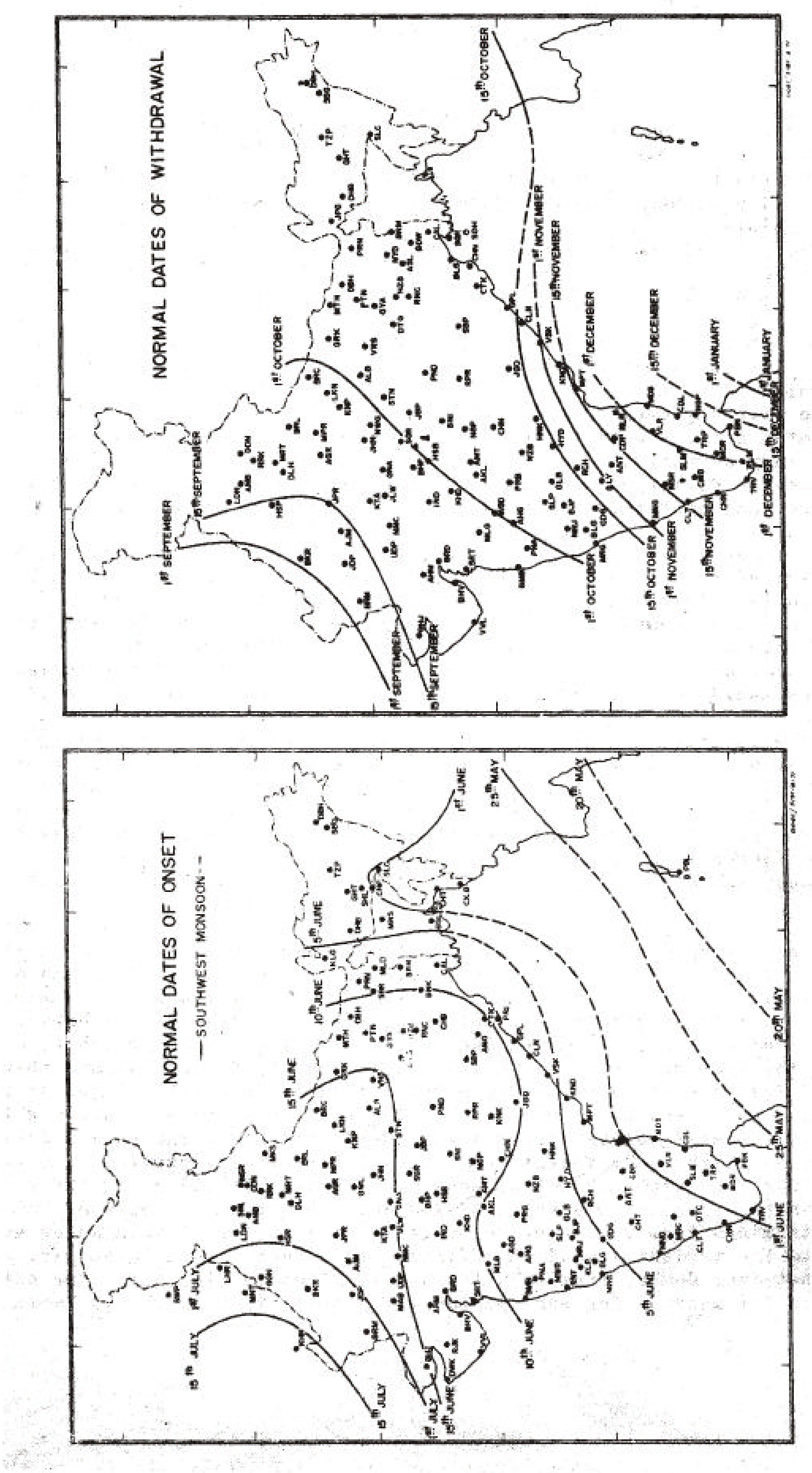

NORMAL DATES OF WITHDRAWAL
NORMAL DATES OF ONSET
— SOUTHWEST MONSOON —

Fig. 3.1

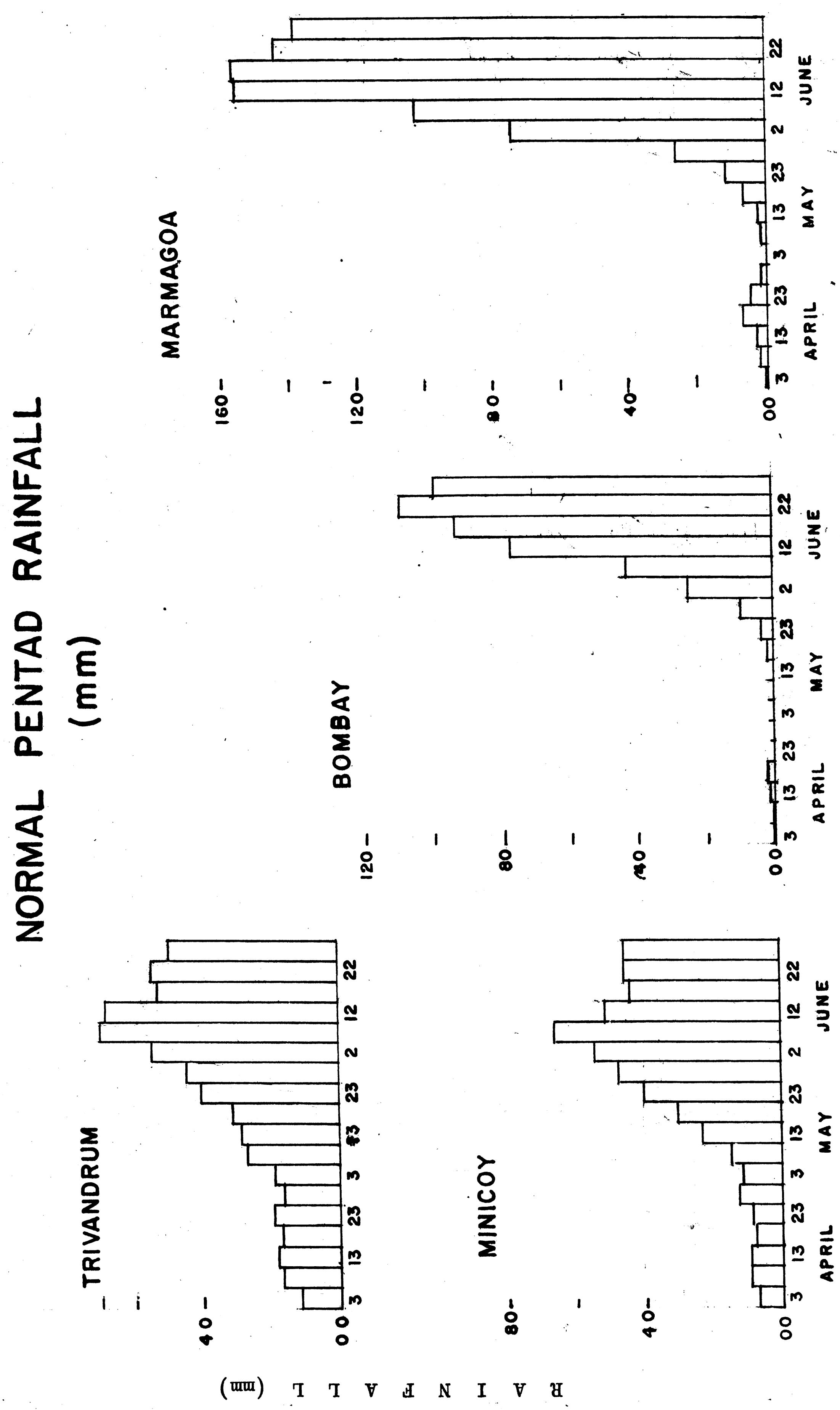

NORMAL PENTAD RAINFALL
(mm)
TRIVANDRUM
MINICOY
BOMBAY
MARMAGOA
RAINFALL (mm)
APRIL
MAY
JUNE
3
13
23
2
12
22
00
40
80
120
160

Fig. 3.2

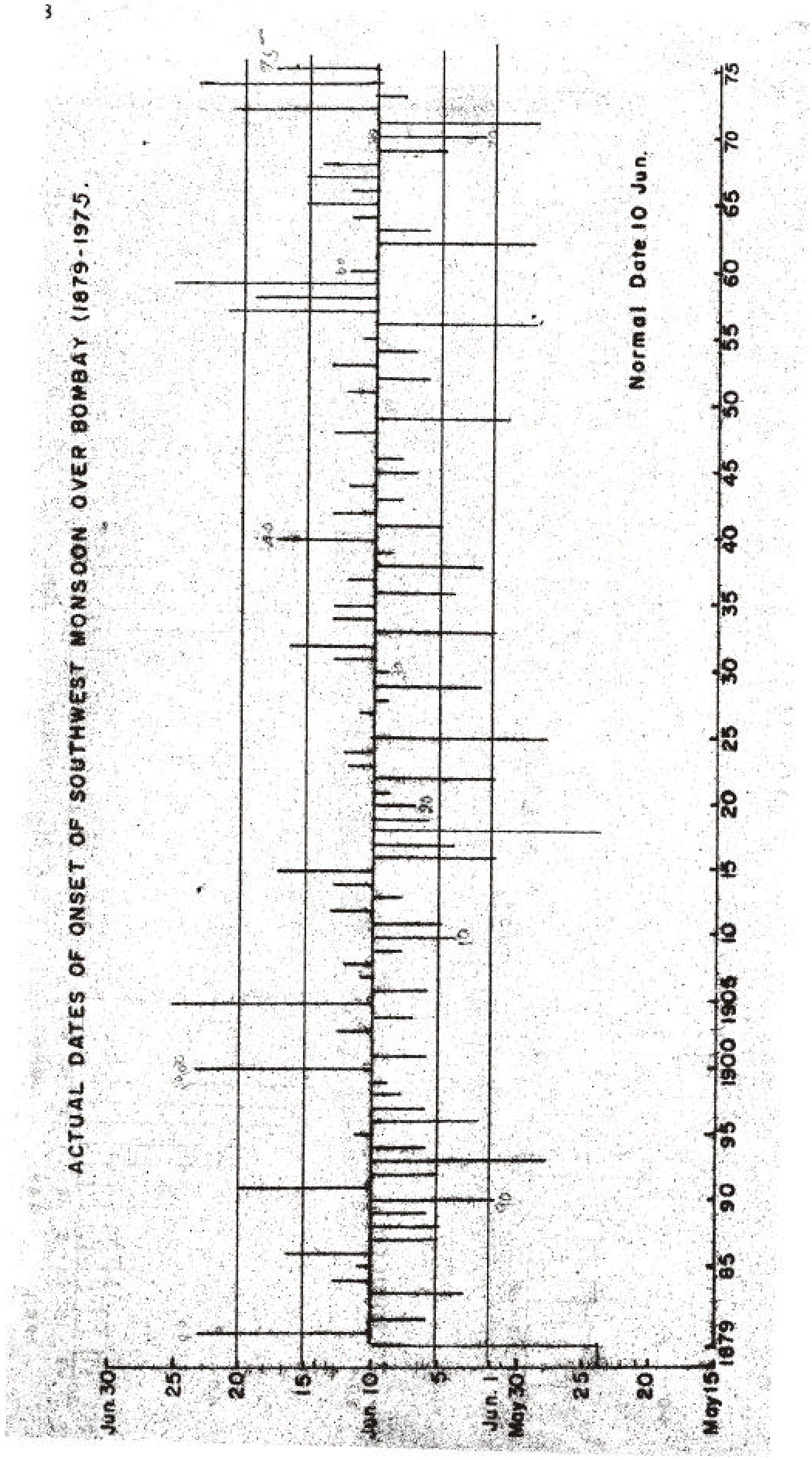

ACTUAL DATES OF ONSET OF SOUTHWEST MONSOON OVER BOMBAY (1879-1975.
Normal Date 10 Jun.
Jun 30
25
20
15
Jun 10
5
Jun. 1
May 30
25
20
May 15
1879
85
90
95
1900
1905
10
15
20
25
30
35
40
45
50
55
60
65
70
75

Fig. 3.3

Table 3.1

Advance of Monsoon

Area	**Normal date**	**Earliest**	**Most delayed**
Coastal Karnataka	4 June	19 May, 1962	14 June, 1958
North Konkan	8 June	29 May, 1956	25 June, 1959
West Bengal	7 June	27 May, 1962	23 June
Vidarbha and most parts of Madhya Pradesh	12 June	First week of June	Last week of June
Bihar	12 June	6 June	1 July
East Uttar Pradesh	15 June	5 June	3 July
West Uttar Pradesh	25 June	10 June	9 July

From the normal dates given in Fig, 3.1, it should not be concluded that the advance takes place progressively, once monsoon has set in the Andamans or Kerala. The activity often weakens after an advance of about 500 km and a fresh surge is needed to spread the monsoon air mass further„ The synoptic patterns ushering in the monsoon are the same as those that cause increased rainfall in subsequent period. The changes in circulation at the onset and precursors which have forecasting values have received considerable attention

3.5 The manner of the disappearance of upper tropospheric westerlies and shift of westerly jet to the north of Tibet with the advance of the monsoon, has been extensively studied. The sequence appears to vary from year to year depending on the time of onset of the monsoon. Ananthakrishnan et al (1968) have summed up that in years of normal or delayed monsoon over Kerala there was a sudden weakening of the upper tropospheric westerlies over north India at the time of onset, In years of early monsoon, westerlies over north India persisted in strength for about a fortnight after the onset. During the week before onset, westerlies at New Delhi have ranged from 30 to 80 kt. Out of 13 years studied, there was weakening in nine years during the onset of monsoon in Kerala and no change or even strengthening in the remaining years. In the week after onset, they either weakened further or continued at the same speed. In all years the west wind maximum shifted northward at the time of the onset or a pentad earlier. Ramage (l97l) has pointed out that 300 mb westerlies at Kuche (41° 45' N, 83° 04' E) increase from June to July by 15 kt, the highest speeds being in July and August, in support of the westerly jet shifting to north. Yin (1949) recognised the westward shift of the quasi–stationery trough in westerlies at 500 mb, from 90° to 80° E at the time of the burst of the monsoon in 1946 and interpreted it as a phase–shift induced by the orography of the Himalayas consequent on the displacement of the low–latitude westerly jet to the north of the mountains. As mentioned earlier, the weakening of the westerlies does not have a unique relationship with the time of burst of monsoon and in any case the consequent displacement of the trough, though interesting, may not have a direct relationship with the monsoon.

3.6 The five to sixteen day interval between the first appearance of easterlies at 200 mb at Aden and onset of the monsoon over Kerala pointed out by Sutcliffe and Bannon (1954) has been found by Ananthakrishnan and Ramakrishnan (1965) to be much more variable, e.g. a month in 1964 (late onset by a week) and simultaneous in 1960 (early onset by a fortnight). While the appearance of easterlies over Aden is a pre cursor to the summer circulation, it has little utility to predict the date of onset of monsoon. During the week before the onset of monsoon into Kerala, upper tropospheric winds over Bombay are still mostly westerly (10/20 knots). Reversal takes place usually either during onset or immediately thereafter, but in a few years before also. Generally speaking, south of 20^o N, easterlies appear before the onset and further north after the onset in Kerala. The interval between them shows large variations.

3.7 In the monsoon circulation, at 500 mb, the sub–tropical high extends from northeast Africa to northwest India. Pant (1964) and Ramamurthy and Keshavamurthy (1964) pointed out that the shift to this position from near about 20^o N is associated with the advance of the monsoon. This change was also found to be simultaneous with the onset of monsoon rains over the south Peninsula by Ananthakrishnan et al (1968).

3.8 A lower tropospheric trough between tropical easterlies and equatorial westerlies is not infrequently seen in the southern parts of the Bay of Bengal and the Arabian Sea at the time of advance of the monsoon. This is also seen in the mean flow at 700 mb in May. Riehl (1954) identified the forward edge of the monsoon with such an equatorial shear–line at 700 mb, Nataraja Pillai (1965) regards the onset and further advance of the monsoon is governed by the cyclonic and anti– cyclonic cells at this level.

3.9 Ananthakrishnan and Thiruvengadathan (1968) studied the reversal of thermal gradients as derived from thermal winds (10 days means) at Trivendrum, Nagpur and New Delhi in relation to monsoon onset. The onset takes place when the meridional thermal gradient has reversed at all levels between 200 mb and 700 mb. The reversal starts in upper troposphere six weeks before but takes place between 700 and 500 mb almost simultaneously with the onset of the monsoon.

3.10 Flohn (1960) put forward that the heating of the Tibetan Plateau which is at mid–tropospheric level, produces an anticyclone aloft. The effect of this cell is the reversal of the normal temperature and pressure gradients in the layers between 600 and 300 mb. This reversal acts like a switch for the atmospheric circulation over the southern half of Asia and produces nearly simultaneously in the first half of June, advancement of the ITCZ to northern India together with a rapid extension of the equatorial westerlies (burst of the monsoon) and migration of tropical easterlies (Bay monsoon) towards northern India. Ananthakrishnan (1965) has emphasised that the first onset of monsoon over Kerala has no relation with the Tibetan Plateau acting as a heat source in middle troposphere. At that time, the sub–tropical anticyclone in the upper troposphere lies over north India between 20^o and 25^o N, well to the south of the Tibetan Plateau. It takes five to six weeks for the monsoon to be fully established over India and only then the sub–tropical ridge shifts to about 30° N. In a later paper, Flohn (l965a and 1965b) reckons that the latent heat released by the premonsoon rains over northeast India as contributory to the establishment of Tibetan anticyclone at 30^o N, 88^o E. However, this may not be such an important factor, as the rise in upper tropospheric temperatures at Gauhati and New Delhi from March to June are not significantly different, in spite of the difference in rainfall. Flohn (1960) observes that in years with a high hemispherical zonal index (35° – 55° N) in June, the onset of monsoon in Delhi occurs cm the average on June 28, but in years with low index on July 4. No similar relationship could be found for latitudes south of 25° N. While Mooley (1957) regards western disturbances moving across northernmost parts of India as favourable for at least temporary extension of monsoon current into Punjab, west Uttar Pradesh and Kashmir, Chakravorty and Basu (1957) find them having a retarding effect on the advance of monsoon into northeast India.

Ramaswamy (l97l) has presented 500 mb contour patterns over Asia averaged over the second halves of June 1965 and 1966, illustrating respectively the difference between late and normal advance of the monsoon (Fig. 3.4). In 1965, monsoon had not advanced into northern India to the west of 78^o E by the end of June, Westerly circulation was dominating the sub–continent at 500 mb. A well–marked mean trough in the westerlies was extending from the middle latitudes into northern India. In 1966, monsoon had reached 75^o E by the end of June and a high was over Tibet with an extensive low centred near Allahabad.

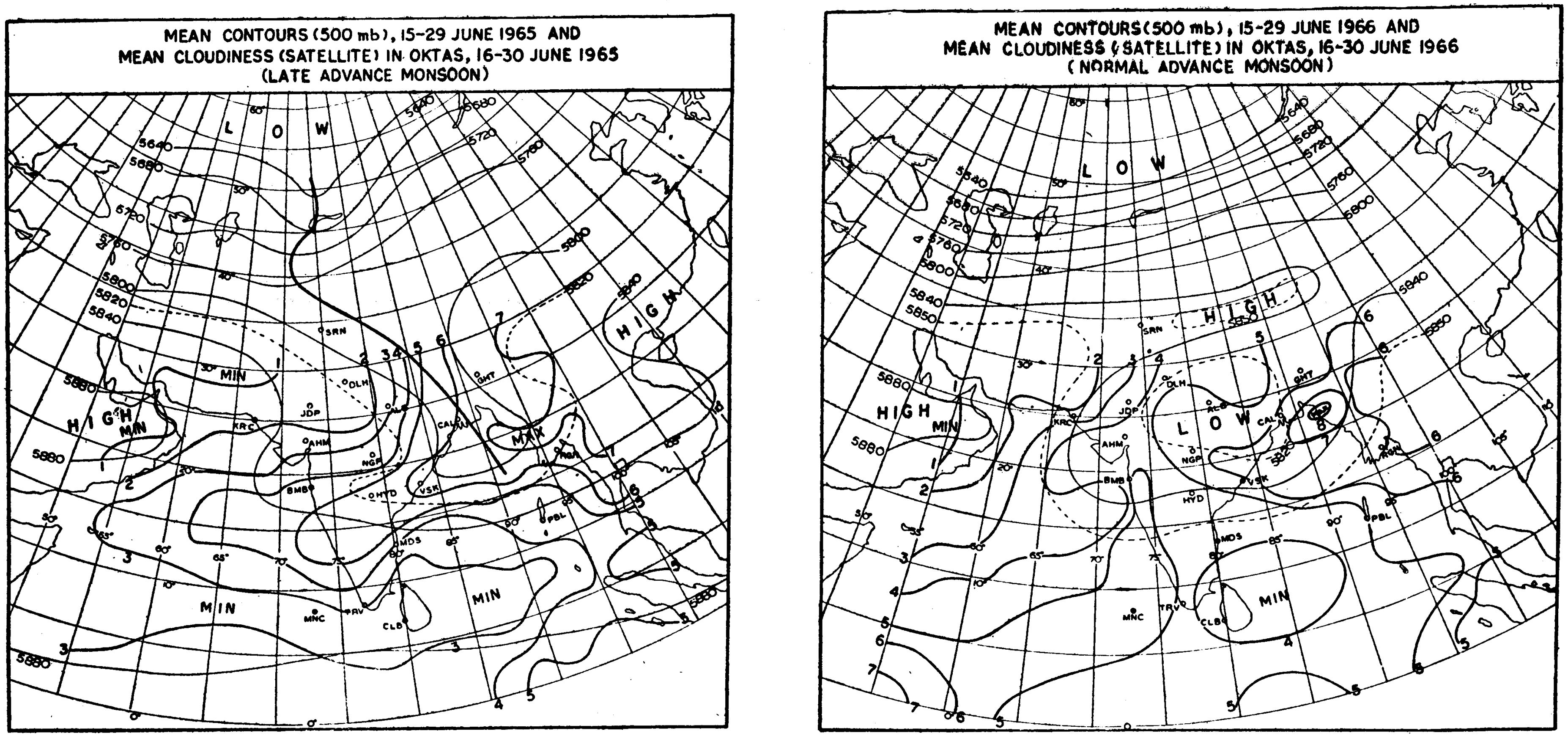


Fig.3.4 - Thin continuous lines are contours drawn at intervals of 40 gpm and dashed lines are odd contours. Thick continuous lines are isopleths of mean cloudiness in oktas. Very thick continuous line extending from 50° N to 20° N approx. is a trough line. Maximum and Minimum refer to cloudiness.

3.11 Koteswaram (1958 and 1960) regards that the setting in of the easterly jet stream over south India coincides with the burst of the monsoon. Upper divergence in the left exit sector of the easterly jet induces convergence in the lower levels and onset of the monsoon.

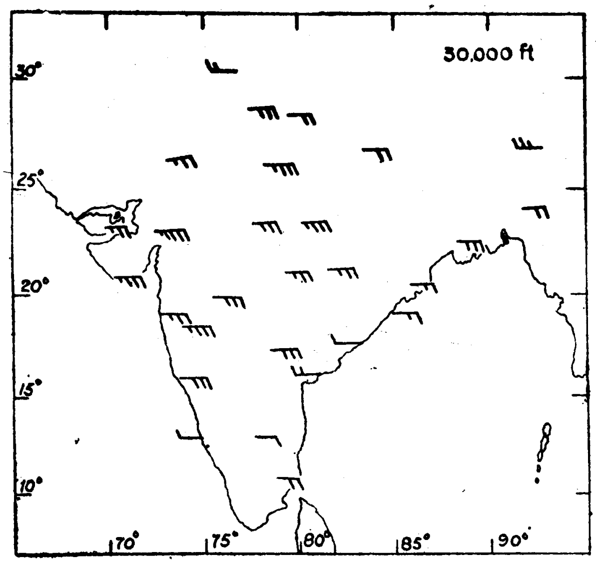
3.12 Ramaswamy (1965) found that in the stratosphere, at 50 mb, circulation was much less cellular in 1956, a year of very early monsoon, than 1957 when monsoon was delayed,. He does not explain the relationship with lower troposphere.

3.13 An interesting feature of the monsoon is that it reaches by 1st June upto 20^{o} N in east Bay and along Burma coast, before it commences to advance along the west coast. According to Ramage (l97l) 'the near Equatorial Trough intensifies and moves north sufficiently between April and May to be activated by the upper divergence east of the trough in the sub–tropical westerlies. The first disturbance so to develop, starts the summer monsoon rains of Burma and over northeastern India, expands and intensifies to Bengal the rains previously confined to Assam'. Desai (1967) seems to have rightly pointed out that in May the heat low has still an extension into the eastern part of the Peninsula, As such the air in the trough just south of the equator gets into the circulation on the eastern side of the Peninsular trough and causes early occurrence of monsoon in the east Bay, At this time over the Arabian Sea, to the west of the Peninsular trough, there is north westerly flow upto the tip of the Peninsula which does not favour the advance of the monsoon along the west coast. The Peninsular trough should disappear for •monsoon to advance along the west coast.

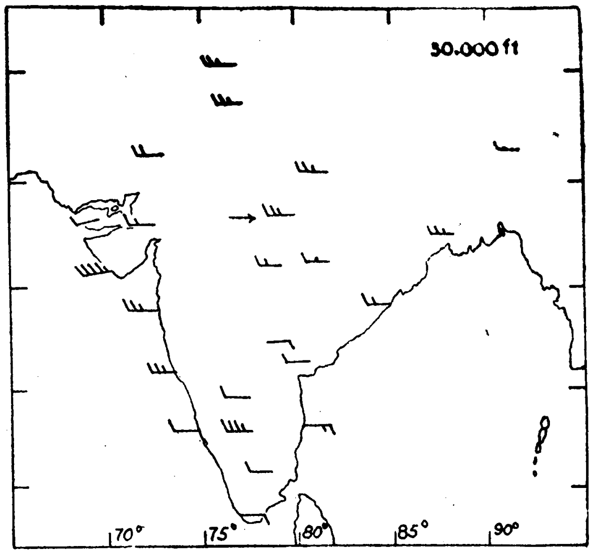
3.14 Rai Sarcar and Patil (1962) have presented charts (Fig. 3.5) showing that changes of winds at 9.0 km from the beginning of April to the end of that month give an indication of early or late onset of monsoon. Decrease of westerly component indicates early onset and increase of west winds the opposite. Differences in these changes between the Peninsula and north India can even show early onset in one area and delay in the other. In another paper (1961) they show that there are marked differences in the thermal pattern in May at 300 mb indicative of the time of onset of monsoon. The normal thermal patterns of January, May and August (at 300 mb) are given in Fig, 3.6a, The thermal high south of 10^{o} N in January shifts to the Tibetan Plateau by August, In May, it is over the central Bay. In May 1965, a year when monsoon came by due date over most parts of the country, the thermal pattern was like the normal. The thermal pattern of May 1956 (Fig. 3.6b),with a major part of the country having come under the sway of the monsoon, was already like normal August, Even in the first half of this month when the monsoon had not come over the country, the thermal high had shifted to the Tibetan region. On the other hand, in 1957, the monsoon was late by 10 to 15 days and the highest temperatures persisted south of 10° N like January (Fig, 3.6c), Out of all the upper levels, 300 mb showed most marked changes which have forecasting values. These results are to be confirmed by a study of more years.

3.15 Ananthakrishnan and Ramakrishnan (1965) find that the first appearance of westerlies over Bahrein, New Delhi and Allahabad gives a prior indication of the withdrawal of the monsoon although it has little forecasting value.

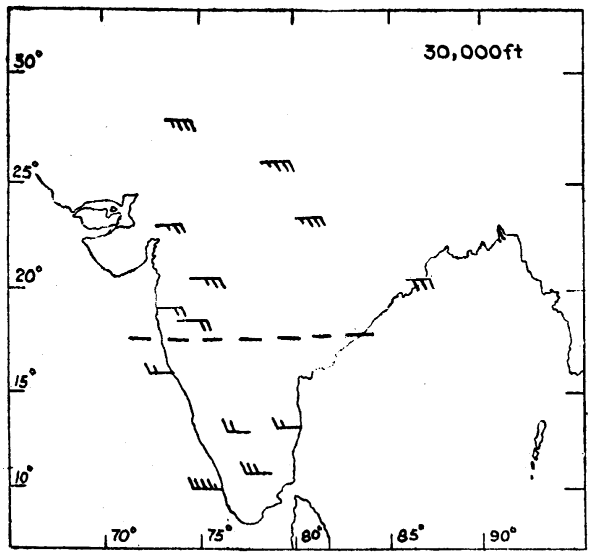
3.16 The burst of the monsoon over Kerala has received considerable attention, being the start for further advance over the rest of the country, FMU Report No, IV–18.2 by Ananthakrishnan et al (1968) presents a detailed study of the subject. There is a pronounced tendency for the formation of low pressure systems at the leading edge of the monsoon current. In 45 percent of the years, a trough of low pressure or a more intense system (cyclonic storm 8 per cent) is present in the Arabian Sea at the time of onset of monsoon along the west coast. The monsoon may also advance along the west coast with a disturbance in the Bay of Bengal, Still 25 per cent of the monsoon onset is without any surface low pressure system. In these years, the low pressure systems would have been confined to the upper air.



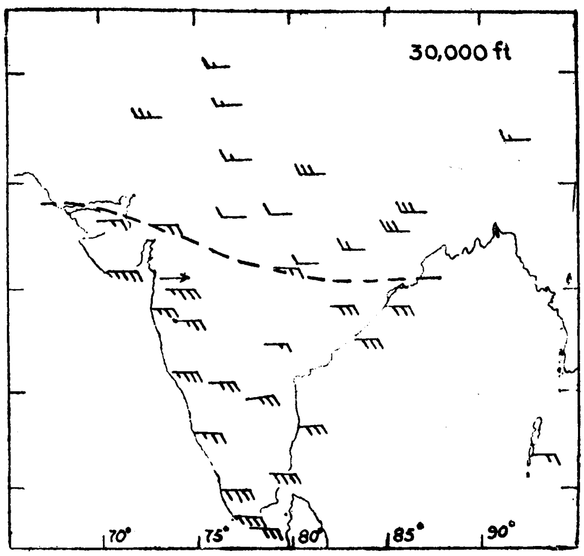
Early monsoon over the country in 1956



Late monsoon over
most parts of the country in 1957

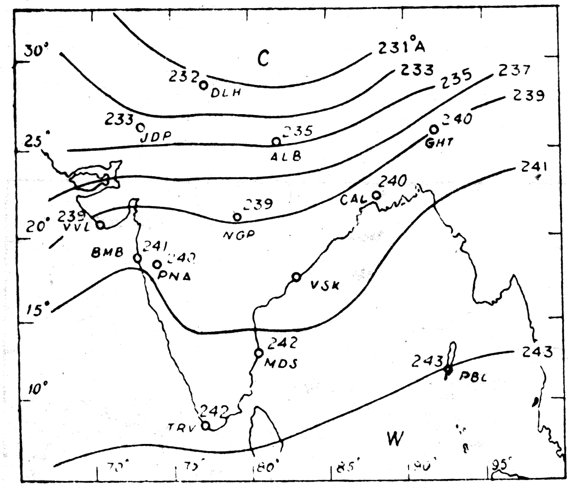


Late monsoon in the south Peninsula but
nearly timely monsoon elsewhere in 1944

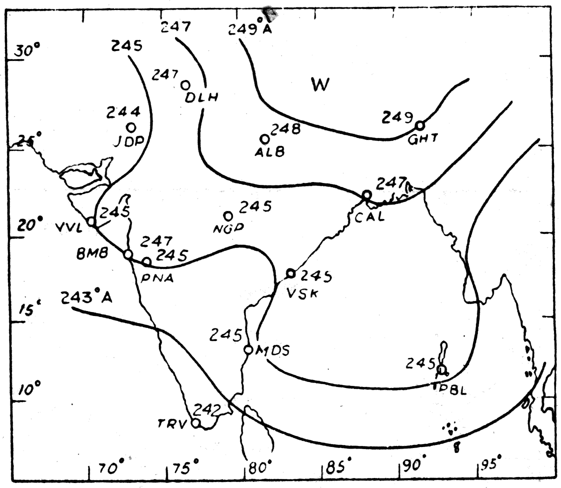


Early monsoon in the Peninsula but
late monsoon elsewhere in 1949

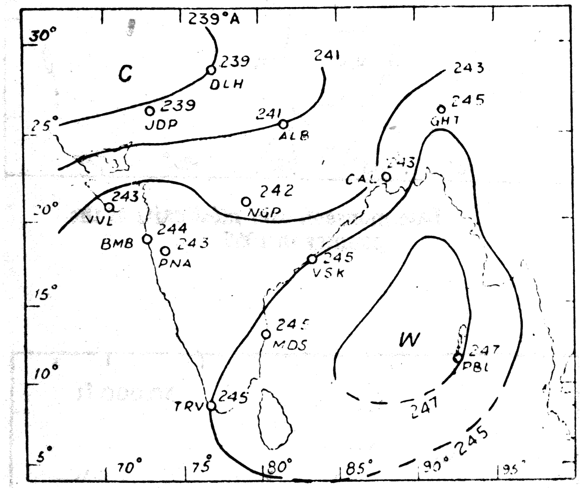
Fig.3.5 - East-west wind vector difference between the first and second fortnights of April.



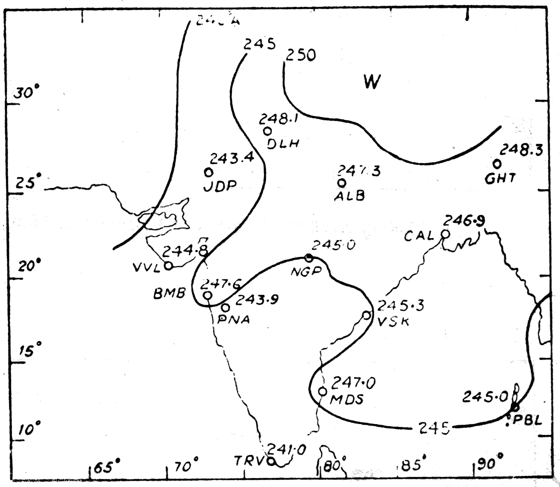
(January)



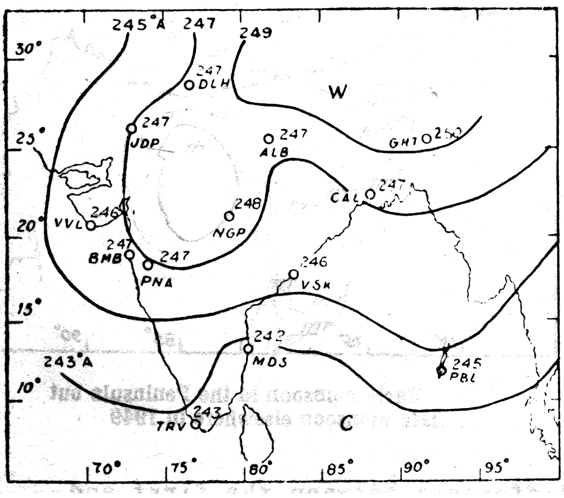
(b) (i) May 1956 (300 mb)



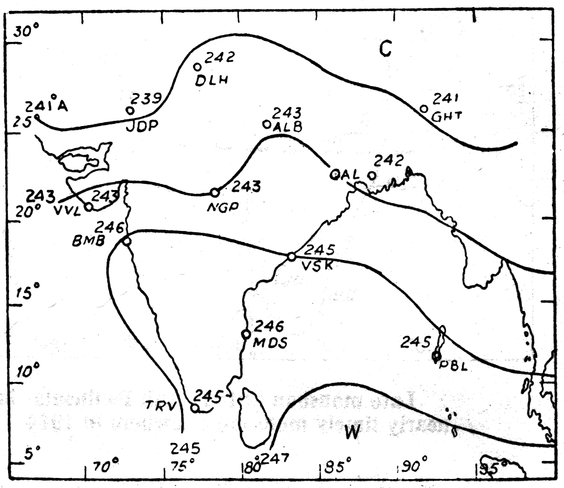
(May)



(b) (ii) 1 - 16 May 1956 (300 mb)



(August)



(c) May 1957 (300 mb)

Fig. 3.6 Normal distribution of temperatures at 300 mb level.

Ships' reports of heavy rain, squalls, strong westsouthwest to west winds, high waves or heavy swell in south Arabian Sea in May and June may be indications. Satellite cloud data have confirmed what was earlier known from sparse ships' observations, that there is a northward movement of organised cloud maximum (though not continuously) from equator to 20^{o} N during the onset of monsoon over the extreme south Peninsula and its progress northwards (Jambunathan, 1974). As the monsoon advances northwards along the west coast, there is a relative decrease in cloud amount near the equator, A pressure rise at the southern stations like Colombo. Trivendrum or Minicoy with a fall to the north is favourable. Moisture content does not show any increase. The precipitable water at Trivendrum is 4.57 gm in June when the monsoon is in sway, compared to 4.38 gm in April and 4.85 gm in May. Increase in depth and strengthening of the lower tropospheric westerlies are other features at the time of onset. Upper tropospheric easterlies at 14 or 16 km generally reach 40 kt in the preceding week, Trivendrum recording 60 kt in most years. While the deepening and strengthening of westerlies is rather abrupt in the course of two or three days, appearance of upper tropospheric easterlies at 14–16 km and their extension downward and strengthening is more gradual. Generally deep and strong westerlies are associated with good monsoon rainfall and a weak and shallow field with poor rains. In more than half the number of years, western disturbance activity persists over northwest India and Pakistan at the time of onset of monsoon over Kerala, which can be so even in years of late onset.

3.17 Figure 3.7 shows the fluctuation of upper winds at Trivendrum in May and June 1966 while monsoon set in on 4 June. The features discussed earlier are well illustrated in this sequence. Lower tropospheric westerlies increased in depth and strengthened before the onset but weakened two days after. There were one or two phases of westerlies strengthening earlier. The upper easterlies which were only 20 kt at the beginning of the May, built up gradually to 100 kt a few days after the monsoon advance. The shallow trough along–off west coast at the time of onset of monsoon along Kerala is illustrated in the chart of 1 June 1957 presented in Fig. 3.8.

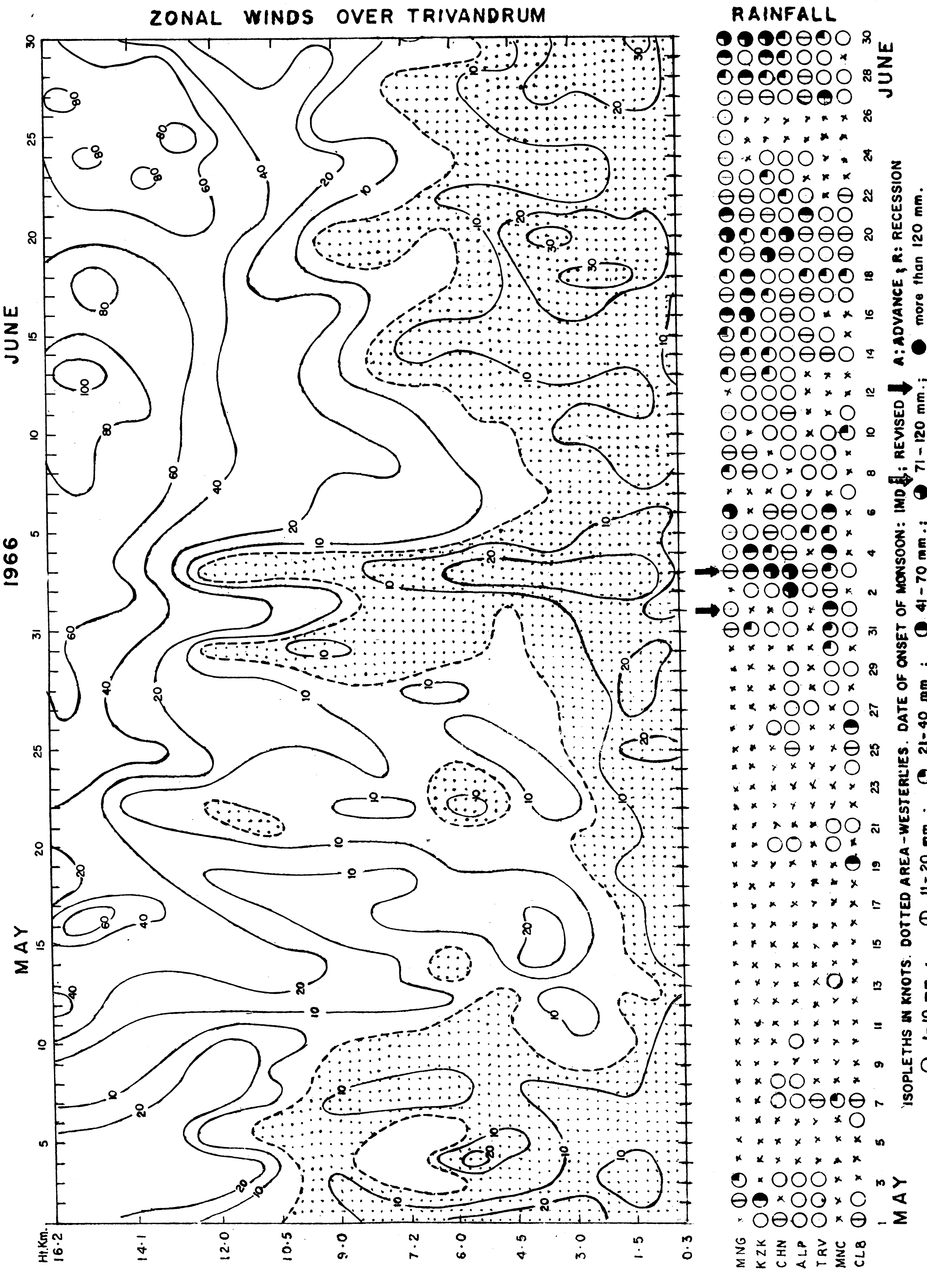
3.18 From a study of the onset of monsoon in Kerala from 1961 to 1968, George (1970) is of the view that given a pre–existing lower tropospheric feature causing convergence at sea level in the area, upper divergence in the western portion of an advancing trough in the easterlies or the left exit portion of an accelerating east wind maximum over Madras, usually precedes and causes the burst of monsoon over Kerala. His case of June 1967 is shown in Fig. 3.9. Monsoon burst in Kerala on 8th June while an 'upper easterly trough (300–150 mb) passed over Madras on 7–8 June and over Trivendrum on 9th,,

3.19 The FMU Report No. IV–18.2 by Ananthakrishnan et al (1968) sums up as follows the synoptic indications for the imminent onset of the monsoon over Kerala

i) Any disturbance in the Arabian Sea/Bay of Bengal. The most common initial form of the disturbance is a trough of low pressure in south– east Arabian Sea.

ii) Reports from ships and island stations in the south Arabian Sea, of heavy convection, squally weather and rough seas or swell from south west with moderate to strong winds from some southerly to westerly direction.

iii) The strengthening and deepening of lower tropospheric west winds over extreme south Peninsula and Sri Lanka and strengthening of upper tropospheric easterlies to 40 kt for a few days at 14 to 16 km; at the time of onset, the easterlies reach a maximum speed of about 60 kt.



ZONAL WINDS OVER TRIVANDRUM
RAINFALL
1966
MAY
JUNE
Ht.Km.
MNG
KZK
CHN
ALP
TRV
MNC
CLB
ISOPLETHS IN KNOTS. DOTTED AREA-WESTERLIES. DATE OF ONSET OF MONSOON: IMD ; REVISED ; A: ADVANCE ; R: RECESSION
1-10 mm.; 11-20 mm.; 21-40 mm; 41-70 mm.; 71-120 mm.; more than 120 mm.

Fig. 3.7

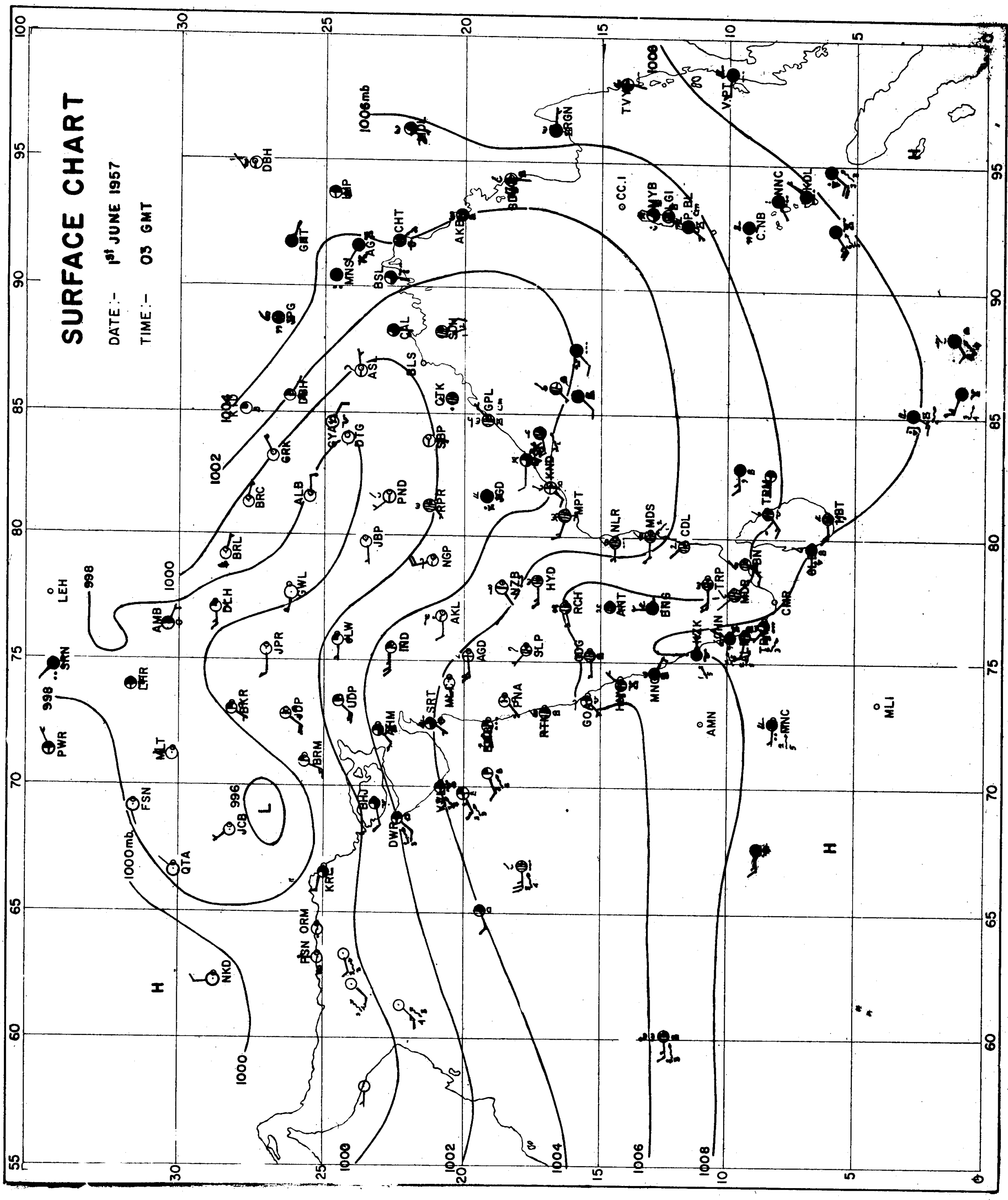


Fig. 3.8

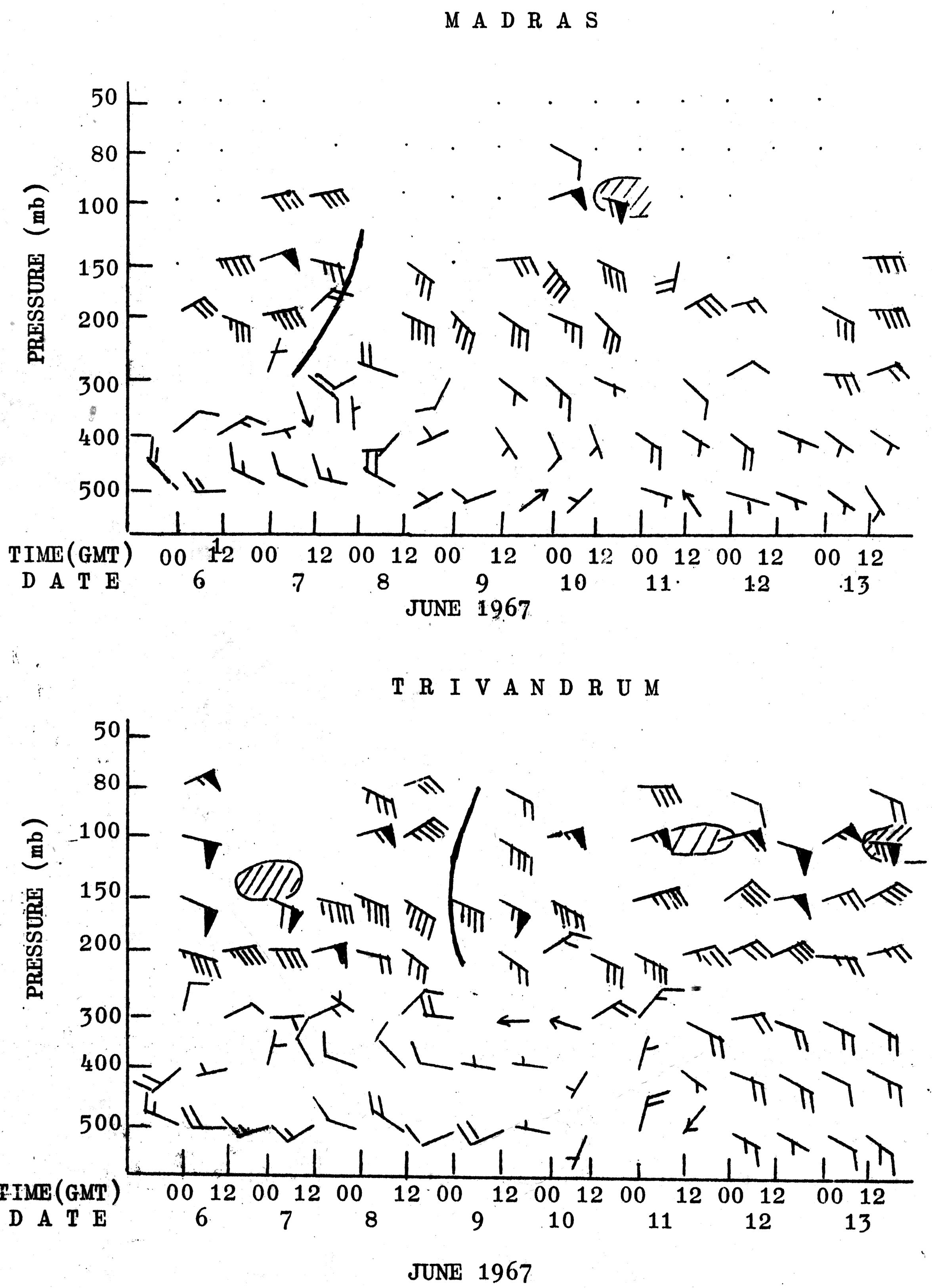


Fig. 3.9 Hatched area indicates wind speed 60 knots and above

iv) The tendency of the strong westerlies of the upper troposphere over north India to break up or to shift northwards.

v) Persistent moderate to heavy clouding in the south Arabian Sea shown by satellite pictures and its tendency to shift northwards.

All these features may not always be present simultaneously. The reorganization of circulation to the monsoon patterns extends over an internal ranging from a few days to one or two weeks.

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CHAPTER 4

FLOW ACROSS THE EQUATOR

4.1 Simpson (l92l) recognised that air from south of the equator in the Indian Ocean is drawn into the southwest monsoon circulation. As seen from the wind reports of ships, the southwest air motion over the north Indian Ocean is a continuation of the southeast trades over the south Indian Ocean. Inconsequence, the air which reaches India would have travelled 6,500 km over the ocean and therefore might be expected to be highly charged with aqueous vapour.

4.2 Rao (1964) showed that averaged over the globe, there is an interhemispheric circulation across the equator from the winter to the summer hemisphere in the lower troposphere. It is least marked in the mid–troposphere. In upper troposphere there is reverse flow. Napier Shaw's computations show that the mass of air in the northern hemisphere has a maximum at mid–winter and minimum at mid–summer (Shaw, 1936). Compared to the total amount of air in a hemisphere of about 2,700 billion tonnes, the range of variation is about ten billion tonnes. In a month, there may be a net flow of about 1.6 billion tonnes of air across the equator. This is equivalent to a uniform airflow across the equator of about 2 nautical miles in a month throughout the vertical extent of the atmosphere or 1 kt across one degree length. Hence marked flow across equator should be compensated by reverse flow at some other level or at some other point.

4.3 However, at Colombo, in spite of its location in the central longitude of the monsoon area, the meridional component is 5 kt from north at 800 mb. But the flow at 150 mb is 9 kt from north, like the global average, Singapore and Nairobi at the eastern and western ends of the monsoon region fit in with the global pattern. The results of Tucker (1965) while confirming the cross–equatorial circulation, indicated that at Gan (0° 41 ' S 73° E) the flow in the lower troposphere was generally weak in July and sometimes even from the northern hemisphere. These cast a doubt whether in the central longitudes there was significant flow into the monsoon area from the southern hemisphere above the surface layer.

4.4 The work of Findlater (1969 a and b) identified the manner of feed from the south of the equator into the Indian Southwest Monsoon in the Arabian Sea. Apart from the flow at the surface into the summer hemisphere, strong flow from south with a mean speed of about 30 kt at the equator prevails over eastern Africa, Figure 4.1 shows the mean meridional flow at the equator in July between longitudes 35^{o} E and 75^{o} E, as presented by him. This is the strongest cross–equatorial flow observed so far, compared with the global average of not more than 5 kt, in the lower troposphere at any level. This mean flow across equator from 35° E to 75° E from surface to 600 mb amounts to 77 x 10^{12} metric tonnes day^{-1} as against the total mean flow in lower troposphere estimated by Rao (1964) as 16.2 x 10^{12} metric tonnes day^{-1} without taking into account this strong flow off eastern African coast. 0.0015 of the earth's atmosphere is thus moving in this belt from one hemisphere to the other, each day.

4.5 As speeds between 40 kt and 100 kt are frequently reached at one point or the other in this cross–equatorial flow below 3 km, Findlater calls it 'low–level jet stream near the equator'. Low level southeasterly jet streams flow intermittently from the vicinity of Mauritius, over the northern tip of Madagascar, to reach Kenya coast as southerlies. Sometimes, this southeasterly jet from the Mauritius area is joined by or even replaced temporarily by low level jet streams moving northward through the Mozambique Channel after bursts of cooler air come around the tip of southern Africa. The jet stream is not always a single core but made up of a series of segments. Fig.4.2 (a and b) illustrates these features. The core of the jet stream, as defined by the 50 kt limit, is usually 200–350 km wide, 600–900 km long and 1 km deep, Findlater quotes a case of travel of the jet segment reproduced in Fig. 4.3. The core moved at a speed of 50 kt along the stream-line .

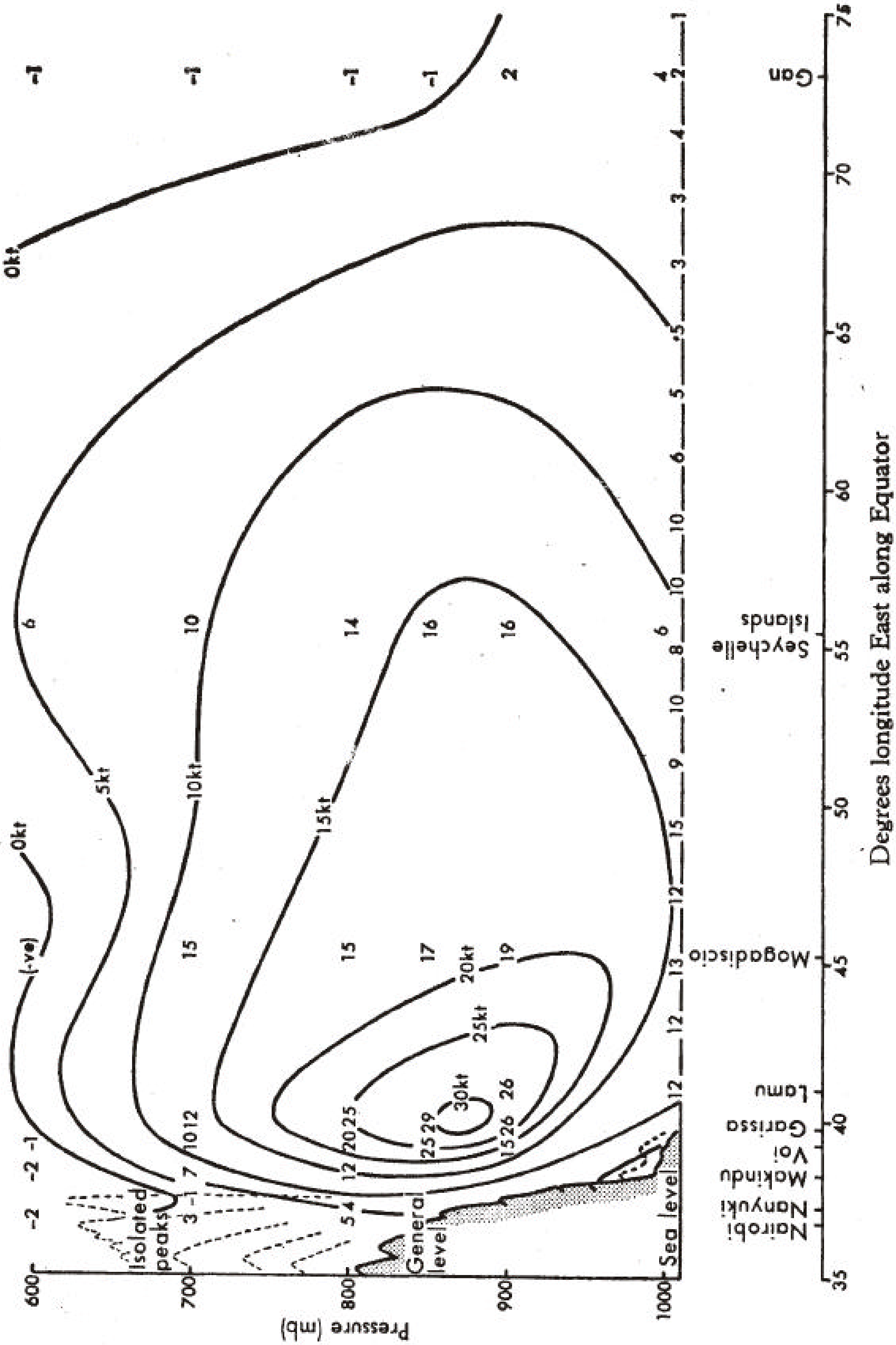

Fig. 4.1 Mean meridional flow at the Equator in July, (Findlater, 1969).

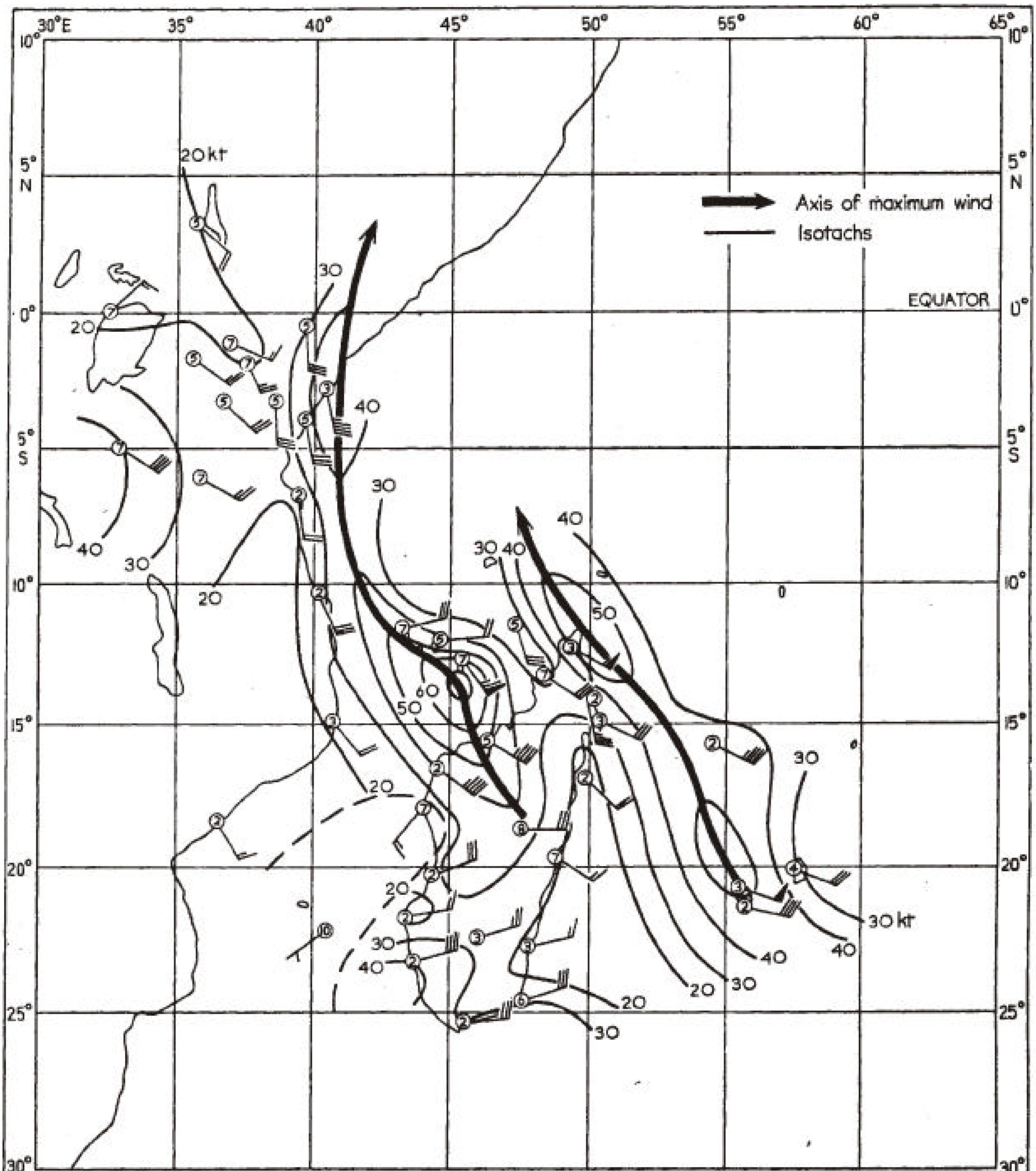

Fig.4.2 (a) Low level jet streams from the southeast, 26 June 1966. (Each plot represents the highest wind speed recorded on each pilot balloon ascent with the height of the maximum wind in thousands of feet, shown at the point of the wind arrow). (Findlater, 1969).

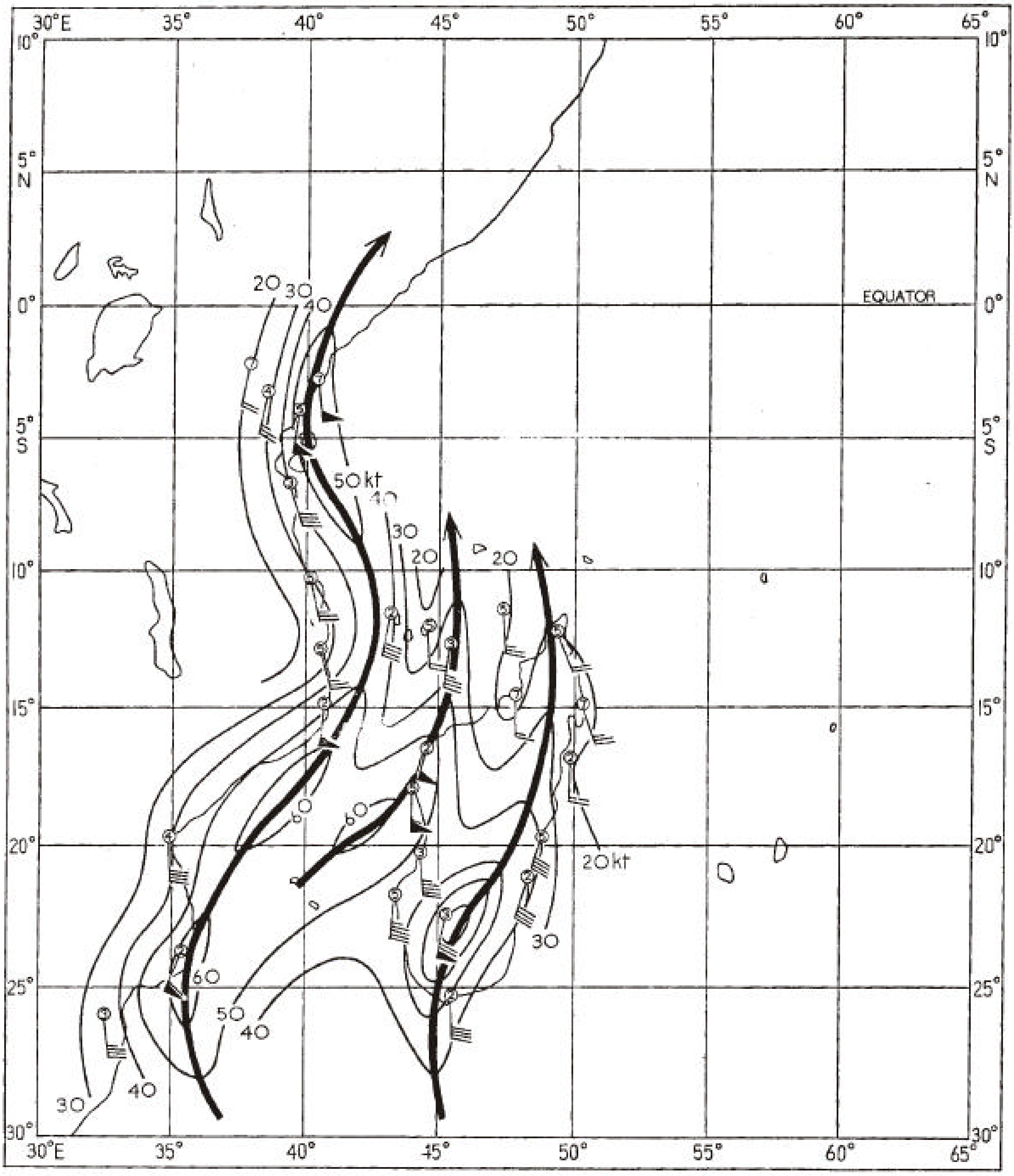

Fig.4.2 (b) Low level jet streams from the south. 26 July 1966. (The method of plotting is the same as in Fig. 4.2(a)). (Findlater, 1969).

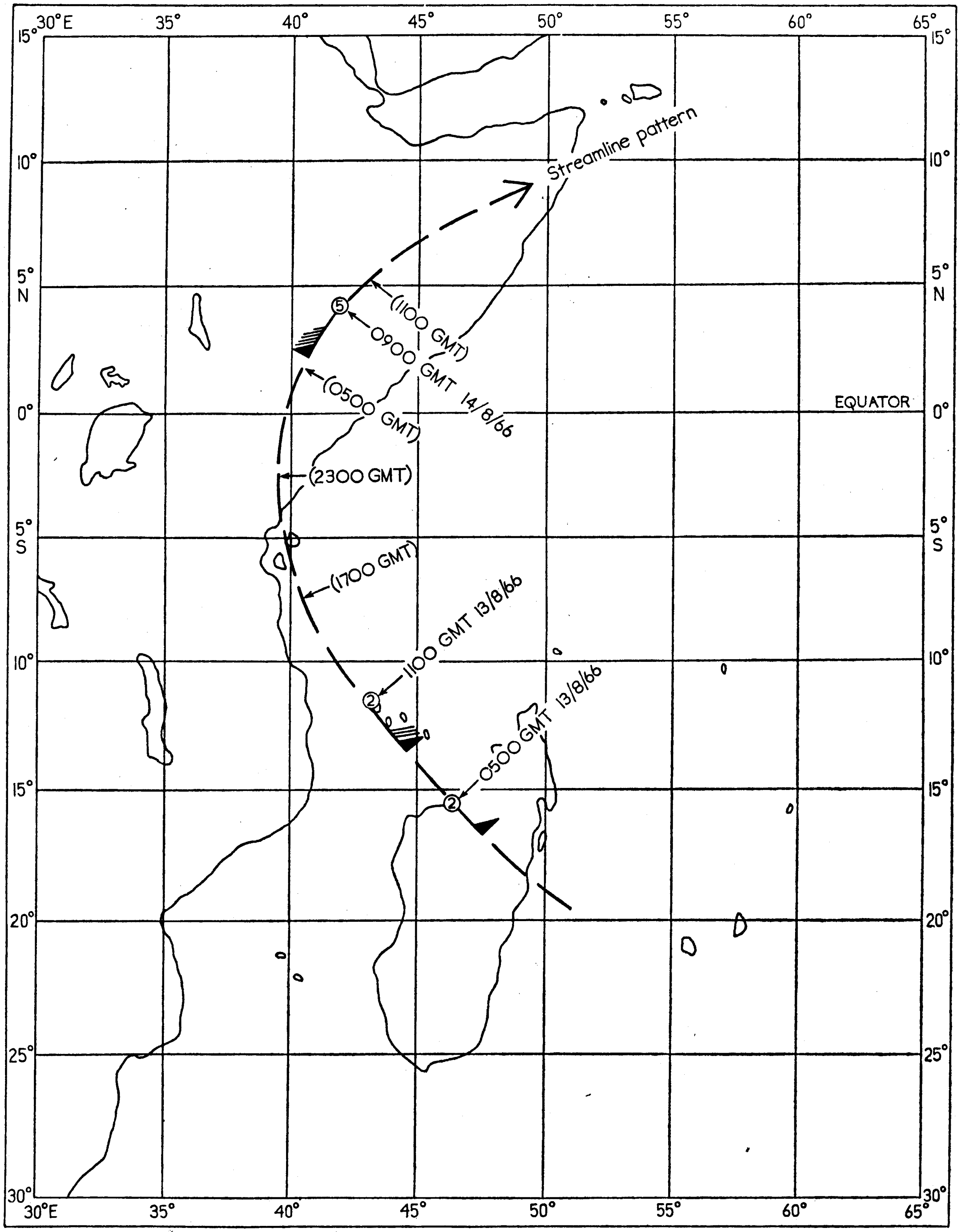

Fig.4.3 Movement of a jet core, 13-14 August 1966.
(The height of the core, in thousands of feet, is shown at the point of the wind arrow). (Findlater, 1969).

4.6 As mentioned in the earlier section, strengthening of the lower tropospheric westerlies in the Arabian Sea is a sign of increase of monsoon rains along the west coast of the Indian Peninsula (south of 20° N) both at the time of onset and later. Ships' reports of 30 kt or more in the east Arabian Sea have been taken as an indication for a spurt in monsoon rains along the coast. Probably the strong wind regime of the cross–equatorial jet travels along the mean streamlines with a speed of the order of 50 kt to reach the west coast and increases the rains. Wind speed convergence ahead of such surges could be the cause of the rains. Findlater (1969 a) studied the relationship of the cross–equatorial flow over Kenya on the rainfall of western India. Cross–equatorial flow at Garissa (0 29' S 39 38' E) was calculated for each day by adding the wind speeds at 0.9, 1.2 and 1.5 km. This was compared with the total rainfall at the, three coastal stations, Veraval, Bombay and Mangalore and the inland station Poona. The overlapping five–day means of both the elements are as represented in Fig. 4.4. The strong association between the two parameters is quite apparent, rainfall fluctuations lagging by a day or two.

4.7 Findlater's work has clearly established that southern hemispheric air of differing origin is accelerated into a well–defined stream which crosses the equator in a limited zone of longitude and becomes the southwest monsoon flow of the Arabian Sea. Equally strong compensatory flow into the southern hemisphere in any narrow belt has not been detected so far. Most probably the general flow in the upper troposphere from the summer hemisphere compensates for this also. At Gan, the cross–equatorial flow in the upper troposphere is about the global average, though it is insignificant in the lower troposphere.

4.8 Analysis during the IIO E (Ramage and Raman 1972) brought out the existence of a clockwise circulation in the lower troposphere between Gan and Diego Garcia, affecting flow near the equator and responsible for the northerly flow in Gan– Colombo longitudes.

4.9 Malurkar (1958) visualised that air crossing the equator in different longitudes could strengthen the monsoon over the Indian region in certain corresponding belts based on trajectories. Shallow low pressure areas moving westwards south of the equator facilitated such crossing which he picturesquely called 'pulses'.

4.10 Gordon and Taylor (1970) derived equations to compute trajectories of air parcels in low latitudes, assuming steady state pressure fields. They concerned themselves with flow at sea level and assumed a frictional force (FV) acting against the wind (F = 2.5 x 10^{-5} sec^{-1} and V is the wind speed). Knowing the pressure field in any area, trajectories were computed in one hour time steps. Climatological means of initial velocities could be the starting point for computation of subsequent trajectory. The divergence of mean resultant surface winds for July 1963 computed by him for the Arabian Sea (shown in Fig. 4.5) is not unrealistic.

4.11 A cross section of the winds observed by the U S SR ship OK EA N along the equator is shown in Fig. 4.6 as presented by Pant (1974). In the period covered by the observations, monsoon was generally weak over the country. The strength and depth of the flow across equator into the northern hemisphere decrease from 47° E to 57° E. These observations did not extend upto 40° E where Findlater observed jet–like flow. But still OK EA N's observations confirm considerable flow across equator into the Indian monsoon area, in the lower troposphere.

4.12 Fujita, Watanabe and Izawa (1969) made a case study of cross–equatorial flow in September 1967 in the eastern Pacific, using AT S–I photographs. Cross— equatorial flow from the southern hemisphere curves anticyclonically in the northern hemisphere until an equatorial anticyclone forms under favourable conditions, without any pre existing high pressure centre. Air parcels carry their negative absolute vorticity of the southern hemisphere across equator which corresponds to anticyclones of the northern hemisphere. These authors compute the trajectories

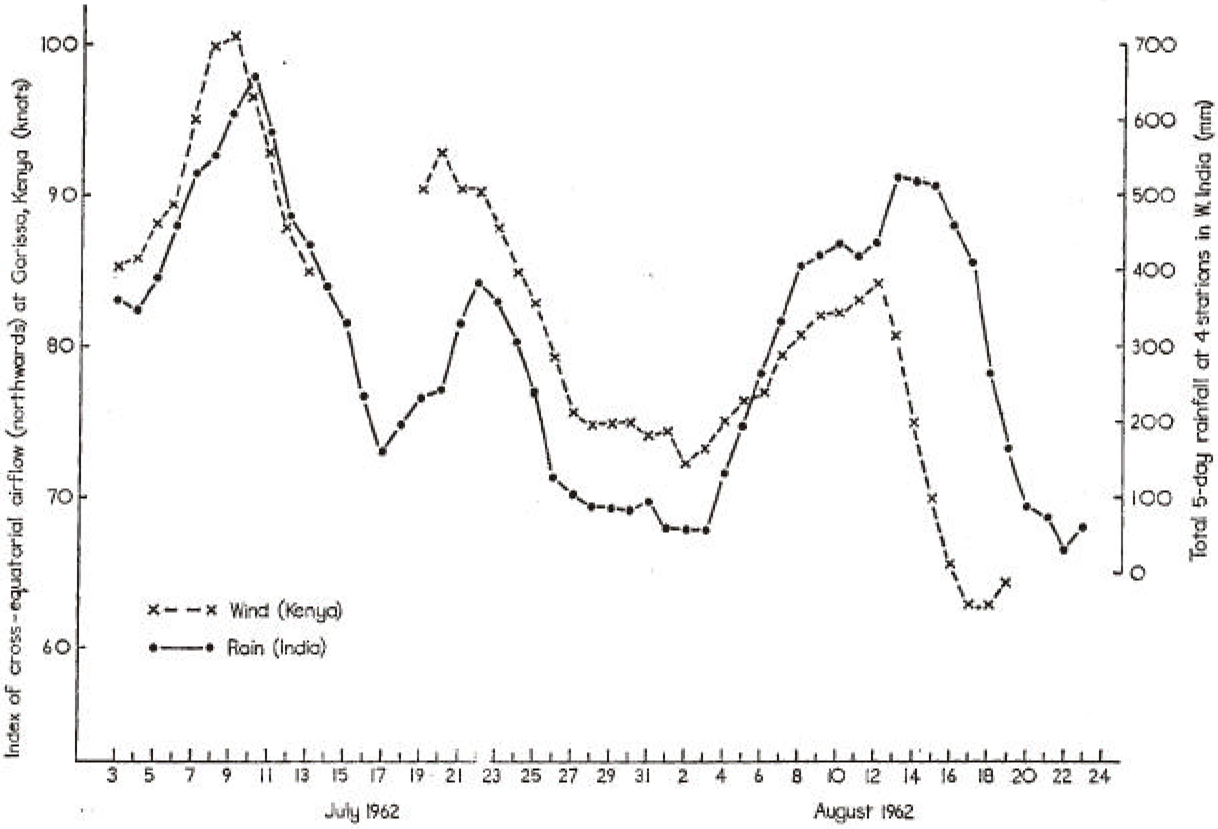

Fig.4.4 Cross equatorial air flow at Garissa (Kenya) in relation to rainfall at four stations in western India (Smoothed values). (Findlater, 1969).

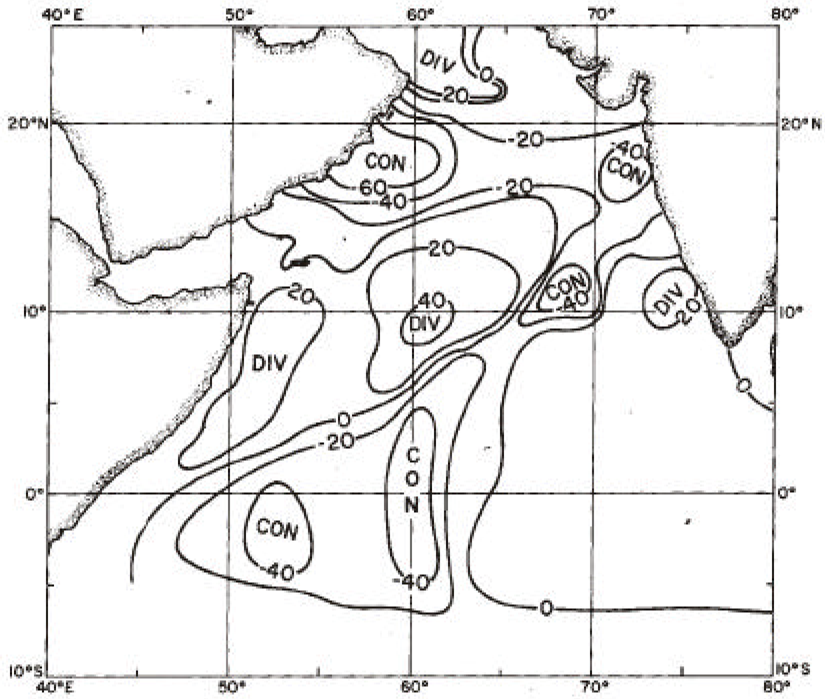

Fig.4.5 Computed divergence of mean resultant surface winds in units of 10^{-6} sec^{-1} for July I963 (Gordon and Taylor, 1970).

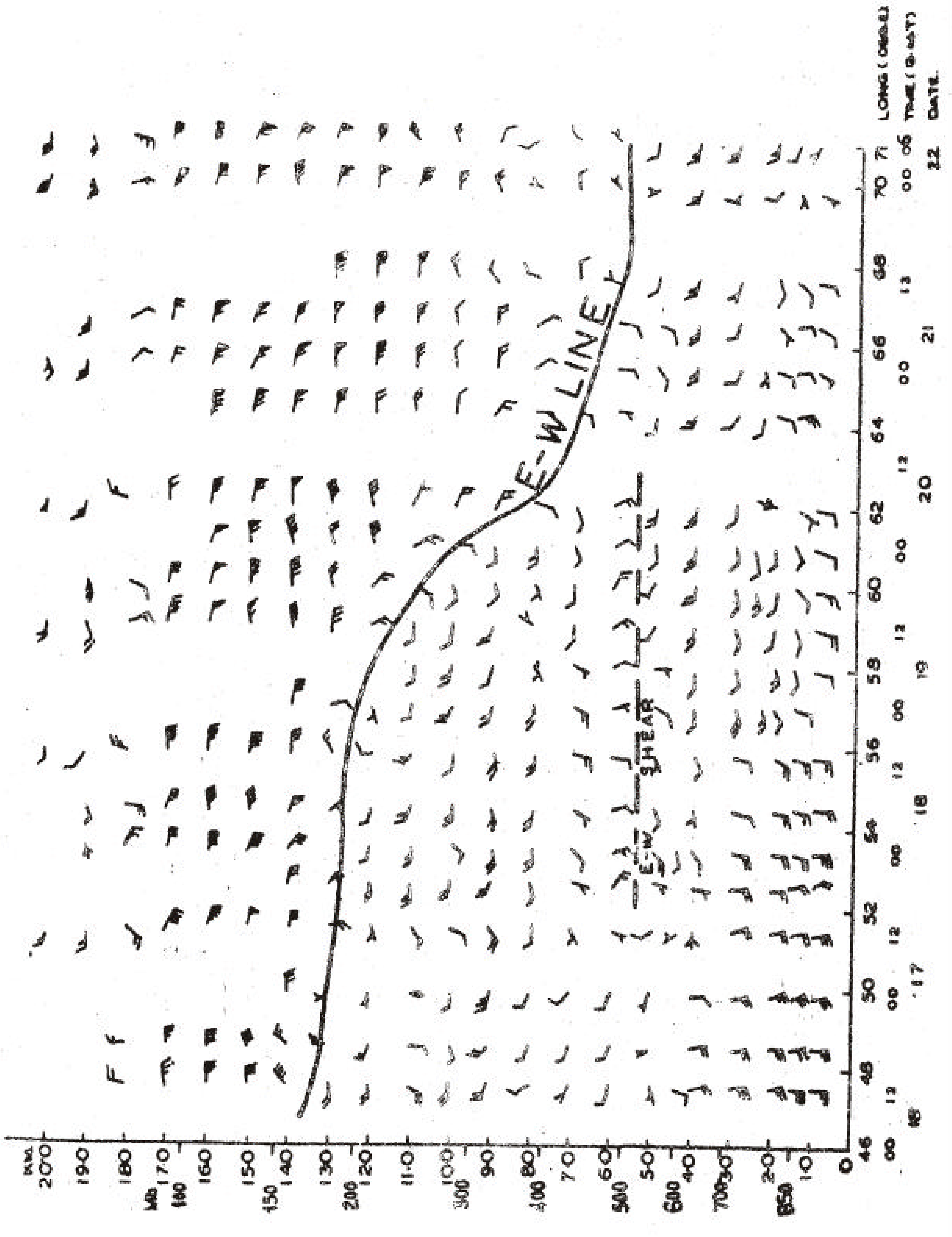

Fig. 4.6 Vertical distribution of wind at the equator during 16-22 June 1973.

north of the equator, in the frictional layer, under some simplifying assumptions, taking into account frictionally induced divergence due to vorticity and dissipation of vorticity due to friction. Computed trajectories show anticyclonic curvature.

From the cloud pattern and velocities over the eastern Equatorial Pacific, Fujita et al (1969) have presented a five–stage model for the migratory equatorial anticyclone. In the 'pushing stage', a large–scale flow from the southern hemisphere pushes northward, producing a convex band of inter–tropical cloudiness which may reach 10^{o} N. Formation of tropical storms is sometimes observed along the band, in which northern and southern hemisphere air interact with large horizontal wind shear and cyclonic relative vorticity. Within one to three days, the flow from the southern hemisphere gains sufficient anticyclonic relative vorticity and starts returning southward, which is called the 'recurving stage'. The inter–tropical cloud band shows little change. Tropical depressions which have formed in the 'pushing stage' tend to move out of the region of cloud band. After a day or so, in the 'cut off stage', the equatorial anticyclone will be characterised by an enclosed circulation. Then, the anticyclone centre is encircled by air from the southern hemisphere. The next is a 'mixing stage' as the northern air keeps pushing the cloud band along the leading edge of the anticyclone. Such a joint push both by the northern trades and the flow from south of equator, often results in a very intense zone of convergence with cyclonic vorticity. The intense band of inter–tropical cloudiness located in this zone may be called the 'burst band' and this stage the 'burst stage'. This lasts only one to two days, the 'burst band' then disintegrating quickly into fragments. Even then, rather intense flow continues to the south of the equatorial anticyclone centre which is moving north to northwest. This is called the 'interacting stage' to denote the interaction taking place between a cold front of higher latitudes and an equatorial anticyclone which has moved into middle latitudes. This entire sequence covers a period of about two weeks. Similar equatorial anticyclone associated with cross–equatorial flow has not been identified during the southwest monsoon in the Indian Ocean. Whether the cyclonic vortex south of equator near Gan is subject to periodical changes of this type requires investigation. Fujita et al consider that the cold ocean surface temperatures near the equator in the eastern Pacific would suppress vertical mixing inside the friction layer which would facilitate gain of anticyclonic relative vorticity as the air mass moves north of equator, according to their model.

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CHAPTER 5

AIR MASSES

5.1 The main characteristic of the southwest monsoon period is the prevalence of a highly moist air mass in great depth over most parts of India; so that even circulation patterns favourable for weak convection are able to produce rainfall. The change is best seen by considering the April conditions first.

5.2 In April, continental air mass (Tc) prevails over northern India west of 85° E and over the western half of the Peninsula north of 15° N. Dew point depression is about 20° C at 900 mb. Lapse rate is nearly dry adiabatic upto 700 mb at 12 Z, though morning inversion is still present. As a result of intense vertical mixing in the adiabatic layer, mixing ratio is nearly uniform and dew point depression decreases with height. The mean ascent of Nagpur (Fig. 5.l) illustrates these features.

Port Blair (Fig. 5.2) represents a maritime air mass (Tm), which is present at Trivendrum and Minicoy also. Dew point depression increases from 5° C at 900 mb to 11° C at 600 mb. Both at Trivendrum and Minicoy, lapse rate is rather unstable upto 900 mb at 12 Z for air of maritime origin, Influenced by the peninsular trough, conditions at Visakhapatnam, Madras and Bangalore are intermediate between the two air masses mentioned above, at 900 mb, but more akin to Tc aloft.

Greater Assam and West Bengal with high frequency of rain and thunderstorms, particularly the former (i.e. Greater Assam), show conditions in the mean, more similar to maritime air mass. The curve of Gauhati is shown in Fig, 5.3.

The Tc air mass warmer than Tm at surface, becomes colder by about 600 mb, due to the higher lapse rate in the former (900 mb – Nagpur 27° C, Port Blair 21° C; 600 mb – Nagpur –1° C, Port Blair 3° C; 500 mb – Nagpur –9° C, Port Blair –6° C).

5.3 In July, the characteristics of the monsoon air mass are (i) lapse rate slightly more than saturated adiabatic, (ii) high humidity (dew point depression not exceeding 7° C below 500 mb), so that ascent required for condensation is not more than 50 mb in lower levels and (iii) convective instability. Tm is the best name for this air mass, as after origin (at least at surface) in the south Indian Ocean, it traverses considerable distances over the Arabian Sea and the Bay and is modified by widespread convection. The ascents at the island station of Minicoy and coastal stations of Trivendrum and Bombay represent the conditions of the Arabian Sea branch before coming over the Peninsula, The ascents available only for evenings at Minicoy (Fig. 5.4) show surprisingly a lapse rate of 7° C km^{-1} in the first kilometer and more than saturated adiabatic by 1° C km^{-1} upto 600 mb. Dew point depression of 3° C at 900 mb increases to 7° C at 700 mb. Trivendrum is similar, except that the high lapse rate upto 1 km is not present in the morning. This increase in lapse in the lowest layer in the afternoon is witnessed at many stations. The interesting feature at Bombay (Fig. 5.5) is the rather stable layer between 800 and 700 mb (4° C km^{-1}) and drop in dew point depression from 3° C at 800 mb to 5.5° C at 700 mb. This may be associated with periodical northwesterly flow when low pressure systems are nearby to northeast. Port Blair is more or less similar to Minicoy. At Nagpur (Fig. 5.6), frequently in depression field, dew point depression does not increase with height, being 4° C or less upto 500 mb. Gauhati and Calcutta ascents are more like Nagpur, though 1 to 2° C warmer.

Madras shows the effect of travel of the air across the peninsula as 2 to 4° C increase in temperature upto 800 mb and dew point depression, particularly in the afternoon. Compared to Trivendrum, dew point depression is less at 700 mb and 600 mb at Madras. Similar differences are present between Bombay and Visakhapatnam.

Over west Rajasthan, the mean flow brings monsoon air below 1.5 km after travel over hot land and aloft northerlies flow around the western anticyclone.

Mean tephigrams, April 1961-65 (00 GMT)

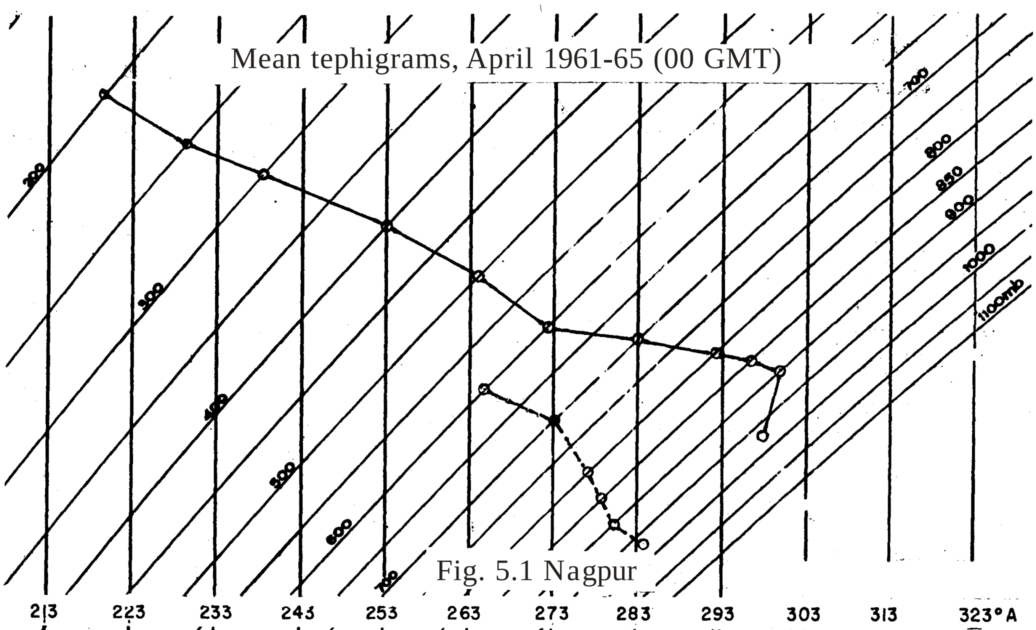


Fig. 5.1 Nagpur

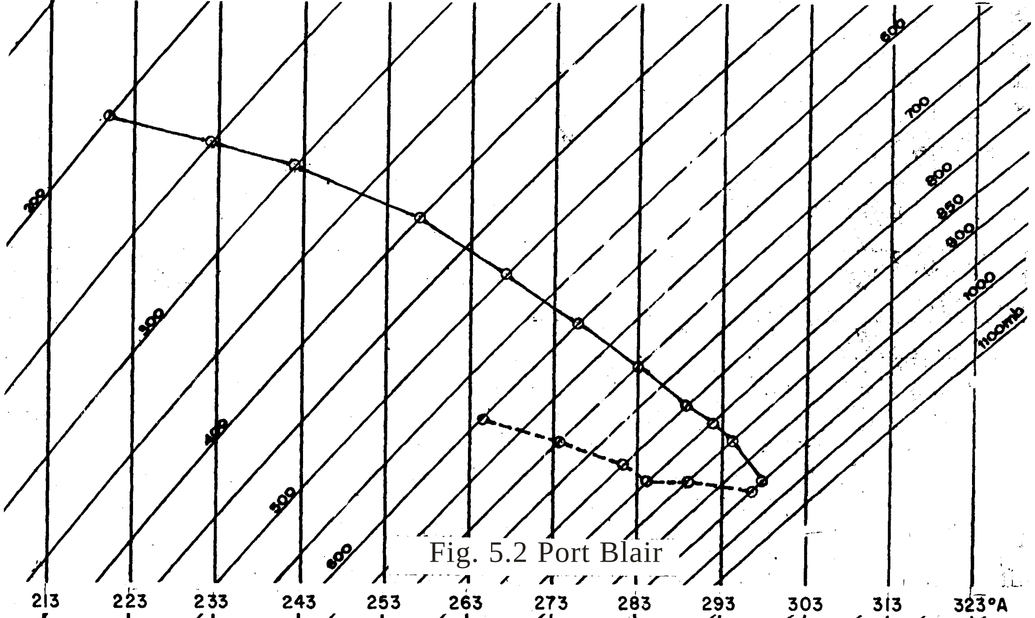


Fig. 5.2 Port Blair

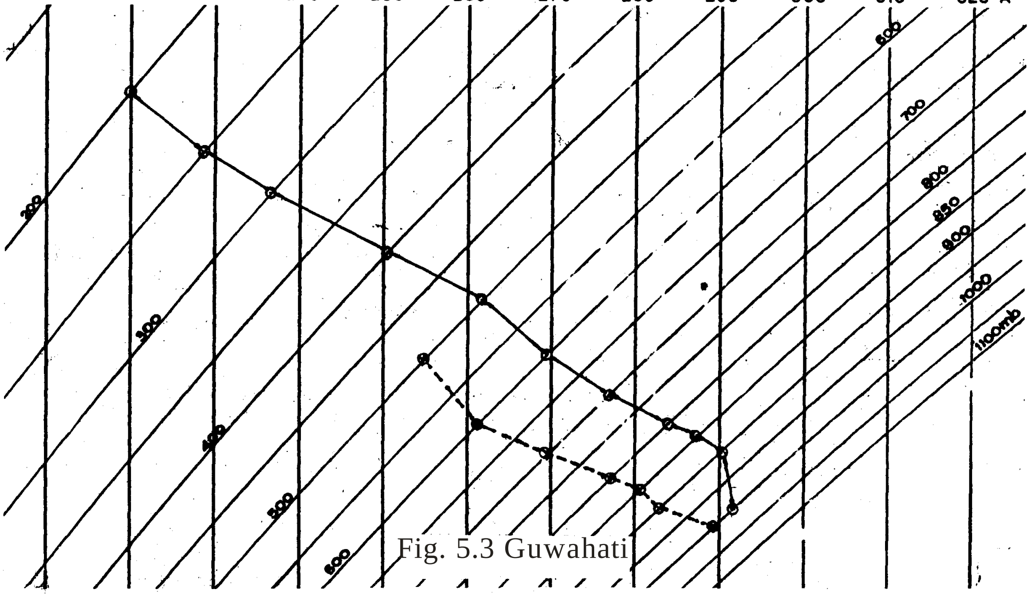


Fig. 5.3 Guwahati

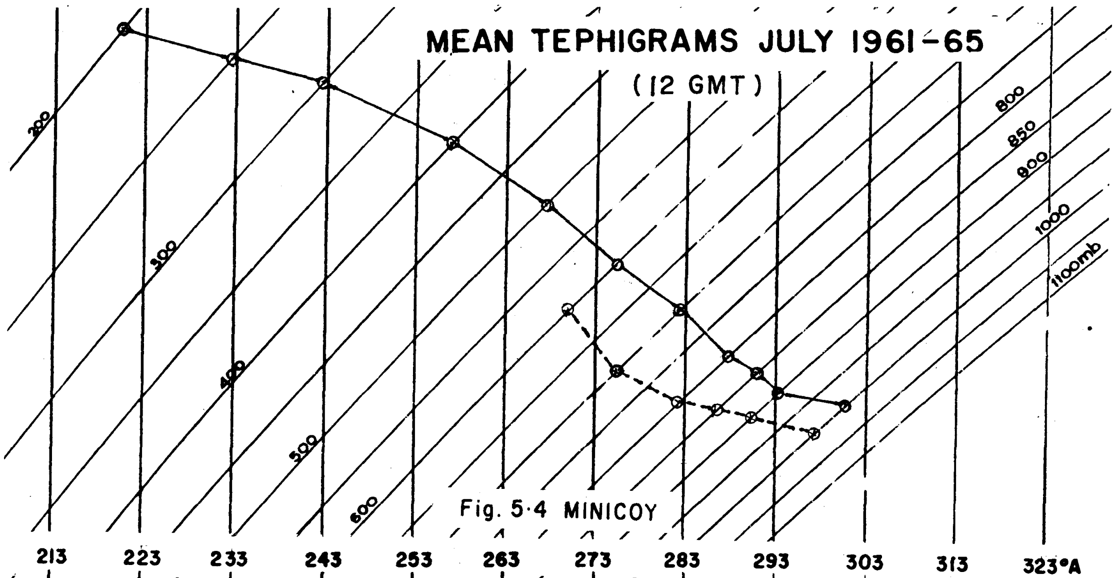


Fig. 5·4 MINICOY

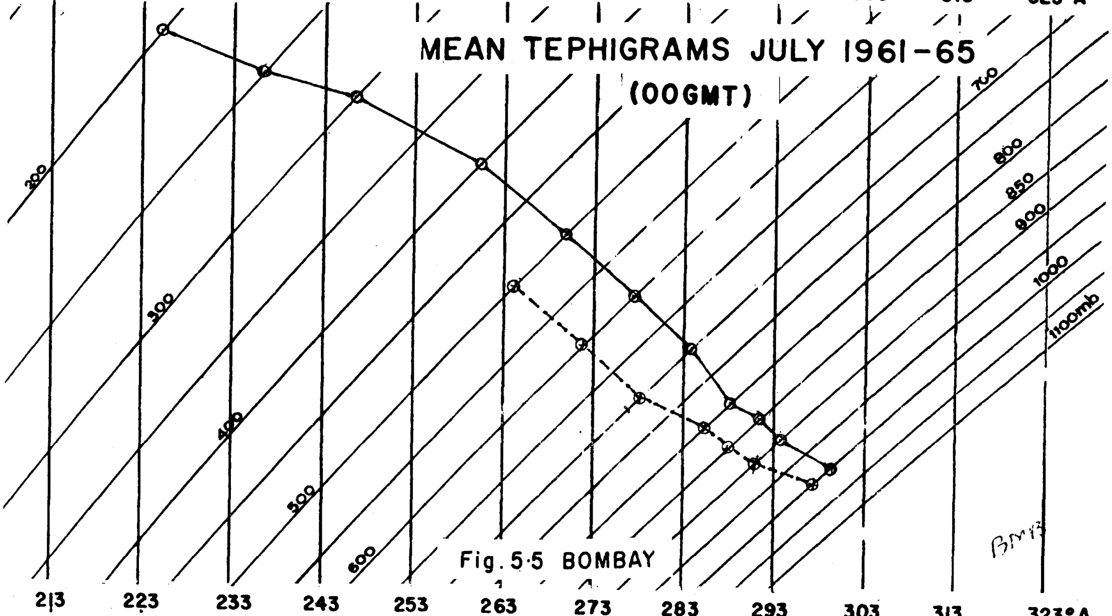


Fig. 5·5 BOMBAY

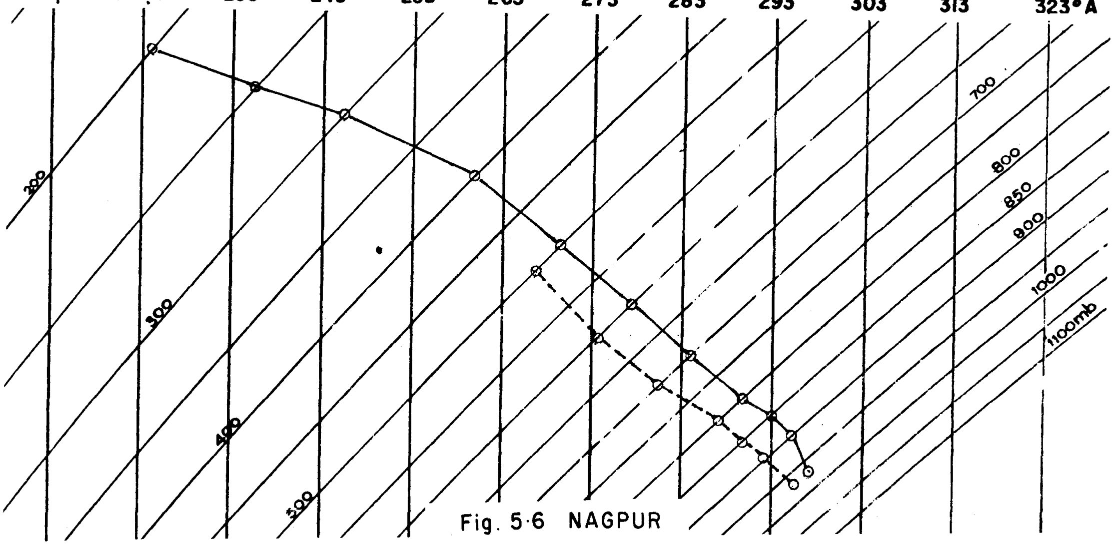


Fig. 5·6 NAGPUR

However, with favourable synoptic systems, monsoon air mass reaches this part now and then in depth so that the mean conditions are less significant. The mean ascent of Jodhpur (Fig. 5.7) is warmer than the air to the south or east. The near dry adiabatic conditions of May are replaced by lapse rate more than saturated adiabatic. Dew point depression at 00 Z is 5° C at 900 mb but increases to 10° C at 600 mb. At 12 Z lapse rate is nearly adiabatic upto 900 mb and dew point depression first decreases in this well mixed layer from 10° C at 900 mb to 7° C at 800 mb before again becoming 10° C at 500 mb. Ahmedabad shows the decrease in lapse rate between 800 and 700 mb as at Bombay. Apart from more afternoon heating, Allahabad shows conditions similar to the east and south, with dew point depression not exceeding 5° C. The characteristic of Delhi (Fig. 5.8) is dew point depression between 5^{o} and 7° C

5.4 Variations in the air mass with the activity of the monsoon rains have been studied. There is very little change in the mean temperatures from active to weak monsoon or dry weather, differences being within 1 to 2° C. Dry weather or weak monsoon shows slightly (<2° C) higher temperatures, though at Bombay and Visakhapatnam this is 1° C lower in mid – and upper troposphere. Dew points are better connected with monsoon activity. In active monsoon, dew point depression is generally 4° C or less. At the surface, there is little variation in dew point temperatures with monsoon rain intensities. Mostly above 900/850 mb, dew point depression increases with decreasing activity of monsoon, though it is not in many cases very marked between normal and strong monsoon. Fall in dew points is more with weak monsoon or dry weather. Average dew point depression in such situations is 6° C (500 mb) at Gauhati, 8° C (600 mb) at Trivendrum, 10° C (650 mb) at Bombay and 5° C (600 mb) at Calcutta. Ascents of individual days during weak monsoon or dry weather show subsidence characteristics. A few typical curves of active and weak monsoon are shown in Fig. 5.9 to 5.14 from the Forecasting Manual Series.

5.5 Krishna Rao (1952) pointed out that at Bombay every fresh strengthening of monsoon is associated with 1 to 2° C fall in temperature upto 800 mb. Break in monsoon shows decrease in lapse rate between 800 and 700 mb and marked fall in temperature above 600 mb, probably due to advection of continental air. Comparing Bombay and Poona, 80 miles apart, but on the windward and leeward sides respectively of the western Ghats, Poona is warmer by 1° to 2° C upto 500 mb but nearly the same aloft, irrespective of the activity of the monsoon. Average relative humidity is 5 percent lower upto 600 mb. These differences are probably due to the descent of air on the leeside.

5.6 Ananthakrishnan and Rangarajan (1963) have analysed the frequency of stable layers at Trivendrum, Port Blair, Madras, Visakhapatnam and Bombay using data from 1956 to 1960. Lapse rate of 2° C km^{-1} or less has been taken as a stable layer. Their results are given in Table 5.1. (on next page)

Stable layers in the first one kilometre even during this period of high cloudiness, more so in the mornings, show effect of radiational cooling. Even Port Blair has this feature while Bombay displays the least. This may be controlled more by local topography. The presence of stable layers in lower troposphere indicates that the more frequent and conspicuous stable layers noticed in west Arabian Sea and near equator are not completely effaced on a few occasions by the time the air mass reaches the Indian Peninsula and the Bay of Bengal. Subsidence prevalent in weak monsoon may also be the cause of decrease of lapse rate.

MEAN TEPHIGRAMS JULY 1961–65 (00GMT)

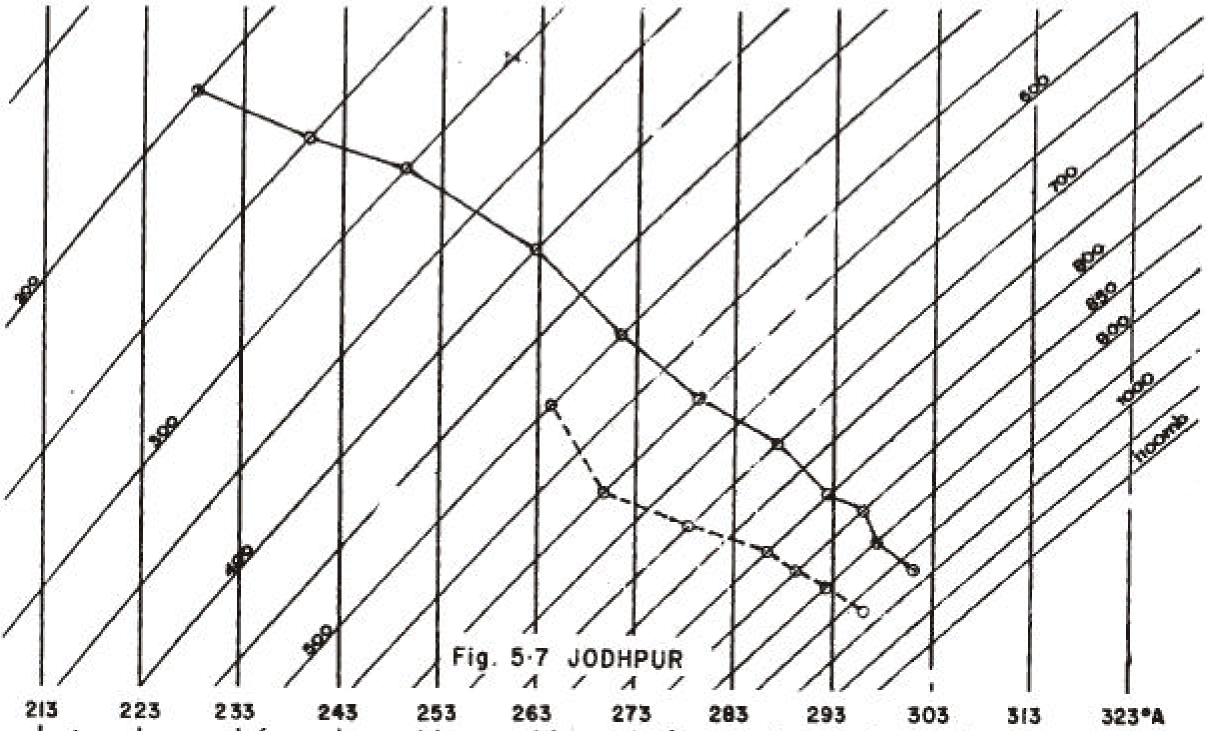


Fig. 5·7 JODHPUR

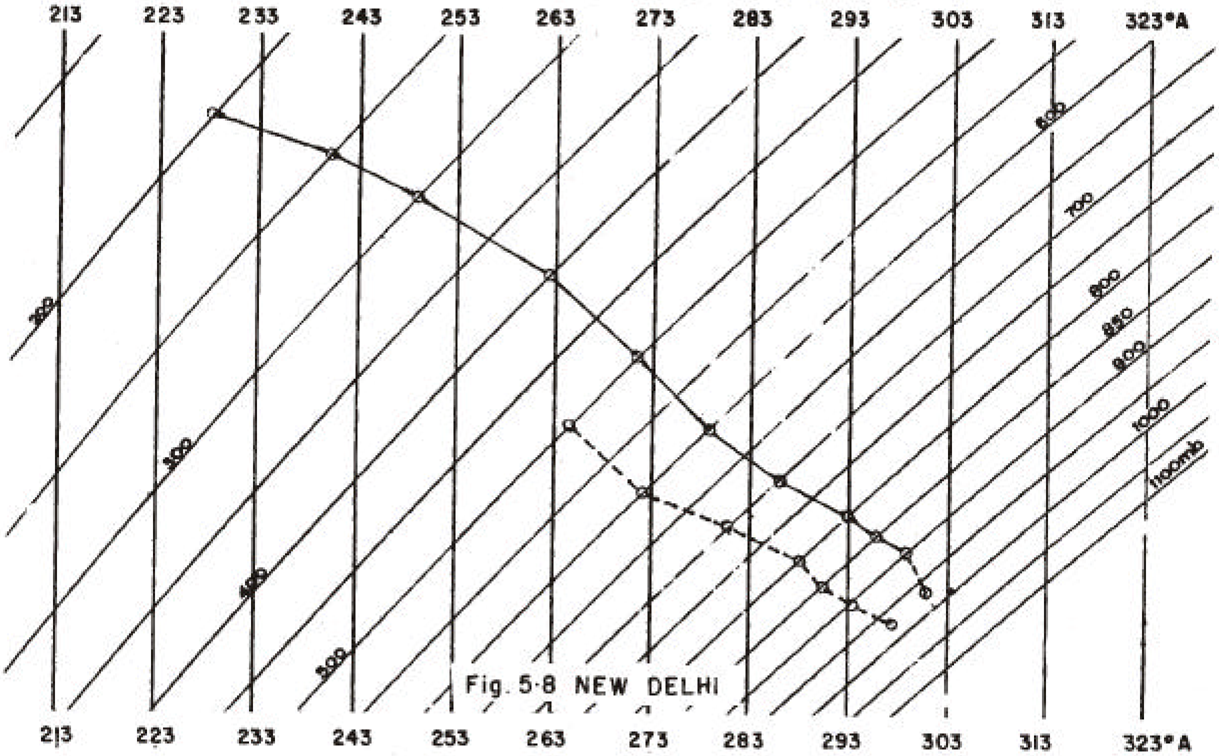


Fig. 5·8 NEW DELHI

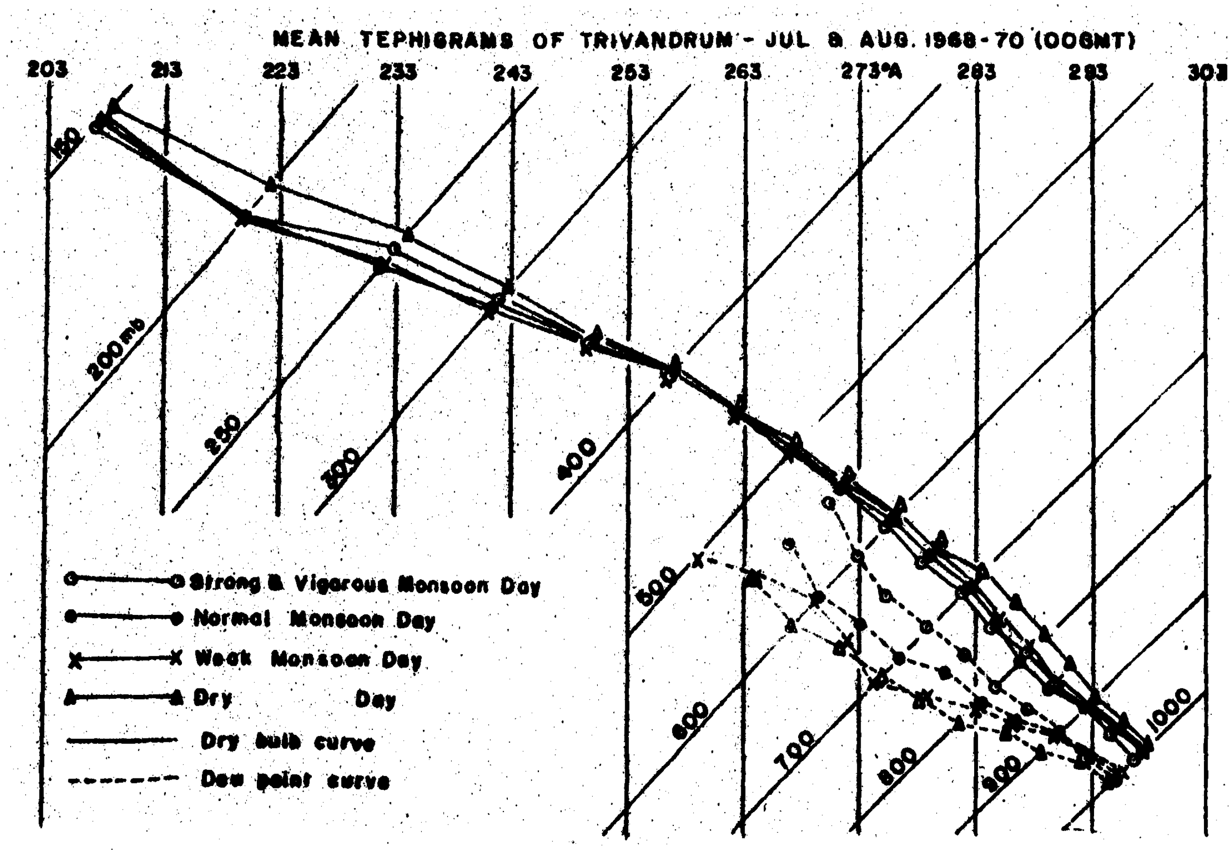


Fig. 5.9

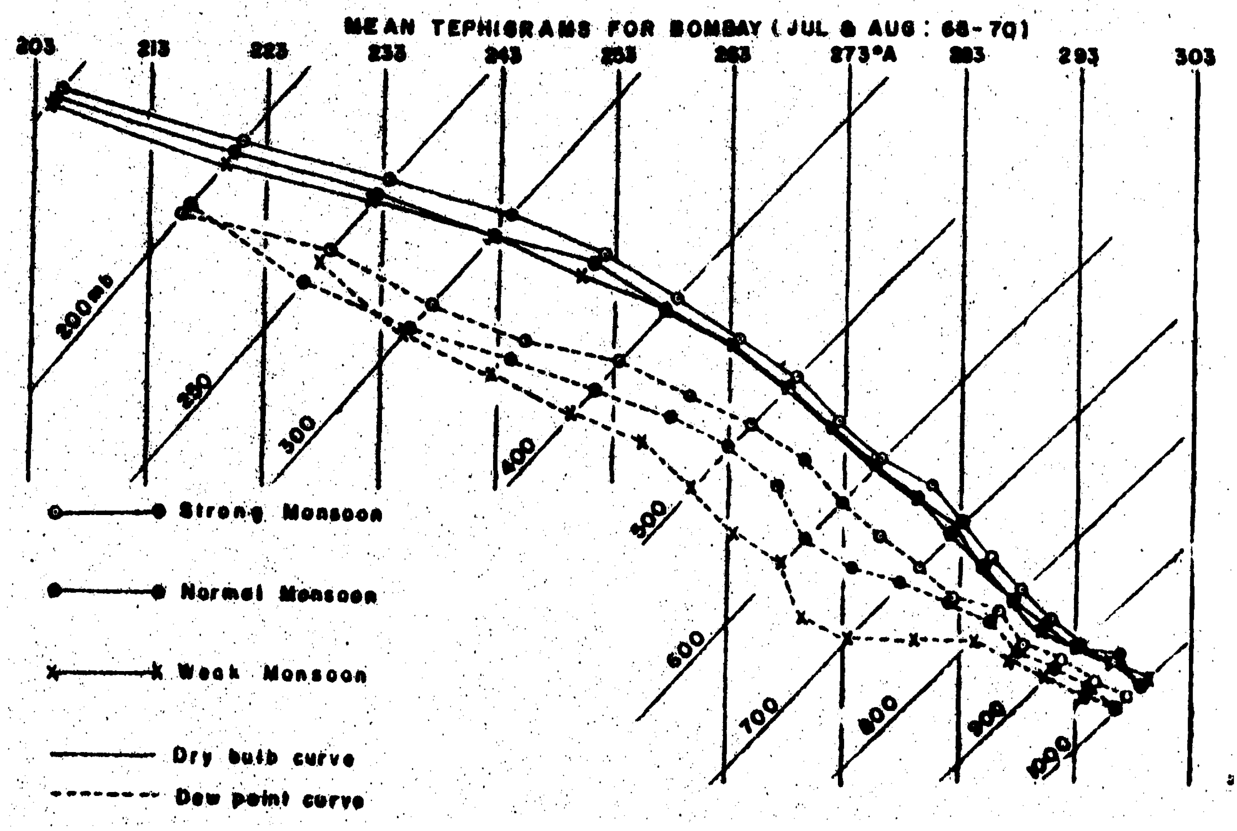


Fig. 5.10

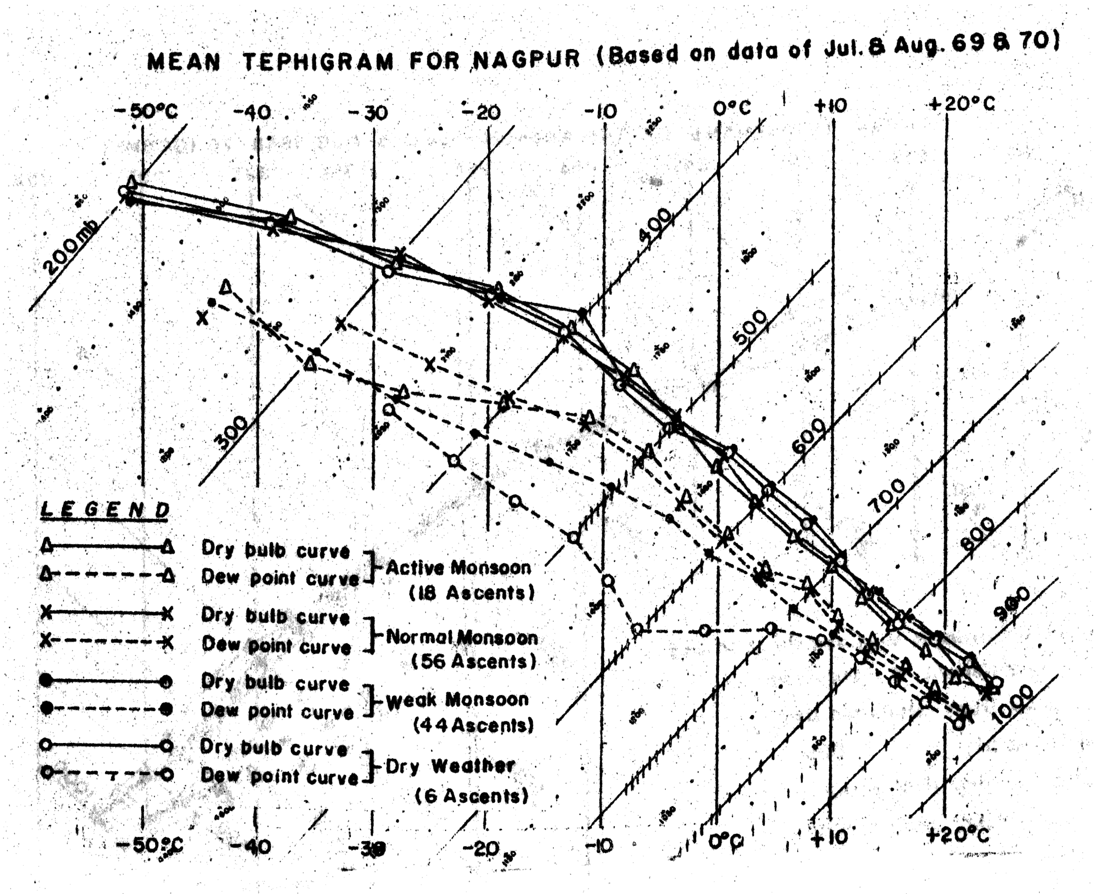


Fig. 5.11

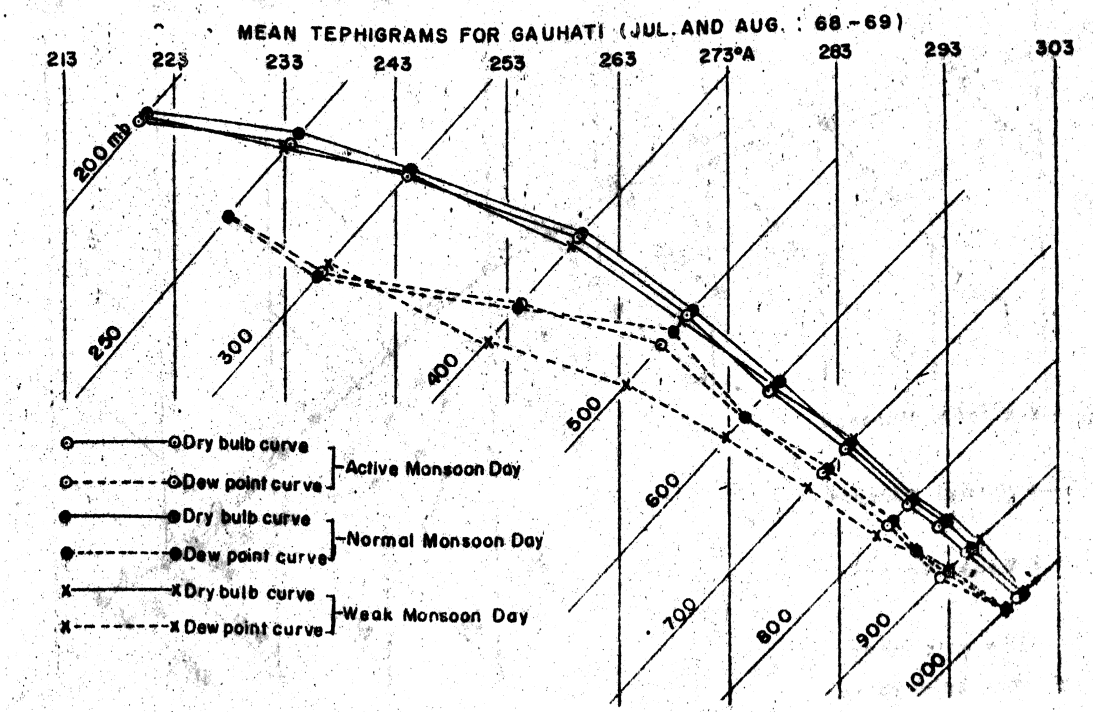


Fig. 5.12

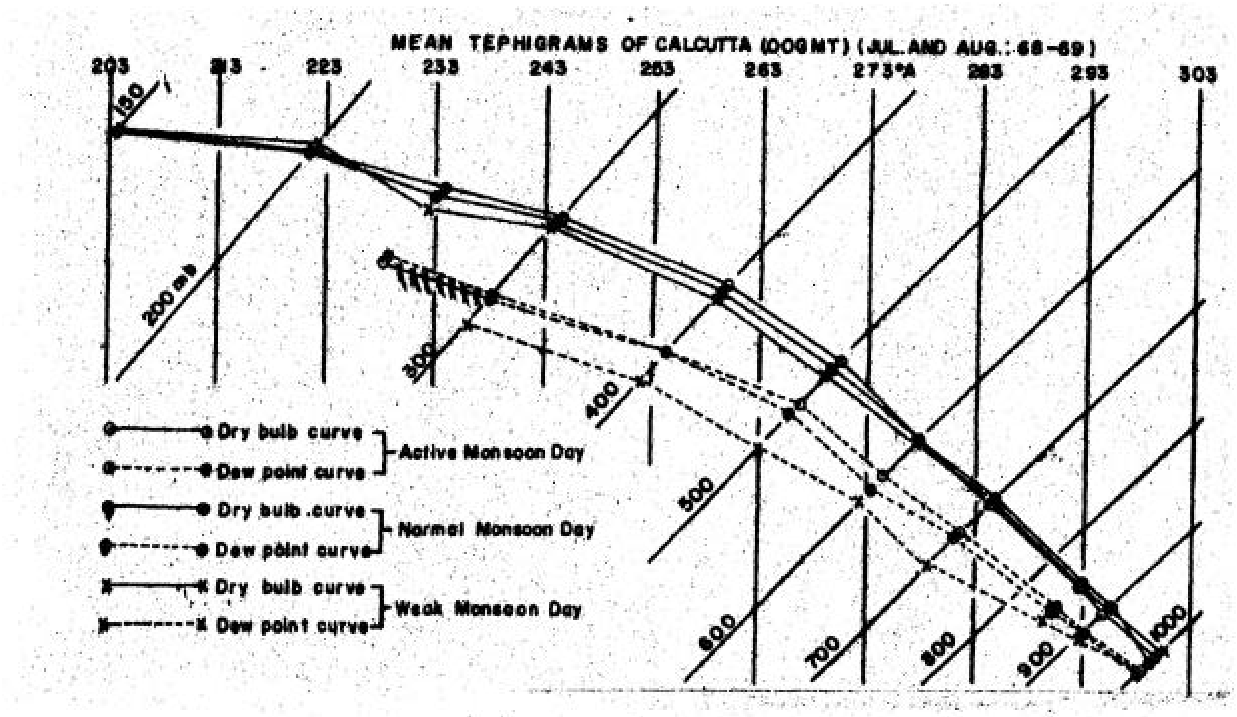


Fig. 5.13

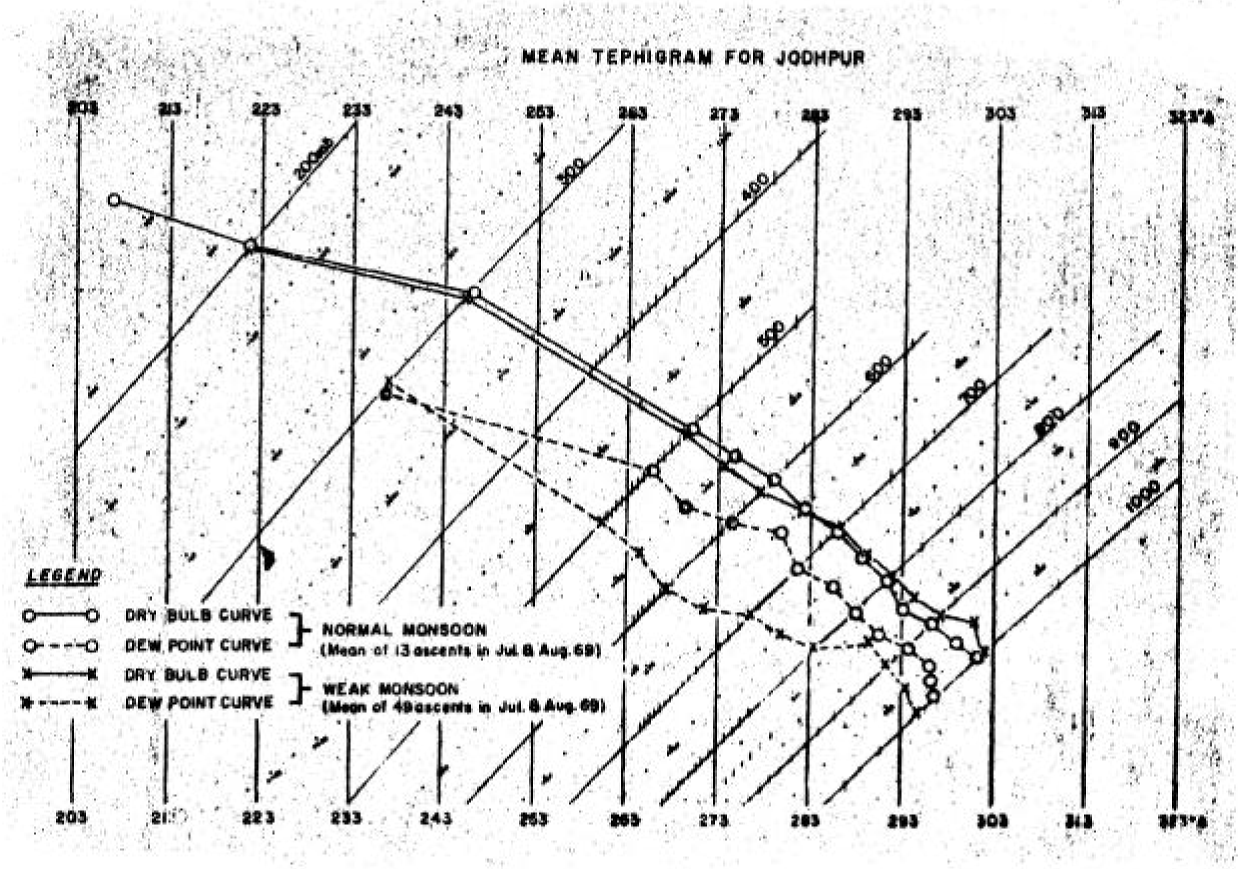


Fig. 5.14

Table 5.1
Percentge Frequencies of Bases of table Layers
(upper figures 00 Z and lower figures 12 Z)

Layer km.	Trivandrum Jun	Trivandrum Jul	Trivandrum Aug	Trivandrum Sep	Port Blair Jun	Port Blair Jul	Port Blair Aug	Port Blair Sep	Madras Jun	Madras Jul	Madras Aug	Madras Sep	Vishakhapatnam Jun	Vishakhapatnam Jul	Vishakhapatnam Aug	Vishakhapatnam Sep	Bombay Jun	Bombay Jul	Bombay Aug	Bombay Sep
19-20			5/0							0/14					17/0				5/6	0/11
18-19				0/3	8/0		0/50		8/0	6/10	0/25		17/0				17/21	12/17	8/13	22/0
17-18	4/22	6/9	8/5		18/25				11/38	19/11	17/0	0/5	0/21	19/26	7/29	6/3	23/30	14/44	31/28	24/28
16-17	42/30	17/22	28/28	33/12	15/10	0/23	20/40	33/43	19/28	13/9	18/13	29/31	20/22	28/26	20/38	4/34	20/19	25/19	16/29	22/25
15-16	14/16	28/22	20/27	25/34	12/9	5/4	12/0	6/11	21/3	23/12	18/21	23/14	9/8	13/18	9/13	28/28	3/9	9/6	8/9	12/20
14-15	9/11	14/18	18/13	19/15	0/7	6/6	4/5	10/14	3/4	3/7	5/5	4/9	2/2	9/4	10/10	14/9	5/7	0/2		2/3
13-14	2/1	3/2	2/3	4/2		2/0			0/2			2/2		2/0	4/0					
12-13	1/2			0/3	0/2	2/0											1/2	0/2		
11-12								0/2						0/2						
10-11																	0/2			0/3
9-10							2/0	2/0											0/2	
8- 9					0/2								2/0					1/3	1/3	
7- 8	3/1			0/2	2/3	2/0							3/3	2/4			1/3	3/1	1/2	
6- 7	1/3	2/2	3/1	4/1	0/2		2/2	3/1					7/4	3/4			2/4	4/6	2/1	2/1
5- 6	2/2	9/1	6/3	3/3	4/3	5/1	7/5	5/1		3/2	3/4	0/2	5/7	4/3	3/3	1/2	7/8	5/5	7/4	4/4
4- 5	4/1	5/5	4/5	3/4	0/2	3/2	5/4	5/6	0/4	1/3	3/7	3/4	4/4	5/3	3/4	1/2	8/8	3/5	5/5	10/4
3- 4	10/5	13/7	9/7	12/12	3/6	7/3	2/7	3/5	3/2	2/7	3/5	5/10	2/1	5/5	2/3	5/4	3/3	5/3	5/6	7/9
2- 3	16/13	13/11	14/11	13/12	5/5	1/5	6/5	4/9	4/0	3/2	0/4	2/3	2/2	6/9	1/5	2/6	14/9	14/11	21/12	10/16
1- 2	8/6	7/5	16/11	15/8	8/11	7/6	7/12	4/3	1/2	5/0		2/1	7/6	9/8	5/8	1/4	20/12	4/0	3/3	4/3
Surface to 1	30/11	22/5	25/12	24/18	42/23	44/22	40/23	50/14	28/26	14/21	17/20	16/17	71/57	38/26	32/17	43/12	13/11	5/3	3/3	8/4

5.7 Sawyer (1947) studied the boundary between the monsoon air mass and the continental over northwest areas of the sub–continent with the data then available. He demonstrated that the continental air warmer than the monsoon air upto 700 mb, lies above the latter at these levels. Above 700 mb, the monsoon air being warmer will overlie the continental, This peculiar variation with height of the difference in temperature between the two air masses due to the different lapse rates had been pointed out earlier by Roy and Roy (1930). Ramanathan and Ramakrishnan (1933) had noticed that the continental northerlies extended farthest to the south at 700 mb. A schematic cross section of the frontal surface between the two air masses is shown in Fig. 5.15. The slope of the lower part of the frontal surface appears to be between 1/100 and l/300. Sawyer presented the ascent of 25 August 1945 of Veraval (Fig. 5.16) showing the protrusion of continental air in mid–troposphere.

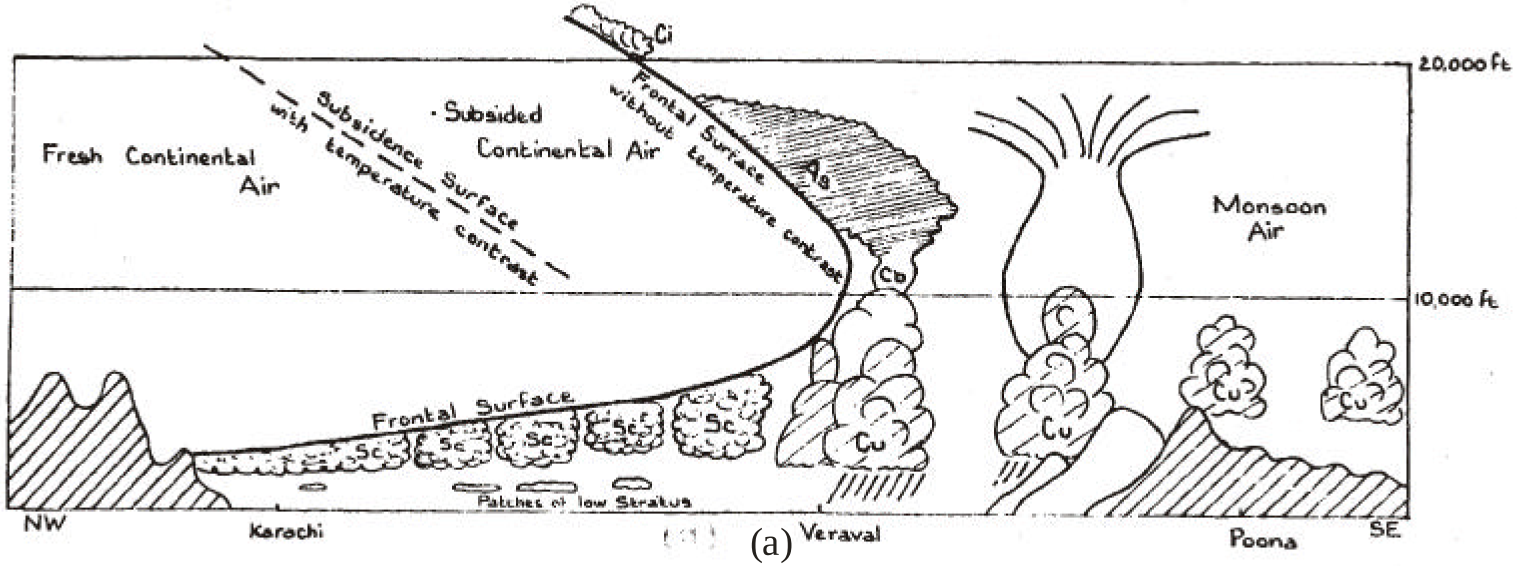
5.8 The IIOE provided radiosonde data from ships and dropsondes from aircraft over the Arabian Sea and some data also from the Bay of Bengal, Information from near the equator was interesting to know the effect of the long sea–travel that is inevitable in air coming across equator from the south Indian Ocean. The surprising features noticed at the equator and over many parts of the Arabian Sea were the shallowness of the moist layer, lapse rates much in excess of saturated adiabatic lapse rate and very dry air mass conditions and adiabatic lapse rates in mid–and upper troposphere with deep stable layers in between. Though these ascents were mainly to the west of Long. 70° E, in a few cases considerable deviations from the coastal characteristics were found to the east also.

5.9 The dryness of the air, except in a shallow layer, and the dry adiabatic lapse rates near the equator caused surprise in relation to the concept that long sea–travel would enable the air to pick up moisture. But the Northeast Trades display these features even in vast oceanic expanses and the Southeast Trades, south of the equator should be no different. In this connection, the observation of Flohn (1958) that ships' observations in the Atlantic between 10° N and 10^{o} S showed a high correlation between zonal and vertical components of wind as well as a weaker but significant correlation between meridional and vertical components, appears relevant. He explains these as due to vertical component of Coriolis force ($2wCosf$) and spherical shape of the earth (also see Rao and Chelam 195l)–The flow from the southwest over the sea to the north of the equator is more favourable for the transport of moisture into upper layers than that from southeast to the south of the equator.

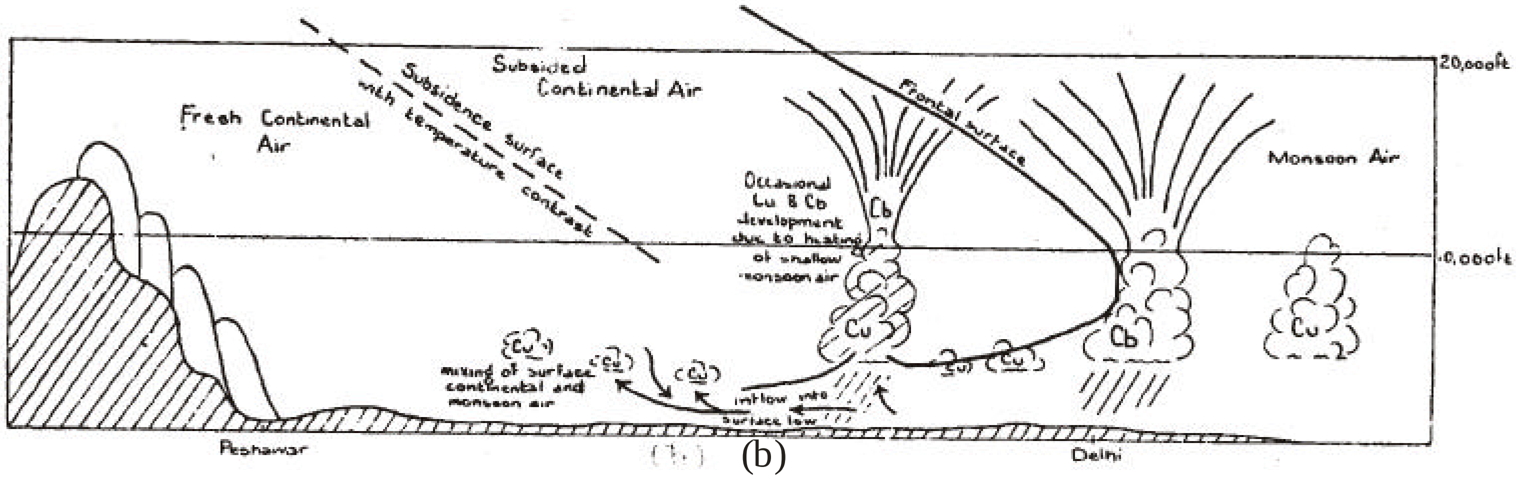
Riehl (1954) summarises the results of the sounding of the Meteor Expedition in the Atlantic, On the average, there was an inversion at 2000 m at equator between 500 and 1500 m near 15° latitude in both the hemispheres. The lapse rate within 5 degrees of the equator was 1.45° C to 0.70° C km^{-1} upto 200 m, 1.30° to 0.62° C km^{-1} between 200 and 500 m and 0.82° to (h26° C km^{-1} from 500 to 1000 m. Riehl remarks that in the equatorial trough zone inversion does not exist as a mean condition, though stable layers appear in specific weather patterns. Lapse rate is nearly moist adiabatic through the bulk of the troposphere and the moisture content is higher than in regions with inversion. But there is often a level above which relative humidity drops to low values. Occurrence of stable layers in lower troposphere with–dryness aloft, no doubt less so in equatorial trough, is a common characteristic of all equatorial oceanic regions.

Findflater (1969) presents the ascent of Dar–es–salaam (7° S) of 4 July 1966 as typical of the monsoon flow on the east African coast, The nearly dry adiabatic lapse rate upto 850 mb, inversion at 700 mb and extreme dryness aloft are very much unlike the monsoon air mass over India, particularly during active monsoon conditions

5.10 Desai (1970a) has presented the dropsonde ascents of IIO E on 2nd and 4th July 1963 over the Arabian Sea (Fig. 5.18) during a phase of active monsoon conditions along Konkan–Karnataka coast and Gujarat. On 4th July, Bombay was exhibiting nearly saturated adiabatic lapse rate and near saturation in great depth but



(a)



(b)

Fig. 5.15 (a) Schematic cross section through Intertropical Front along line joining Karachi and Poona. (Sawyer, 1947).
(b) Schematic cross section through Intertropical Front along line joining Peshawar and Delhi. (Sawyer, 1947).

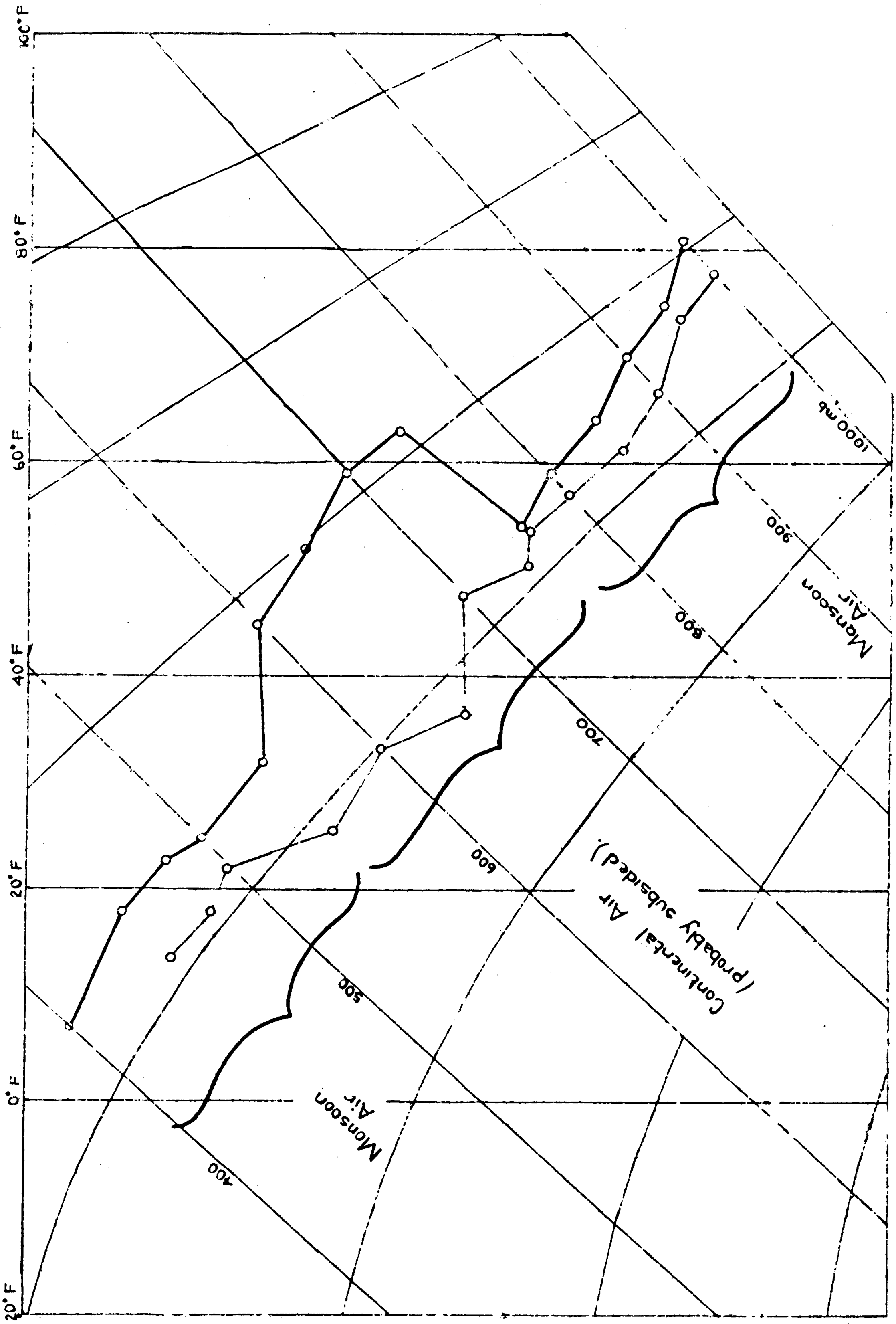


Fig.5.16 Tephigram for Veraval 25 August 1945 1400 GMT. (Wet bulb included), (Sawyer, 1947).

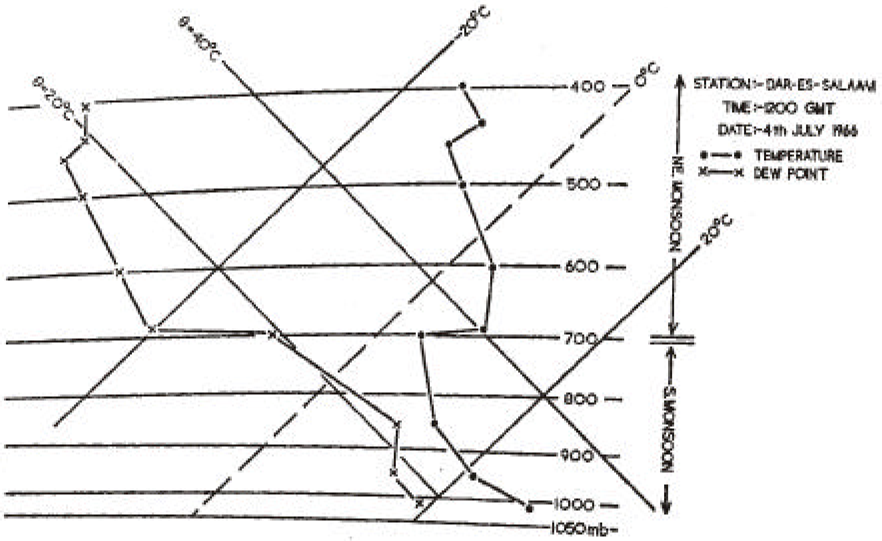
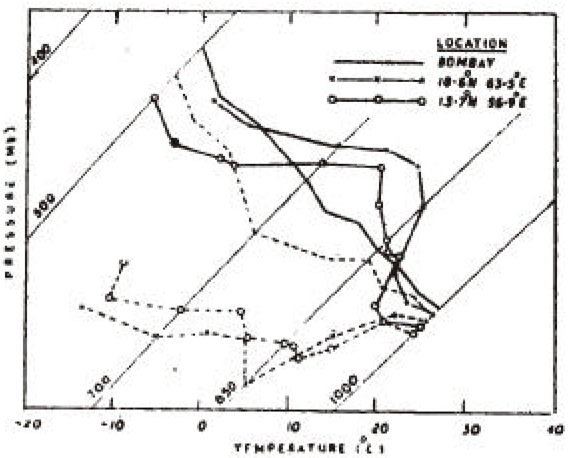
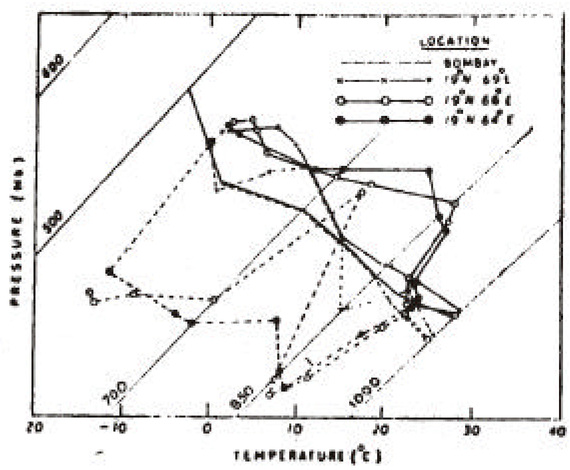


Fig.5.17 A typical upper air sounding during the southwest monsoon on the east African coast. (Findlater, 1969).



(i) Tephigram for 2 July 1963



(ii) Tephigram for 4 July 1963

Fig. 5.18

in the latitude, 7^o to the west, the similarity was only upto 900 mb. Aloft there was a sharp inversion in a layer of 50 mb and a dew point depression of $20\,^o$ C at 850 mb. Overlying this was air of about $8\,^o$ C km^{-1} lapse rate upto 600 mb. (The slight increase in humidity at 750 mb due to a cloud layer is not significant). More or less similar features were noticed at $64\,^o$ E also. The dropsonde at $69\,^o$ E in the same latitude is similar to Bombay upto 770 mb but thereafter becomes warmer, right upto 550 mb where the ascent stops. An interesting point is the stability between 770 and 650 mb or so, indicating that the transition form cold moist to warm dry air occurring at 900 mb at $66\,^o$ E is at 770 mb at $69\,^o$ E. A slopping boundary is possible between the cold moist air to the est and warm dry air in the west. The accents of 2nd July at $18.6\,^o$ N, $63.5\,^o$ N, $56.9\,^o$ E show same differences from Bombay which was less moist this day than two days later. Other interesting features are (i) dry adiabatic but moist layer at $56.9\,^o$ E between 900 mb, (ii) air warmer than Bombay by $7\,^o$ C at 770 mb becoming colder above 680 mb , by much as $7\,^o$ C at 500 mb.

Two dropsonde ascents of 9th July 1963 at $12\,^o$ N, $67\,^o$ E and $13\,^o$ N, $72\,^o$ E (Fig. 5.19a discussed by Desai (1970b) are interesting for their similarity to Dar-es-salaam ascent (Fig. 5.17) presented by Findlater. That these conditions could occur at 72^o E within 3^o of the Indian coast, would show that the air mass with features noticed near the equator could travel sometimes without much modification almost upto the Indian coast. This was not a period of strong monsoon along the west coast, Desai (1966, 1967, 1968a) considered the increase of the depth of the moist layer in the ascent nearer the coast as the effect of the Western Ghats. At many levels there is warming by $2\,^o$ - $3\,^o$ C at $72\,^o$ E compared to $67\,^o$ E. He is also of the view that for any specified strength and direction of westerlies along the coast, rainfall would be more on the coast when about 500 mb to the west there are deflected trades in the lower levels and drier unstable air above with an inversion between the tow than with moist air of near saturated adiabatic lapse rate above the deflected trades.

5.11 Colon (1964) had earlier discussed the air mass structure upto 500 mb along the trajectory from near $0\,^o$ N, $47\,^o$ E to Bombay. Four typical dropsonde over the ocean air compared with Bombay relating to 1st and 2nd July 1963 (Fig. 5.19b) All the four oceanic soundings show near dry adiabatic conditions upto 850 mb or lower, an inversion or lapse rate less than saturated adiabatic above and again a dry adiabatic layer aloft. The lows dry adiabatic layer is caused by intense vertical mixing induced by strong winds and flow over warmer sea waters. Colon recognised that the cause of the inversion is air mass differences between the moist relatively cool air mass with a long oceanic trajectory near the surface and the hot dry air mass of land origin form northeast Africa, Arabia, Pakistan, Iran etc.

The unstable layer is a characteristic of the continental air mass in the above areas. Colon is of the view that the presence of this unstable layer is of great importance to the rain-producing potential of the monsoon current, since once the inversion is destroyed there is a favourable stratification for rapid release upward of moisture leading to condensation and precipitation.

Colon discussed the processes by which the air mass dry in mid-troposphere in the western parts of the Arabian Sea becomes moist as it comes close to coast as witnessed in the radiosonde ascents of Bombay. His description of clouds as observed in the aircraft flights of IIOE is that near the equator. lines of cumuli oriented parallel to the low level flow were observed. There was a decrease in convective cloudiness northwards. The predominant cloud deck from about $7\,^o$ N to $15\,^o$ N was a low level stratocumulus layer with base around 500 m and top near 900 m. North of 10^o N, there was a layer of haze near the surface. In the west central sections, there was no middle or high clouds. From about 65^o E to Bombay , the amount of middle and high clouds and weather activity increased. Widespread cloudiness and rainy conditions predominated through the oceanic area just west of Bombay. Convergence associated with synoptic scale perturbations are essential to account for the weather to the east of $65\,^o$ E, particularly towards the coast. The effect of such systems may be accentuated by the orographic effects of the Western Ghats.

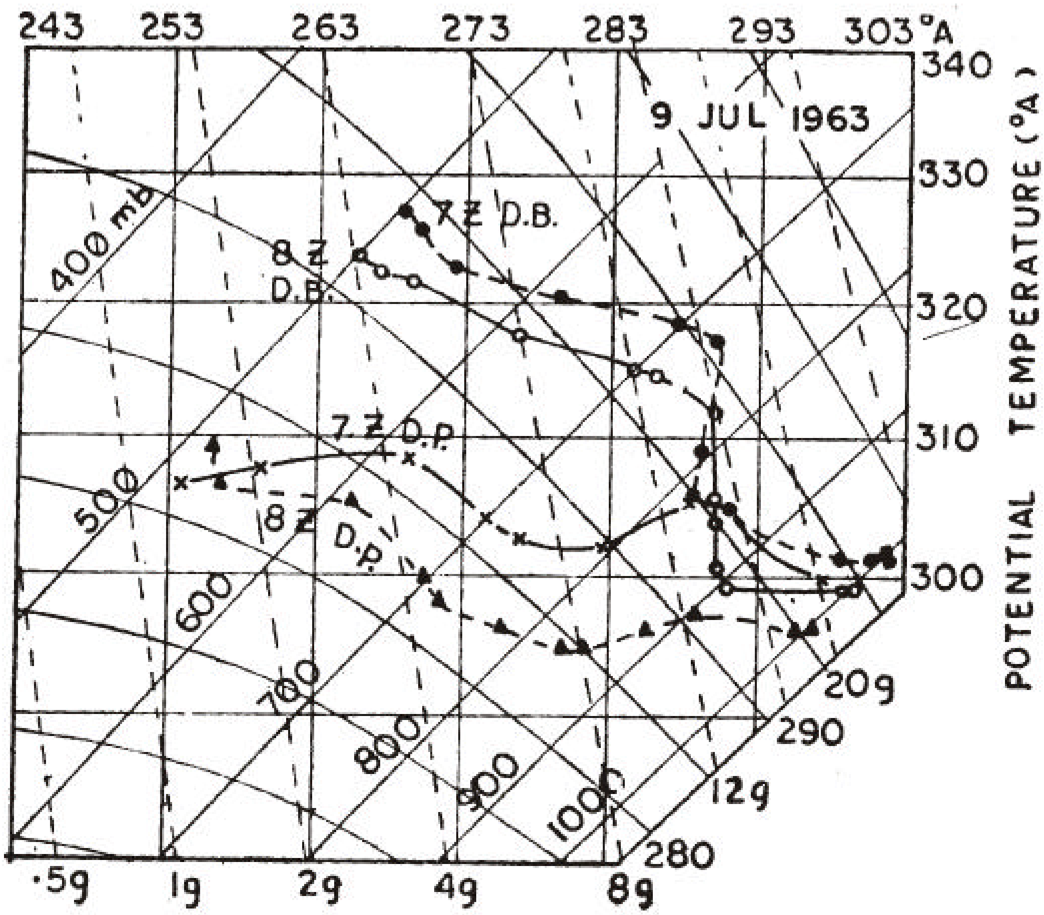


Fig. 5 19 (a) Dropsonde ascents at location 12°N 67°E (0800 GMT) and 13°N 72°E (0700 GMT) of 9 July 1963.

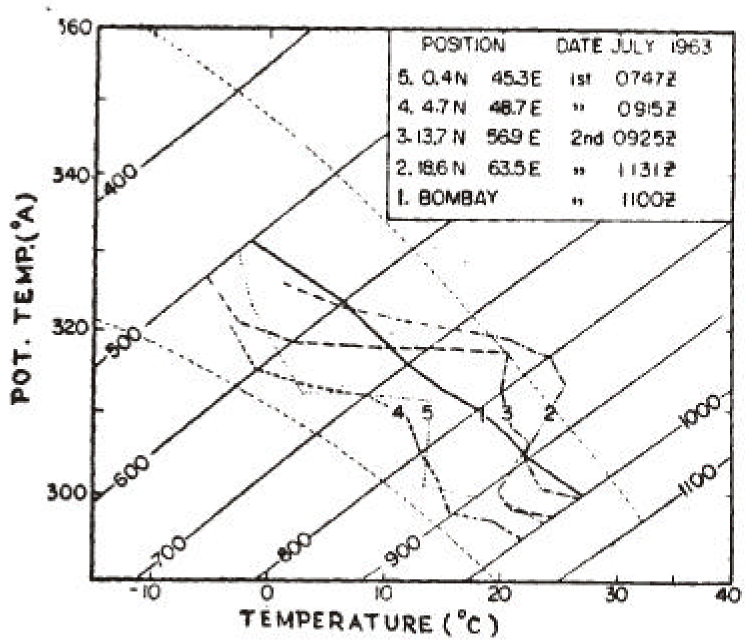


Fig.5.19 (b) Tephigrams with temperature soundings at various positions downstream along the monsoon current.

running almost at right angles to the air stream in lower levels, the influence of Ghats probably extending some distance upstream. However, weather extends too far upstream for orographic effects to be solely responsible.

5.12 Rama (1966) proposed a totally different approach to identify the origin of the monsoon air over the Arabian Sea. This was by assessing the radon content of air. Radon has a half life of 3.8 days and is exhaled continuously out of the land surface from where it is transported upwards by convection and mixing. Un like most other radio-neucleides, radon is not scavenged from atmosphere by precipitation. Radon concentration of sea water is very low and therefore the amount of radon transported from ocean into atmosphere is only one–fortieth that over land. Hence air which has spent a few days over the ocean would have a very low radon content while air of recent continental origin would show itself by the high radon concentration.

His diagrams of radon measurement on cruises from Bombay to Mombaza and Aden are shown in Fig. 5.20. In Gulf of Aden, the concentration was 140 to 160 dpm/ m3 (hereafter mentioned as units) showing the continental air mass. In the Indian Ocean to the south of 5 to 10^{o} N when not too near the coast, radon was 2–4 units confirming that the air had remained over ocean for a long period. Off the coast of Africa and Arabia, values ranged between 15 and 110 units. In the rest of the Arabian Sea radon varied between 4 and 35. He associated concentration of 20–35 units over the open Arabian Sea with continental air masses from north Africa. The radon content of 4–8 units observed as far north as 16–17^{o} N between 55^{o} and 65^{o} E is no doubt regarded as intrusion of maritime air mass.

The radon measurements by Rama are in the air at the surface. At this level the seasonal wind flow is known without doubt and the streamlines over most of the Arabian Sea are a continuation of the flow south of the equator. Whatever may be the deductions about the origin of air masses aloft, in the first one kilometre, the air over most of the Arabian Sea is of maritime origin and the radon measurements have to be reconciled with this established fact. Findlater has shown that strongest cross–equatorial flow is to the west of 45^{o} E. This air may have had a previous flow over the African continent, before flowing over the Indian Ocean parallel to the African coast and becoming southwest monsoon current over the Arabian Sea. This type of history has been visualised by both Findlater (1969) and Desai (1966). What meteorologists regard as origin for any air mass is not whether at some past time the air was over land or seas to affect its radon content, but its recent history determining its temperature and moisture content. From this criterion, the air mass over the Arabian Sea in the lowest kilometre is maritime and has come from south of equator.

Besides, the effect of mixing at lateral and upper boundaries with other air has to be given careful consideration in interpreting radon measurements as recognised by Rama (1968) and elaborated by Desai (1969) . If the drier air noticed at upper levels in western and northern parts of the Arabian Sea and adjoining parts is of subsided origin it would have only low radon content.

5.13 Recent Monex has again provided data of temperature and humidity over the Arabian Sea. Pant (1974) has presented data near equator. On the whole, lapse rates between 900 mb and 400 mb were near saturated adiabatic with tendency for stable layers, of depth between 20 and 100 mb to occur between 850 and 700 mb (some times even an inversion). Above these, there is a layer with a high lapse rate more than saturated adiabatic, indicating subsidence. Relative humidity decreased in these layers. Typical ascents of 18th June 1973 at 55.8° E and 19th at 57° E are shown in Figs. 5.21 and 5.22. At these locations there was no convection as seen from satellite cloud pictures. The ascent at 51.4° E on 17 June (Fig. 5.2l) shows subsidence characteristics being absent when there is convection, as the satellite cloud pictures showed a cloud area across equator from south near 52 ° E on l6th and 17th. Below 900 mb, the lapse rate increases from 6.4° C km^{-1} at 47.5° E to 9.1° C km^{-1} at 71.0° E. The high relative humidity at the surface and rather high values in upper troposphere can be seen from Fig. 5.23. Ascent at 1 50' S, 45.5° E of 15th June.

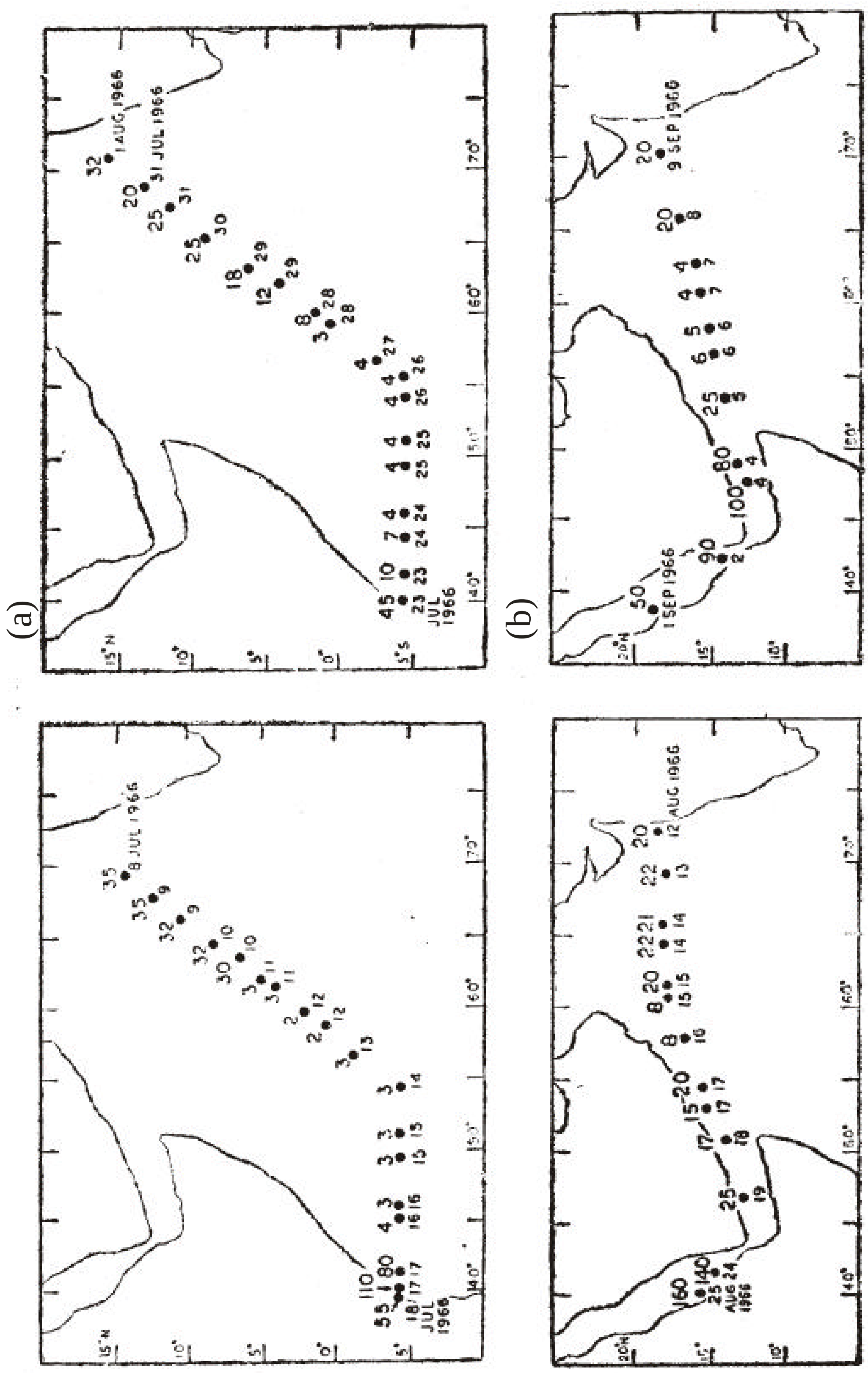


Fig. 5.20 Concentration of radon in surface air over the Arabian Sea and the Indian Ocean during monsoon period (Unit dpm/m^3)
(The upper figure indicates the concentration of radon and the lower one the date of observation

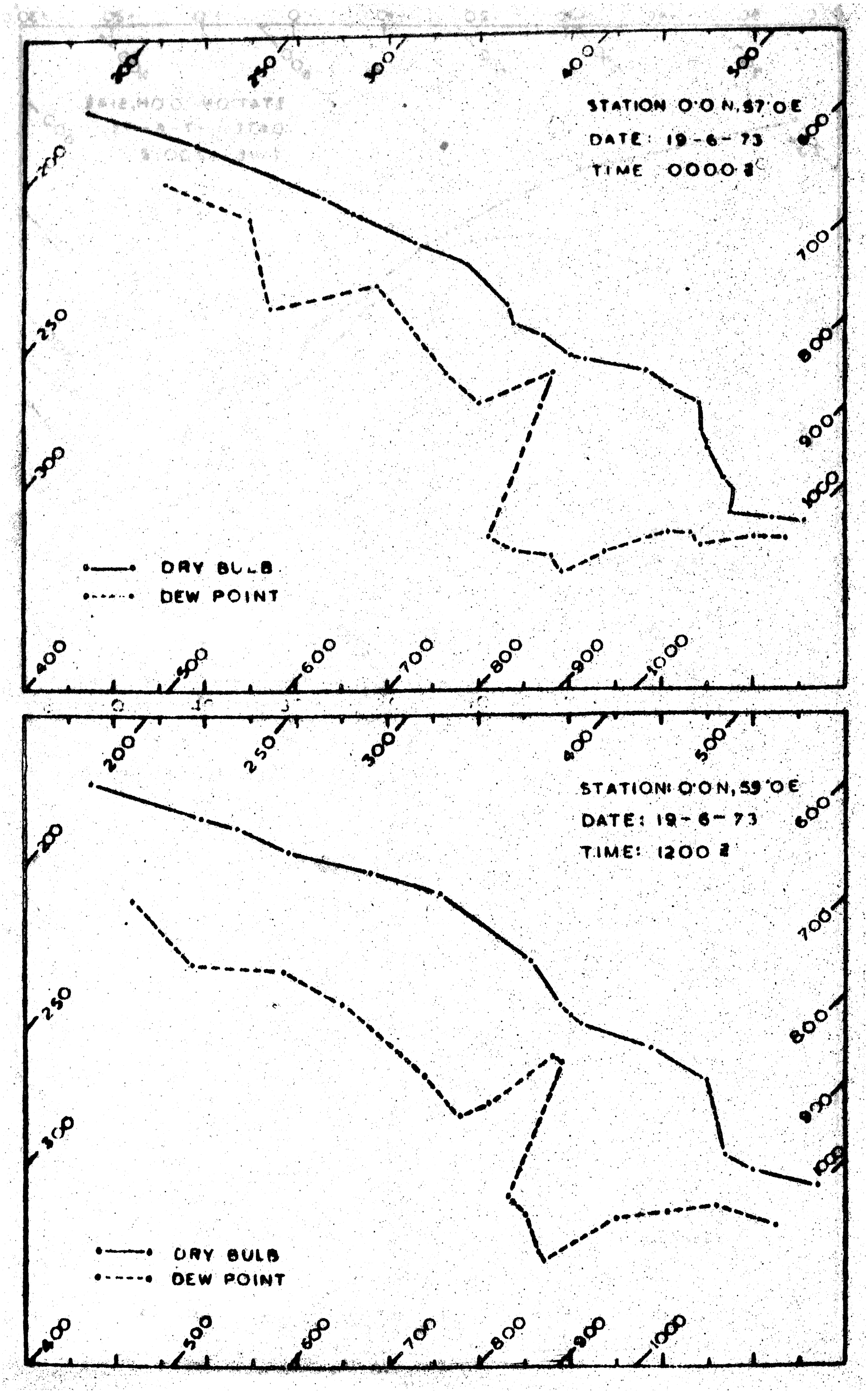


Fig. 5.21 typical ascens along the Equator

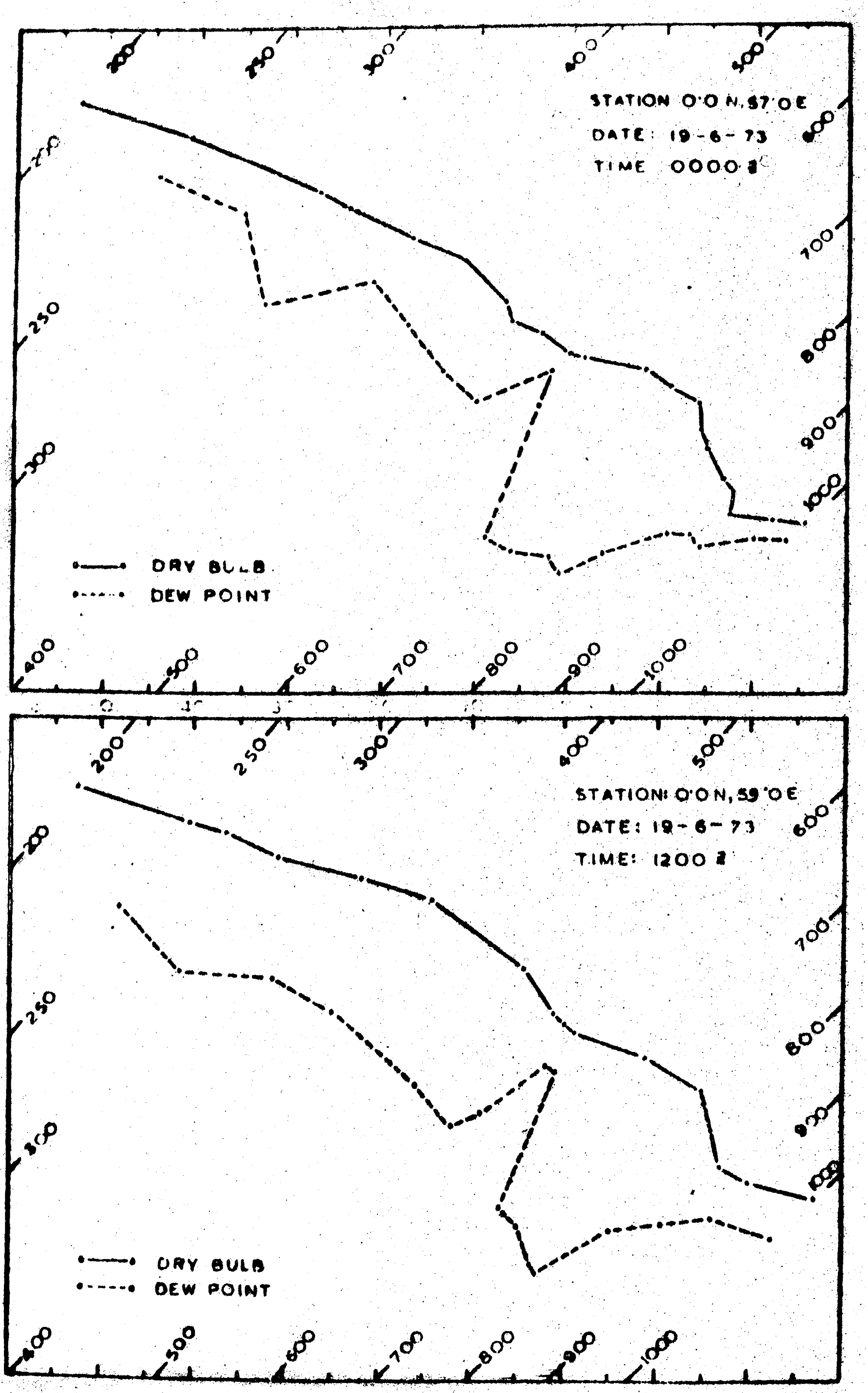


Fig. 5.22 Typical ascents along the Equator

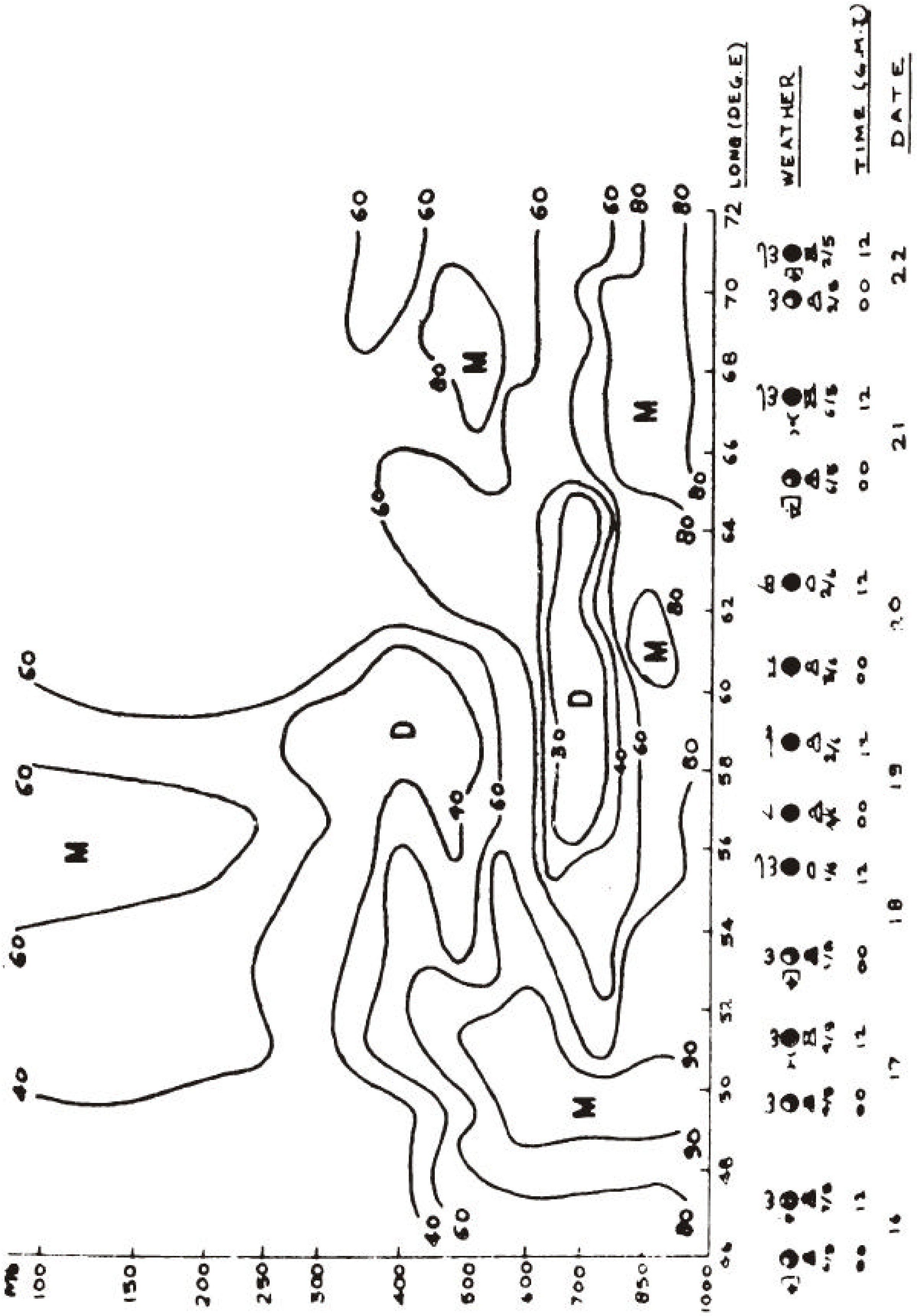


Fig. 5.23 Vertical distribution of relative humidity (%) during 16-22 June 1973 at the Equator

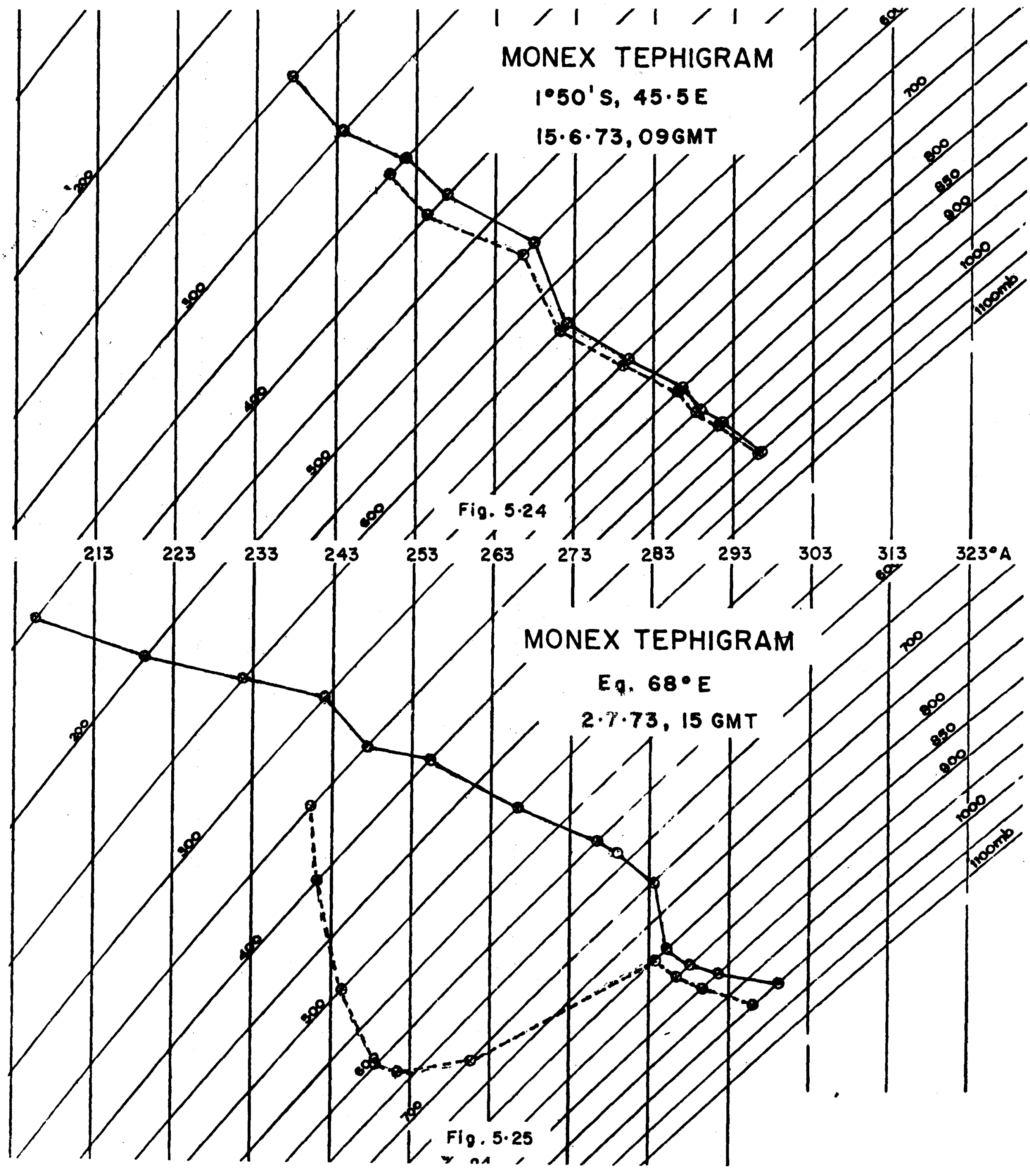


Fig. 5.22 Typical ascent along the Equator

(Fig. 5.24) shows characteristics like monsoon air mass along the west coast of India.

Subsidence characteristics in some of the ascents at equator need not lead to the conclusion that subsidence is actively taking place at that location. In the lower troposphere the trajectory of the air is from the southeast trades further south and the subsided air in the trades would retain the characteristics in the travel toward equator unless convection in the equatorial trough changes the features. In some situations further subsidence could take place as seems from the extreme dryness of ascents as of 2nd July at 0° N, 60° E (Fig. 5.25).

5.14 Though the upper air ascents over the Arabian Sea and equatorial Indian Ocean are insufficient, certain broad conclusions can be drawn from the data so far available. The type of temperature and humidity characteristics seen along the west coast south of 20° N, of lapse rate slightly more than saturated adiabatic upto at least 500 mb and very high humidity in that layer is a rarity in the Arabian Sea and near equator except 200 or 300 km from the west coast to the south of 20° N. In the rest of the area, the lowest 100 to 150 mb shows a moist layer with near dry adiabatic lapse rate. This is caused by turbulent mixing due to strong winds. To the east of the dividing line running through 20° N, 70° E to 10° N, 55° E this layer may have not infrequently near saturated adiabatic lapse rate, convection having come into play. Only in rare cases, this prevails to the west of the dividing line. Usually there is a dry layer of some depth between 850 and 600 mb. The transition to the dry layer is always as an inversion or isothermal layer in the western part, also associated with marked change in wind direction and dry air with dry adiabatic lapse rate lies above. This is continental air overrunning the maritime air below. The change to dry air in the eastern part is also associated with increase in stability with more unstable air above. Except rarely, this upper layer does not develop adiabatic lapse rate. This is the subsided part of the southeast trades, which characteristics is carried along to north of equator unless modified by convection. The most important question is how this dryness is removed in the air by the time it reaches the west coast.

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CHAPTER 6

SEMI-PEBMANENT SYSTEMS

6.1 Heat Low

6.1.1 The progressive development of a heat low over the sub-continent and its location over central parts of Pakistan in July is perhaps the most important causative factor of the monsoon. This low is a part of the low pressure belt extending from Sahara to central Asia across Arabia, Iran, Afghanistan, Pakistan and northwest India with off-shoots of troughs in various directions. The centres of the low in different months are - over east Madhya Pradesh in April, Punjab in May and Pakistan in June and July. There is a little retreat to east and south in August and September. The centres of the heat lows over land areas are located near regions of maximum heating, out of reach of maritime air mass. For, the development of heat low causes the circulation which brings cooler air mass in favourable area Offsetting the effect of solar heating. The low centres develop depending upon the balance between these two factors. The tapering shape of the Peninsula and the Himalayan barrier to north and Assam hills to northeast which make the maritime air mass pervade over most of India displace the centre of the low to the extreme northwest of the sub-continent. Blocking of cold air incursions from the north by the Himalayas in the lower troposphere must have made the heat low over the Indian sub-continent more intense.

6.1.2 The lowest pressure for the whole belt from Africa to Asia is over Pakistan and neighbourhood. This is not the place of the highest temperature for the lowest pressure to develop. Sahara records much higher temperatures. Banerji(1930, 1931) considered that the configuration of mountain ranges in the northwest corner of the sub-continent and the adjustment of wind field and consequently the pressure pattern throws up the low pressure centre over the central parts of Pakistan. As far as the Indian monsoon is concerned, what is important is the existence of low pressure centre in the northwest of the sub-continent and not whether the pressure in it is the lowest for Africa and Asia.

6.1.3 This heat low is only in the first 1.5 km and is over-lain by a well marked ridge extending upto upper troposphere, which is a part of the subtropical high pressure belt. There is frictional convergence in the heat low and weak ascent in it while subsidence is likely in the ridge. A more stable layer may occur in the lower part of the column of subsidence due to vertical shrinking. In mean ascents, this kind of stable layer does not show itself prominently. This is the case even in trade wind inversion. Examining the mean ascents of Multan and Quetta (Fig. 6.l) in the area of heat low, a layer of increased stability seems to occur between 500 and 400 mb as shown in the table below :-

Table 6.1

Lapse rates ($^{O}C\ Km^{-1}$)

		600-500 mb	**500-400 mb**	**400-300 mb**
Quetta	July 46	0.75	0.51	0.65
	July 47	0.71	0.57	0.67
Multan	July 45	0.66	0.54	0.71

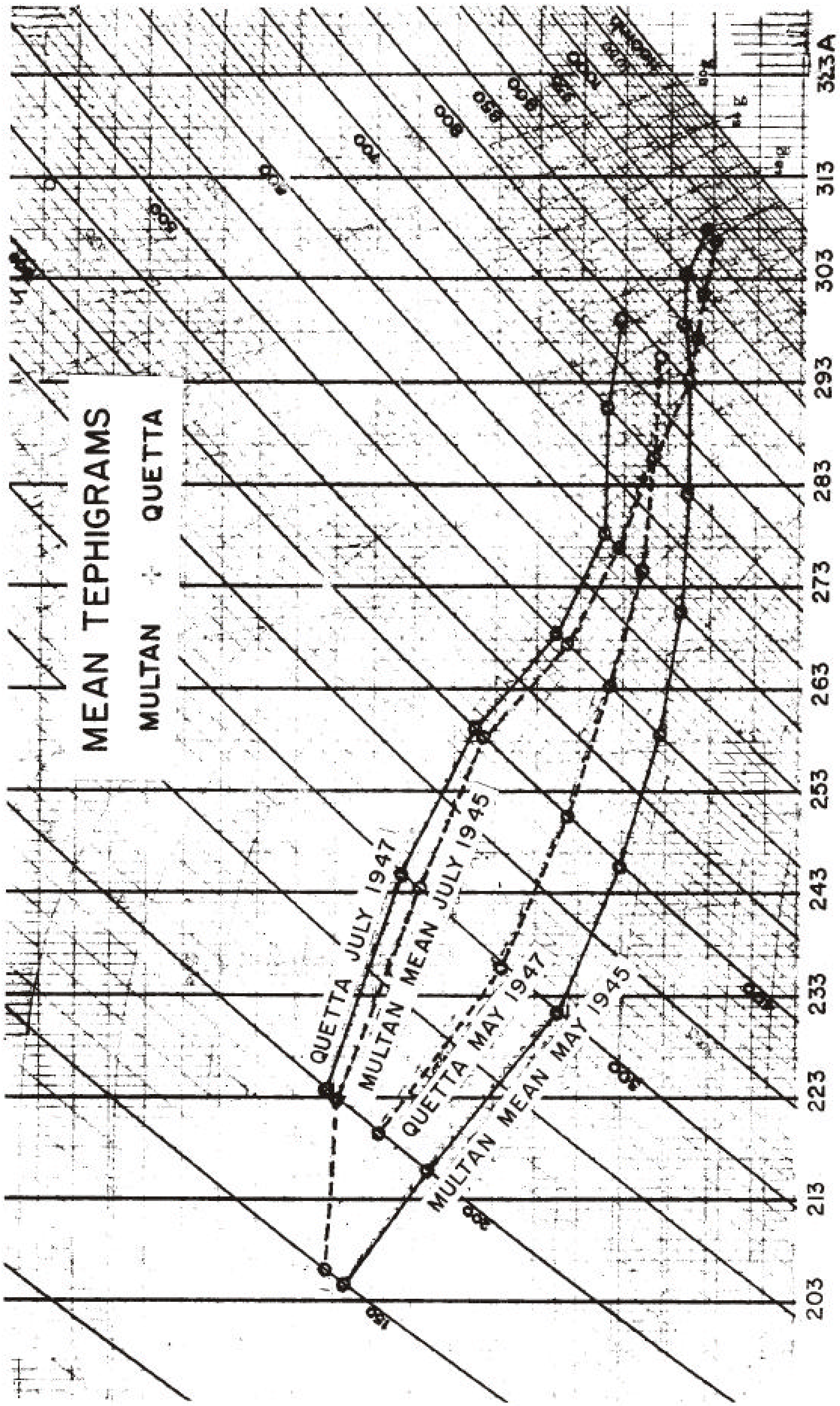

MEAN TEPHIGRAMS
MULTAN + QUETTA
QUETTA JULY 1947
MULTAN MEAN JULY 1945
QUETTA MAY 1947
MULTAN MEAN MAY 1945
203
213
223
233
243
253
263
273
283
293
303
313
323A

Fig. 6.1

Subsidence seems to extend upto 500 mb in the mean. At Quetta, dry adiabatic lapse rate prevails from 800 mb (surface level) to 600 mb. Multan has a slightly less lapse rate, perhaps due to occasional incursion of moist air. Sawyer (1947) found cold outbreaks of continental air across Iran and Afghanistan into India causing lowering of temperatures over Punjab between 700 and 500 mb. The warming which follows, suggests subsidence in the continental air above 700 mb over northwest India. This, according to Sawyer, appeared to be confirmed by some of the soundings in the vicinity of the partition between the continental air and the moist monsoon air which showed presence of air above 700 mb with temperatures similar to monsoon current but with humidities too low for the monsoon air mass.

6.1.4 The intensity of the heat low has been correlated with monsoon activity with some success. Departure of pressure from normal in this region and gradient of departures are taken into account, Below normal pressures in the heat low region and above normal pressures in the Peninsula are regarded as favouring monsoon activity over the country. Pressure gradient would then be strong over the Peninsula which, is conducive to monsoon rains. The heat low may also strengthen when the ridge aloft weakens under the influence of westerly troughs moving further north. Some westerly troughs can cause formation of weak lows over northern India and lead to increase of rainfall.

Ramage (l97l) has shown (Fig, 6.2) that the surface pressure at Jacobabad, in the heat low, is inversely related to the intensity of monsoon rains over a strip of the sub-continent between 18° and 27° N, He quoted the observation of Dixit and Jones (1965 Prepublished) that upper tropospheric temperatures over the heat low are warmer by 2^{o} - 6^{o} C during rains than in monsoon lull and concludes that subsidence aloft raises temperature of middle and upper troposphere and reduces surface pressure in heat low. As would be seen in later sections, rainfall in this strip is not only influenced by tropical systems approaching from the east but also by the trailing edge of the troughs in westerlies in mid-and upper troposphere. Both the systems even interact in causing rains. The effect of such systems would be to weaken the mid-and upper tropospheric ridge above the surface heat low and show that strengthening of the heat low is positively correlated to increase of monsoon rains.

6.1.5 As a ridge overlies the heat low above 1.5 km, the question arises whether the ridge has a role in maintaining the surface low. As the axis of the ridge is close to the axis of the heat low, the pressure distribution of the ridge is inimical to that of the low. While it may be argued the subsidence in the upper ridge is maintaining the clear skies required for the heat low, this is not the case in the months before the monsoon sets in. In April and May, the sub-tropical ridge is well outside the heat lows over the sub-continent.

6.1.6 The desert area of the Indian sub-continent comes under the heat low with sub-tropical ridge aloft. Bryson and Baerries (1967) have pointed out that the desert coincided with the extent of the divergent sinking of air at 700 mb. Further, according to them, if the atmospheric subsidence over the area was less, the moist monsoon layer which is ordinarily shallow, would be deeper and the slight summer rainfall maximum would be considerably larger. Their discussion inclines to the view that if subsidence could be reduced by lessening the quantity of dust suspended in the atmosphere in that area, it may be possible to increase rainfall in the desert areas. As pointed out by Desai (1968) the shallowness of the monsoon air over the desert is due to the continental air brought in by the sub-tropical ridge. This circulation pattern is not due to dust being carried aloft from the surface of the desert. Even over the Arabian Sea, west of 68°E, monsoon air mass is only 1 to 1.5 km deep, with continental air aloft, brought by the anticyclonic flow which cannot be ascribed to any surface dust, unless the argument is built around advection of dust from neighbouring land areas. There appears to be so little justification to expect a deepening of the moist layer over northwest India if the dust is somehow removed.

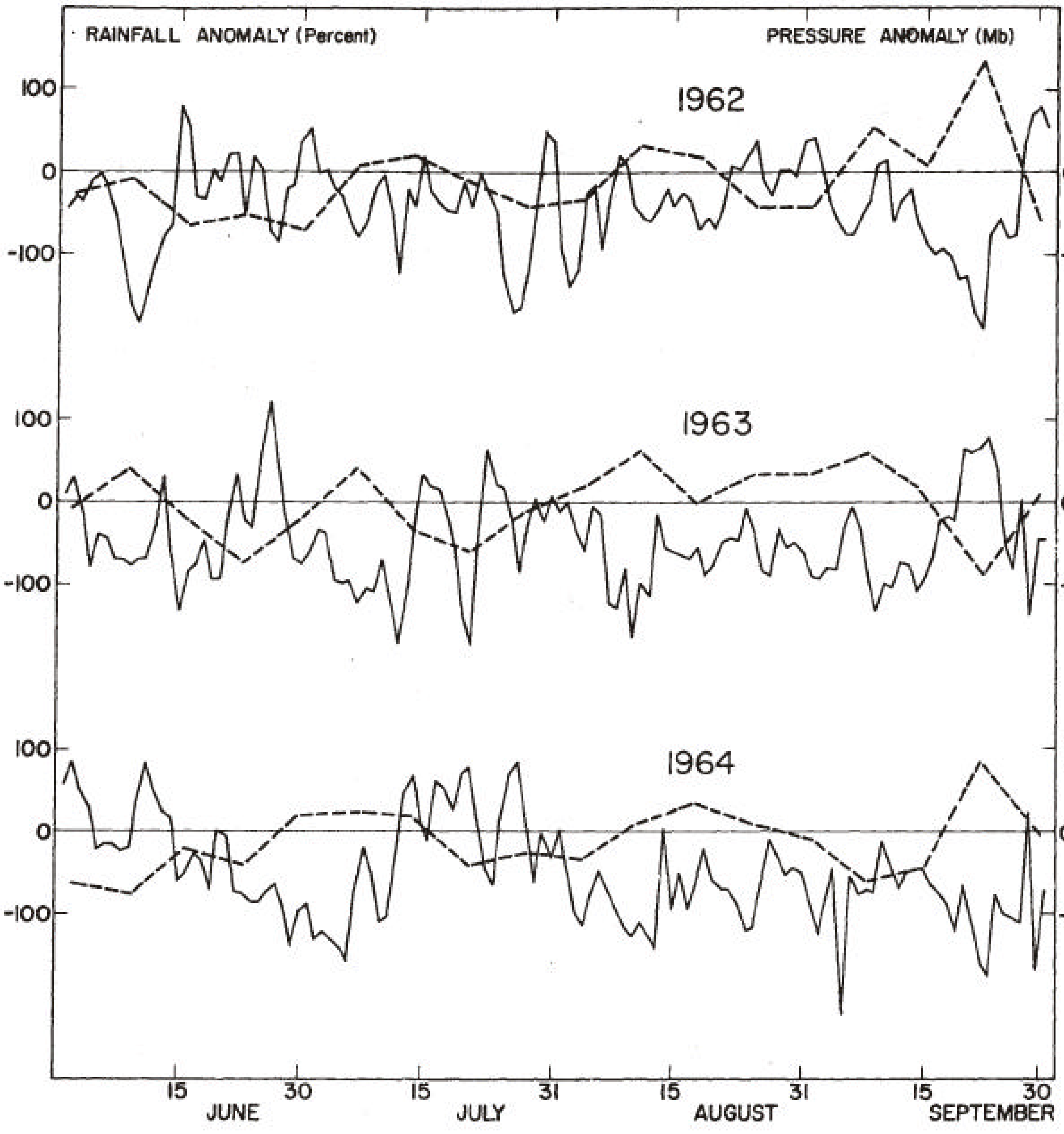


Fig 6.2 Daily anomalies (mb) of mean sea level pressure measured at 0800 L.T. at Jacobabad (Full line) and weekly percentage anomalies rainfall Indian sub division lying roughly between 18 ^{O} and 27 ^{O} N (dashed line) for th summers of 1962 1963 and 1964 (Ramgae, 1971).

6.2 Monsoon Trough

6.2.1 From the seasonal low over Pakistan and neighbourhood, a trough extends southeastwards to Gangetic West Bengal. The trough line runs at surface from Ganganagar to Calcutta through Allahabad, with west to southwest winds to south and easterlies to the north of the trough line. Mean surface wind at Calcutta is mainly from south in spite of the depressions which form in the north Bay of Bengal. This trough is seen in the upper air also upto about 6 km, the trough line sloping southwards with height. At 4 km it runs from Bombay to Sambalpur. The air mass to the south of the trough line is the Arabian Sea monsoon while the air to the north may have had some travel over the Bay. About 2° C higher temperature to the north, and the slope of the trough line in the right direction, indicates resemblance to a quasi-stationary front. Srinivasan, Raman and Mukherji (1971) have pointed out that in the western sector the trough slopes rapidly southwards with height in the lower troposphere but the slope becomes less marked in mid-troposphere. In the eastern sector, the southward slope is less in the lower troposphere and increases appreciably in the mid-troposphere. These variations are consistent with thickness gradients in the two portions, as given below :

Table 6.2

Mean Thickness Gradient near Monsoon Trough

Section	Isobaric Layer (mb)	Mean thickness gradient (gpm per degree latitude)
Jodhpur to Bombay	1000-850	3.4
	850-700	2.9
	700-500	2.0
Allahabad to Visakhapatnam	1000-850	1.3
	850-700	1.4
	700-500	3.4

The monsoon trough is regarded as the equatorial trough of the northern summer in the Indian longitudes. In this season, a weaker trough persists within five degrees south of the equator. Riehl (1954) shows that the pressure profile about the equatorial trough (in summer hemisphere) averaged over the globe is quite symmetrical in northern and southern summer. In spite of the existence of a weak second trough, in the Indian longitudes close to the equator in the winter hemisphere, Riehl's delineation may still be applicable to the more dominant system. In the Indian monsoon trough, the pressure gradient equatorward of the trough is more than the average for the whole globe.

6.2.2 The position of the trough line varies from day to day and has a vital bearing on the monsoon rains. No other semi-permanent feature has such a control on monsoon activity. This is not so much due to upglide along the slope of the trough or convergence in it, as due to the different synoptic systems that prevail with the different positions of the trough. Position of trough line close to the foot-hills has come to be referred as 'break in monsoon' on account of drastic decrease in rains over the country, though the Himalayan mountain belt experiences heavy falls which can cause floods in the rivers originating there. Swing of the trough to the central parts is with monsoon depressions from the north Bay moving west to westnorthwest across the country. The run of the trough line cannot be uniquely identified when such systems are embedded.

6.2.3 Riehl (1954) discusses the distribution of rainfall relative to the equatorial trough. Heavy rainfall and equatorial trough coincide seasonally as well as in the annual mean over the oceans. The rainfall maxima over land lie equatorward of the trough wherever its position deviates markedly from the equator. Especially over Africa in July and Australia in January, the trough line marks the ,poleward margin of the rain belt, in agreement with synoptic experience. Along the convergence line itself, Tc air begins to flow over Tm air at the ground. Because Tc air is relatively dry, the skies are entirely clear at the trough, even with occasionally extreme convergence. The precipitation belt remains well to the south, where Tm air is present in depth and where the temperature difference between the ascending Tc and the Tm has disappeared.

When we average the rainfall relative to the equatorial trough, we obtain symmetrical profiles over the oceans. During the southern summer, the continental peak (in rainfall) lies 2^o latitude north of the trough; in the northern summer, it departs much farther. Solot (l950) also finds that with ITC over Sudan located between 15^o and 17^o N in June, July and August, precipitation is not at the boundary itself but commences only 300 km to the south.

At least over the central and eastern parts of the Indian monsoon trough, the air mass is always maritime and 2^o C warmer in the north than the south. The trough slopes southwards with height. Mean rainfall is more to the south on account of the heavy rains in the southwest quadrant of depressions which travel westnorthwestwards, a little to the south of the mean position of the monsoon trough line. Along the mean position of the trough itself, rainfall is a minimum and again increases towards the foot of the Himalayas.

6.2.4 Normally some rain occurs near the surface trough and to south on account of the convergence in it and southward slope with height. When the trough rapidly shifts north or south, which can be even 5 latitude in a day, monsoon activity is enhanced in that area. The latent instability of the air mass enhances rainfall in the trough when it is over Bihar and east Uttar Pradesh, even without a low.

The charts of 7 July 1962 (Fig. 6.3 a and b) from Rao, Srinivasan, Ramakrishnan and Raman (1970) show the rainfall due to the monsoon trough. The trough was rather diffuse earlier but became marked that day.

Ramakrishnan (1972) finds that if the tilt of the trough is large and it is at 15^o N at 500 mb, west coast gets moderate to heavy falls, but if the trough at that level is along 20^o N, west coast rain is scanty.

6.2.5 The eastern portion of the monsoon trough shifts southwards into north Bay before a depression forms. In the rear, the trough swings back northwards over northeast India. The monsoon trough having shifted to a northerly position, can cause increased rains over Assam; a fresh trough may form near the normal position. In September, the monsoon trough takes northnorthwest/southsoutheast orientation, when a depression is forming at a lower latitude.

Srinivasan and Ramakrishnan (1970) have tabulated the position of the trough line at 1 km (from 1961 to 1969) which is given in Table 6.3.

The near normal position is only on 30 to 47 per cent of the occasions. At 77^o E, it is more frequently to the north of the mean position, indicating the influence of middle latitude systems travelling to the north of India. Over northeast India the trough is more often to the south than to the north due to monsoon depressions. The above table also brings out that normal rainfall need not be a maximum near the mean position of the trough. The increase from July to August in the frequency of the position to the north in the western and central parts is interesting, as the breaks in monsoon are more in August. The monsoon trough is not so well defined at the beginning of the season, while towards the end, the western end becomes diffuse. In September, the eastern end shows wide fluctuations in position and orientation.

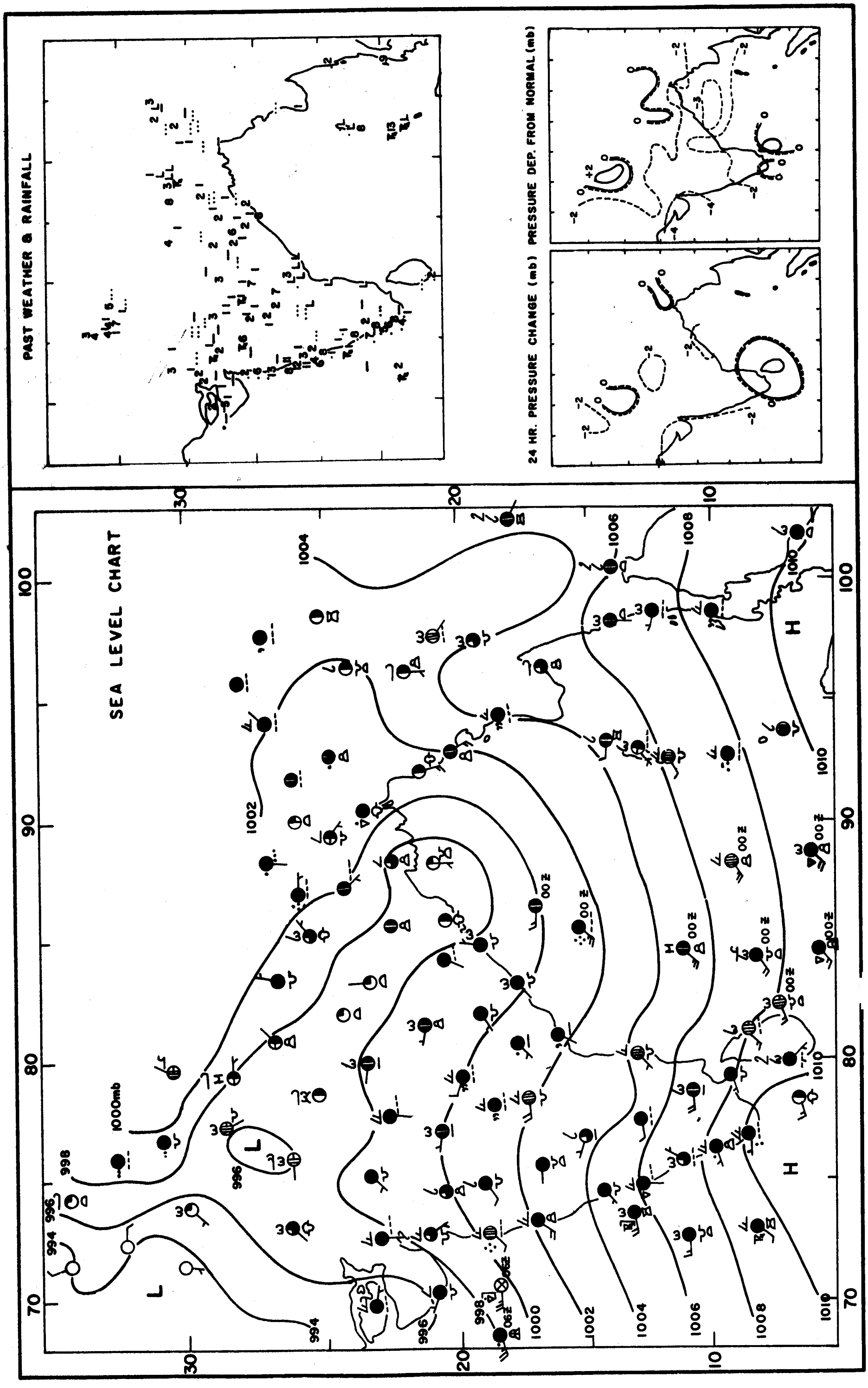


Fig. 6.3 (a) Synoptic charts, 0300 GMT : 7 July 1962

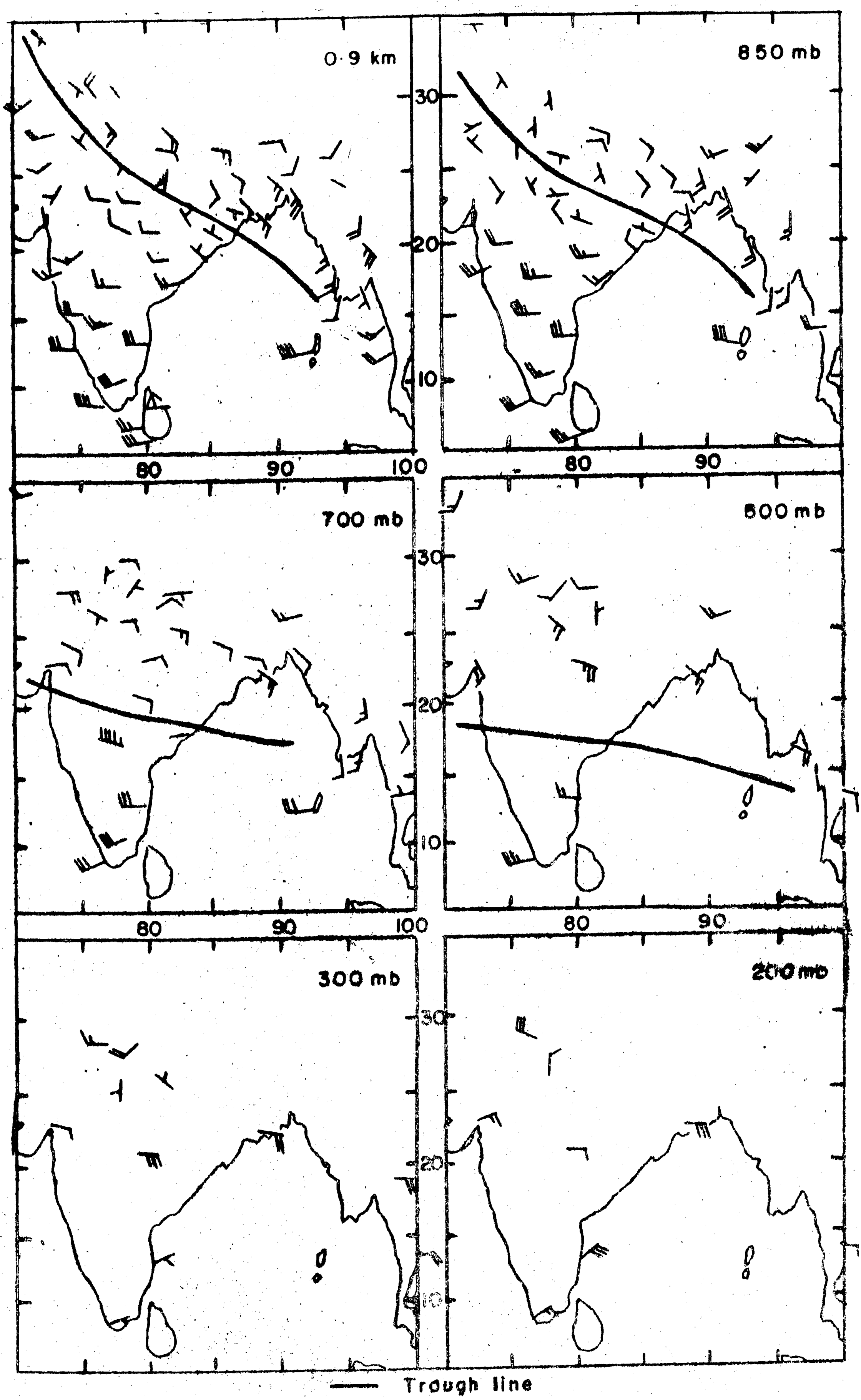


Fig. 6.3 (b) Upper Winds 7 Jul. 1962 00 GMT

Table 6.3

Position of Monsoon Trough

Month	Longitudinal section (^{O} E)	Normal position of trough (° N)	Percentage frequency of location of trough		
			South of normal position	Normal position	North of normal position
	77	27-29	27	30	43
July	81	25-27	27	45	28
	87	23-25	37	35	28
	77	27-29	20	33	47
August	81	25-27	20	47	33
	87	23-25	34	39	27

6.2.6 According to Banerji (1931) the orientation of the Assam-Burma hills and the Himalayas, makes the streamlines over northeast India take a hyperbolic shape, with the Burma hills as an asymtote. With the westerlies over the Peninsula, this explains the formation of the monsoon trough. That westerlies can sweep across northern India without forcing a trough is shown in break monsoon situations. Many a time, the position of the trough is so far south that it cannot be attributed to the influence of the mountain ranges bordering the region. Another interesting point is that when the trough is close to the northern mountains at sea-level, it may not be seen at 1.5 km, indicating that the usual south-ward slope is not present.

Ramage (l97l) explains the position of the monsoon trough and the less rains in its mean position as due to subsidence. Anabatic winds on the southern slopes of the Himalayas are part of the large scale summer circulation as made out by Flohn (1968). Though strongest in the afternoon, they prevail throughout the twenty-four hours in a day. They coincide with the northern rainfall maximum and presumably form a part of a local vertical circulation in which air returns southwestward in the middle troposphere and tends to sink over the northern plains. Subsidence warming diminishes surface pressure. The southward slope of the trough and the existence of a weak Hadley cell between the trough and the Himalayas with ascent in the trough may not support such a picture.

6.3 Easterly Jet Stream

6.3.1 Koteswaram (1958a and 1958b) studied in detail the easterly jet stream over India. South of the sub-tropical ridge over Asia, the easterly flow concentrates into jet stream centred near about the latitude of Madras at 100 mb in July. The jet stream runs from the east coast of Vietnam to the west coast of Africa. Over Africa, the location is at 10° N. Normally, the jet is at an accelerating stage from the south China sea to south India and decelerates thereafter. Consequent upper divergence is regarded as favourable for convection upstream of 70° E and subsidence downstream. Position and speed fluctuate from day to day. On the anticyclonic side of the easterly jet stream, constancy of absolute angular momentum is observed, quite similar to the anticyclonic side of the westerly jet stream. Over most of the Atlantic Ocean, continental America and Pacific Ocean, the easterly jet is not generally found. According to Troup (l96l) upper easterlies over Australia in southern summer are considerably weaker, besides being less extensive in latitude.

6.3.2 Mokashi (1970) studied the easterly jet over Trivandrum, Madras and Vishakhapatnam for the period 1961 to 1965. He finds that the percentage frequency of high speeds decreases from Trivandrum to Vishakhapatnam, higher speed occurring at greater height as seen from Table 6.4. This is the opposite of the sub-tropical westerly jet where stronger jet maxima are at a lower altitude.

Table 6.4

Percentage Frequency of Speeds in Easterly Jet and Height of Maximum

Station	Percentage of cases maximum wind (kt)			Mean height of max. (km) when max. wind (kt) is		
	< 80	= 80 to <100	=.100	< 80	= 80 to <100	=100
Vishakhapatnam	63.6	31.5	4.9	15.3	16.0	16.1
Madras	54.5	37.7	7.8	14.8	15.3	15.9
Trivandrum	42.2	45.5	12. 3	14.1	14.8	15.2

Mokashi (1974) made a fresh study of the easterly jet from a re-analysis of the rawin trajectories of Madras and Trivandrum (at vertical intervals of 0.5 km) for the period -1961-65. He compared the data of standard levels at Gan and Colombo for the same period. His results are summarised in Table 6.5. The core of the jet appears to be between Colombo and Trivandrum in June near 14.5 km, between Trivandrum and Madras in July at 14.5/15.0 km, near Trivandrum in August at 14.5 km and near Colombo in September at 14.2 km. The height change is rather small and the mean vector wind speed has varied between 62 and 69 knots only. This is by averaging the wind at each level over a month. On the other hand, it is interesting to average at each station the winds at the level of maximum speed. The mean speeds at the maximum wind level are naturally greater. This approach shows that highest speeds are at Trivandrum near 14.6 km. Speeds from June to August are near 85 kt but weaker in September, being 77 kt. Maximum northward position may be in August near Trivandrum and in other months perhaps between Trivandrum and Colombo.

Table 6.5

Statistical Parameters of upper winds for Gan, Colombo, Trivandrum and Madras

Station	**Mean zonal wind**		**Mean Scalar wind**		**Mean Vector wind**		**Average of max winds**	
	Speed (kt)	**Level (km)**	**Speed (kt)**	**Level (km)**	**Speed (kt)**	**Level (km)**	**Speed (kt)**	**Level (km)**
				JUNE				
Gan	51.1	13.5	55.5	13.5	54.0	13.5	78.7	14.55
Colombo	64.6	14.2	66.2	14.2	66.0	14.2	84.1	14.69
Trivandrum	67.4	14.5	68.8	14.5	67.6	14.5	84.6	14.71
Madras	57.9	15.0	59.2	15.0	57.9	15.0	80.0	15.04
				JULY				
Gan	50.5	13.5	55.5	13.5	55.0	13.5	80.4	14.40
Colombo	62.4	14.2	64.3	14.2	64.0	14.2	84.4	14.54
Trivandrum	64.3	14.5	65.9	14.5	65.1	14.5	84.7	14.61
Madras	65.1	15.0	65.8	15.0	65.1	15.0	81.7	15.05
				AUGUST				
Gan	52.0	13.5	56.0	13.5	56.0	13.5	79.0	14.56
Colombo	62.3	14.2	64.4	14.2	63.2	14.2	81.4	14.61
Trivandrum	68.6	14.5	70.0	14.5	69.O	14.5	85.0	14.63
Madras	63.3	15.0	64.7	15.0	63.3	15.0	81.3	15.01
				SEPTEMBER				
Gan	48.6	13.5	51.6	13.5	49.5	13.5	74.8	14.49
Colombo	61.1	14.2	62.3	14.2	62.0	14.2	75.4	14.55
Trivandrum	54.3	14.0	55.9	14.0	54.4 54.3	14.5 14.0	77.0	14.58
Madras	49.9	15.0	51.5	15.0	49.9	15.0	72.9	15.16

In the absence of detailed analysis of rawin trajectories of Gan and Colombo, like those of Trivandrum and Madras, the study may have underestimated the winds of the two southern stations. The conclusion of Mokashi that Colombo-Trivandrum is the core of the easterly jet seems to be fully justified (See also Ramage and Raman, 1972).

Mokashi (l974)also compared the upper tropospheric winds at Minicoy, Trivandrum, Madras, Goa, Hyderabad, Bombay, Nagpur and Ahmedabad on each day in July and August, 1971. The maximum wind was at Minicoy on 60 per cent of occasions and north of Madras in 23 per cent of cases. The latter was at a higher altitude. He concludes that the idea of two cores of maximum winds existing simultaneously in the rather diffuse zone of easterly jet is not unreasonable.

The mean meridional wind components at the level of the highest mean winds are given in Table 6.6. They increase from Madras toward the equator.

Table 6.6
Mean meridional wind components (kt) (northerly +ve)

Station	June	July	Aug.	Sep.
Madras	-0.8	0.0	0.8	-1.4
Trivandrum	2.5	8.0	4.7	-1.2
Colombo	11.8	10.9	9.8	2.9
Gan	16.5	17.7	17.5	12.5

The standard deviations of the meridional (σy) and zonal (σx) winds at the level of the maximum mean vector wind are shown in Table 6.7.

Table 6.7
Standard deviations (kt) of meridional and zonal winds, at the level of maximum mean vector wind

Station	June		July		August		September	
	s x	**s y**	**s x**	**s y**	**s x**	**s y**	**s x**	**s y**
Madras	15.7	11.8	16.9	10.9	16.7	12.2	17.5	11.7
Trivandrum	20.6	13.4	15.0	12.2	20.2	13.2	19.8	12.4
Colombo	13.3	7.7	13.7	11.1	16.4	13.4	15.1	11.6
Gan	18.3	12.4	14.1	14.9	15.3	12.8	16.3	12.1

s x - Zonal **s y** - Meridional

s x / **s y** varies between 1 and 2, unlike 0.6 for the sub-tropical westerly jet stream over Delhi (Mokashi 1969). Perhaps troughs and ridges are not present in the easterly jet stream.

The vertical profile of the jet stream at Madras and Trivandrum composited around the level of the highest wind is shown in Fig. 6.4. While it is symmetric at Trivandrum, wind shear is more at Madras below the jet than above.

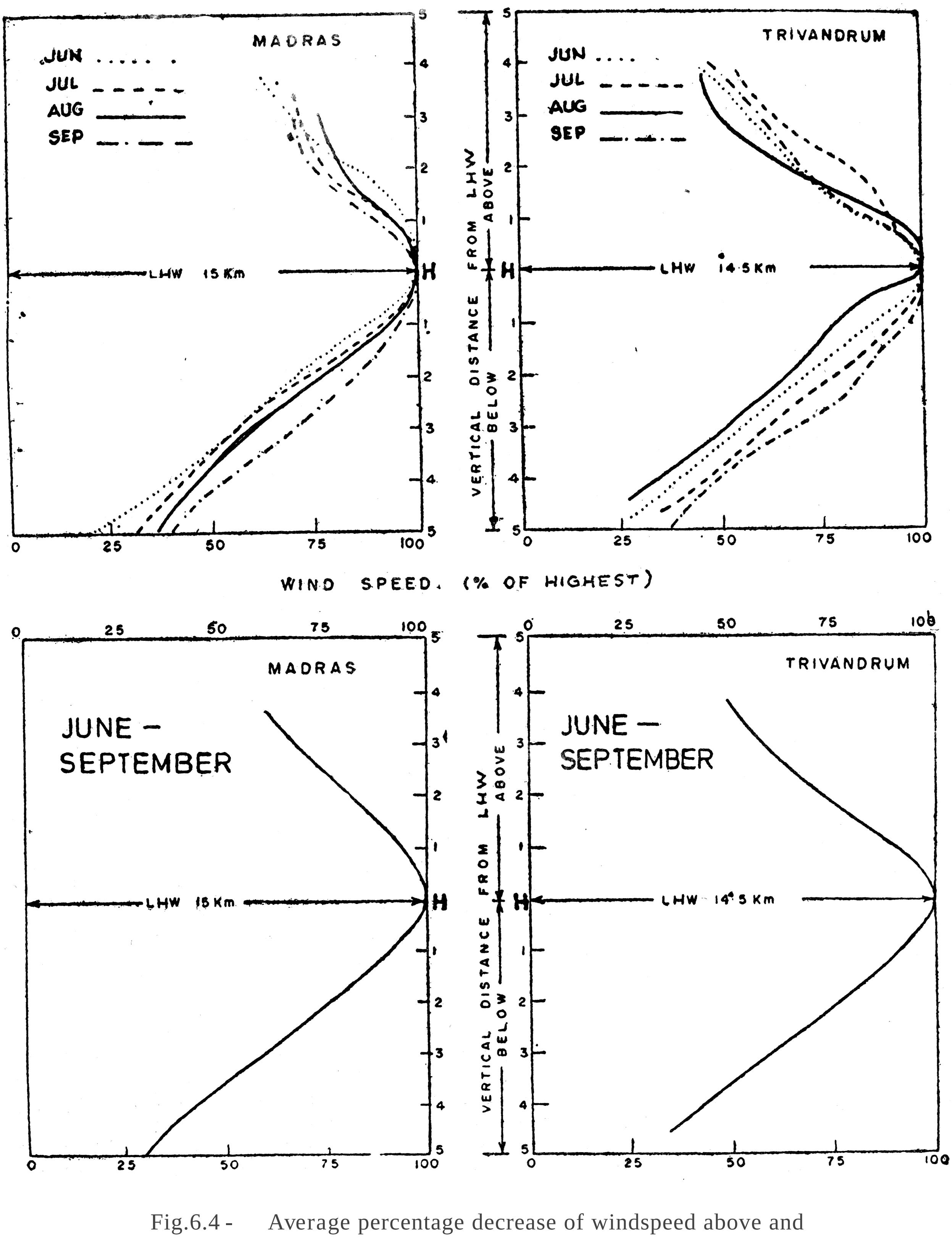


Fig.6.4 - Average percentage decrease of windspeed above and below the level of highest mean (vector) wind (LHW).

These shears may be more representative than values worked out from mean winds at standard levels.

6.3.3 Thiruvengadathan and Ananthakrishnan (1965) report some finer details of the easterly jet over Trivandrum and Minicoy from the data of 1963. The characteristics above the level of the maximum seem to fall into three types -(a) maximum near 14 km and minimum speed at 16 km, (b) maximum between 14 and 16 km and minimum at 18 km, and (c) no minimum between the level of maximum speed and 21 km. These are illustrated in Fig. 6.5.

6.3.4 From the rawin data of 1961 to 1966 (from June to September). Thiruvengadathan (1972) studied the features of the easterly jet stream for Colombo, Trivandrum, Madras, Vishakhapatnam and Nagpur. On most occasions, maximum had been reached at 150 mb at Colombo and Trivandrum but winds were still increasing upto 100 mb at Madras, Vishakhapatnam and Nagpur. Median values of vertical wind shear are given in Table 6.8.

Table 6.8
Vertical shears of zonal winds (mps/km)
(Median Values)

Station	**June**			**July**			**August**			**September**		
	300-200 mb	**200-150 mb**	**150-100 mb**	**300-200 mb**	**200-150 mb**	**150-100 mb**	**300-200 mb**	**200-150 mb**	**150-100 mb**	**300-200 mb**	**200-150 mb**	**150-100 mb**
Colombo	5.1	4.4	-5.9	4.6	3.7	-4.2	5.2	4.7	-6.1	4.3	4.7	-7.9
Trivandrum	5.2	4.8	-5.2	4.9	4.3	-1.6	5.2	4.3	-3.0	4.5	3.6	-4.3
Madras	4.5	4.6	2.4	4.3	5.6	0.2	4.6	3.5	1.1	3.8	3.1	-0.6
Vishakhapatnam	2.5	3.2	1.5	3.3	3.8	1.7	3.1	3.3	1.7	1.8	2.0	l..5
Nagpur	1.8	2.4	3.0	2.7	3.1	2.1	2.7	2.6	2.6	1.3	1.3	2.7

Positive : Easterlies increase with height.

Horizontal wind shears of zonal winds between neighbouring stations are shown in Fig. 6.6. At 150 mb, cyclonic and anticyclonic shears are equally frequent between Colombo and Trivandrum, cyclonic being slightly more frequent in July. In other sectors anticyclonic shear is generally prominent. At 100 mb, cyclonic shear predominates between Colombo and Madras and anticyclonic shear to north. But in all sectors and in both levels (150 mb and 100 mb), cyclonic and anticyclonic shears occur from time to time. The maximum winds at the four stations were compared. The level of maximum wind slopes upwards gradually from Colombo to Nagpur. The core of the jet lies near Trivandrum in June and September and between Trivandrum and Madras in July and August.

6.3.5 Srinivasan (1960) found upper tropospheric easterlies (>50 kt) over Calcutta to be of two types. While sometimes strong winds persist for a week or ten days, at other times wind maxima transit quickly, being noticed only for a day or two. The former type is associated with the monsoon trough lying north of the normal position, towards the Himalayas and causes large defect in rainfall. Fig. 6.7 illustrates such a case. The number of jet maxima passing over the Gangetic West Bengal is small compared to the number of troughs in upper easterlies. Sometimes both of them moved simultaneously. Temperatures seem to be lower below wind maxima also.

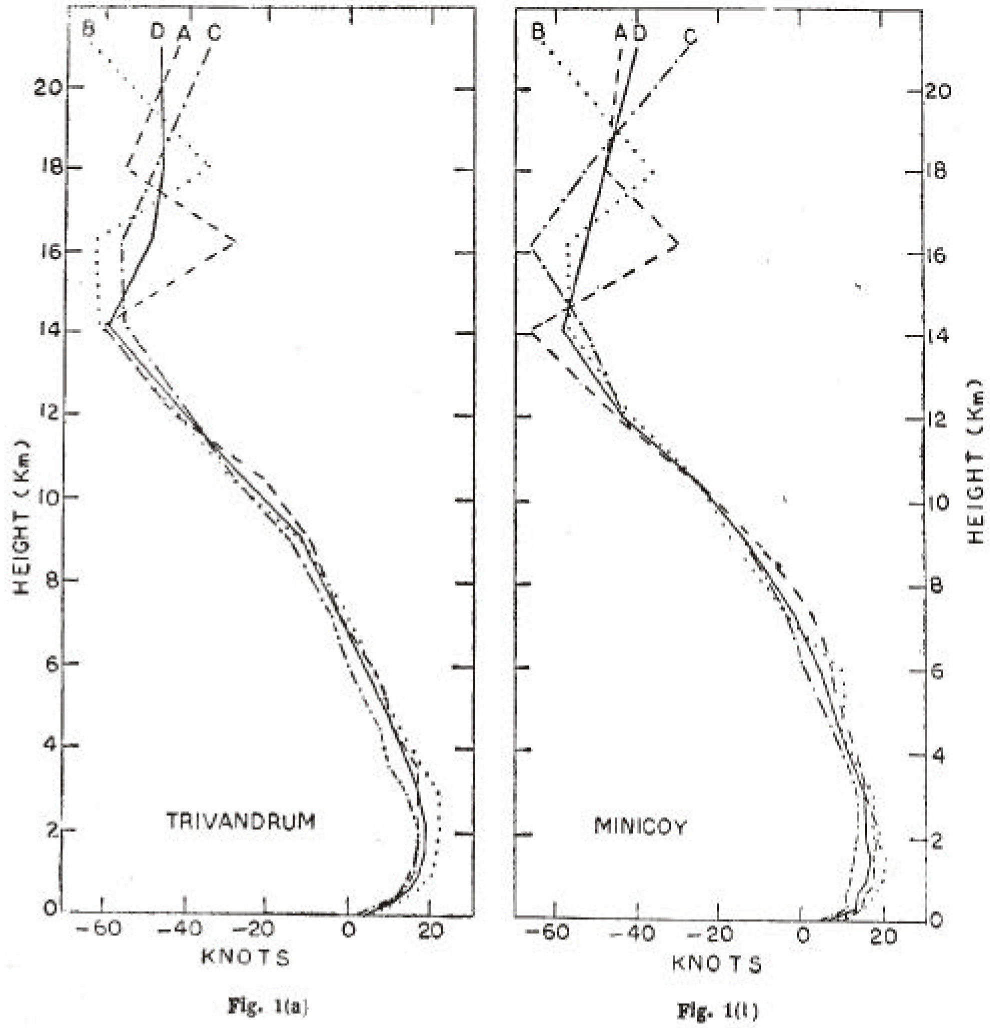
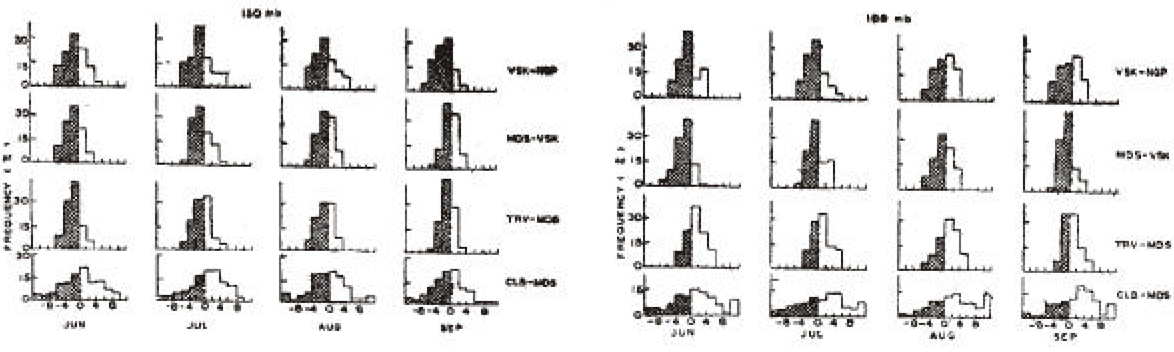


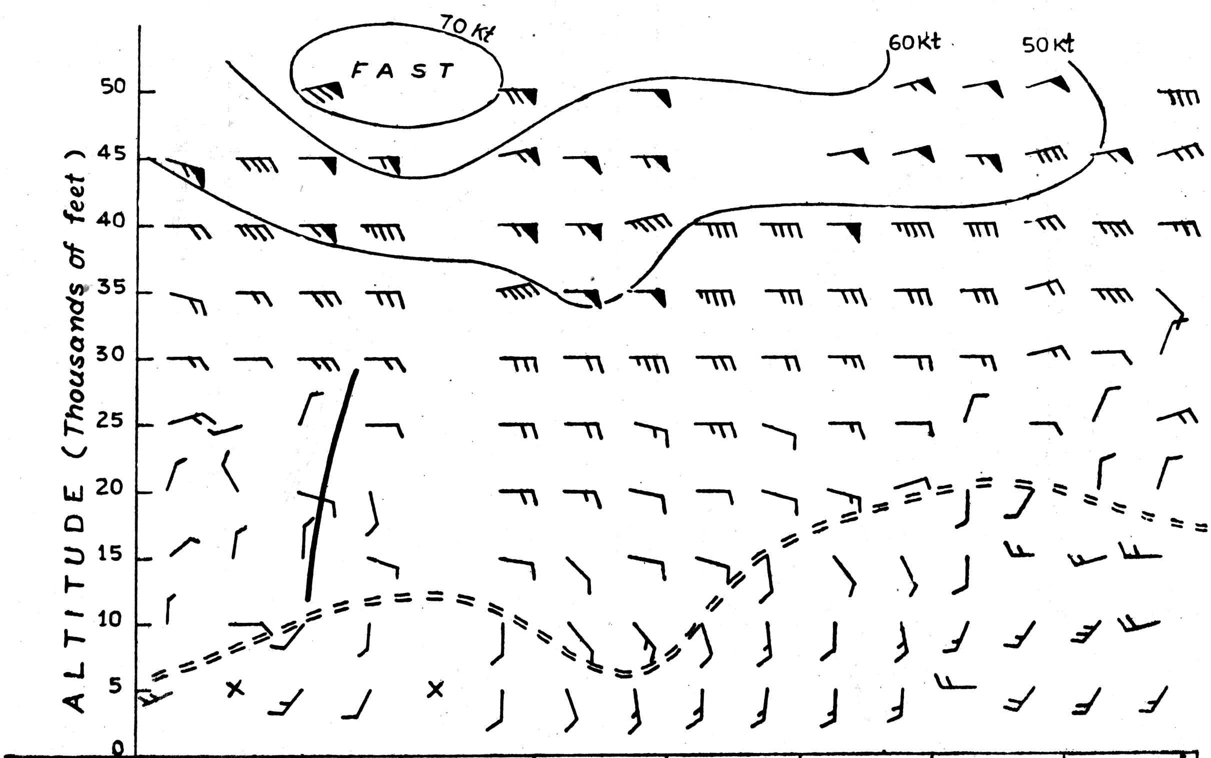
Fig.6.5 - Mean zonal winds (For A, B, & C see text; D represents mean zonal wind for all the categories A, B, & C).



(a) Horizontal shears of zonal winds at 150 mb.
(Unit 10^{-5} sec $^{-1}$; Anticyclonic shears shaded).

(b) Horizontal shears of zonal winds at 100 mb.
(10^{-5} sec $^{-1}$; Anticyclonic shears shaded).

Fig.6.6



		29-7-1955	30-7-1955	31-7-1955	1-8-1955	2-8-1955	3-8-1955	4-8-1955	5-8-1955
Rainfall over Calcutta	(Inches)	0·27	0·01	0·01	0·06	0·25	0·06	0·54	0·36
Average Rainfall in Gangetic W. Bengal	(Inches)	0·1	0·1	0·1	0	0·1	0·1	0·3	0·3

Thin continuous lines - Isotachs; Thick line - Trough line. The normal daily rainfall in Gangetic West Bengal during this period is approximately 0.4" to 0.5".

Fig.6.7 - Vertical time section for Calcutta, 29 July - 5 August 1955.

6.4 Tibetan High

6.4.1 In July, at about 700 mb and aloft, a ridge lies over Pakistan and northwest India to the west of about 75° E, with its axis along 30° N. Another high appears to the east of 80° E at 500 mb with axis near about 28° N. According to Ramage and Raman (1972) and Chin and Lai (1974), this high has its centre at 28° N, 98° E, distinct from the Pacific High at 140o E. At this level, the high covers the Tibetan Plateau while the centre of the high is at its eastern periphery. More marked at 300 mb, its extent is between 70° E and 110° E, with centre near 30° N, 90° E, while the Pacific High has weakened very much and is centred near 120° E. At 200 mb (Fig. 6.8). the only centre is at 30° N, 88° E and the high extends from 78° E to at least 140° E. There is only a broad high pressure belt at 100 mb from even 30° E to 150° E along 35° N over Indian region. Interestingly, there is also a ridge at 700 mb with its axis along 40° N, between 75° E and 95° E, just north of the Tibetan massif. The high over Tibet from 500 mb upward, centred near Tibet at 500 mb and over Tibet at 300 mb and 200 mb, has come to be known as the 'Tibetan High'. Flohn (l950) first described the formation of this permanent warm anticyclone. In June the axis of the anticyclonic belt is at about 25° N, at 300 and 200 mb, and near 30° N only at 100 mb, at the southern periphery of Tibet. August is similar to July, but the anticyclone is a little more to the north and slightly more intense. In September, the anticyclonic belt is near about 26° N upto 200 mb and 30° N at 100 mb. Thus in all these months the 100 mb position is at 30° N or further poleward but extending well outside the limits of the Tibetan Plateau. Between 500 mb and 200 mb, the high pressure belt is well to the south of Tibet in June and September but over Tibet in the other two months in between. The anticyclonic shear north of Tibetan High is larger than the shear on the equatorward side of the high (Krishnamurthy, 1971). Whether the heating of the plateau by the incoming solar radiation is responsible for the development of anticyclonic cell over Tibet between 500 and 200 mb has not been clearly established. For development of a 'high' colder air should occur at some higher level rather than warmer air below. The core of the easterly jet is at least 15 south of the centre of the Tibetan High and easterlies are weaker nearer Tibet. Hence the role of this system in causing higher speeds in the easterly jet is a debatable point. (See also Raman and Ramanathan, 1964). However, the heating over Tibet at mid–tropospheric level, accentuates the seasonal north–south temperature gradient. The decrease in heights of 500 and 200 mb surfaces from Tibet is more towards east in January and to the west in July.

Radiation balance over the Himalayan ranges and Tibet should be important for the dynamics of the Tibetan High. Bishop, Angstrom, Drummond and Roche (1966) reported some measurements of total incoming solar and sky radiation and reflected component, by the Himalayan Scientific and Mountaineering Expedition of 1960–61 and the American Mount Everest Expedition of 1963, for April and May, before the onset of the monsoon. These are the only observations that provide any guidance. 1961 measurements were at an altitude of 5,720 m on the 'Silver Hut' glacier on the flank of Everest. Weather was predominantly fine and clear, lowest temperature –27° C and average wind speed 40 kmph. In this area, the precipitation from December 1960 to May 1961 was 13 cm, while the average precipitation in the rest of the year is estimated as 33 cm. The 1963 observations were on the Khumbu glacier at an elevation of 5,425 m.

The atmospheric turbidity over the Himalayas as estimated from these radiation measurements is rather high compared to isolated mountains like Mount Whitney (California, USA) and Mauna Loa (Hawaii, USA). But this high turbidity over the Himalayas is about the same as the average reported for Poona (550 m, 18.5° N, 74 E) by Mani and Chacko (1963) which is usually low for its latitude. Bishop et al (1966) speculate whether the high ridges of the Himalayas also form a kind of barrier protecting the atmosphere above the Indian plains from being injected by more dust polluted air from the desert regions of the north.

Average values established for the total short wave albedo during these Everest expeditions are very much lower than those generally measured elsewhere . The Everest values are characteristic of snow–ice surface in melt zones during ablation periods.

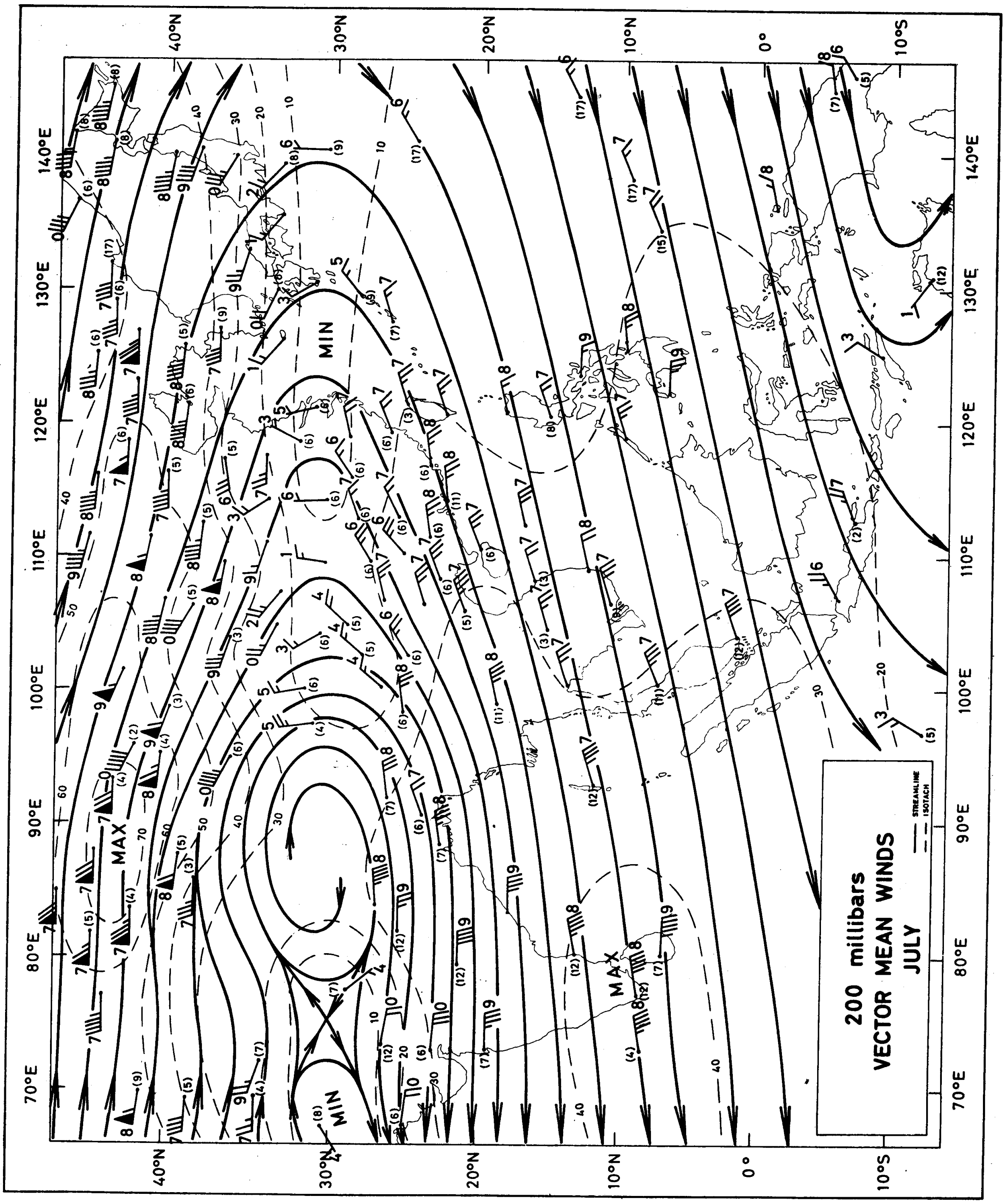


Fig. 6.8 Vector mean winds July 200 mb. (Chin and Lai, 1974)

The very low water vapour content of the atmosphere over Mount Everest area leads to the conclusion that the noctural radiation out to space from the snow surfaces must be exceptionally high, about 0.22 – 0.26 cal cm^{-2} min. This is equivalent to a cooling of 20° C in 12 hours in a 500 m thick layer at that altitude. This, along with the high intensity of sun radiation during day time, causes large diurnal range (20 – 25° C) of temperature as reported by the expedition. How far these conditions are modified during the monsoon has to be studied.

6.4.2 Koteswaram (1958b) pointed out that the ridge of the upper troposphere over Asia is at a more northerly position than over Africa. The anticyclone over Tibet is only from 500 mb upwards, while over Africa, Arabia and further east, it is even from below 700 mb.

6.4.3 An interesting aspect of the position of the sub–tropical high in mid–and upper troposphere over India from June to September is that it is about 10 more poleward than in the southern hemisphere during its summer. These positions are given in the table below :

Table 6.9

Positions (Lat.) of the sub–tropical ridge in Indian longitudes

500 mb	India	Jun. 22	Jul. 28	Aug. 29	Sep. 27
	South Indian Ocean	Dec. 18	Jan. 18	Feb. 23	Mar. 17
300 mb	India	Jun. 23	Jul. 31	Aug. 28	Sep. 25
	South Indian Ocean	Dec. 10	Jan. 15	Feb. 18	Mar. 15

This peculiar northward displacement of the sub–tropical high characteristic of this area must be having an important influence on the monsoon which requires further study.

6.4.4 Ramaswamy (1965) points out that well–distributed rainfall over India is associated with well–pronounced and east–to–west oriented anticyclone over Tibet at 500 and 300 mb levels, and a pronounced high index circulation over Siberia, Mongolia and north China. The Tibetan 'High' may sometimes shift much to the west of its usual position. In such a situation, the monsoon may extend further westward into Pakistan and in extreme cases into north Iran, though such a westward position of the Tibetan 'High' would be against its having origin in the heating effect of the Tibetan Plateau.

6.4.5 At the time of the burst of monsoon over Kerala, Tibetan 'High' is not established. The position of the sub–tropical high is to the south and even the mid–latitude westerlies may be prevailing in northern India in upper troposphere. Ramamurthi and Keshavamurthy (1964) found, in 1963 the Tibetan 'High' was established much later, after the onset of the monsoon.

6.4.6 Krishnamurthy (l97l) regards planetary scale ascent and divergence of air in the ridges can account for generation of negative relative vorticity and maintenance of anticyclonic systems.

6.4.7 Thiruvengadathan (1972) concludes that between prolonged periods of strong and weak monsoon over the Konkan, there is no difference in the position of sub–tropical ridge. Ramamurthi, Keshavamurthy and Jambunathan (1965) find that during weak monsoon the thickness values between 300–200 mb over central parts of the country are relatively higher and lapse rate lower compared to strong monsoon. This may be due to large scale subsidence over the central parts associated with sub–tropical anti–cyclone.

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CHAPTER 7

MONSOON DEPRESSION

7.1 Monsoon depressions are the synoptic features that cause most of the monsoon rains. These are low pressure areas with two or three closed isobars (at 2 mb intervals) covering an area of about five degrees square, which form in the Bay of Bengal north of 18^{o} N, move westnorthwest at least upto the central parts of the country before weakening or filling up, and give widespread rains in the southwest quadrant with many heavy falls.

7.2 These low pressure systems are referred to as depressions when surface winds are upto 33 kt (while over the sea) and cyclonic storms when higher speeds prevail. Weaker systems with only one closed isobar and wind speeds less than 17 kt, are called lows. While they are over the northern parts of the Bay of Bengal, assessment of their intensity is rather uncertain, being based on the availability of ships' observations and the system extending upto the coast. Table 7.1 shows the number of depressions and cyclonic storms that formed from June to September in the Bay of Bengal, Arabian Sea and over land in a period of eighty years from 1891 to 1970.

Table 7.1

Number of cyclonic disturbances (1891–1970)

	June		**July**		**August**		**September**	
	D	**S**	**D**	**S**	**D**	**S**	**D**	**S**
Bay of Bengal	71	35	107	38	132	26	141	32
Arabian Sea	18	15	9	3	2	2	9	5
Land Area	12	l	39	1	42	0	21	1

D – Depressions S – Cyclonic Storms

(from "Tracks of Storms and Depressions in the Bay of Bengal and the Arabian Sea" – India Met. Dept. – 1970 – under publication)

The above statistics is according to the region of formation; disturbances that formed over the Bay and later entered the Arabian Sea have been reckoned separately. The systems in early June, usually associated with the advance of monsoon, are not typical of monsoon depressions, particularly in the Arabian Sea. So also some of the disturbances of September. The frequency of cyclonic disturbances in August is about the same as other months, in spite of the incidence of large number of 'breaks'. Compared to the average of two to three disturbances per month, five or six may occur in a few years, even two systems being present simultaneously. Between 1891 and 1970 there were six years without a depression or a cyclonic storm in July and four years in August. The number of cyclonic disturbances (depressions and cyclonic storms) in a year ranges from 4 to 14 as seen in Table 7.2.

7.3 Bhalme (1972) finds a significant decrease in the depressions and cyclones in the southwest monsoon season from the beginning of the present century. Oscillations of about 40 years and a quasi–biennial oscillation of 2.4 years are also noticed from spectral analysis. Raghavendra (1973) reports a long term period of 30 to 45 years and weak indication of a 5–year period. However, Rao and Jayaraman (1958) find that there have been no changes in recent years and there is no periodicity.

Table 7.2

Number of cyclonic disturbances in different years in the Monsoon season

Year	No.	Year	No.	Year	No.	Year	No.
1891	10	1911	11	1931	7	1951	10
1892	9	1912	7	1932	8	1952	7
1893	11	1913	10	1933	12	1953	7
1894	12	1914	9	1934	10	1954	9
1895	12	1915	7	1935	8	1955	6
1896	11	1916	6	1936	11	1956	8
1897	13	1917	12	1937	11	1957	4
1898	12	1918	7	1938	4	1958	5
1899	15	1919	10	1939	12	1959	11
1900	10	1920	8	1940	9	1960	6
1901	11	1921	8	1941	12	1961	11
1902	11	1922	12	1942	9	1962	6
1903	10	1923	9	1943	10	1963	6
1904	14	1924	10	1944	12	1964	9
1905	11	1925	10	1945	8	1965	6
1906	10	1926	10	1946	12	1966	8
1907	10	1927	14	1947	9	1967	7
1908	14	1928	10	1948	9	1968	8
1909	8	1929	13	1949	7	1969	8
1910	8	1930	9	1950	11	1970	7

(from "Tracks of Storms and Depressions in the Bay of Bengal and the Arabian Sea" – India Met. Dept, 1970 – under publication).

7.4 In June, July and August, depressions and storms generally form in the Bay north of 18° N and west of 92^{o} E, and a few upto 15° N. Most of the formation is to the north of 20° N. In September, however, the formation extends upto 14^{o} N, In the Arabian Sea, the systems originate in June within five degrees of the coast north of 12^{o} N and in rare cases very close to Gujarat coast in later months. Land depressions develop mostly over northeast India. Weaker lows may appear in all these areas and over land north of 20° N; in stray cases the low over land may intensify into a depression. Ananthakrishnan and Bhatia (1958) pointed out that the place of formation of monsoon depressions is over the sea area nearest to the southern periphery of the upper tropospheric anticyclonic cell in the north Bay, the same position relative to the anticyclonic cell as in pre–and post monsoon months.

7.5 Depressions and storms of July move mainly west to westnorthwest over the Bay and across the country upto 25° N, and westnorthwards in August. In higher latitudes , the movement becomes more northerly. In June, the movement is more spread out; besides, the other differences are, change to a northerly course at lower latitudes (even 20^{o} N), recurvature and north to northeast course while over the Bay. The last is a residual of the May characteristics. Depressions early in June, with advancing monsoon, may take a northnortheast track and usher the monsoon into West Bengal and Assam and adjacent states. Even in July and August, some initial northerly track may occasionally be seen before systems change to the usual westnorthwest movement. September is more or less like June. While the tracks of Bay depressions in July and August are within a narrow belt, they are very much spread out in June and September.

A few westward moving systems may emerge into the Arabian Sea and keep on to that course. The rather more frequent depressions and storms that form in the Arabian Sea in June move between northwest and north. The rare depressions in other months that form off Saurashtra move northwards.

In July, the average speed of the monsoon depressions is 5–10 km hr^{-1} to the east of Long. 85^o E but more (10–20 km hr^{-1}) to the west. Movement may be very slow at the formative stage over the Bay; and depressions speed up as they come over land. It is faster over the central parts of the country where system have travelled to 30 kmph. West of 80^o E, speeds are rather less to the north of 25 ^{o} latitude, perhaps due to change in course. These general characteristics hold good in other months also, except that speeds are slower. The Arabian Sea systems of June have speeds of 5–10 kmph, slow nearer the coast but faster as they move westward.

Tracks of depressions and cyclones from 1891 to 1960 shown in figures 7.1 to 7.4 illustrate the above features.

7.6 The proportion of cyclonic storms in the Bay as far as the monsoon period is concerned, is greatest in June, decreases in July and is still less in August and September. But all of them weaken on crossing coast and no storm survives 200 km beyond, so that all systems moving across the country are only depressions. 70 percent of systems cross coast between 20^o and 22^o N. They also weaken and dissipate as they travel westward. About one third of depressions and storms fill up before crossing 85° E and about half before 80° E. Chance of these systems reach ing beyond 75^o E is a little less than 10 percent in June and July but more in August (18 percent) and September. Similarly, the probability of depressions reaching north of 25^o N progressively increases in August (40 percent) and September (50 percent). The probability of depressions becoming deep is more in September 50 percent) than in July and August (30 percent).

7.7 According to Srinivasan, Raman and Mukherji (1971) the life periods of Bay depressions and storms are as given in Table 7.3.

Table 7.3
Life period of Bay depressions and storms (1901–1960)

Life Period	**Percentage of depressions and storms**		
No. of days	**July**	**August**	**September**
1	9	7	4
2	11	17	11
3	37	17	18
4	16	21	20
5	13	17	18
6	7	10	10
7	5	6	7
8	2	1	8
9	0	4	4

When depressions weaken, the remnant lows persist for a day or two before filling up. Number of days of such lows are more in August and September than in July. In August, they have a preferred location over north Madhya Pradesh and central Uttar Pradesh, but not in July and September.

7.8 The average number of closed isobars at sea level in these depressions and cyclonic storms is three at interval of two millibars. The latter may have even six isobars while depressions at the stage of filling up may have only one. The shape of the isobars is often roughly elliptical rather than circular with elongation in westnorthwest direction. The average extent of closed iso-bars along the meridian through the centre is about 4.5^o of latitude, varying between 2^o and 10^o.

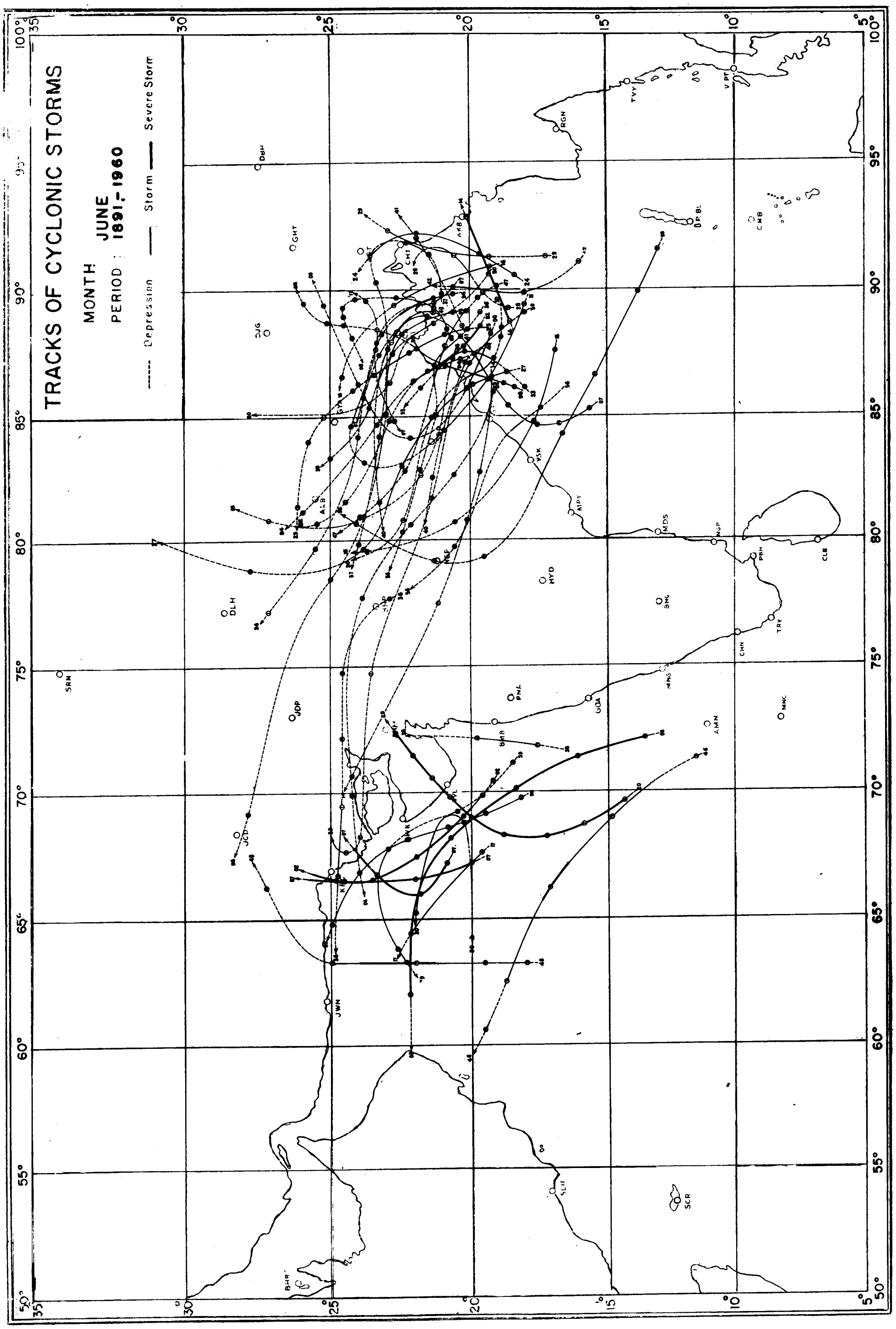

TRACKS OF CYCLONIC STORMS
MONTH JUNE
PERIOD : 1891-1960
Depression
Storm
Severe Storm

Fig. 7.1

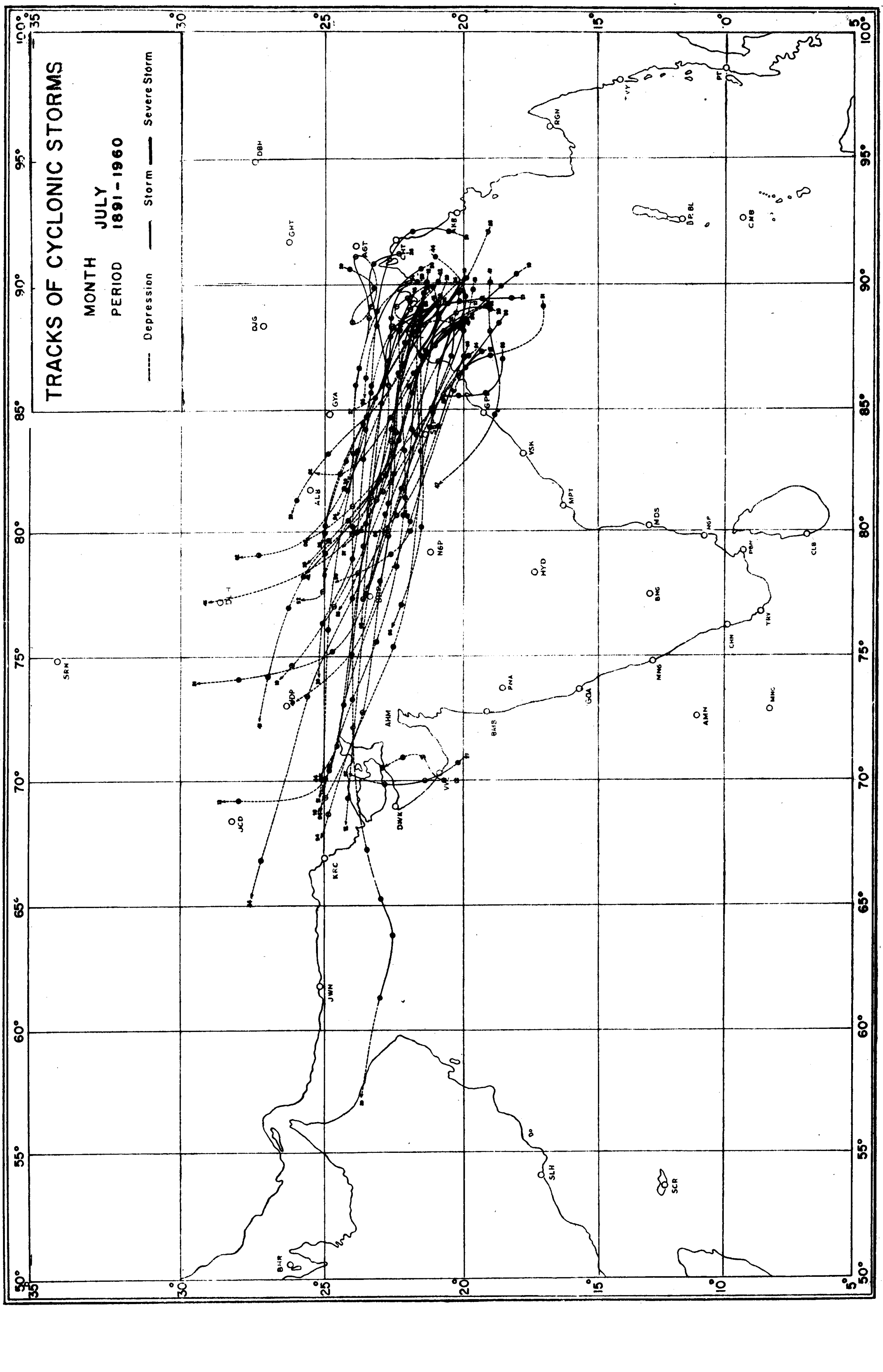

TRACKS OF CYCLONIC STORMS
MONTH JULY
PERIOD 1891-1960
Depression
Storm
Severe Storm

Fig. 7.2

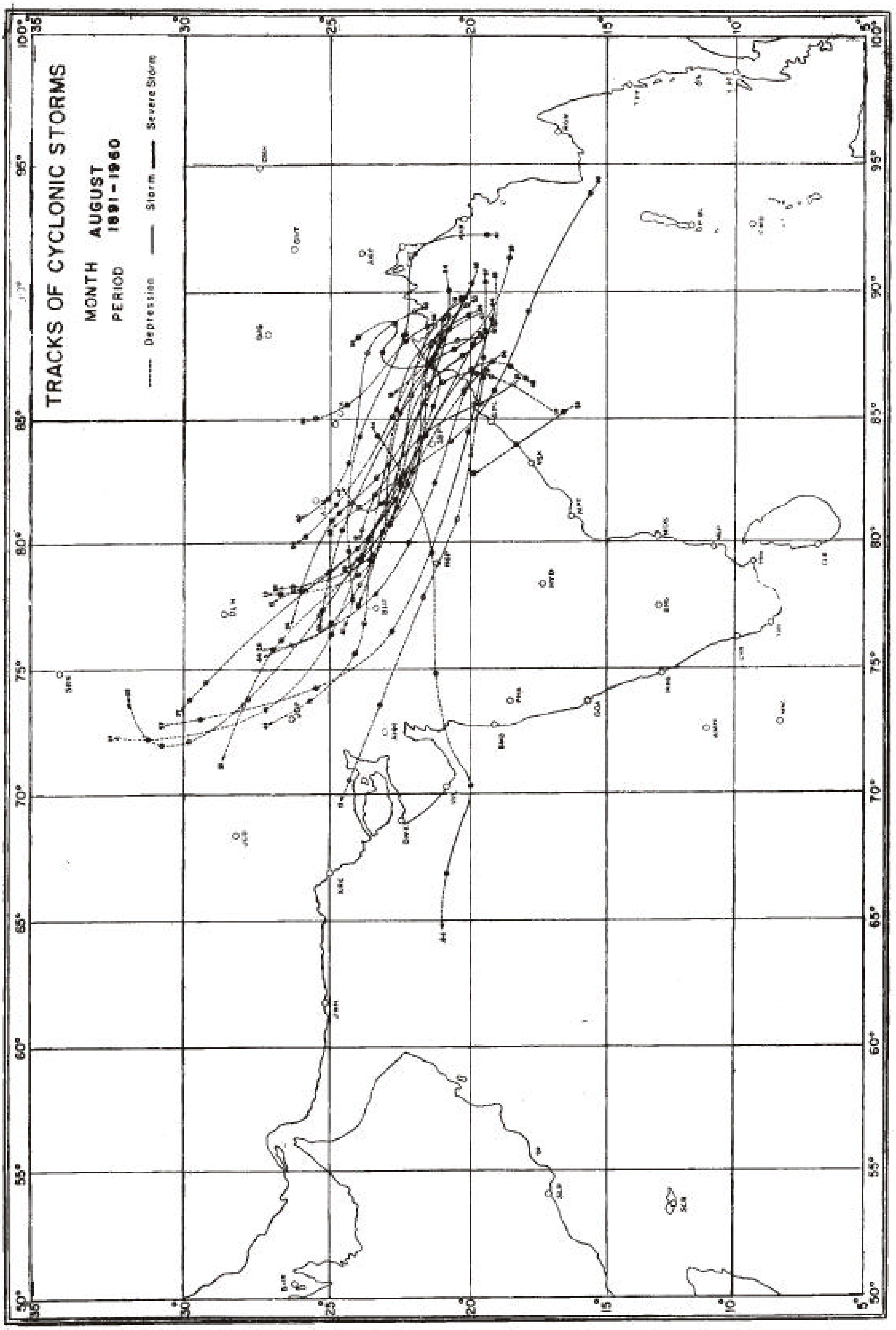

TRACKS OF CYCLONIC STORMS
MONTH AUGUST
PERIOD 1891-1960
Depression
Storm
Severe Storm

Fig. 7.3

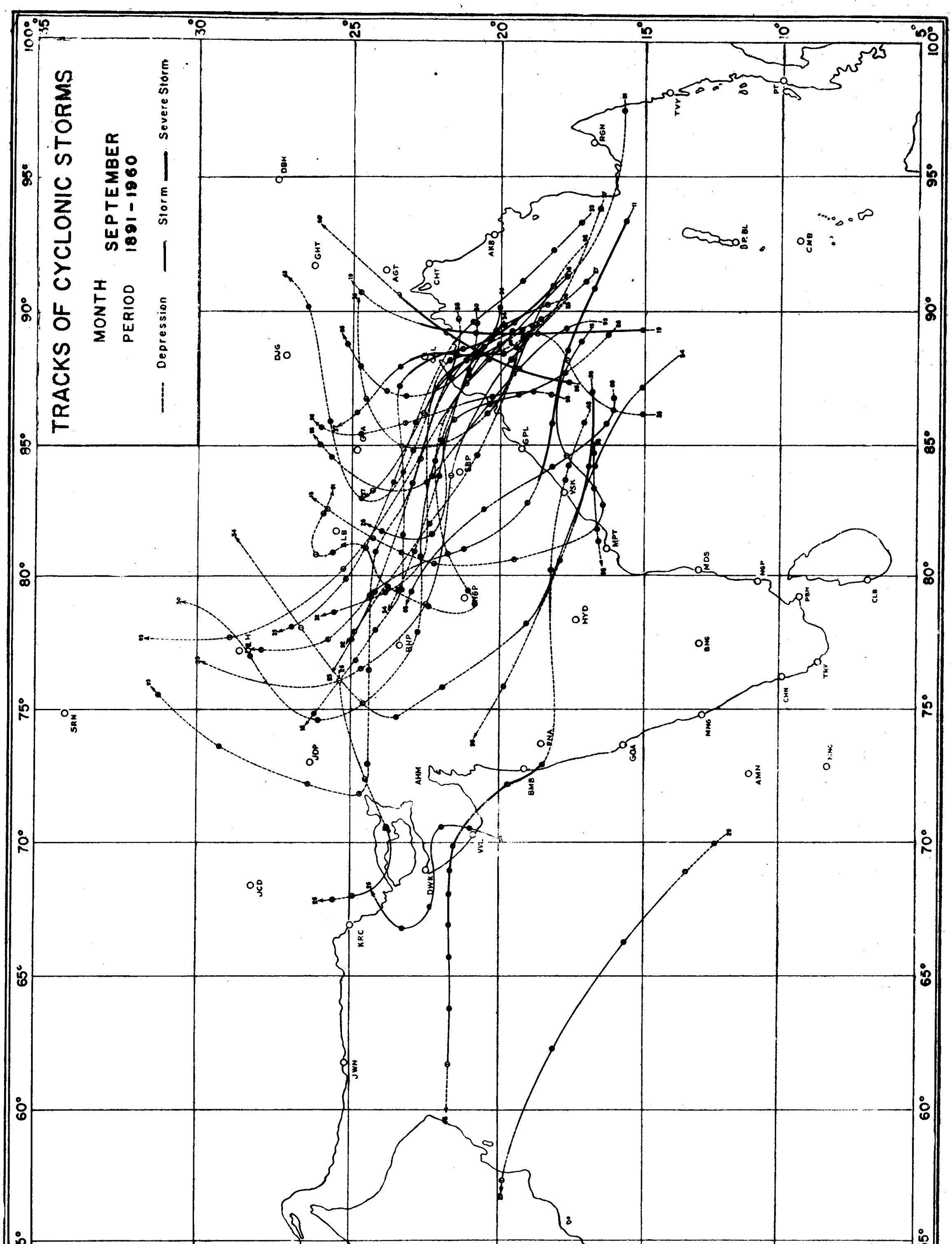

TRACKS OF CYCLONIC STORMS
MONTH SEPTEMBER
PERIOD 1891 - 1960
Depression
Storm
Severe Storm

Fig. 7.4

In the direction of the latitude circles it is 5.5° of longitude, limits being 2° and 12° . In terms of latitude and longitude, systems are elongated by about 1.2 times along the latitude circles, perhaps a little more so along the westnorthwest axis. In about 20 percent of cases, they are nearly circular and 10 percent have elongation along the meridians. The ratio of the extent along longitude to latitude of the outermost closed isobar has varied between 0.7 and 1.7 over a period of two monsoons. Even allowing for the shorter distance per degree along the latitude circles than the meridians most of the systems are elongated in terms of actual distance in westnorthwest direction. This implies that pressure gradient to south of the centre is more than to the west. Some of the depressions of the later period, particularly September, are of very small extent, though causing intense rains.

7.9 Radiosonde/radiowind stations are too far apart to provide sufficient details about the structure of the monsoon depressions at upper levels. Upper air circulation seems to extend usually upto 500 mb and occasionally even 300 mb. Lowest temperatures are probably in the southwest quadrant upto 700 mb as fresh monsoon air flows in and is subjected to lifting in this zone of heavy rains. Highest temperatures are likely in the northwest sector, some distance from the centre, at least in the depressions of larger extent, as the warmer air mass from the northwest is drawn into the field. Due to the lowest temperatures to the southwest, depression centre slopes towards that direction with height.

Srinivasan, Raman and Mukherji (1972) have presented a cross–section (Fig. 7.5), showing the centre of the depressions shifting southwest with height. The slope is one in forty. Rai Sircar (1956) studied the variations of heights of pressure surfaces and isotherms at Calcutta as a cyclonic storm and two depressions passed to the south. A shallow depression formed 100 km southwest of Calcutta on 1st August 1953 and became a cyclonic storm on 2nd, when it was about 120 km south–southwest of Calcutta. Later it gradually weakened. Its track, the heights of pressure surfaces and the temperature field over Calcutta are shown in Fig, 7.6 (a, b and c) . Figures (b and c) could be regarded as a cross–section through the storm, 80 km north of the centre. Two days before the depression formed, a low in upper troposphere was probably moving westwards across the same area. At the time of the lowest pressure in lower troposphere, a high was present in the upper troposphere. While temperature near the surface had undergone little variation, there was a rise at the upper levels as the storm centre approached, the gradient being marked above 500 mb. Two other systems studied by Rai Sircar are in conformity with this pattern, Whether the preceding upper tropospheric low is a wave in easterly as per the model of Koteswaram and George (1958) is not known. It has to be investigated whether the upper tropospheric high above the lowest pressure at surface is caused by the vertical ascent of air or whether it is independent of the depression.

Srinivasan (1953) found that in the case of a deep depression in September 1951 the centre of which passed within 55 km northnortheast of Allahabad, the contour gradient increased aloft between 850 and 500 mb and was more in the rear.

7.10 Depressions mostly develop out of three types of situations. In half the number of cases, a diffuse pressure field develops over north Bay and adjoining areas, where pressure gradient to the south of the monsoon trough is fairly strong, as a normal feature of the monsoon season. Even if the trough had been north of its normal position, it swings southwards to the head Bay prior to the formation of the depression. In 15 percent cases, the signs appear as an upper air cyclonic circulation at any level upto mid–troposphere. One third of the depressions are born out of diffuse lows that travel across Burma into north and adjoining central Bay, some of them having been remnants of typhoons. Ramanna (1969) finds that more than 17 percent of the monsoon depressions can be regarded as having formed out of the remnants of typhoons of China Seas moving across Burma, assuming that they take 2 to 5 days to reach upper Burma, These statistics are not based on actually following the systems but assuming the possible influence if the time interval is between 2 and 5 days. He does not find any statistically significant evidence for

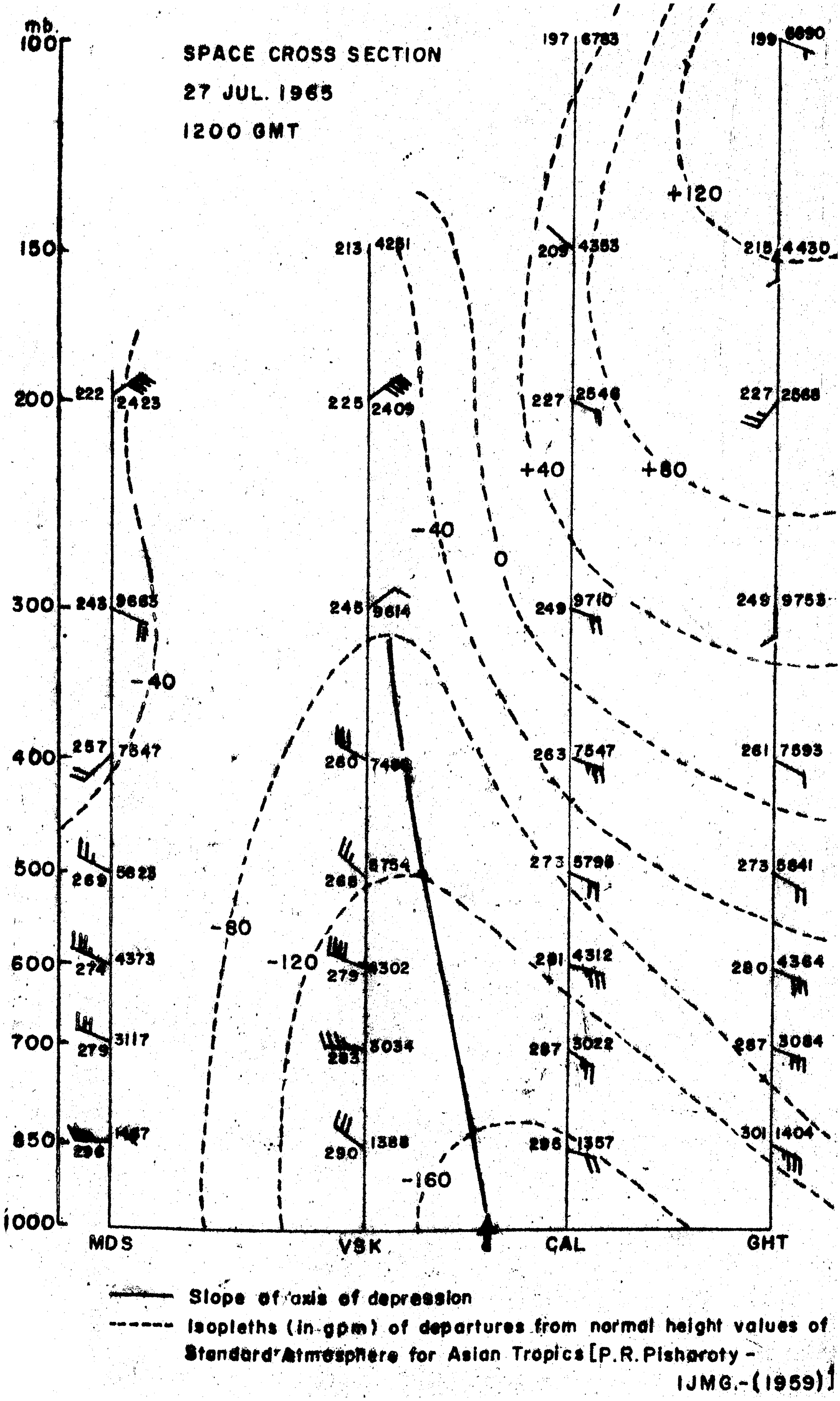
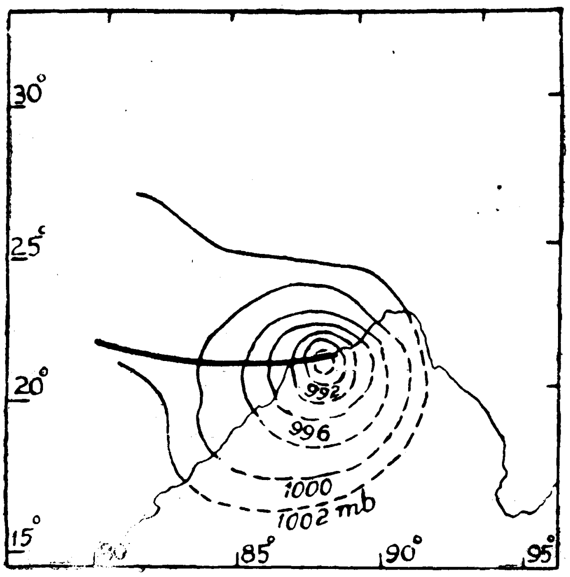
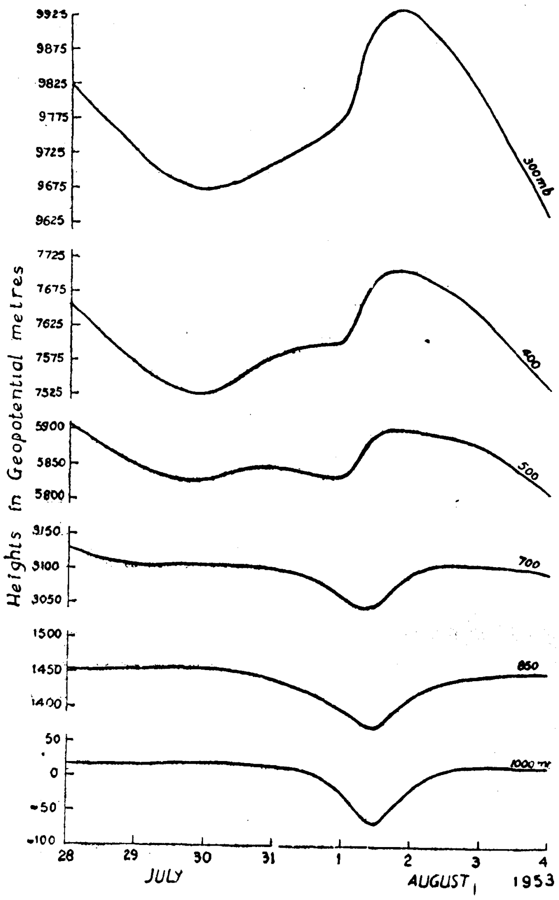

SPACE CROSS SECTION
27 JUL. 1965
1200 GMT
mb.
100
150
200
300
400
500
600
700
850
1000
MDS
VSK
CAL
GHT
+120
+80
+40
0
-40
-40
-80
-120
-160
Slope of axis of depression
Isopleths (in gpm) of departures from normal height values of Standard Atmosphere for Asian Tropics [P.R. Pisharoty - IJMG.-(1959)]

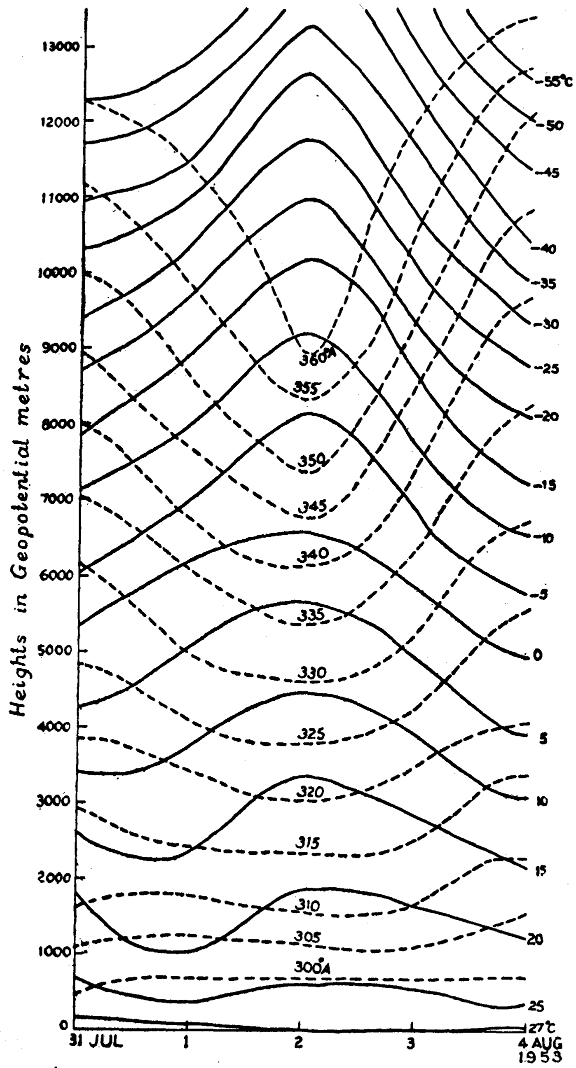
Fig. 7.5



(a) Fig. Cyclonic storm of
2 August 1953 and its track



(b) Heights of various standard
pressure surfaces



(c) T- solenoidal field in the storm
of August 1953
(———- isotherms,........ isontropes)

Fig. 7.6

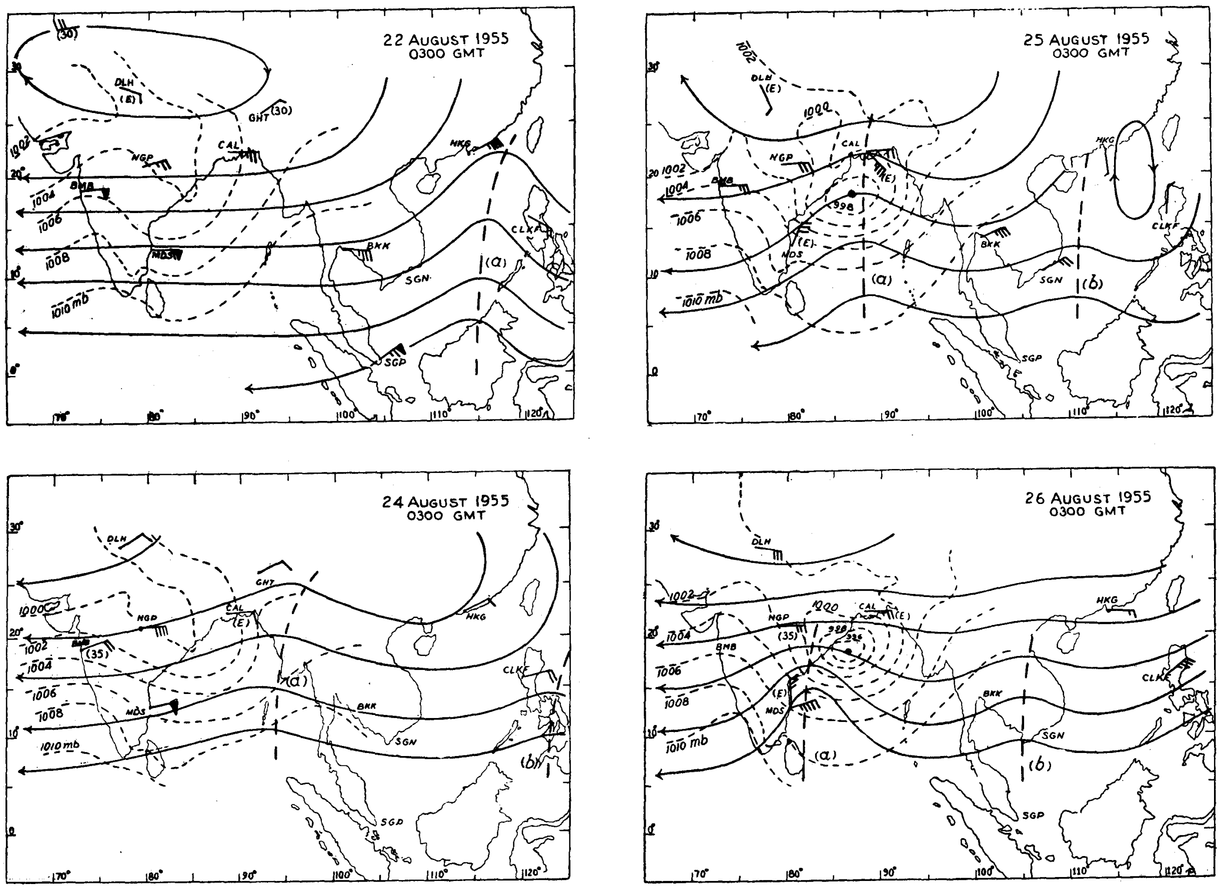
simultaneous formation of typhoons over China Seas and depressions in the Bay. Diffuse lows can be followed while coming into the Bay, as a cloud patch in satellite pictures, heavy rains along Arakan Coast and/or changes in the usual westsouthwest winds along the same coast, apart from identifying a pressure system as such whenever possible. Similarly, in the first two types, winds at surface and in lower troposphere, along Orissa and West Bengal coasts, changing to some northerly direction is a warning sign. These features take one or two days to develop into depression, occasionally longer. Some cases end only as a low which moves inland or dissipates in–situ. Precursor circulations in upper air are more noticed in mid–troposphere as the usual weak winds at those levels make the effect of any superposed perturbations more conspicuous than in the strong wind regime of the lower troposphere. Vorticity development initially in lower troposphere as wind shear, may be as important as the circulation noticed in mid–troposphere in a depression formation, though the former may not be as easily noticeable as the latter in conventional analysis. In all cases, the monsoon trough shifts to north Bay, if it was away. Often, the westerlies to south strengthen causing increase in vorticity.

Just north of the Indian monsoon area, westerly troughs move and their effect sometimes extends southward in the upper troposphere. Attempts have been made to explain formation of monsoon depressions as under the influence of divergence in the forward portion of such westerly troughs or 200 mb high. All the Same, to date, necessary and sufficient conditions to forecast formation of monsoon depressions have not been established.

Koteswaram and George (1958) relate the formation of depression with troughs in upper tropospheric easterlies and/or strengthening of the easterly jet. From an examination of hemispherical charts of 1955, they found that waves in upper tropospheric easterlies between 500 mb and 300 mb could be traced from even 140^{o} E to India. The waves have divergence in advance and convergence in rear on account of vorticity changes. Passage of an upper easterly wave over Calcutta was preceded by the formation of unsettled conditions or the extension of the seasonal trough over the north Bay. With a pre–existing trough or an extension of the normal monsoon trough into the north Bay, the approach of an upper easterly wave resulted in the formation of a monsoon depression. Similar effects are caused by the passage of wind maxima (jetlets) over Calcutta. Development with wind maxima is more rapid than with easterly waves. Easterly waves have a frequency of one in about six days during the height of monsoon. Figures 7.7 (a and b) illustrate the movement of a trough in easterlies and the formation of a depression. Even though the jet core is near 12^{o} N, wind maxima are effective in causing depressions only near north Bay, so also the troughs in easterlies extending from 30^{o} N to 10^{o} N.! Divergence above 300 mb due to systems at those levels can possibly get compensated without linking itself necessarily with lower layers.

George and Datta (1965) report a monsoon depression of September 1963, having formed out of the remnant of a typhoon from the Pacific, under the upper divergence of a 200 mb anticyclone; it was steered by the anticyclone at 300 and 200 mb and later recurved into a warm tongue in the 700–500 mb thickness, simultaneously coming under the influence of an extra–tropical westerly trough. While the sequence is not improbable, the difficulty has been to distinguish between systems behaving differently by assessing the effect of these various factors in an objective manner and in constructing thickness patterns in view of limitation of the relationship to thermal winds, quasi-geostrophic balance not being satisfied.

Desai (1951) is of the view that the presence of Arakan Mountains, Assam hills, eastern Himalayas, the Chota Nagpur Plateau and the northeastern end of the eastern Ghats help the development of cyclonic circulation in northwest Bay. The southeastern end of the monsoon trough runs through this area surface upwards. In contrast, over the northeast Arabian Sea, except for the Western Ghats, there are no mountains to help in setting up a cyclonic circulation. Besides, the western end of the trough axis below 700 mb is also not near this area, Influence like movement of low pressure waves from east across Burma, does not extend to the northeast Arabian Sea. The prevalence of continental air above the low level monsoon air may also be an



Flow patterns at 200 mb level based on the 0300 GMT rawin data
———— Stream lines; - - - - - - - - - . Sea level isobars at 0300 GMT
– – – – – – – Wave Troughs; E - Evening winds.

Figures within brackets indicate the level of the wind in thousands of feet.

Fig. 7.7

inhibiting factor.

Ramage (l97l) considers that the strong vertical shear (≈15 m sec^{-1}) between 850 and 200 mb as against 7 m sec^{-1} in April and 4 m sec–1 in November inhibits development of cyclonic storms in monsoon. Combination of large vertical wind shear and subsidence also prevents tropical cyclones developing west of 70° E

7.11 Srinivasan, Raman and Ramakrishnan (l97l) have studied 70 satellite cloud pictures associated with different stages of monsoon depressions in 1967 – 1969, The most common feature is an extensive heavy overcast mass in the southern sector. This is a distinct bright cloud mass, brighter than others in the neighbourhood. Its extent is 5° – 7° latitude. Sometimes it extends to northwest also. Cumuliform clouds are present in most cases in the northern sector, which get organised into bands as the low intensifies into a depression, more so when the system becomes deep. Associated cloud development is classified into five types. The first type is of a well marked low and the second type a depression. In both these stages the centre cannot be defined from the clouds, In the third type, corresponding invariably to a depression, the centre is located in the cumuliform field close to the northern edge of the overcast mass. In types IV (depression or deep depression) and V (deep depression), the Cumuliform bands spiral towards the centre which is at the northern or northeastern edge of the overcast. Figures 7.8 (a to e) show the illustrations of the types. Sharma and Srinivasan (l97l) found that in a majority of cases, satellite determined centres were displaced to southwest with respect to isobaric centres. These results are in agreement with the finding of Keshavamurthy (1972) and Kulshrestha and Gupta (1964). In the case studied by the latter authors, there was a second overcast further north of centre with a second rain belt developing under the influence of a middle latitude westerly trough.

7.12 Chakraborty (l950) reports five cases of calm centres in cyclonic storms, 4 in monsoon and 1 in pre–monsoon, the period of calm varying between 1 1/2 to 2 1/2 hours and estimates the calm centre as of 10–15 km extent. There was slight increase in temperature but not in wet bulb temperature. The strength of the strong winds around was not symmetrical.

The radar picture of the cyclonic storm of 13–15 September 1958 with a maximum gust speed of 75 kt as observed by De and Sen (1959) is shown in Fig. 7.9. The diameter of the eye varied between 30 and 40 km. The tops of echoes extended to 5 km only.

The spiral structure observed on radar in the cyclonic storm of 21st September 1962, reported by Bhattacharjee and De (1965.) is shown in Fig. 7.10, Recorded winds show a maximum of 50 kt and the speed of movement of storm as estimated from echoes was 29 km hr^{-1}.

7.13 Heavy rains appear mostly confined to the southwest sector of monsoon depression. Elsewhere, it is scanty, surprising for a tropical low in a moist air mass with more than eighty percent humidity. A sign of a depression changing course, is the appearance of rain belt between northwest and northeast, the direction towards which the system may move. No doubt there are some cases, particularly in later months, when rainfall occurs in all sectors. At the time of formation, a depression may have a more uniform pattern of rainfall. Rainfall may occur in the rear also when south/southwest winds in the field are strong. Such rain is more showery as against the steady rain in southwest sector.

Pisharoty and Asnani (1957) composited the rainfall in three monsoon depressions with reference to the tracks and centres of depression. Pressure departure was of the order of –10 mb at the centre and movement was westnorthwest. Heavy rainfall of 7.5 cm and above in the preceding 24 hours was confined to a belt 400 km wide to the left of the track and extending 500 km in advance and an equal distance to the rear along the track (Fig. 7.1l). This does not mean that heavy rain occurs in the rear. The 24 hours past rainfall in the rear of the current morning position

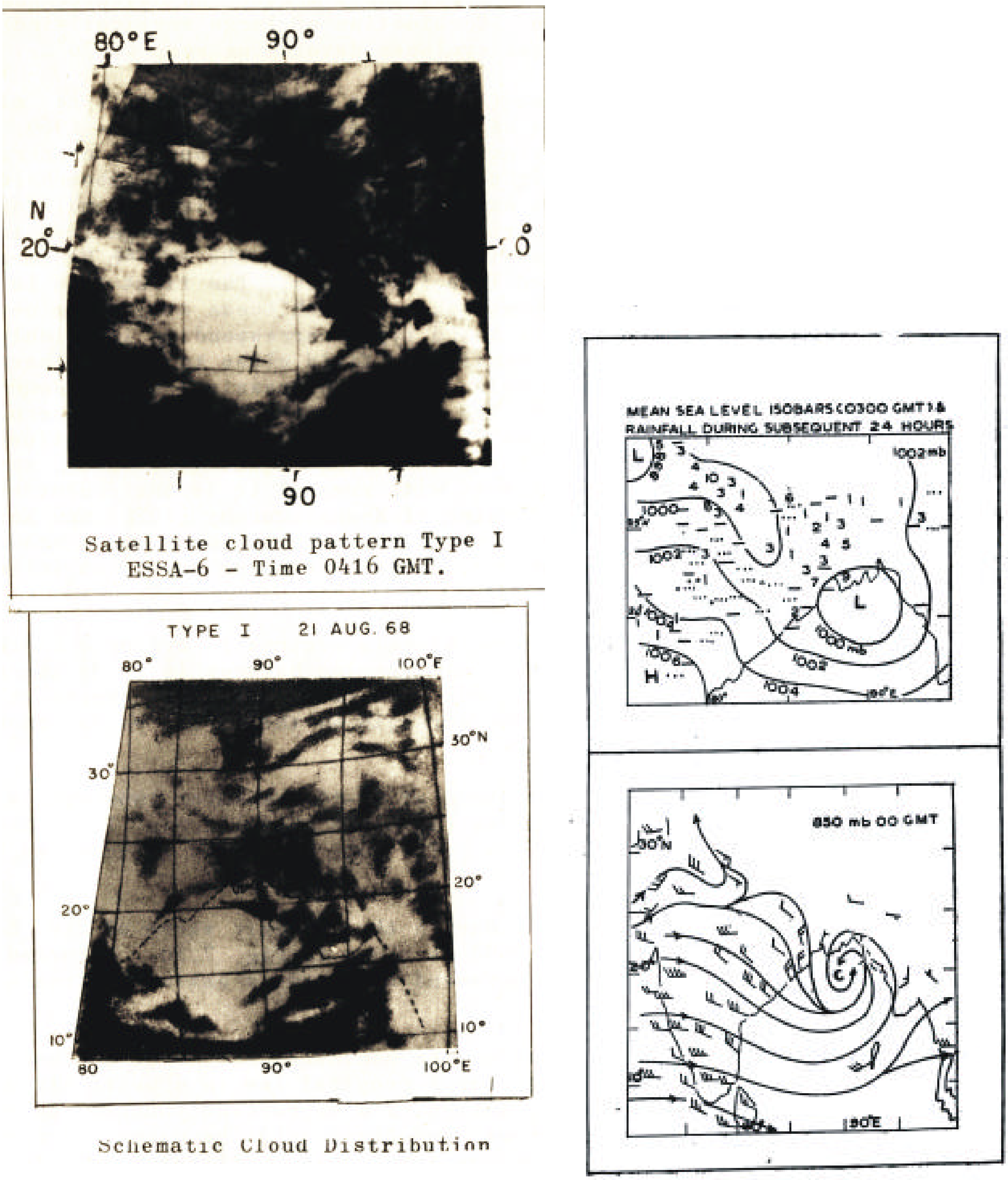
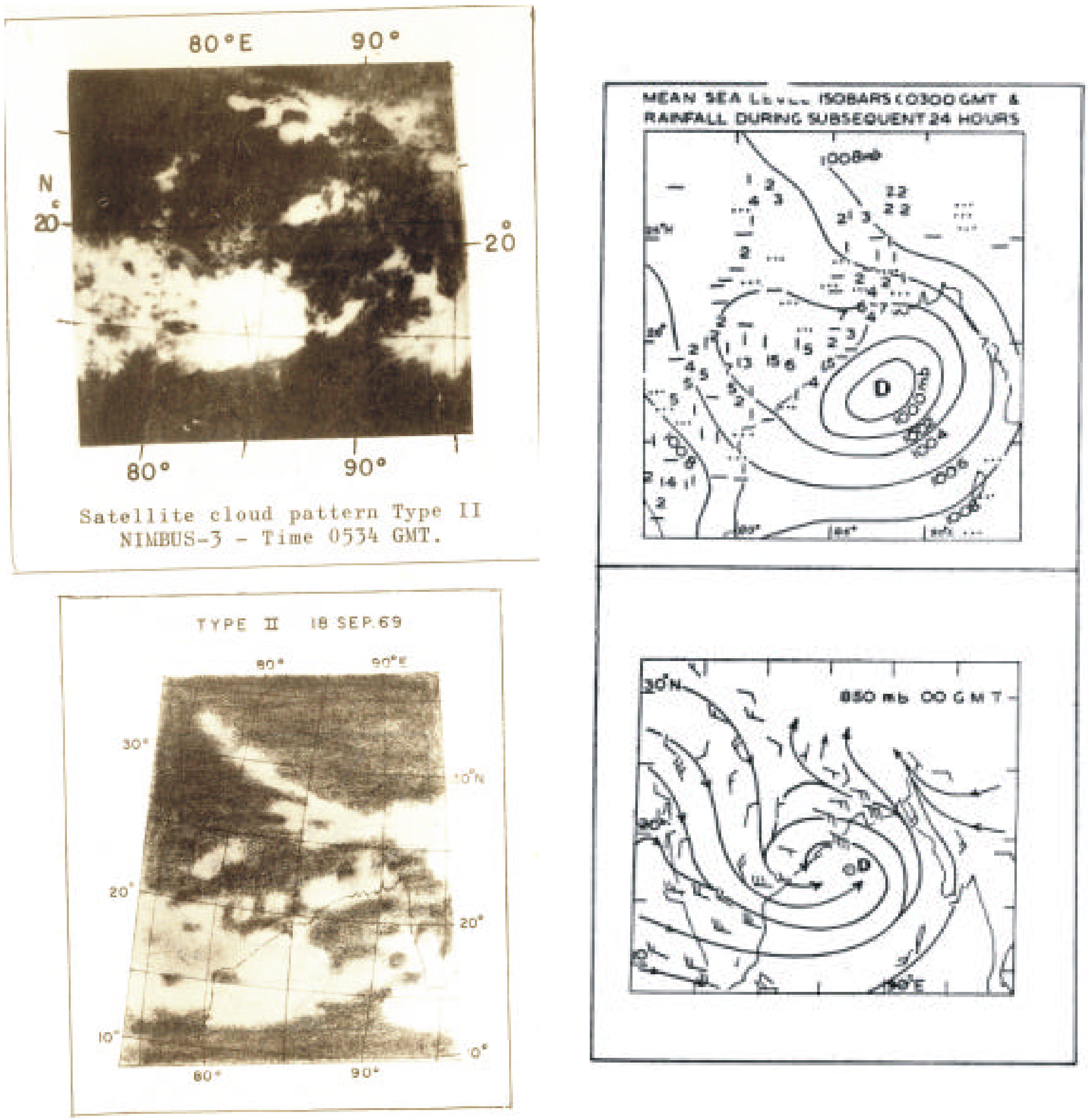
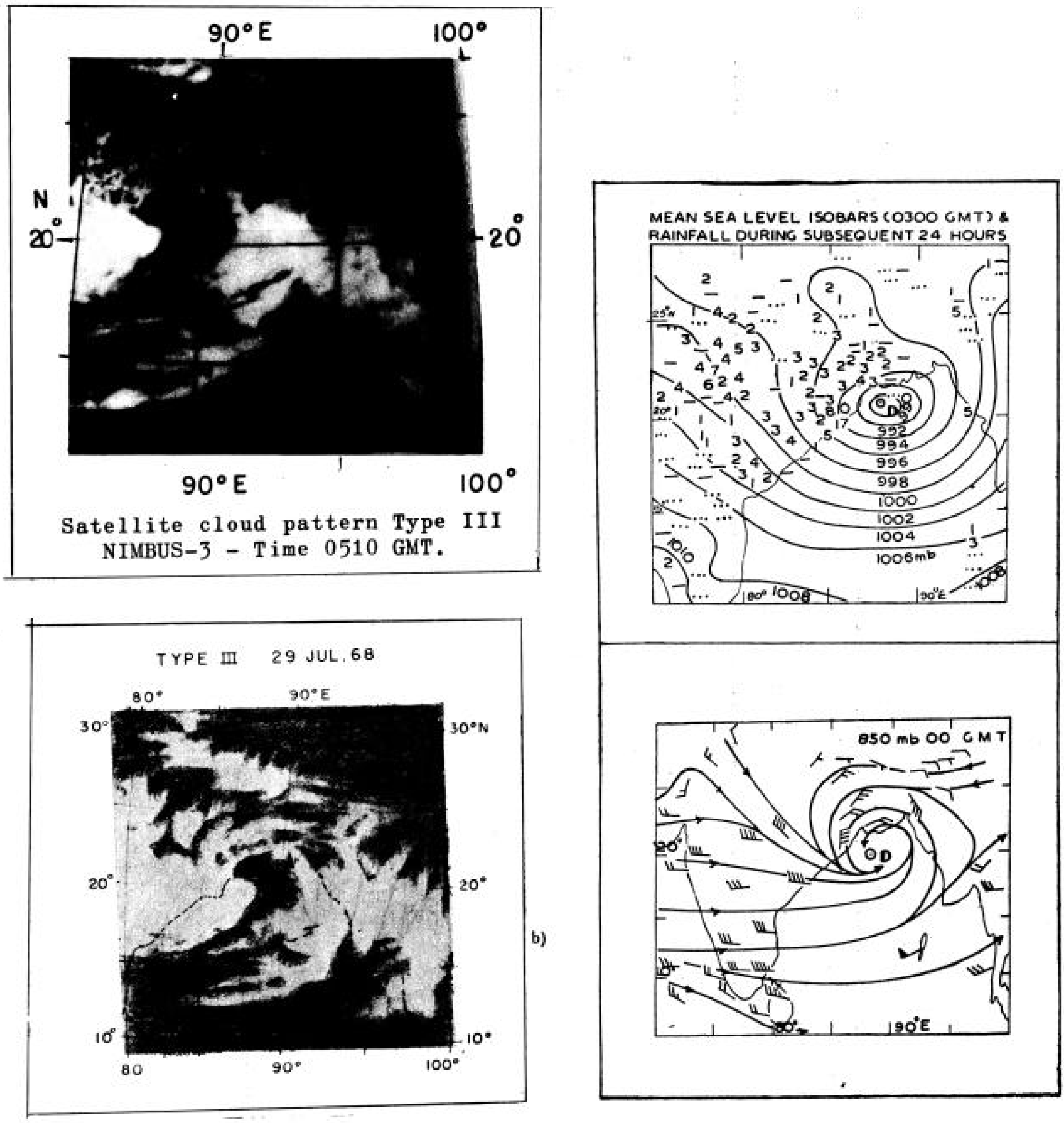


Fig.7.8 (a) 21 August 1968. Low over North Bay of Bengal developing into depression (Central pressure 1000 mb; Pressure departure -4 mb).



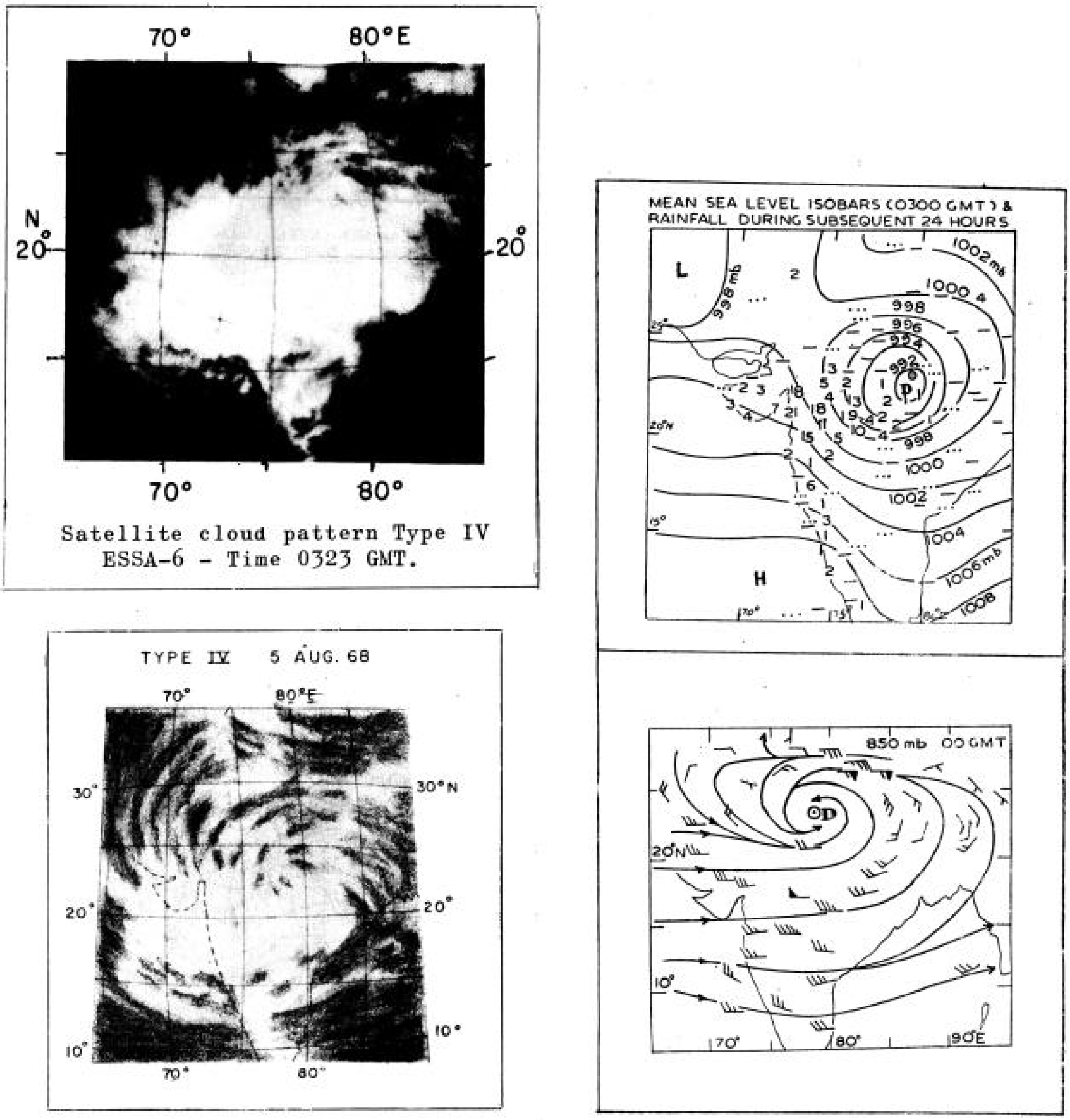
Schematic Cloud Distribution

Fig.7.8(b) 18 September 1969. Depression centred 17.5°N, 87.5°E (Central pressure 1000 mb; pressure departure -4 mb).



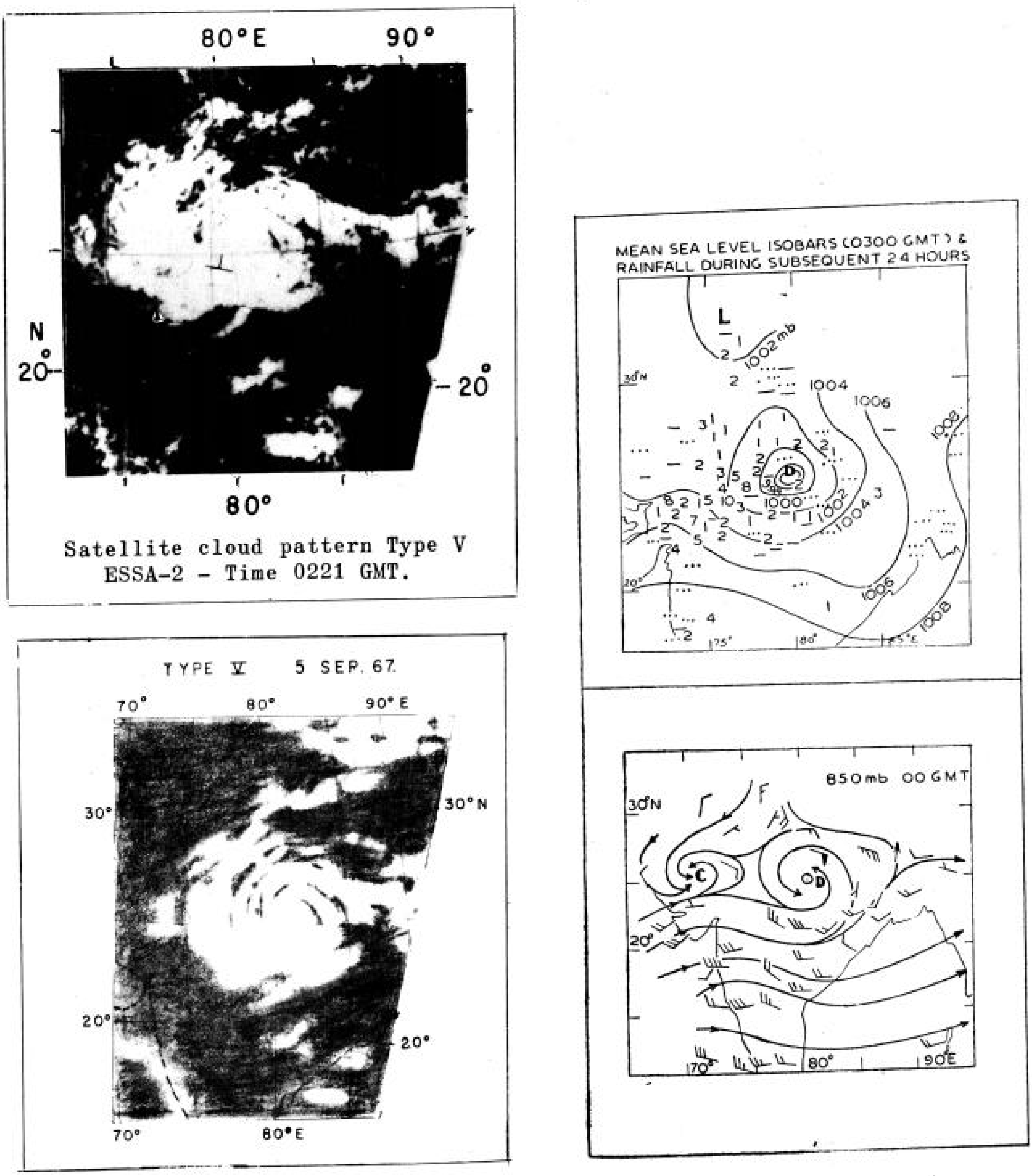
Schematic Cloud Distribution

Fig.7.8(c) 29 July 1969. Deep depression centred 20.5 °N, 88.0°E (Central pressure 990 mb; pressure departure -8 mb).



Schematic Cloud Distribution

Fig.7.8 (d) 5 August 1968. Deep depression centred 22.5°N, 79.0°E (Central pressure 992 mb; pressure departure -10 mb).

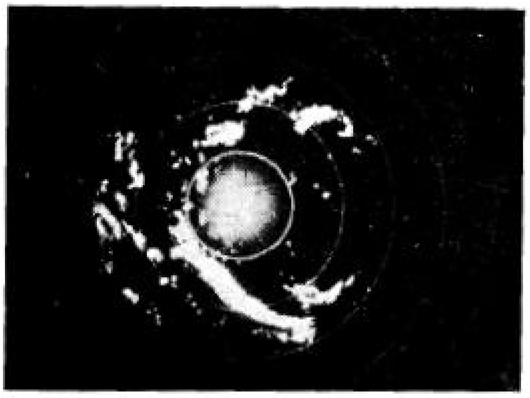


Schematic Cloud Distribution

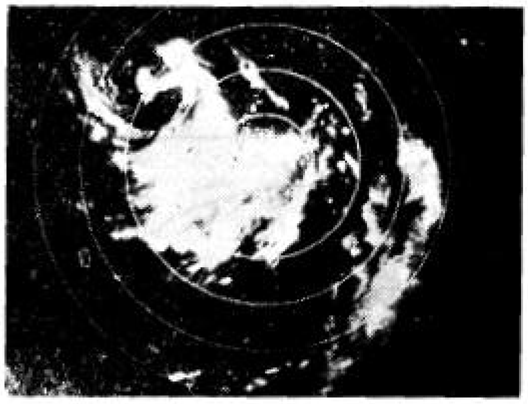
Fig.7.8(e) 5 September 1967. Deep depression centred 25°N, 79°E (Central pressure 994 mb; pressure departure -9 mb)

2113 60 km

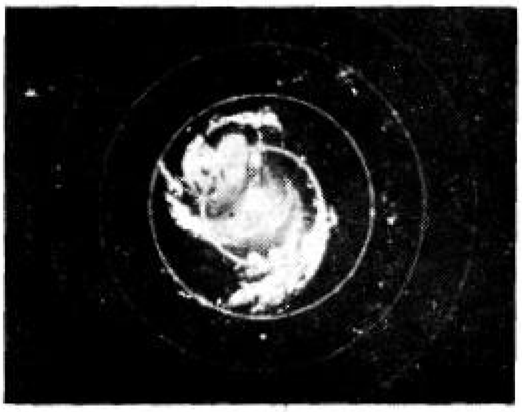
Fig.7.9 PPI presentations of the storm detecting radar at Dum Dum Air Port on 13 September 1958).
(Figures in the left and right hand bottom corners indicate time in 1ST and range rings in km respectively.



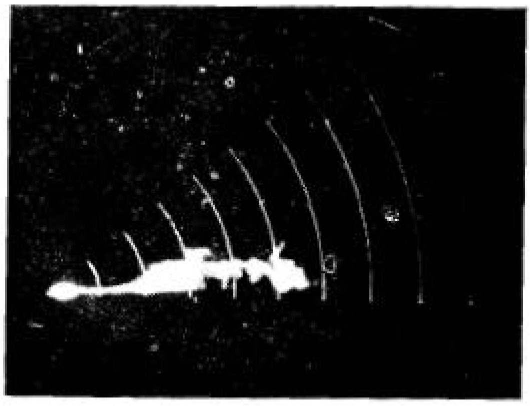
0045 (a) 60



1117 (c) 20



1005 (b) 60



1126 (d) 10
Azm. 310°N

Fig.7.10 PPI and REI presentations of the storm detecting radar at Dum Dum air port on 21 September 1962,
(Figures in the left and right hand bottom corners indicate time in GMT and range rings in km respectively).

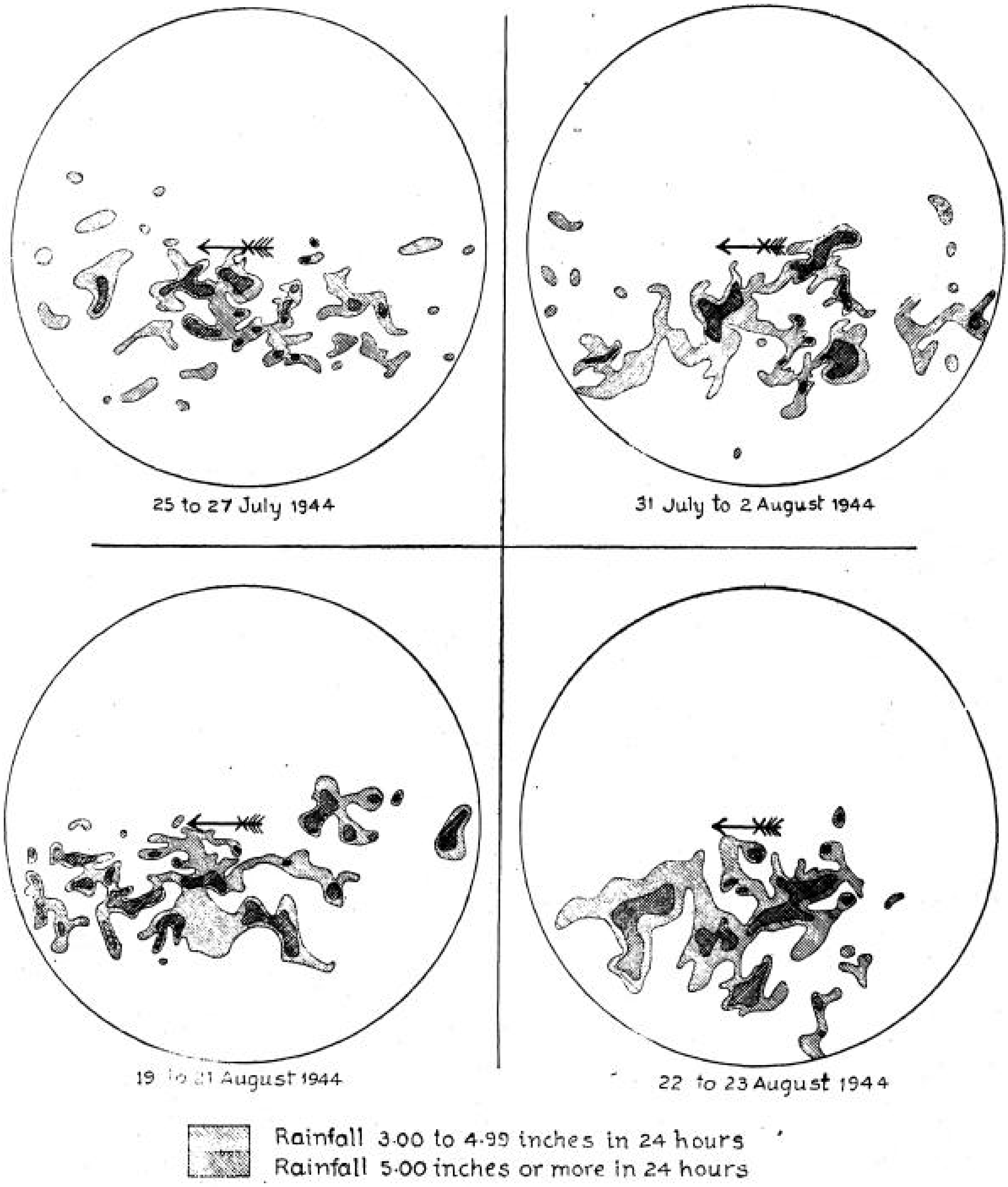


Fig.7.11 - Composite charts of rainfall for area 350 miles around depression centres. (Scale 1" = 140 miles).

had occurred really in advance of earlier positions of the depression. It may be summed up that heavy rain occurs in a belt of 400 km wide to the left of the track for a length 500 km from the centre. The whole of this area does not get heavy rain but is liable to heavy rain. Only 30 percent of the area actually records such rainfall. Almost similar results were obtained by Lal (1958) with slow moving depressions.

Bedekar and Banerjee (1969) composited the rainfall associated with six depressions and found that rainfall exceeding 5.0 and 7.5 cm was invariably concentrated to the left of the track within a narrow belt 2° to 3° wide and 7° length. Heavy rain rarely occurs to the right of the track except at the stage of recurvature. Fig. 7.12 shows the depressions (4–8 mb departure) considered by them and the probable rain areas of 5 cm and 7.5 cm. The probable area is an envelope of region where the rainfall of the specified amount occurred at least in two out of the six cases and the more probable area in at least three cases. Areas of heavy rain (> 7 cm) is somewhat away from the centre, none in these six cases within 60 km of centre. Heavy rain generally occurs in patches. On the average about eight individual patches of heavy rain are observed in an occasion, each covering about 4,800 sq. km. Desai (1972) has shown that in the southwest quadrant heavy rain can occur even within 70 km of the centre under suitable conditions.

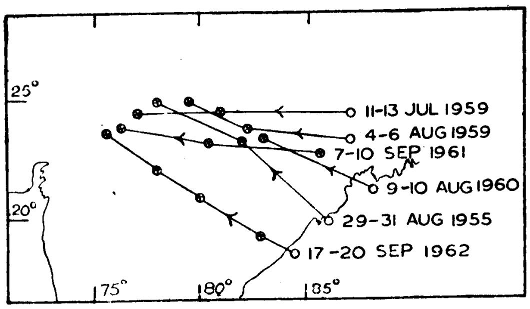
Bedekar and Banerjee have also illustrated from the flow pattern of 0.9 km, the convergence zone that develops between the westerlies and easterlies, away from the central parts of depression. This zone lies in between the flow from the ridge to the west to the trough associated with the depression. In this region, convergence occurs due to curvature correction to geostrophic flow. It shows itself characteristically in the streamlines at 1 km. The convergence zone is 5° away from center. In Fig. 7.13, it will be seen from the rainfall charts of Jabalpur, Raipur and Jagdalpur (situated perhaps in the convergence zone), the fairly heavy rainfall was characterised by one or two spells of very intense rain accompanied by thunder. At Nagpur, perhaps outside this zone, precipitation was rather showery in the beginning but was more of a continuous type later with much diminished intensity. A rainfall maximum between 80° – 81° E in the mean isohyets of July and August, which cannot be associated with any orographic features, is regarded by these authors as due to this convergence zone in advance of depressions.

Dhar and Mhaiskar (1973) have studied in some detail rainfall of ten depressions, two of early October and eight of July to September on the day of their crossing Orissa coast. The mean rainfall profile across the depression track (over a length of 250 km of the track) is shown in Fig. 7.14 a. Maximum of areally averaged rain is between 100 and 150 km to left and decreases to one third at 125 km to right.

Raghavan (1965) reports the case of the depression of 24 August, 1960, which moving from Dum Dum to Dhanbad, showed curved bands of wet and dry zones at 150 km and 300 km from centre. The wet zones were 30 km to 50 km wide and dry zones 50 km to 80 km. This was a case of low rainfall less than 1 cm and not in the heavy rain zone of the southwest quadrant.

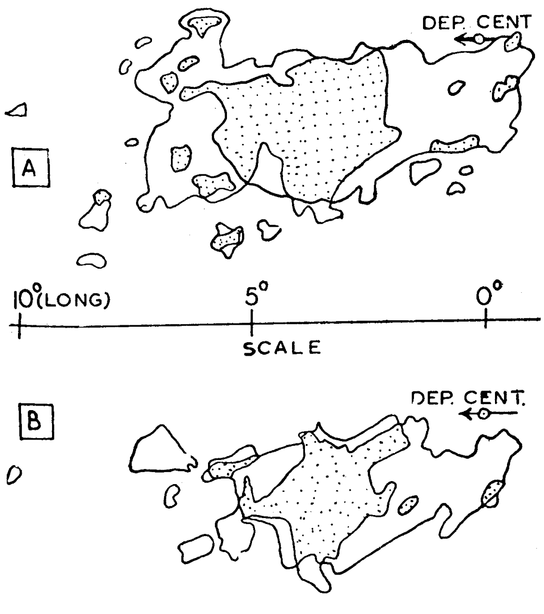
Venkataraman, Chowdhury and Banerjee (1974) composited 33 days relating to 11 monsoon depressions over Madhya Pradesh and Vidarbha moving between west and northwest to study the hourly rainfall distribution. Fig. 7.14 b shows the rainfall in the intervals 0830–1430 1430–2030, 2030–0230 and 0230–0830 IST, Rainfall amounts are least in the first period and increase in area and intensity towards the last period. Area of 100 km around centre has only light rainfall. Most rainfall is between 210° to 300° from the centre. The tendency to become symmetrical around 270° in the last period mentioned by the authors is perhaps due to the slight northward motion of the depression by that time. Their diagrams of highest hourly rainfall are given in Fig. 7.14 c.

A second rain belt often develops outside the depression field to the west an fairly distinct from the southwest sector, due to convergence between northwest flow, northeast/easterly flow around the depression and westerlies to south, in the lower troposphere (Raman and Banerji



Tracks of depressions.

Depression positions at 0300 GMT only have been indicated by small circles. Positions for which rainfall data have been utilised for preparing composite charts are shown with cross within the circle.



Composite maps depicting "Probable areas" around depression centre (shown by continuous lines) for 50 mm of rain (top) and 75 mm of rain (bottom) "More probable areas" are stippled.

Fig.7.12

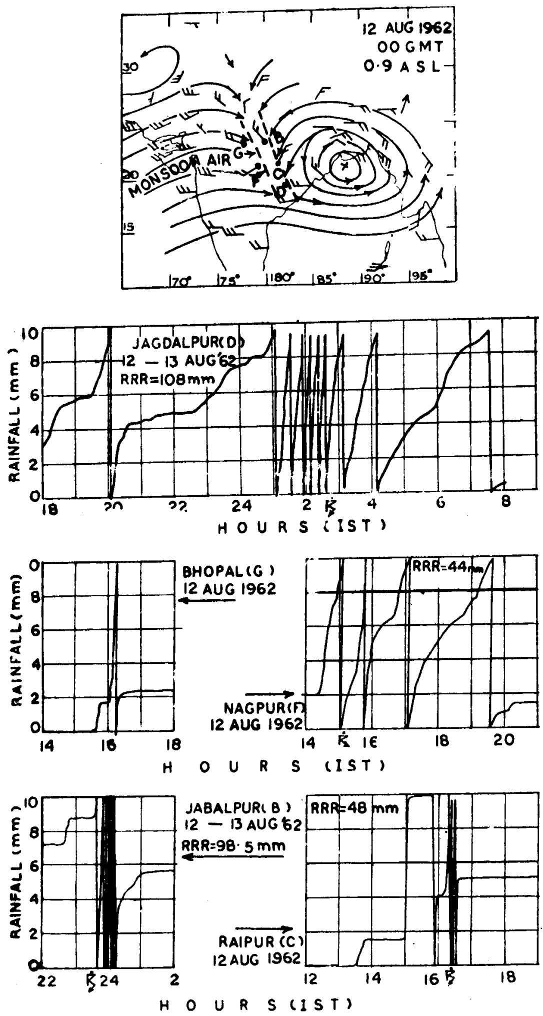


Fig.7.13 - Streamflow chart at 0.9 km a.s.l. for 00 GMT of 12 August 1962 and relevant intensity rainfall charts of Jagdalpur, Bhopal, Nagpur, Jaba1pur and Raipur.

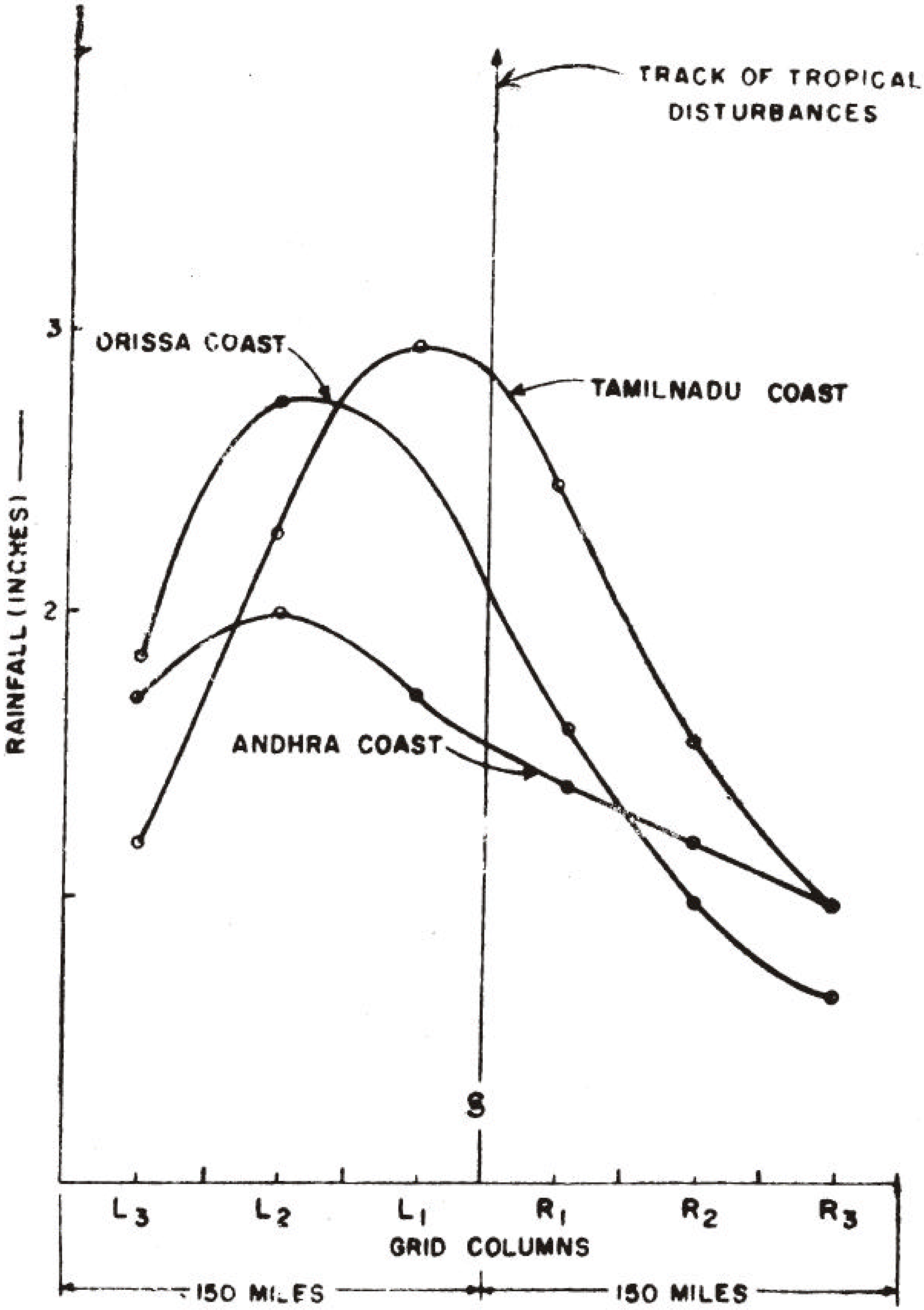


Fig. 7.14(a) Columnwise distribution of rainfall on the day of crossing along the different sections of the east coast. L_1 L_2 L_3 Left' R_1 R_2 R_3 Right

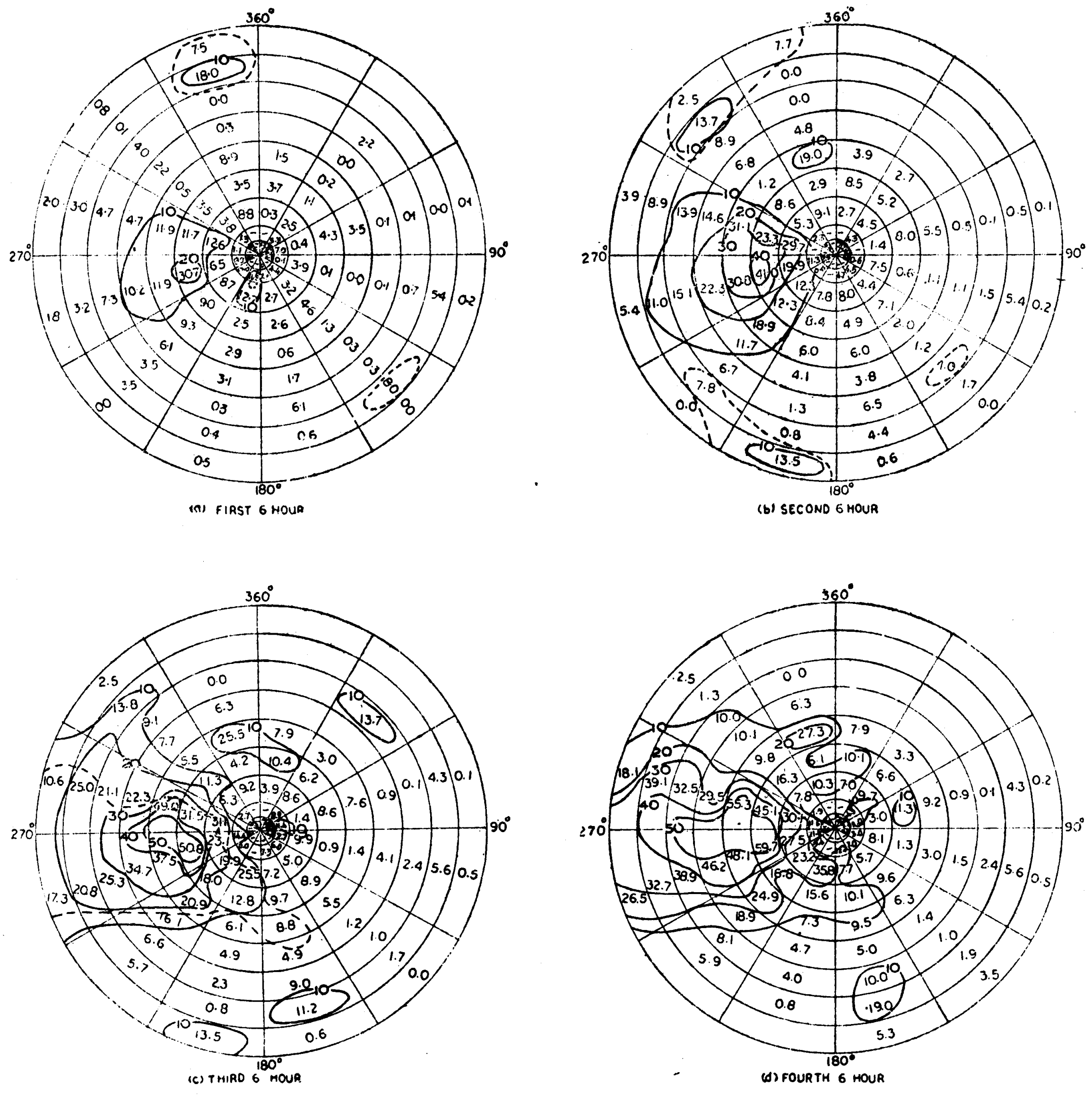


Fig. 7.14(b) Distribution of average rainfall (mm) in different zones around the depression in 6 hr. intervals. The dotted line covers the area where rainfall was 30% of the total rainfall in 24 hour.

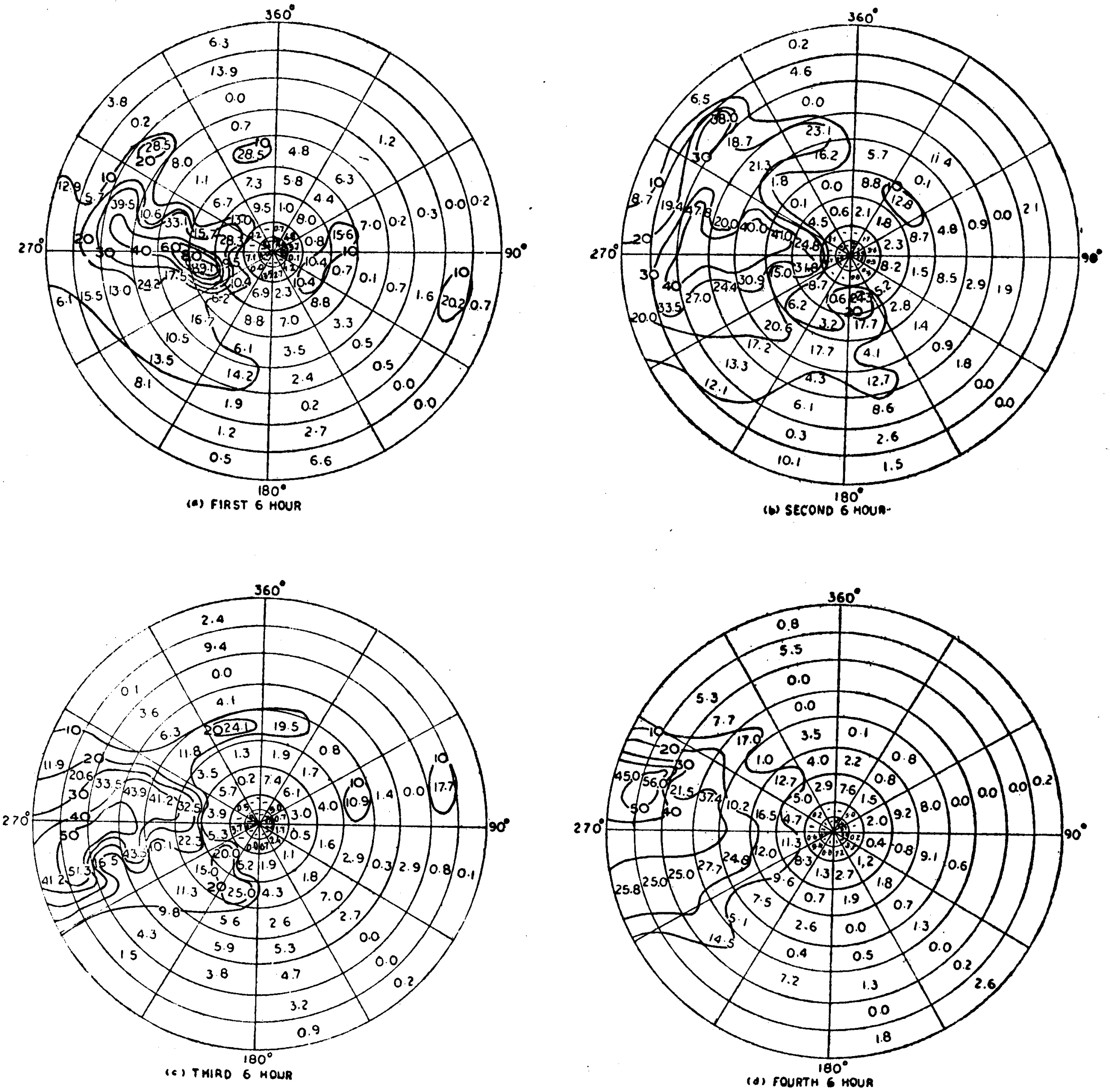


Fig.7.14(c) Distribution of highest hourly rainfall (mm) around the depression in 6 hr. intervals.

1970). This area seems to have been covered in the compositing of heavy rainfall region by Pisharoty and Asnani (1957). In the case of August– September monsoon depressions, this kind of synoptic pattern causes phenomenal rainfall along north Maharashtra coast and southeastern parts of Gujarat. This is best forecast from the wind pattern.

Monsoon depressions while forming in northwest and adjoining parts of west central Bay may strengthen rainfall along Konkan Coast. This happens in a three degree belt to the south of the centre of the depression along west coast. Once the depression has formed, rainfall decreases, unless some other favourable synoptic pattern has developed. A second spell of rains may occur in this area as the depression moves to the west of 77^{o} E but remaining south of 23^{o} N. At this time, rainfall may commence a little to the north of Bombay and extend southwards. These sequences occur frequently but not always.

An interesting effect of the formation of depression in north Bay is the dry weather in Assam, usually liable to good rains. The tephigrams of Gauhati and Visakhapatnam 600–700 km from centre showing subsidence in mid–troposphere in such a case are reproduced in Fig. 7.15 (Srinivasan, Raman and Mukherji, 1972).

Rainfall in September depressions is smaller in areal extent but many become enhanced in area and intensify when they come under the influence of middle latitude westerly troughs which then move at a lower latitude. Initially a separate rain belt may develop further to north and later as the depression moves north, wider and intensive rain belt forms all round the depression. The chart of 4th October 1955 (Fig. 7.16) presented by Parthasarathy (1958) and the case of 20–21 September 1962 analysed by Venkataraman and Bhaskara Rao (1965) illustrate these aspects.

7.14 A satisfactory explanation has to be given for the widespread and heavy rainfall in the southwest quadrant of monsoon depressions. This could be done in relation to the pressure and wind pattern in the lower troposphere. At the surface, the pressure gradient within the field of closed isobars is generally maximum to the south of the centre and least towards westnorthwest in which direction the isobars have an oval shape. Very often there is a pronounced ridge about 7° west of centre and extending to 1° to 2° north of the latitude of the centre. Isobars from this ridge run into a trough south of the closed isobars. The centre of depression slopes to southwest aloft.

Taking account of the difference in temperature between the warmer air in the northwest sector and colder air to southwest, a view has prevailed that a frontal boundary extends from centre of depression to west and the heavy rain in southwest quadrant is due to upglide along this frontal slope (Desai and Koteswaram, 1951). Pramanik and Rao (1948) and Mull and Rao (1949) have regarded that temperature data do not support the existence of a front and the latter authors proposed an explanation based on convergence in gradient wind motion. Pisharoty and Kulkarni (1956) have discussed the convergence on considerations of vorticity variations. In the following para, the subject is discussed from dynamical considerations.

Due to the maximum pressure gradient south of the centre, the maximum vorticity at surface would be in that region and least to westnorthwest, where pressure gradient is minimum. As the depression is moving westnorthwest, curvature of trajectories would be maximum in the left forward sector. As a result, air parcel in the southwest sector would be moving into regions of increasing vorticity and subject to convergence. The slope of depression centre (towards southwest with height would augment this convergence compared to other sectors. From south to westnorthwest of centre (through east), air parcels would be moving towards decreasing vorticity favoring divergence. The flow from the ridge to west of depression to the trough to the south would also be associated with convergence. The whole area of cyclonic vorticity would be subject to frictional convergence but unless this is augmented by other factors through a deeper layer as in southwest sector, rainfall cannot be much.

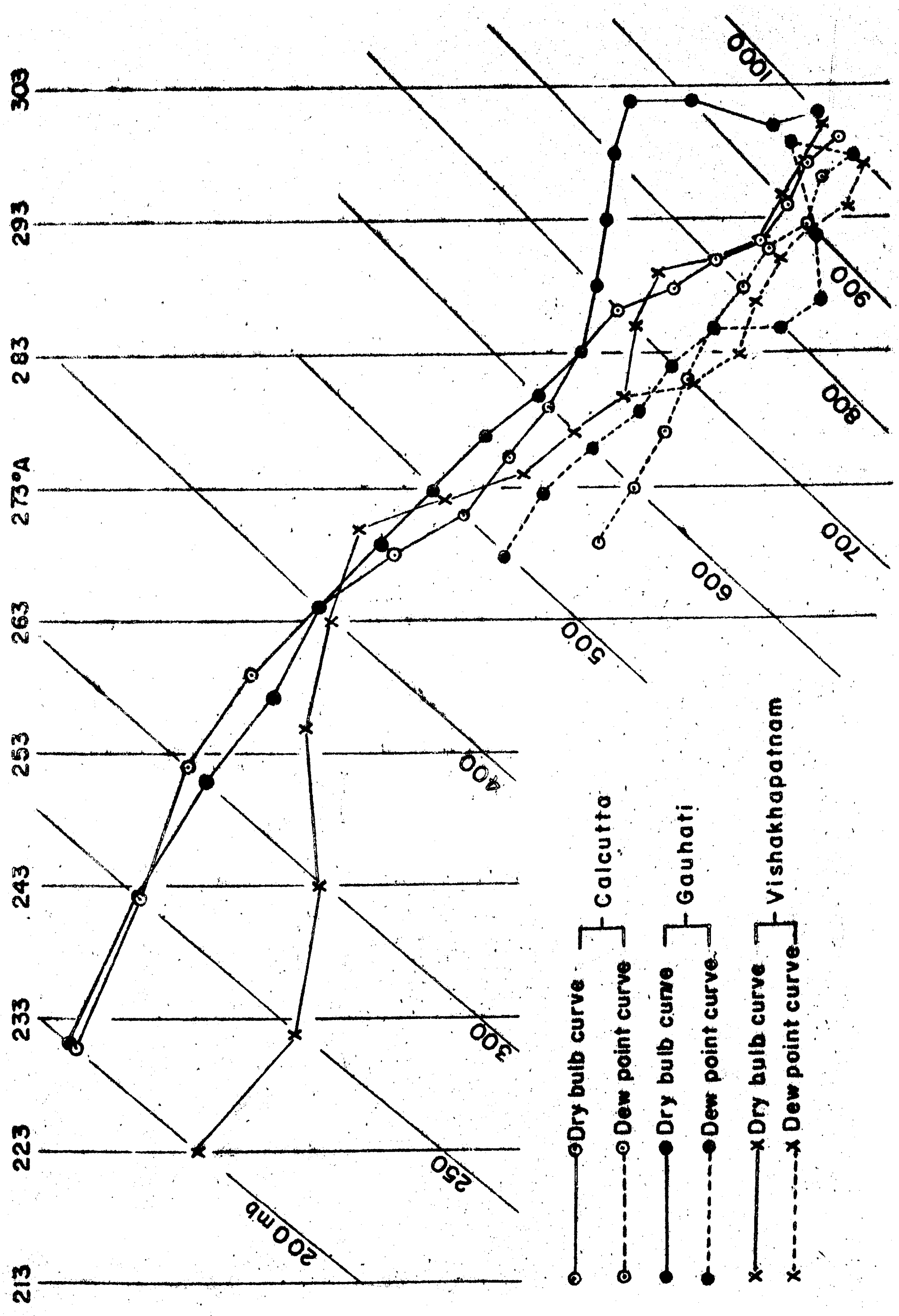


Fig. 7.15 Tephigrams of Calcutta, Guhati and Vishakhapatnam 27 July 1965 00 GMT

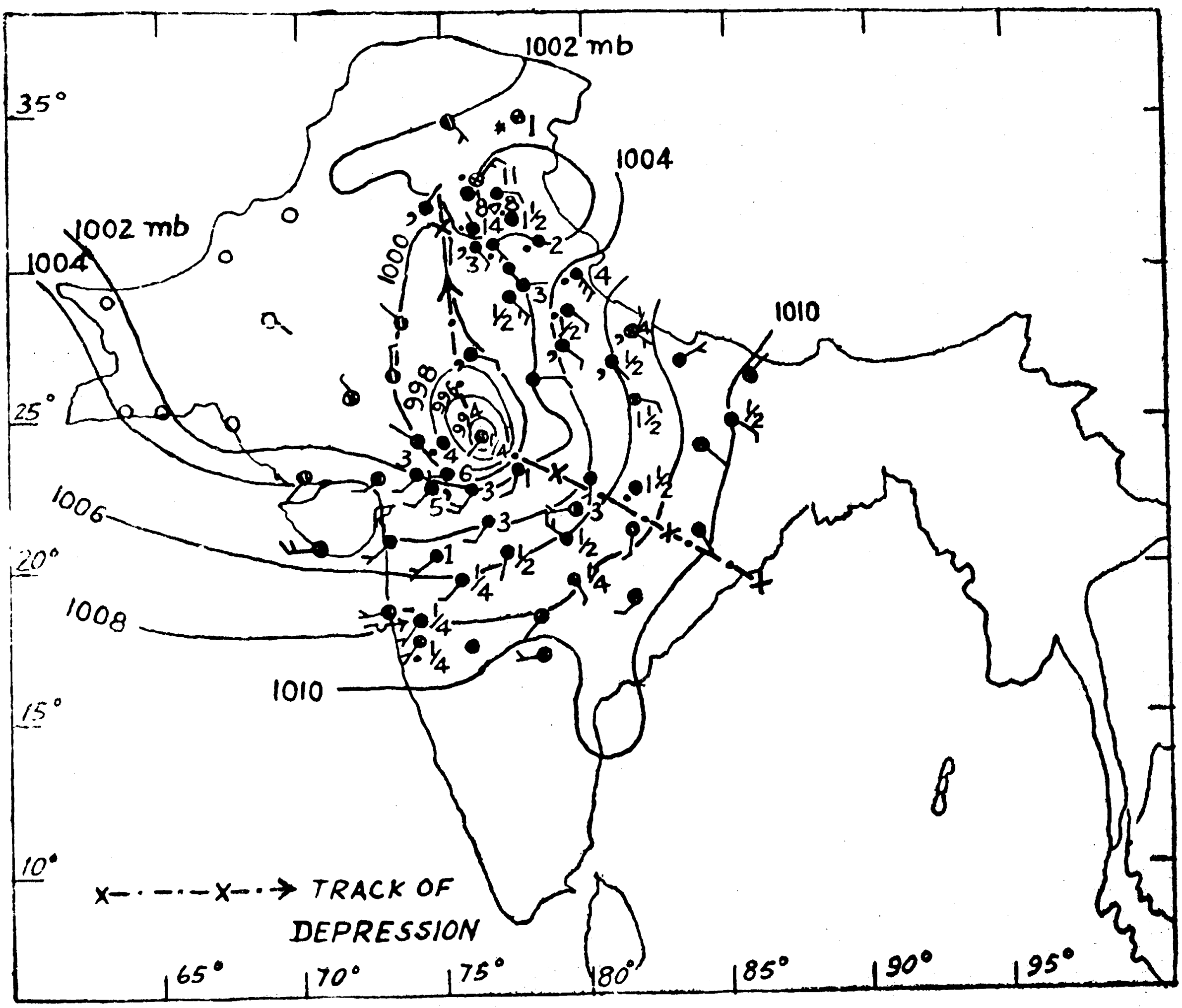


Fig. 7.16 Synoptic chart 4 October 1955 03 GMT.

Pressure gradient may more than make up for greater curvature effect on geostrophic wind towards south and wind may still be maximum at this place and least near about westnorthwest. This would make air flow from south to westnorthwest through east decelerating and accelerating in the rest. A geostrophic component would be outward from centre (i.e. towards high pressure) in the former and towards the centre in the latter. Convergence in the frictional layer would be enhanced in the southwest sector and reduced elsewhere.

If pressure gradient is strong to rear of depressions, rainfall is more therein which would fit with the above reasoning.

7.15 Rao and Rajamani (1970) have conducted a diagnostic study of a weak depression of 25 July 1966 which had only one closed isobar. The case chosen cannot be considered as very typical. Assuming quasi–geostrophic motion and neglecting some lower order terms, the equation

$$s\nabla^2 w + f_0^2 \frac{\partial^2 w}{\partial\ p^2} \approx f_0 \frac{\partial \vec{v}}{\partial p}.\nabla\left(2z\ + f\right)$$

is used to compute vertical motion, taking w = 0 at 1000 mb and 150 mb. Ascending motion occurs in regions where the thermal wind blows from high to low vorticity. Fig. 7.17 shows that marked ascent was taking place in the northern part of the Peninsula where 850 mb flow was from ridge to trough, outside the central part of the depression, generally tallying with rainfall. The vertical velocity was a maximum at 700 mb (2.5 sec^{-1}) and decreased aloft. Patterns of vertical motion at 300 mb differed from 500 mb and below, as the flow pattern changed considerably at this level. Unfortunately the authors have not presented the vertical velocities obtained for the northwest quadrant of the depression.

Figure 7.18 (a and b) show the thickness pattern between 850 mb and 500 mb and geostrophic relative vorticity at 700 mb. The thermal winds are weaker closer to the centre and in the northwest quadrant. Fig. 7.18 (c) shows the vertical variation of w , divergence and relative vorticity at the position of maximum vertical ascent. The maximum w at 700 mb and the level of non–divergence at that level is regarded as in agreement with the theoretical result of Kuo (1953) that for disturbances with wavelength less than 3000 km, the level of maximum w will be lower than 500 mb.

Rao and Rajamani (1972) have computed the relative vorticity, divergence and vertical velocity for a Bay depression of July 1969, using the same equation. Profile along 20^{o} N is shown in Fig. 7.19. Maximum convergence and vertical velocity are 2^{o} west of the centre. The authors make out that as thermal wind is easterly, ascending motion takes place in the western sector. This does not seem to explain the marked difference in rainfall between the southwest and northwest quadrants. The need to have closer radiosonde ascents near the centre is obvious.

Das, Dutta and Chhabra (l97l) have computed the vertical velocity for 1st August 1969 depression on quasi–geostrophic model. The area of computation covers 20^{o} of long. and 12 ½^{o} of lat. Vertical motion due to slope of ground and friction in the surface layer have been taken into account. Diagrams of vertical motion etc. are given in Figs. 7.20 (a, b, c, d, e, f). Maximum w is at 800 mb, of the order of 10x10^{-4} mb sec^{-1}. Marked ascent occurs to the west of the depression centre. The centre of upcurrents at 1000 mb agrees well with centre of heavy rainfall but from 800 to 500 mb maximum ascent is displaced to the north. At 200 mb, subsidence is indicated all over the field. All these computations have the limitation of assuming dry adiabatic ascent of air.

Venkataraman (1955) computed the relative divergence between 1000 mb and 700 mb from contour configuration according to the method of Sawyer and Matthewman (1951). In the depression of 12/13 August 1953, rainfall coincided with areas of maximum relative divergence.

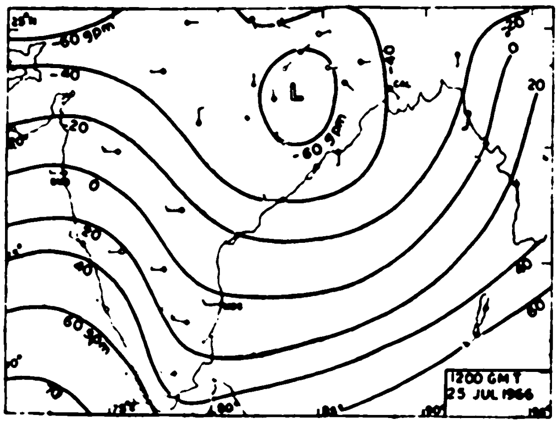


Fig. 7.17(a) 1000 mb. Chart - 25 Jul. 66 12 GMT

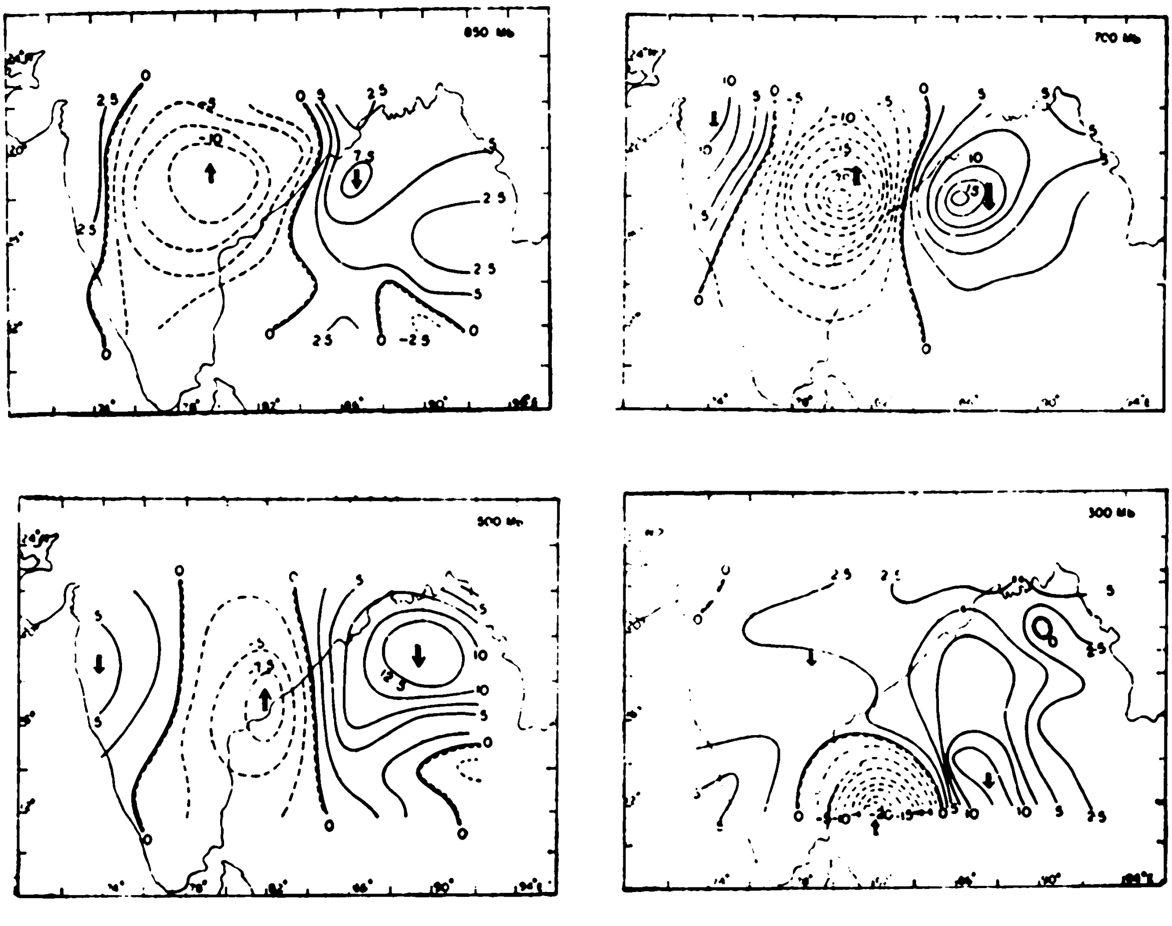
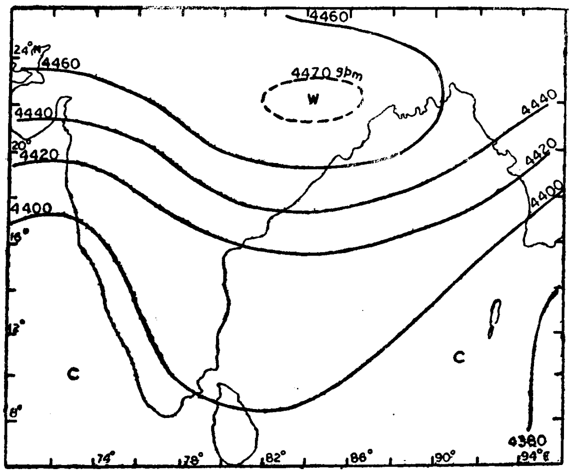
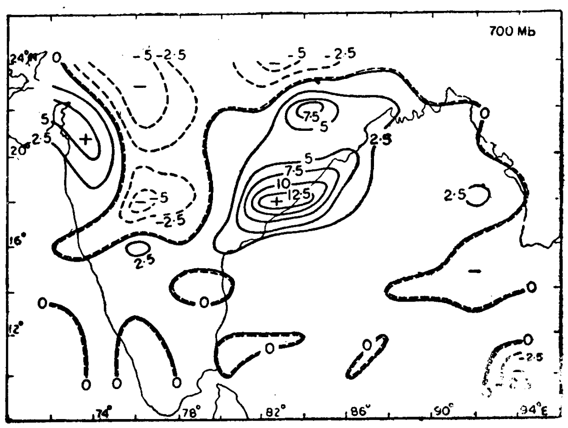


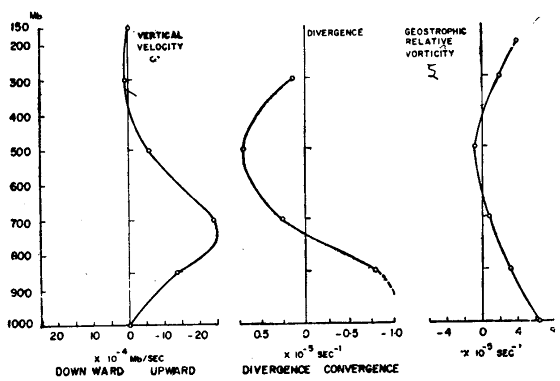
Fig. 7.17(b) to (e) Vertical velocity ω at different surfaces
25 July 1966 12 GMT



(a) Thickness pattern for 850-500 mb
layer at 12 GMT on 25 Jul. 1966.

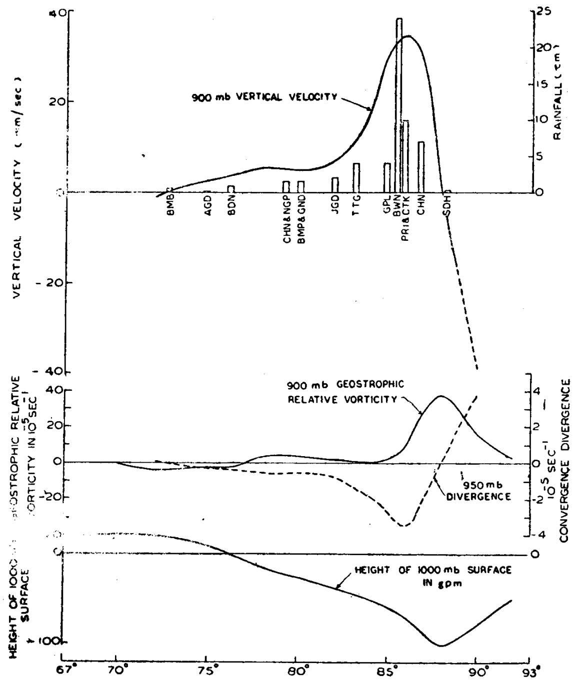


(b) Geostrophic relative vorticity at 700 mb.
surface at 12 GMT on 25 Jul. 1966.

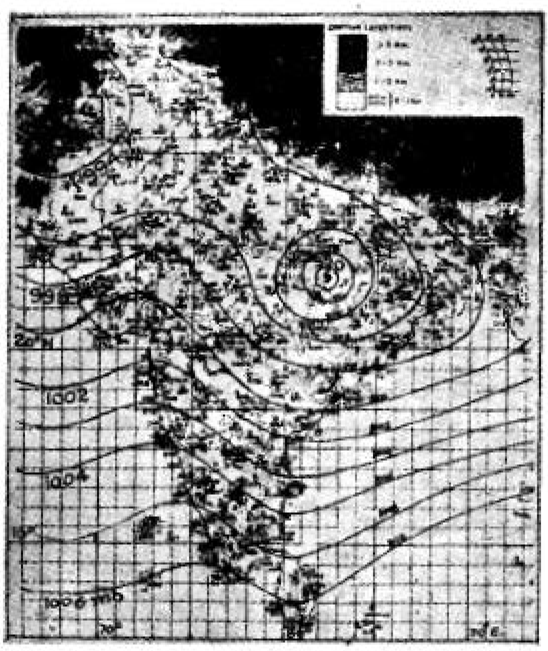


(c) Vertical velocity, divergence and geostrophic
relative vorticity at 18°N, 80°E.

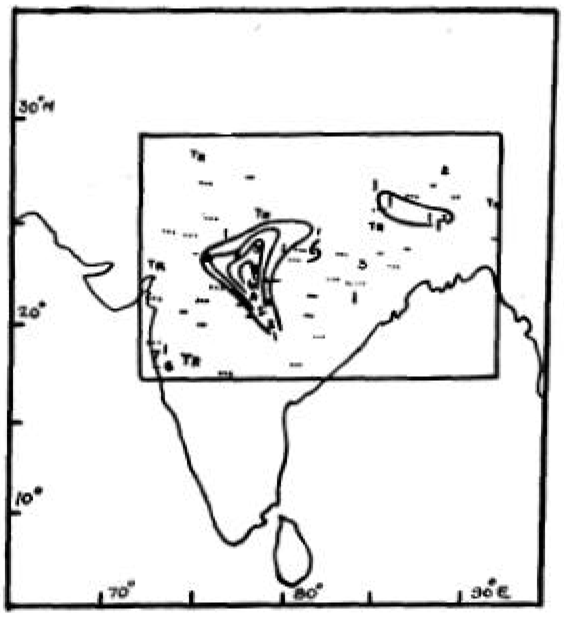
Fig.7.18



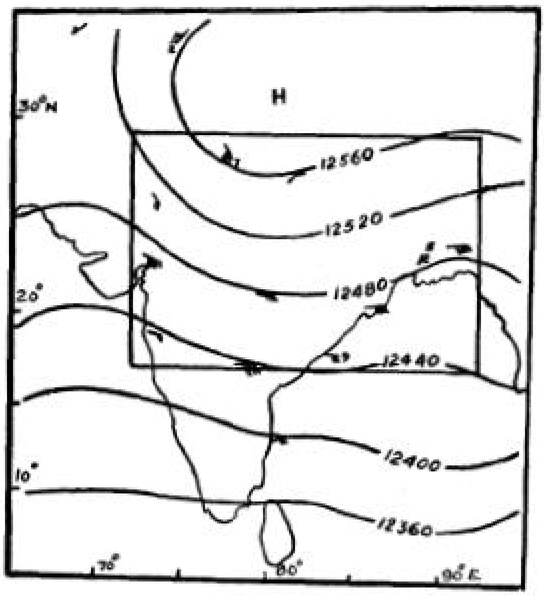
Profiles of contour height, relative vorticity,
divergence, vertical velocity and rainfall along
20° N on 29 Jul. 1969 1200 GMT.
Fig.7.19



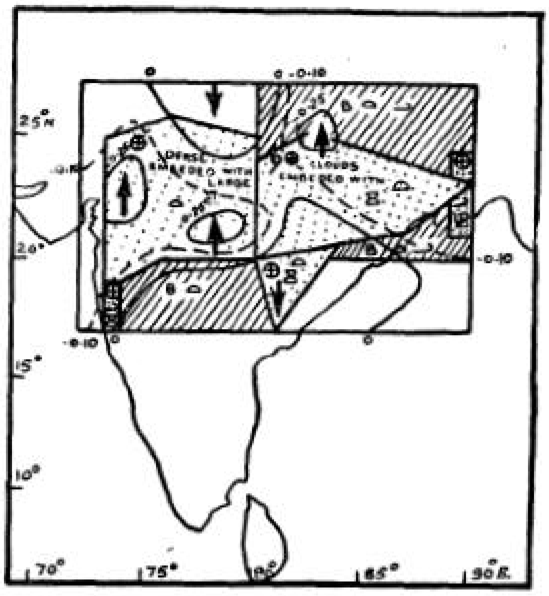
(a) Sea level chart at 00 GMT
on 1 August 1969.



(b) Rainfall distribution as reported
at 1200 GMT on 1 August 1969.

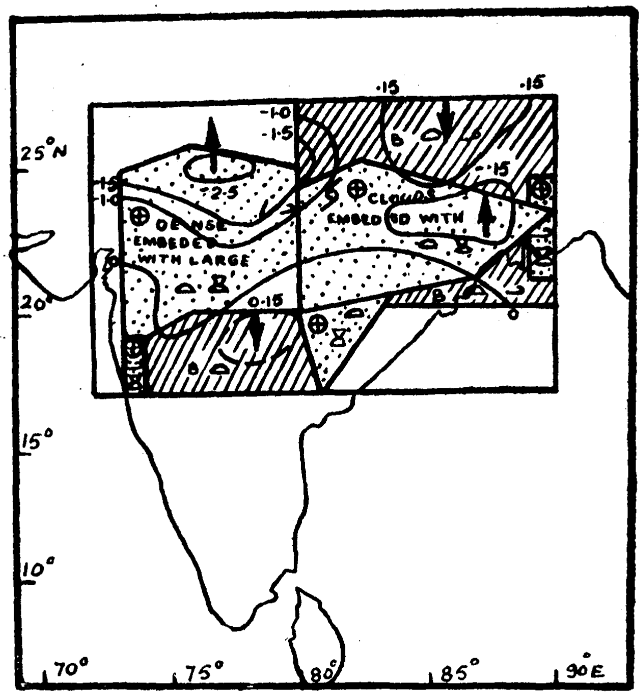


(c) Flow pattern at 200 mb level
at 00 GMT 1 August 1969.

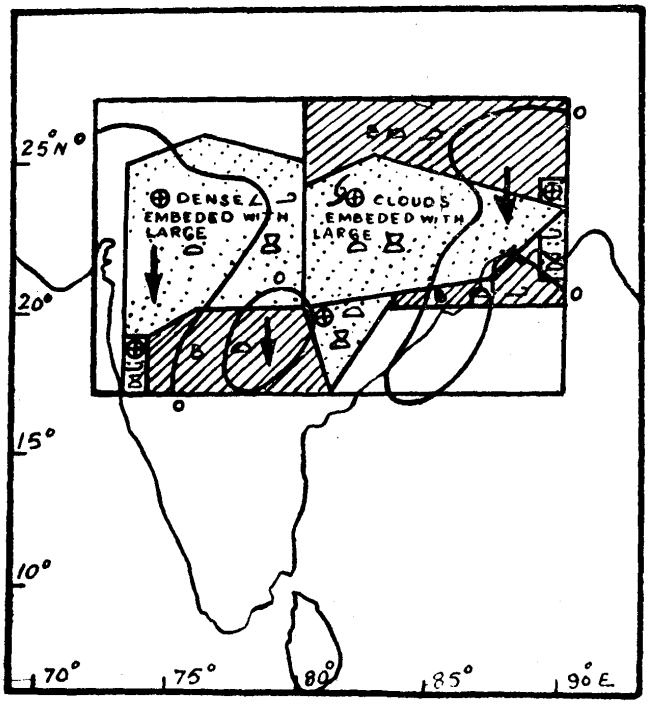


(d) Nephanalysis of 1 August 1969 and
superimposed computed 1000 mb
vertical velocity (Unit 10^{-4} mb/sec),

Fig.7.20



(e) Nephanalysis of 1 August 1969 and superimposed computed 700 mb. vertical velocity (Unit 10^{-4} mb/sec).



(f) Nephanalysis of 1 August 1969 and superimposed computed 200 mb vertical velocity (Unit 10^{-4} mb/sec).

Fig.7.20

7.16 The basic movement of monsoon depressions towards west corresponds to upper tropospheric easterlies. This cannot be a satisfactory explanation when the system is mainly in lower troposphere and is flanked by westerlies to south and easterlies to north. However, in the later part of the monsoon period, depressions recurve in higher latitudes as they get into regions of influence of upper tropospheric westerlies. The movement tends to be more towards west than westnorthwest when the easterlies over north India at 300 mb are stronger. Dynamically satisfactory explanation is to assess the effects of thermal advection and wind divergence at various levels. In the northwest sector there is perhaps warm advection and divergence caused by cyclostrophic effect due to elongation of isobars towards westnorthwest. In the southwest quadrant, in the lower troposphere, cold advection and marked convergence in lower layers are prevalent and the divergence aloft necessary on account of convection, is apparently less than the effect of the former. Thus the usual movement to westnorthwest can be broadly reconciled. At the stage of recurvature, convective belt develops in the new direction of movement and convection seems to achieve the required net mass divergence for movement in that direction. Koteswaram and George (1958) tried to explain the movement on the basis of the Laplacian of thermal advection.

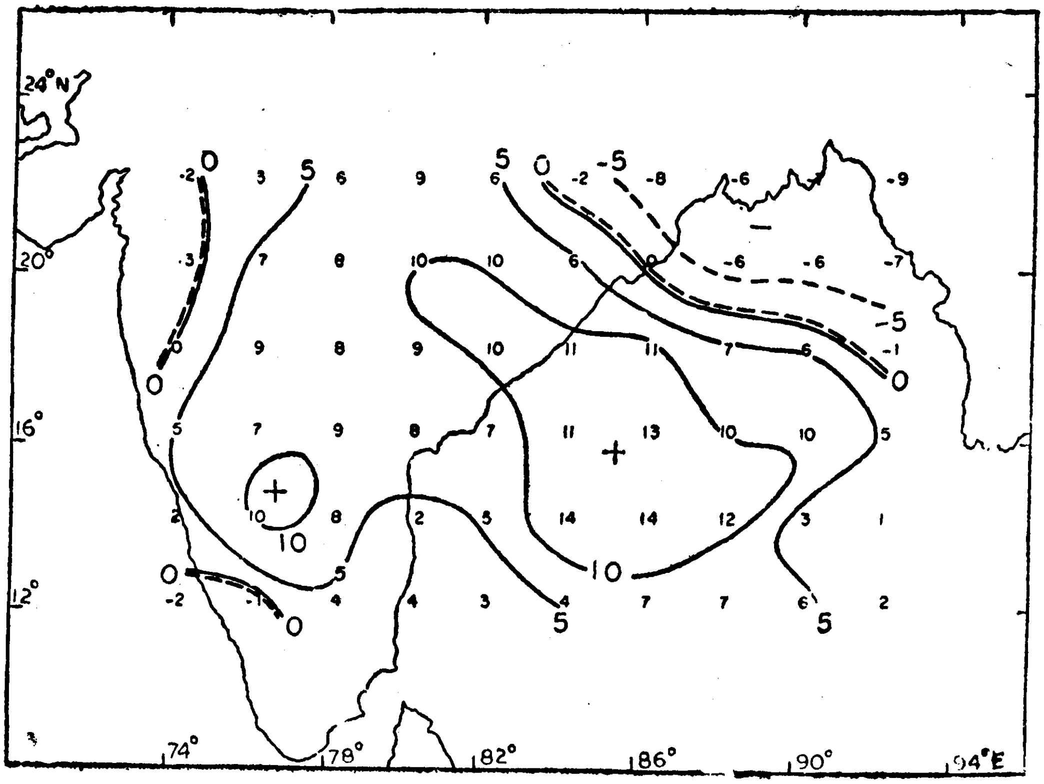
Rao and Rajamani (1970) discussed the movement of monsoon depression from the geostrophic vorticity equation

$$\frac{g}{f_0}\nabla^2\frac{\partial z}{\partial t} = -\vec{n}.\nabla(z + f) + f_0\frac{\partial w}{\partial p}$$

The first term is advection of vorticity and the second contribution of divergence. The westerly current will advect vorticity eastward, south of the centre. Hence divergence has to counteract the effect of advection. From their computations they have computed $\partial z/\partial t$ separately from the two terms and the results are presented in Fig.7.21. In most of the areas, $\partial z/\partial t$ due to the divergence term is opposite in sign to advection. The low of 25/26 July 1966 had only a small northwestward movement and it would not appear that this was towards the areas of maximum fall of contour heights over the Peninsula due to divergence.

7.17 The depression of 5 and 6 August 1964 reported by Rao et al (1970) illustrates many features discussed earlier. Figs, 7.22(a to h)show the sequence of charts. On 3rd August, while lower tropospheric winds over north India were westerly, monsoon strengthened in central Bay where ships reported heavy rain and squalls. Lower and mid–tropospheric westerlies strengthened between 12° N and 17° N and there was considerable cyclonic wind shear in lower troposphere upto 700 mb and a cyclonic circulation at 500 mb between 15^o N and 20^o N. Strengthening of lower tropospheric westerlies to develop cyclonic shear and associated development of heavy rain and squalls in Bay are precursors to formation of depression. On 4th, a low formed over Bay at Lat.19^o N but westerlies were prevailing from west Uttar Pradesh to north Assam in lower and mid–troposphere. A depression (with two closed isobars) developed by 5th, centred at 20^o N, 87^o E. Ridge about ten degrees west of this centre is quite marked. A rain belt in the flow from the ridge to the trough, at the periphery of the depression, can be seen. This may be explained as convergence due to super–geostrophic winds in the ridge and sub–geostrophic flow in the trough. The deep depression on 6th had caused widespread rain to northwest also, not uncommon at this stage. Rain–belt is extending from northwest of centre to southwest. Here convergence is due to increasing cyclostrophic correction to geostrophic flow due to increasing speed downwind. The system weakened the next day.

Formation of depression in Bay due to the movement of a low across Burma is illustrated (Figs. 7.23 a to h) by the depression of 31st July to 4th August, 1967, also reported by Rao et al (1970). On 29th July, pressures were falling over Burma by 1–2 mb and many stations there reported good amounts of rainfall. In the upper a cyclonic circulation was over Bangladesh,



(a)

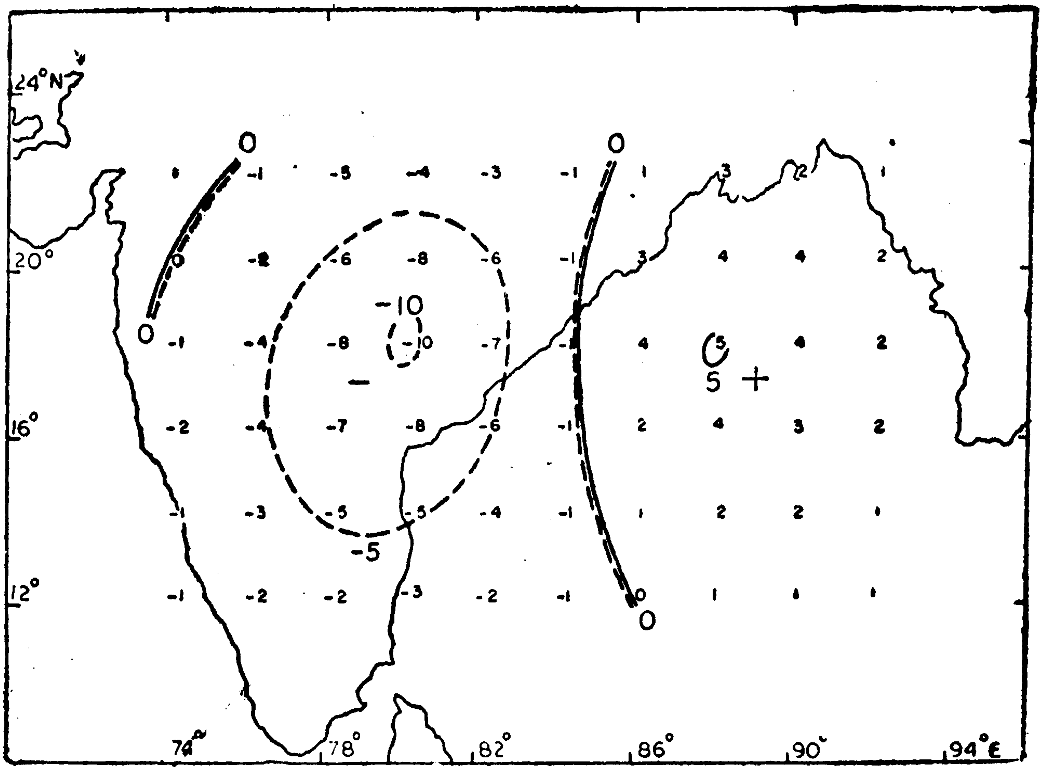


Fig.7.21 - 24 hr. height change (gpm) of 850 mb surface due to (a) vorticity advection and (b) divergence terms.

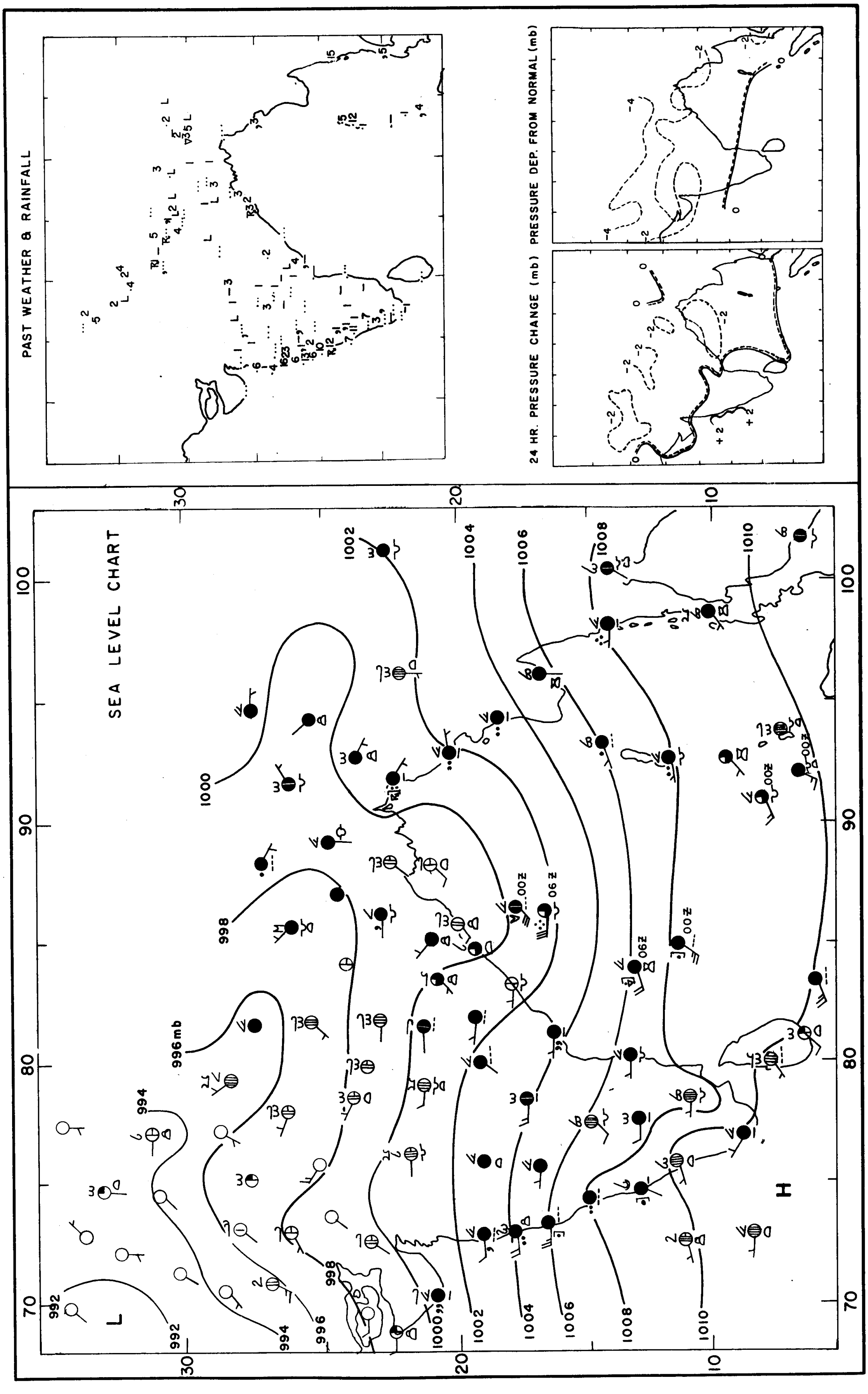
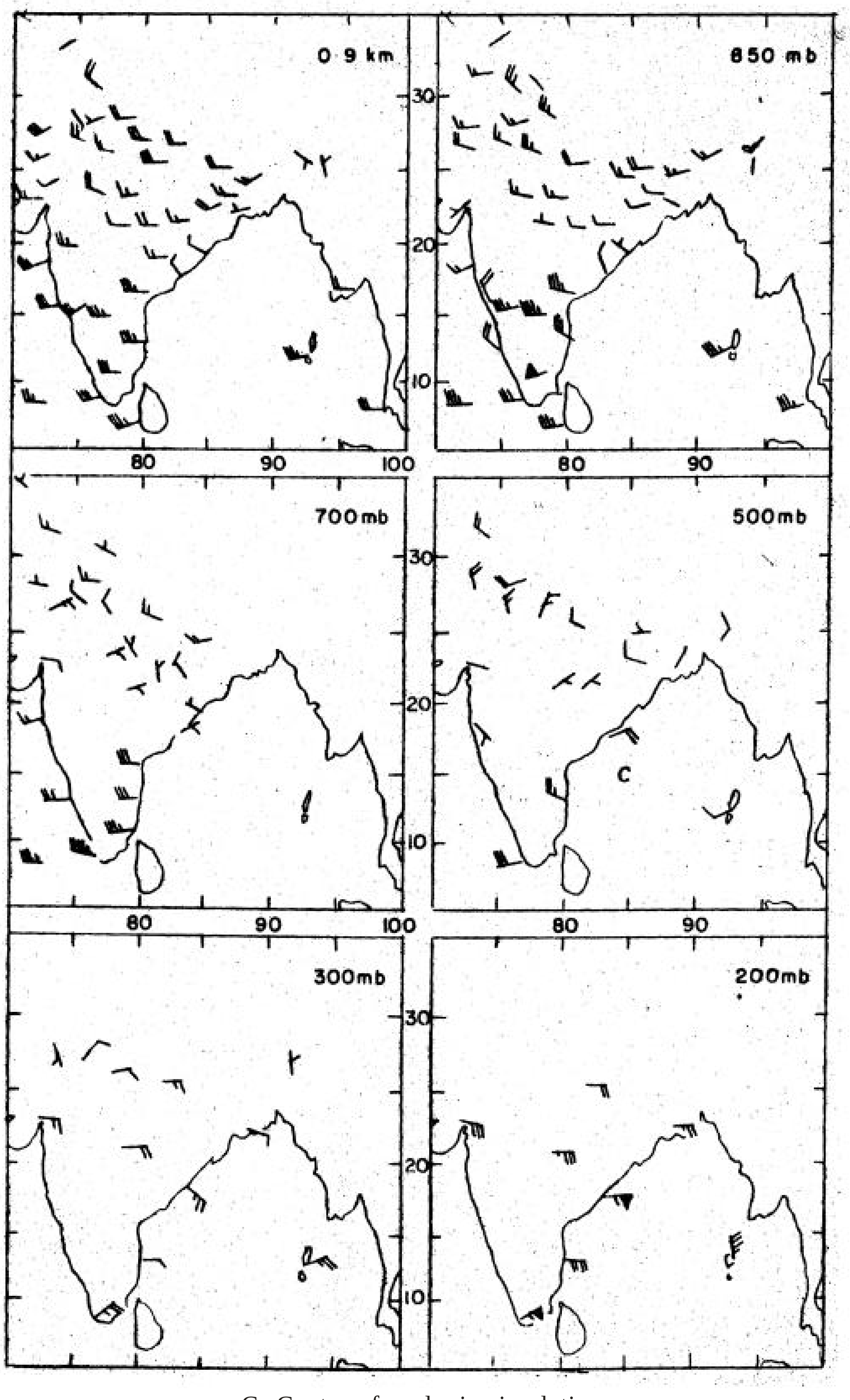


Fig. 7.22(a) Synoptic charts 0300 GMT 3 August 1964



C- Centre of cyclonic circulation

Fig. 7.22(b) Upper Winds 3 Aug. 1964 00GMT

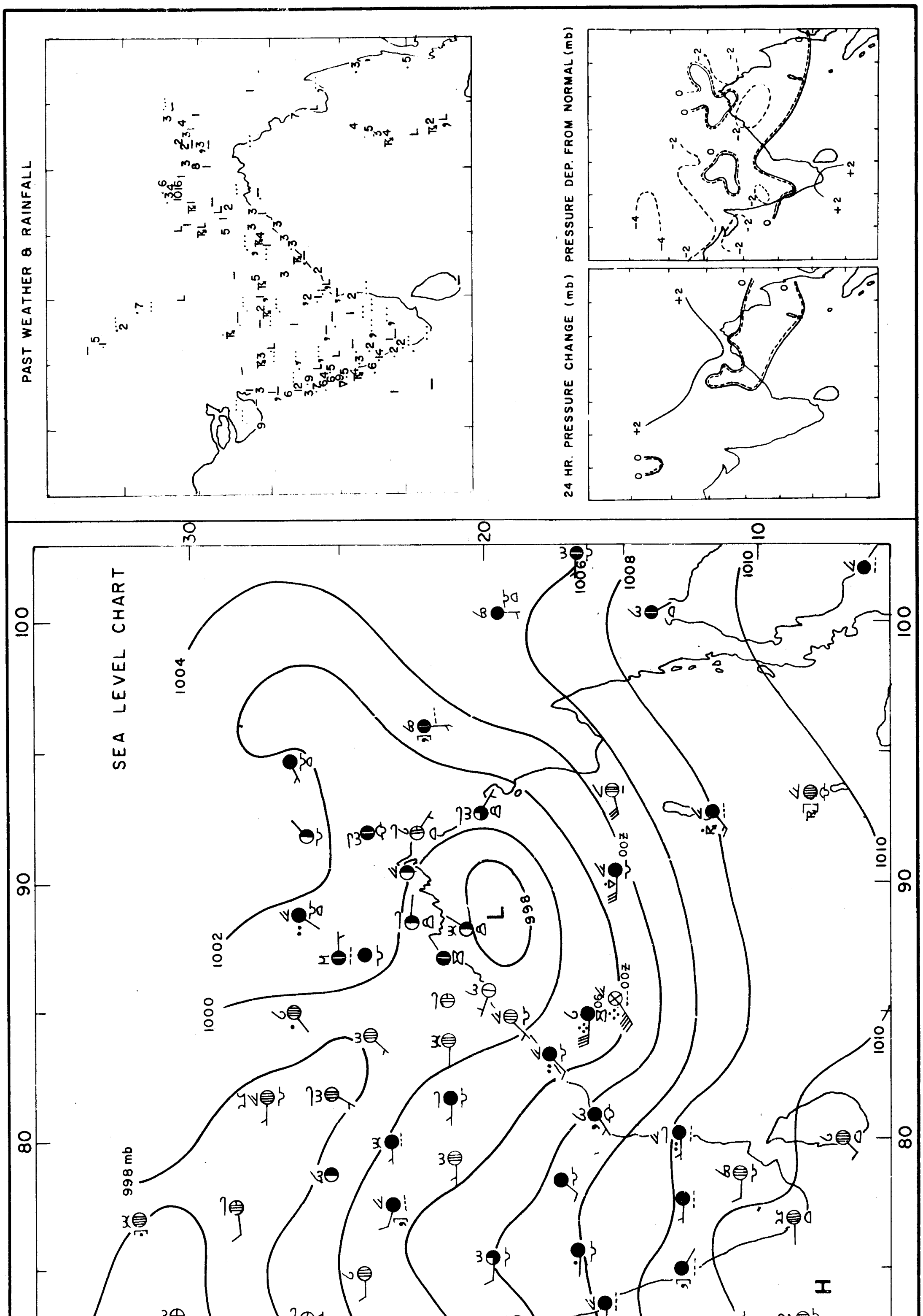
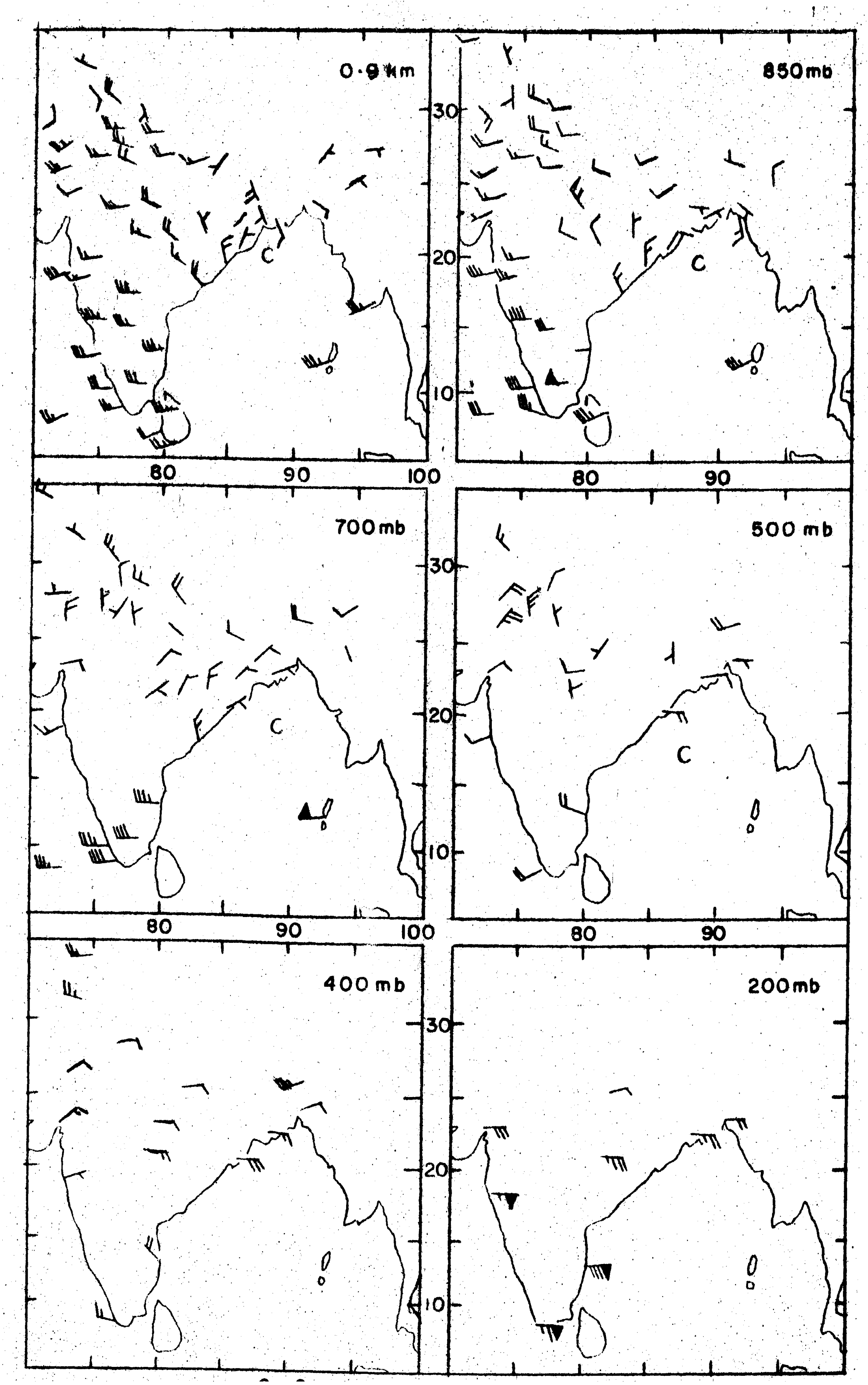


Fig. 7.22(c) Synoptic charts 0300 GMT 4 August 1964



C - Centre of Cyclonic Circulation
Fig. 7.22(d) Upper Winds 4 Aug. 1964 00 GMT

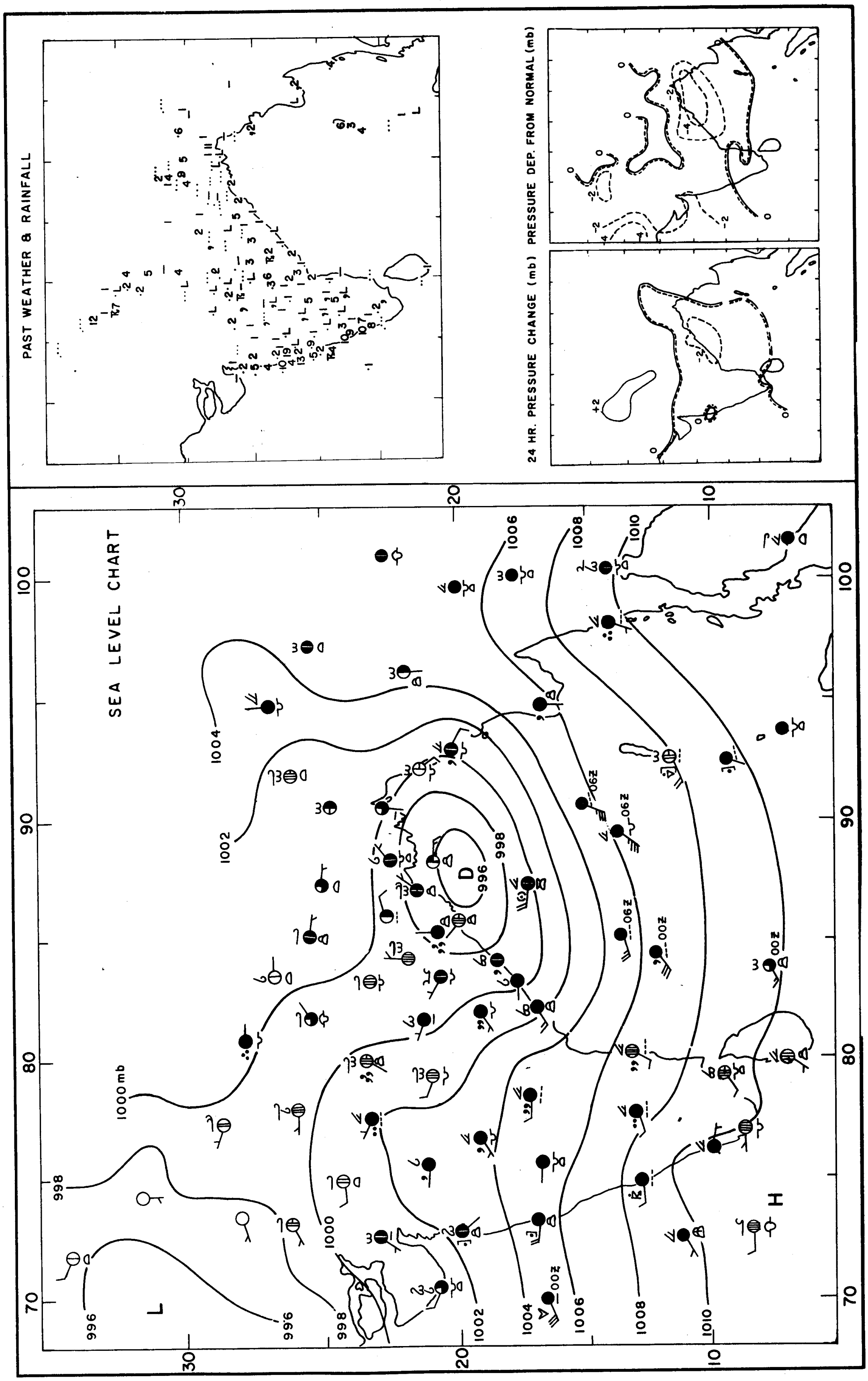


Fig. 7.22(e) Synoptic charts 0300 GMT 5 August 1964

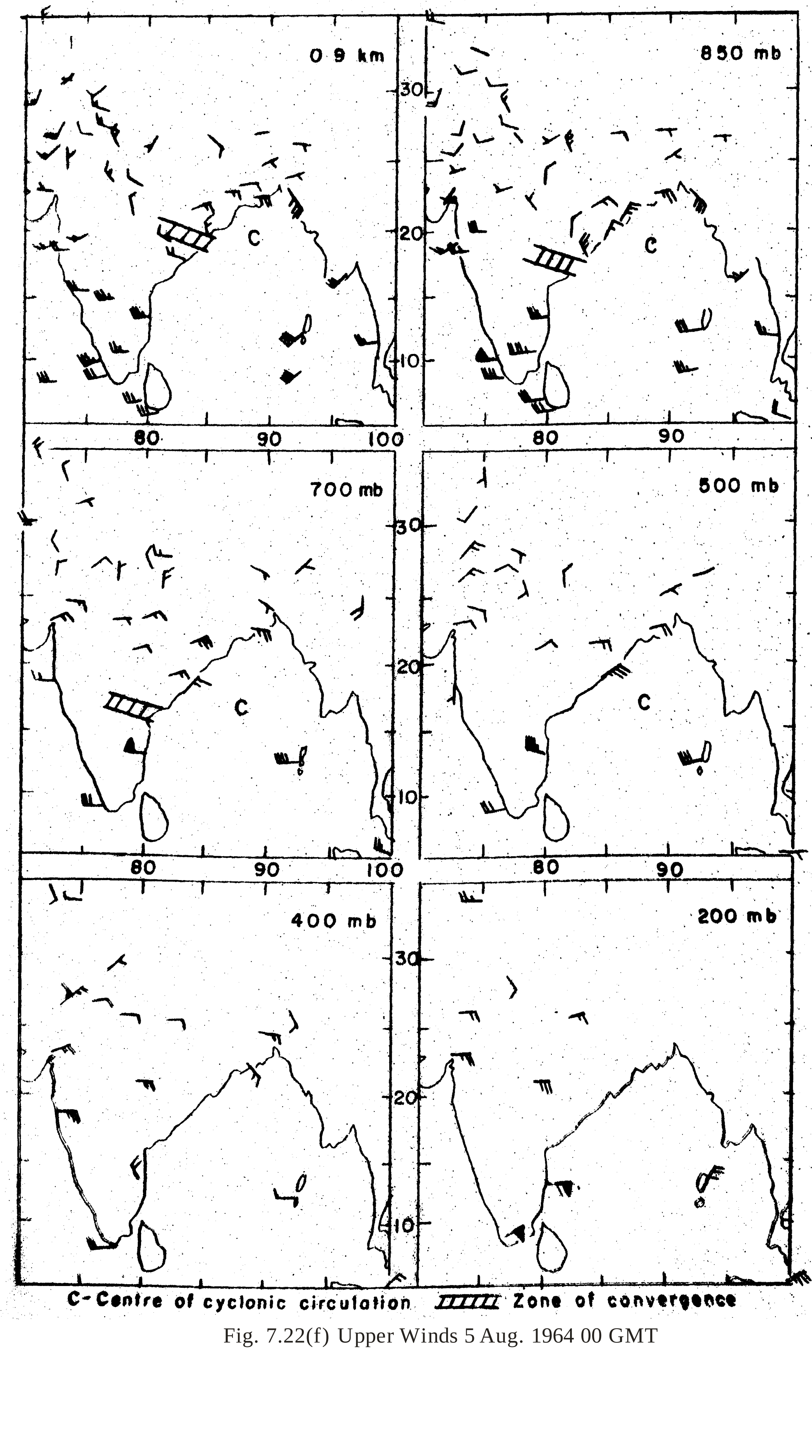


Fig. 7.22(f) Upper Winds 5 Aug. 1964 00 GMT

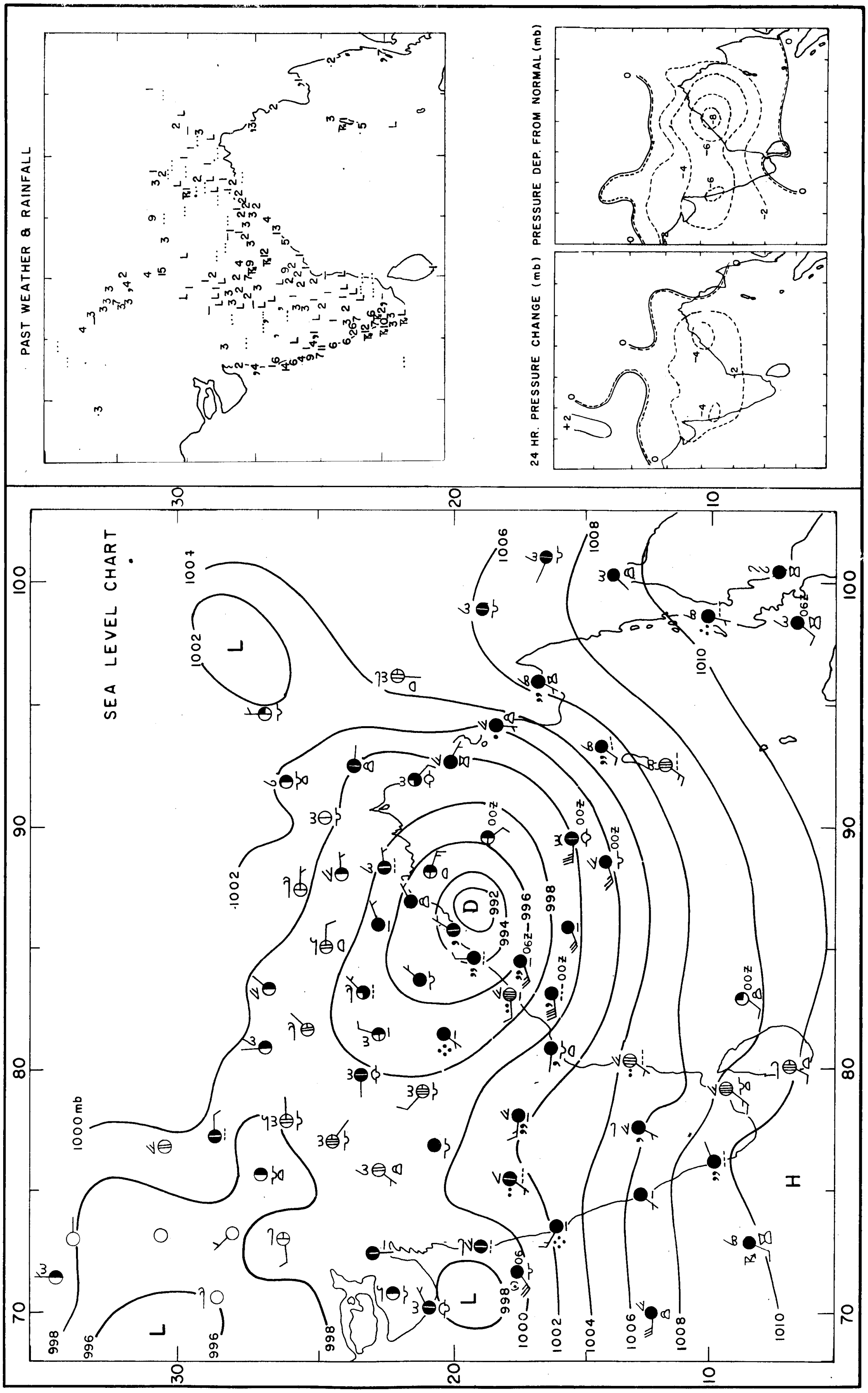
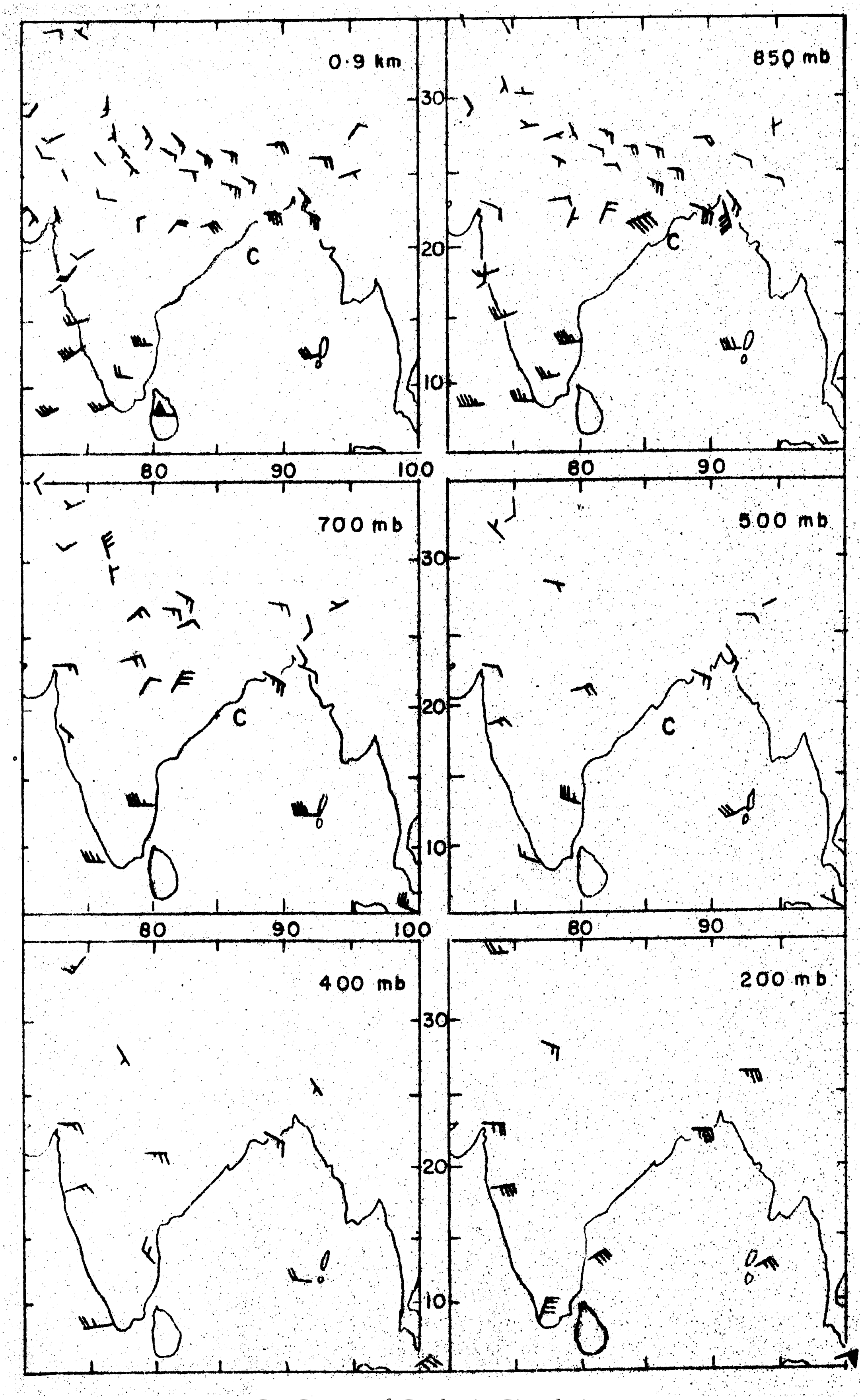


Fig. 7.22(g) Synoptic charts 0300 GMT 6 August 1964



C - Centre of Cyclonic Circulation
Fig. 7.22(h) Upper Winds 6 Aug. 1964 00 GMT

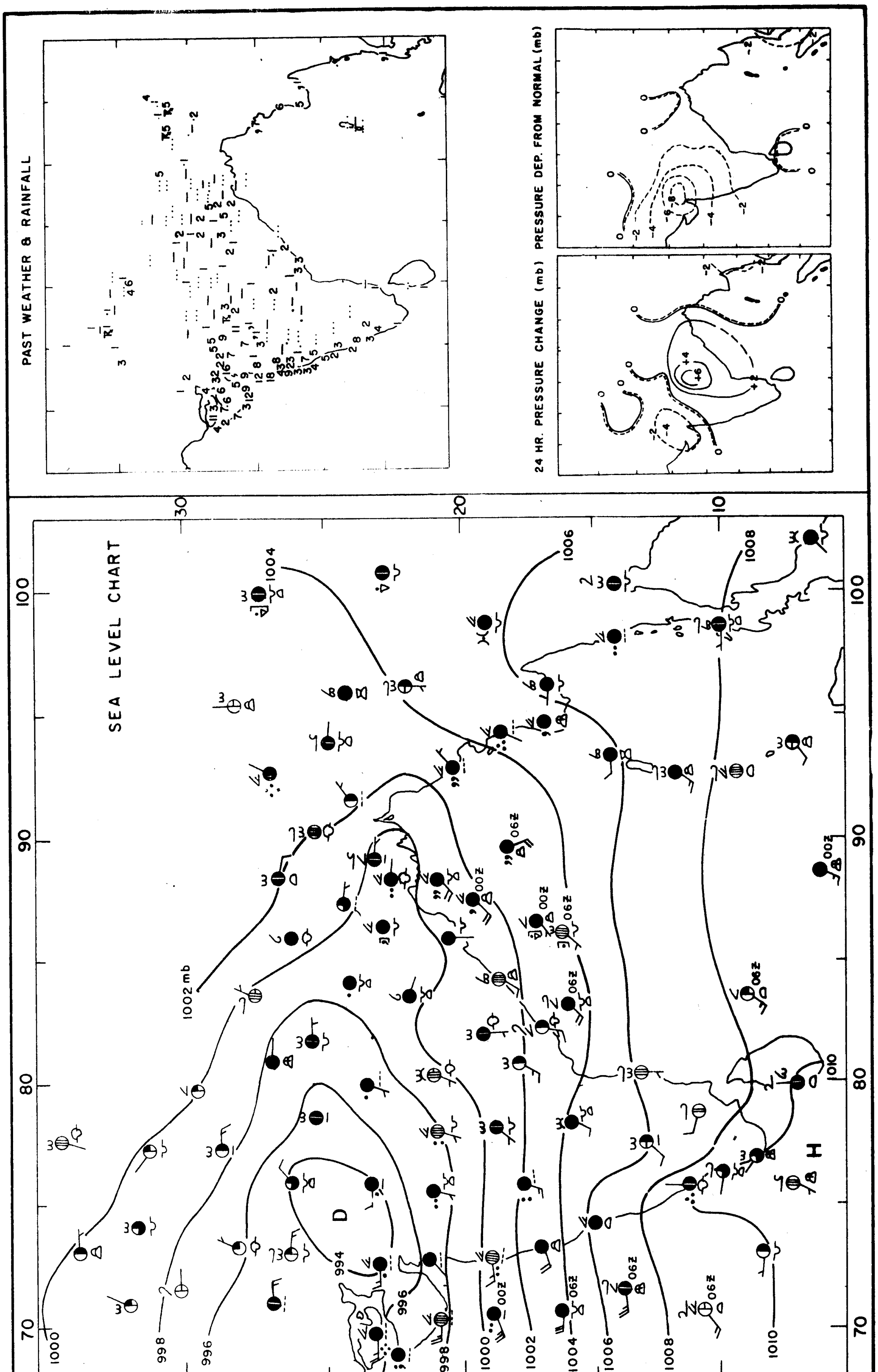
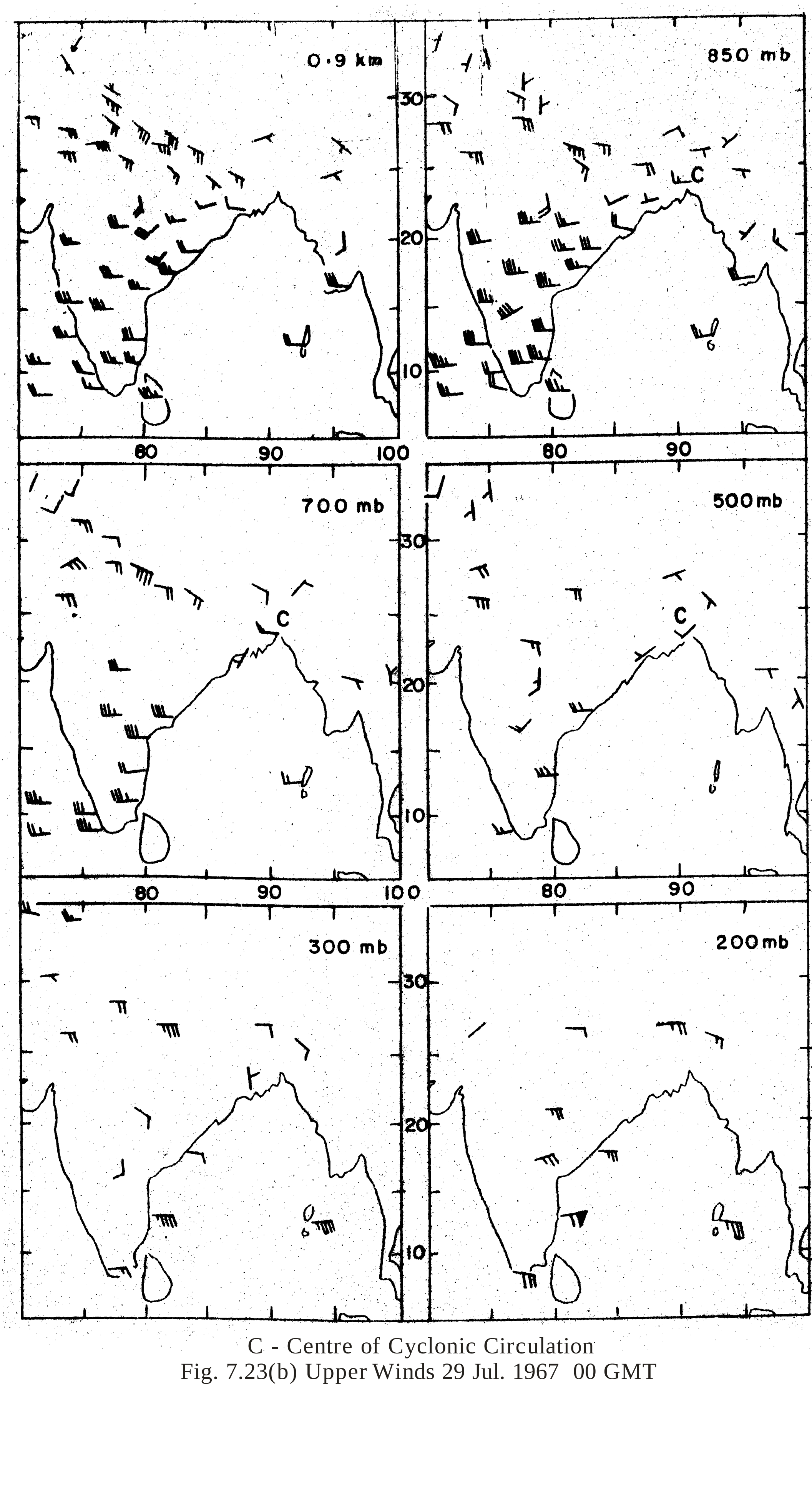


Fig. 7.23(a) Synoptic charts 0300 GMT 29 July 1967



C - Centre of Cyclonic Circulation
Fig. 7.23(b) Upper Winds 29 Jul. 1967 00 GMT

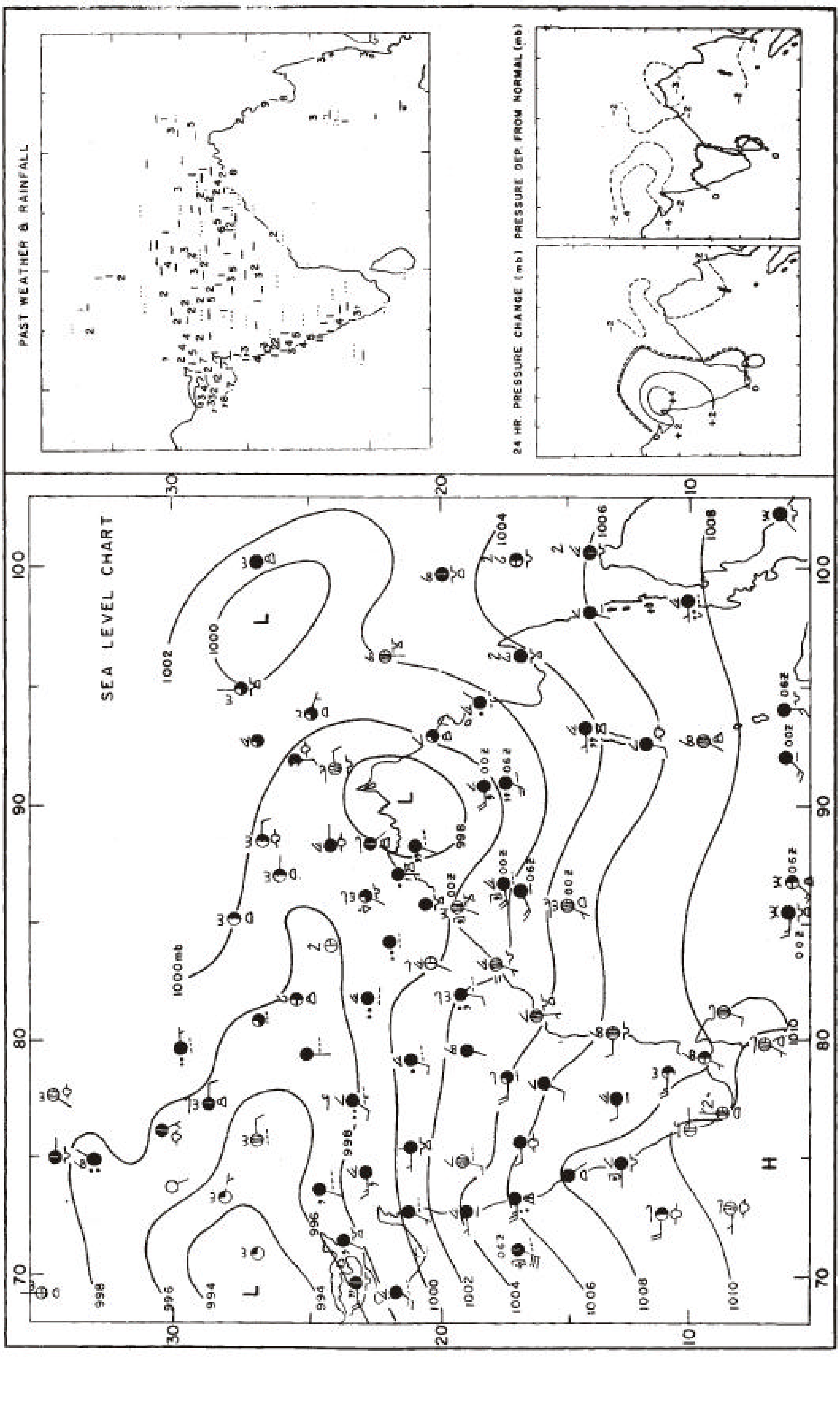


Fig. 7.23(c) Synoptic charts 0300 GMT 30 July 1967

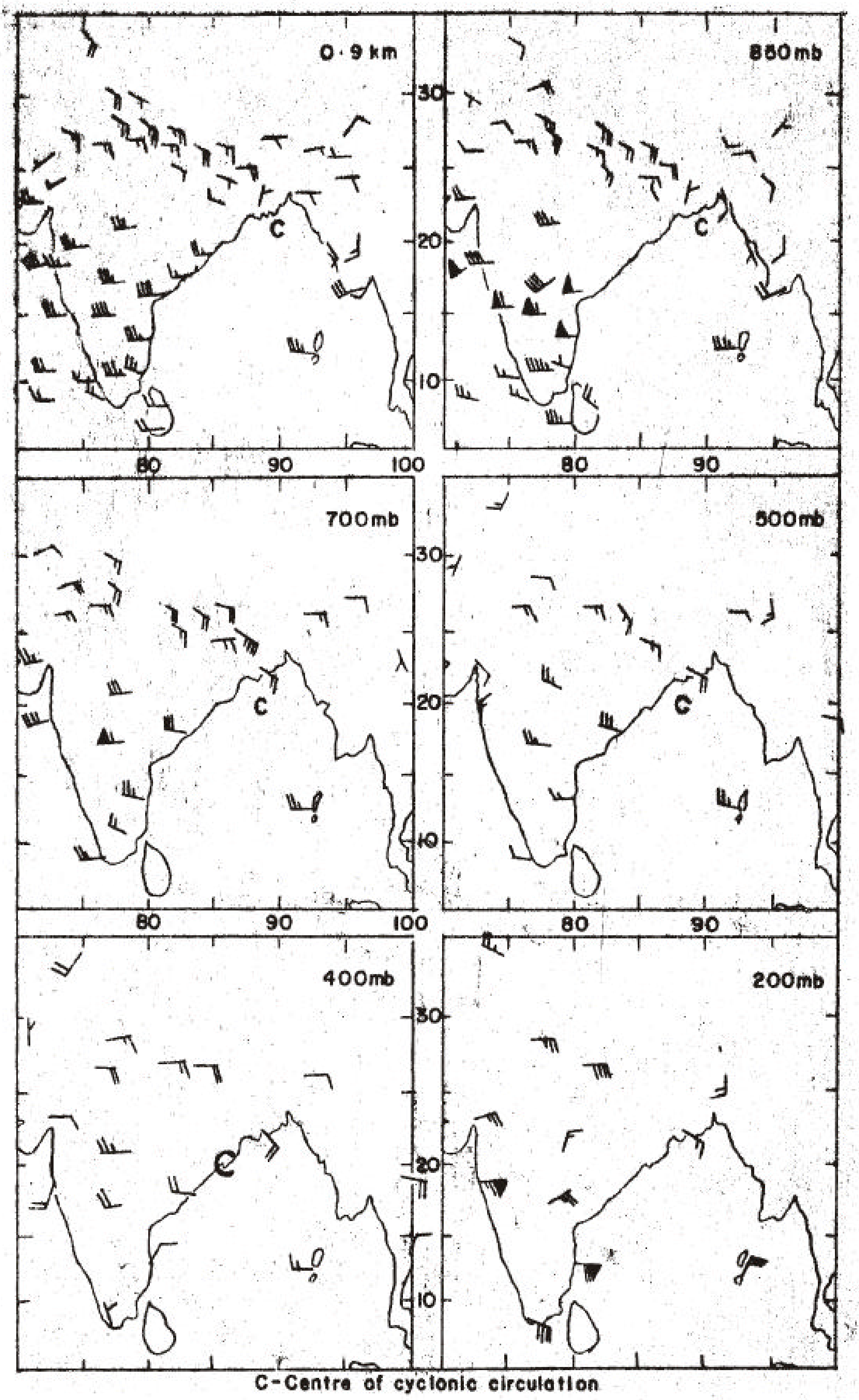


Fig. 7.23(d) Upper Winds 30 Jul. 1967 00 GMT

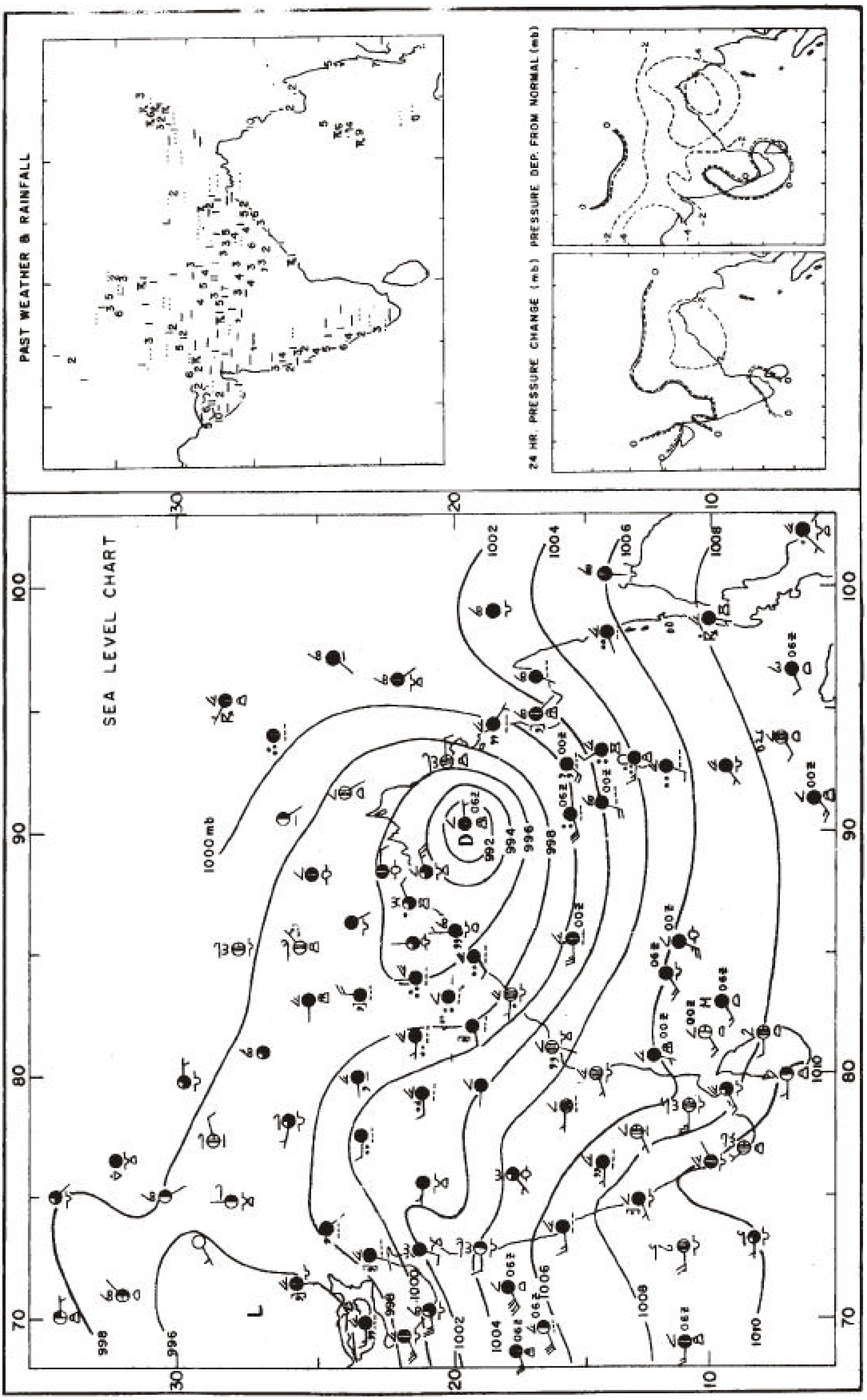


Fig. 7.23(e) Synoptic charts 0300 GMT 31 July 1967

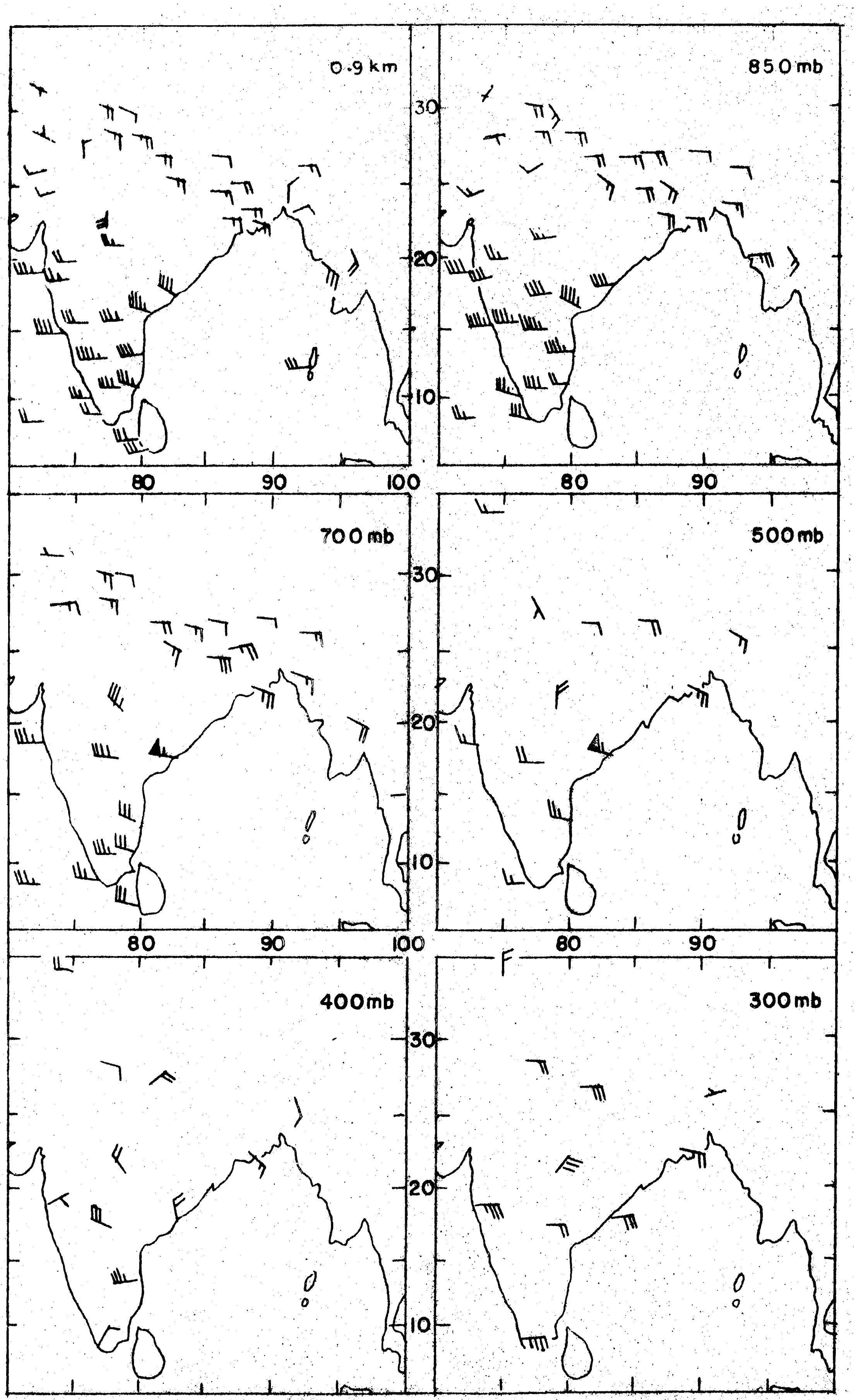


Fig. 7.23(f) Upper Winds 31 Jul. 1967 00 GMT

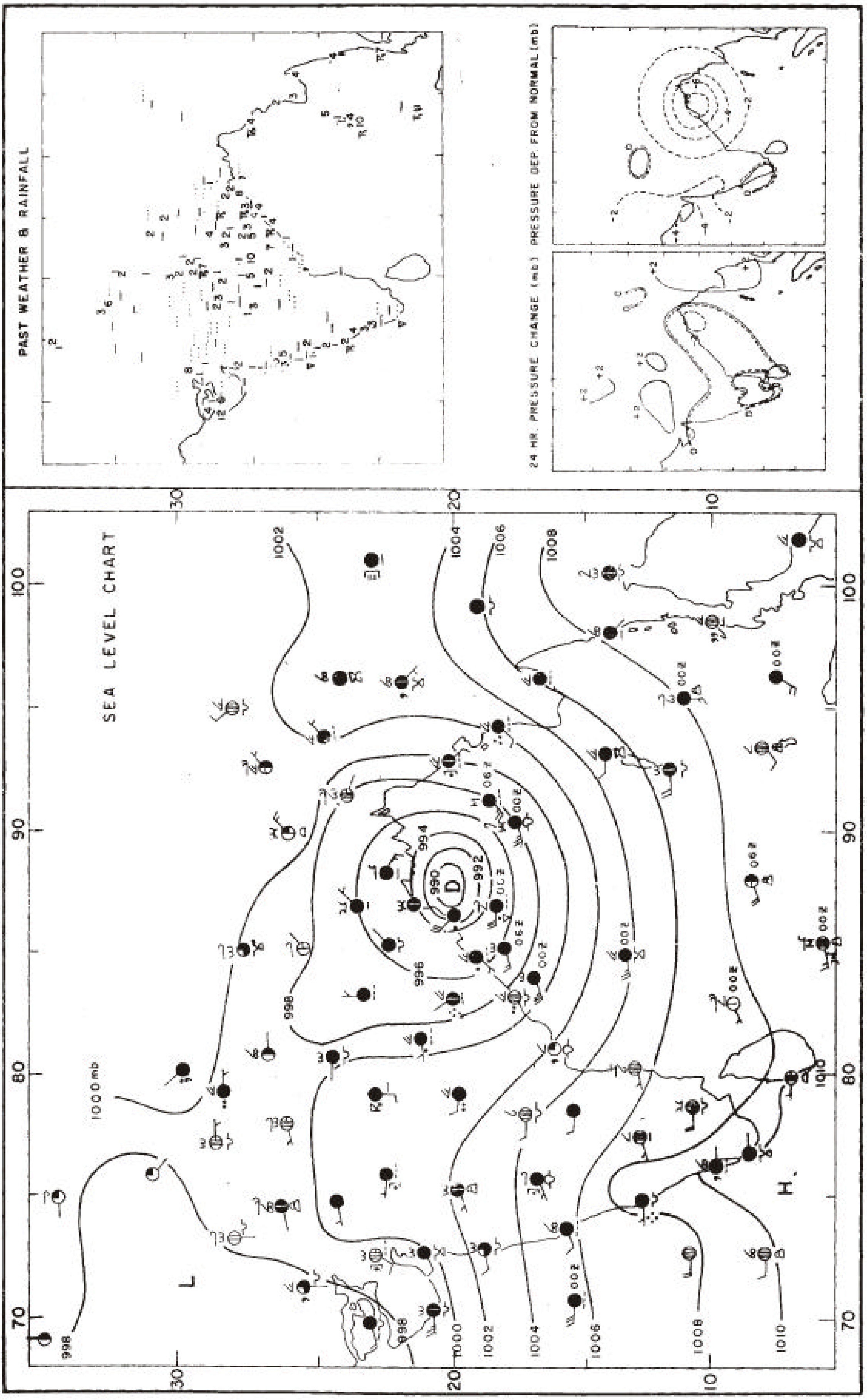


Fig. 7.23(g) Synoptic charts 0300 GMT 1 August 1967

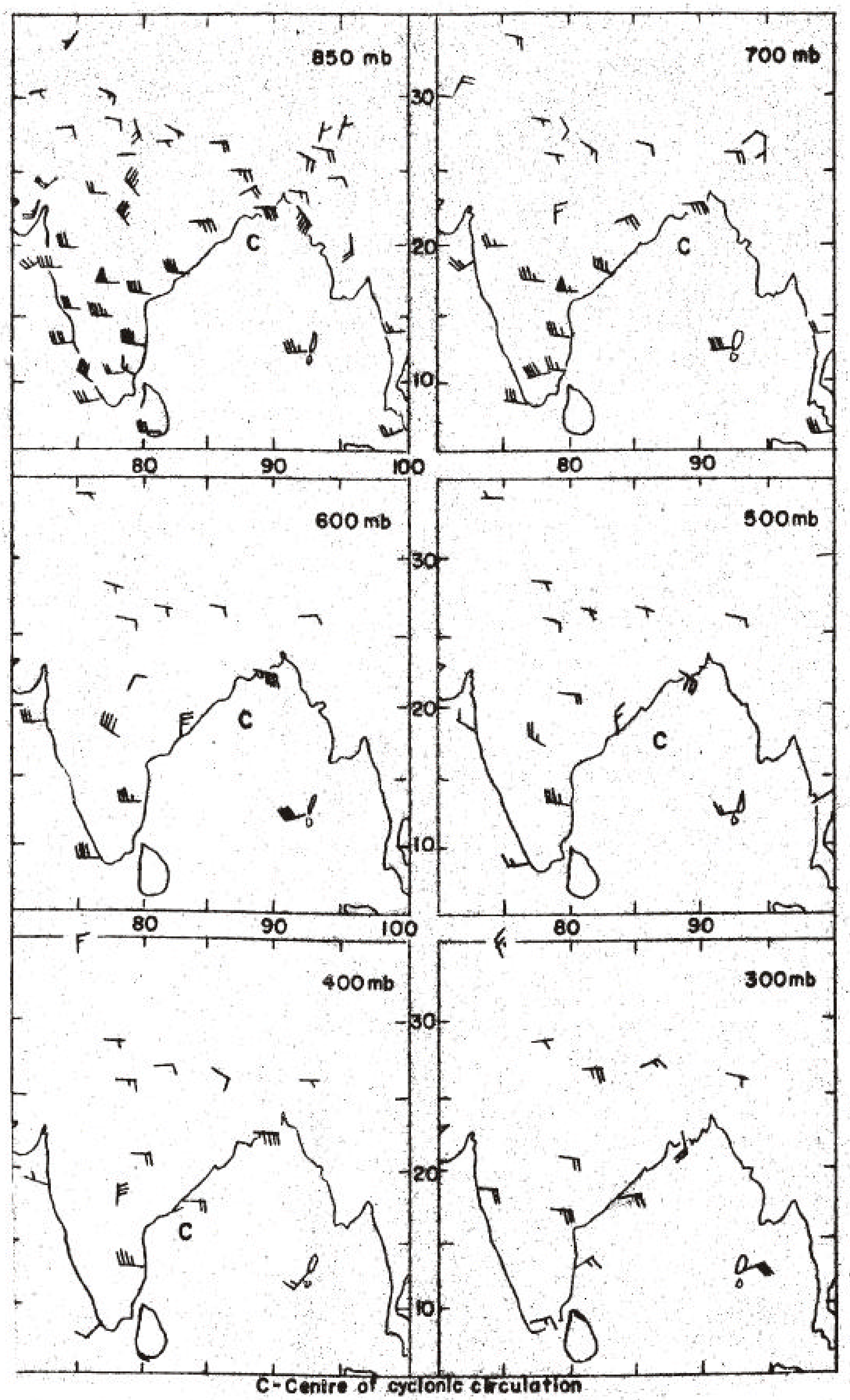


Fig. 7.23(h) Upper Winds 1 Aug. 1967 00 GMT

northeast Bay and adjoining Burma between 850 mb and 500 mb. By evening of 29th, pressure changes over north Burma and adjoining Bangladesh were –3 to –4 mb. A low formed over north Bay on 30th July which concentrated into depression on 31st morning and became deep by evening. The extensive rain belt from the ridge 12° from centre is interesting and the probable cause has been discussed in a previous para. The depression crossed coast on the night of 1st August and weakened.

In both the above cases, the depressions did not move much inland. The deep depression of 25 to 30 July 1967 (Figs. 7.24 a to j) travelled in a more or less westerly direction upto 20th and reached upto the border of Gujarat (Srinivasan, Raman and Mukherji, 1971). A low over northwest Bay on 25th developed into depression on 26th. Fig. 7.24 (a) of this day shows the elongation of axis in the northwest direction, greater pressure gradient to south of centre and a ridge about 13° to west. Two rainfall areas are seen. One is in the southwest sector and the other in the flow from the ridge to the trough at the periphery. Satellite picture (7.24 c) shows the extensive cloudy areas in the depression field, over central parts of the country and in the east coast, The decrease in clouds off the west coast is interesting. On 27th when the centre was 150 km inland, the system seemed to have become a trough at 300 mb. An interesting point of 28th is the change in the winds over Nagpur. from W/l5 kt at 700 mb to ESE / l0 kt at 500 mb, showing southwestward tilt with height. This circulation continued at 300 mb also. The satellite picture (7.24 h) showed Cu bands to the north and heavy overcast to the south of the depression. The increase in monsoon rains between 15^o N and 20^o N along the coast is also an usual feature. Monsoon strengthens not infrequently in the same belt when a depression is forming in northwest Bay. The depressions moved northwest and lay near Ratlam on 29th and weakened by next day.

The 18th to 24th September, 1969 depression (Fig. 25a to m) brings out the differences from the cases of earlier months (Srinivasan, Raman and Mukherji, 197l). Rainbelt far from the centre in the flow from the ridge to the trough of the periphery is noticed. The wind over Visakhapatnam (Fig. 7.25b) changes from southwest at 850 mb to eastnortheast at 500 mb showing southwestward tilt with height. That station had lowest temperatures in lower troposphere. Moving westnorthwest, the depression was 100 km to the east of Jagdalpur on 19th (Fig. 7.25 d). Besides the rain in the southwest quadrant, both 24–hour rainfall and present weather are noticed toward northeast of the depression. A middle latitude westerly trough (Fig.7.25d) was located between 65° E and 72° E, north of 30° N. Upto the next day, the movement was only about 10 km hr^{-1} to northwest. The rain belt on 20th (Fig. 7.25e) is all around, particularly marked to northeast, a little away from the centre. Even in the Satellite picture (Fig. 7.25g), closer to the centre in the northeast, clouding is less but the patch a bit away is quite marked. Movement was still slower upto 21st, hardly 100 km. Rainbelt to north (Fig. 7.25h) is quite prominent. The middle latitude trough (Fig. 7.25i) is also 75° E at 500 mb, and is seen upto 27° N. The depression was between Nowgong and Jabalpur on 22nd, having moved northward and faster than on the previous day. Rainfall is seen for only 150 km all around (Fig.7.25j), but more extensive to the north. The increase in rain towards north in both intensity and area is the effect of northward movement and the effect of westerly trough which was along 80° E at 500 mb (Fig.7.25k). The Satellite picture (Fig. 7.25l) shows the linking up of the clouds of the westerly trough and the monsoon depression. The depression moved north and weakened by 24th. This case brings out the changes in course and rainfall pattern under the influence of the westerly trough and the smaller extent and more uniform distribution of rainfall in September depression.

7.18 Monsoon depressions have a profound influence on the rainfall over most of the country, much more than their numbers would suggest. Monsoon activity in one part or the other is influenced by these systems, starting from the precursors to even during the weakening stage of low. Their travel across a large tract distributes rainfall over a wide area. Such effect of monsoon depressions prevails on the average for about one third of days in the period of the monsoon. The need to gather observations from a close net work of radiosonde/rawin stations in their field,

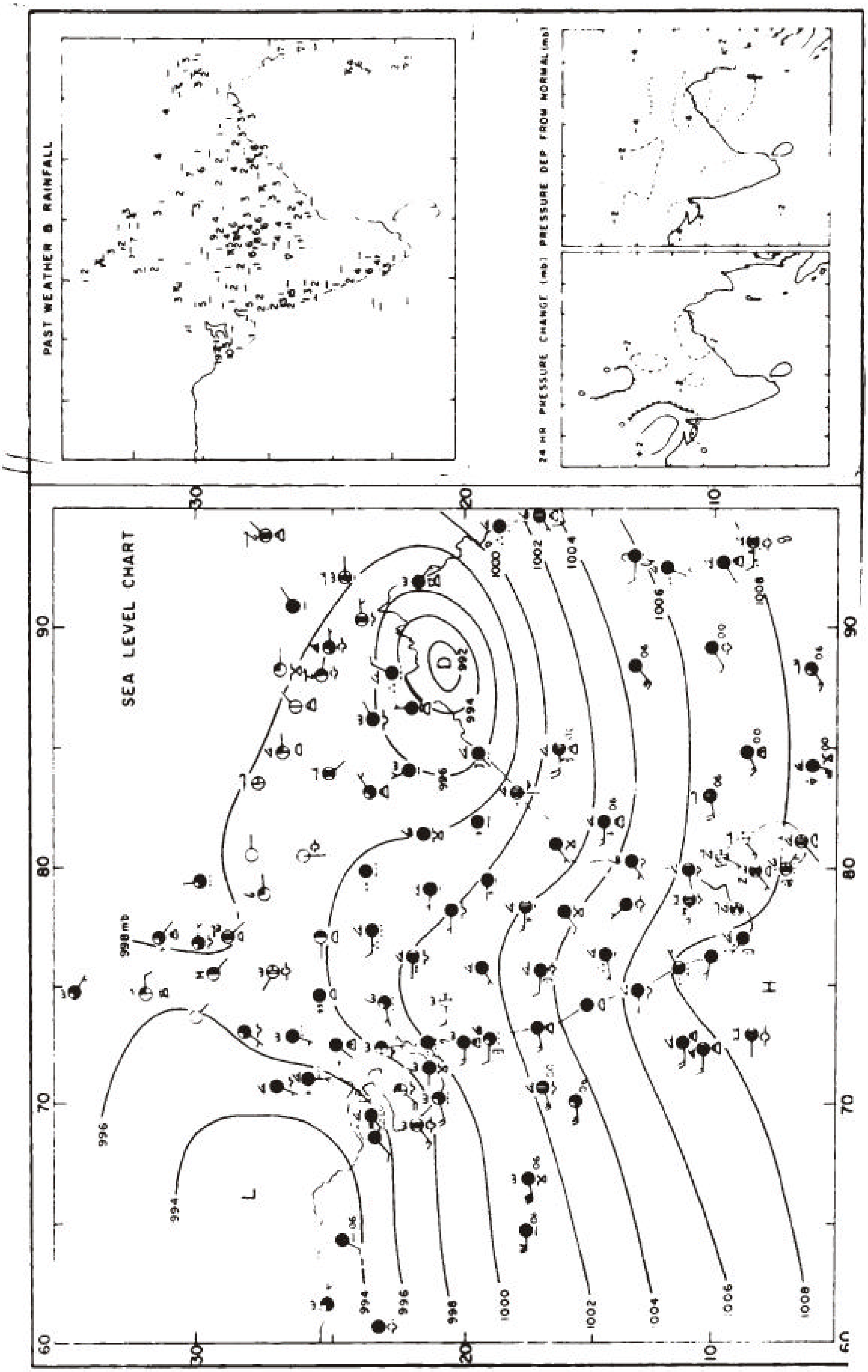


Fig. 7.24(a) Synoptic charts 0300 GMT 26 July 1967

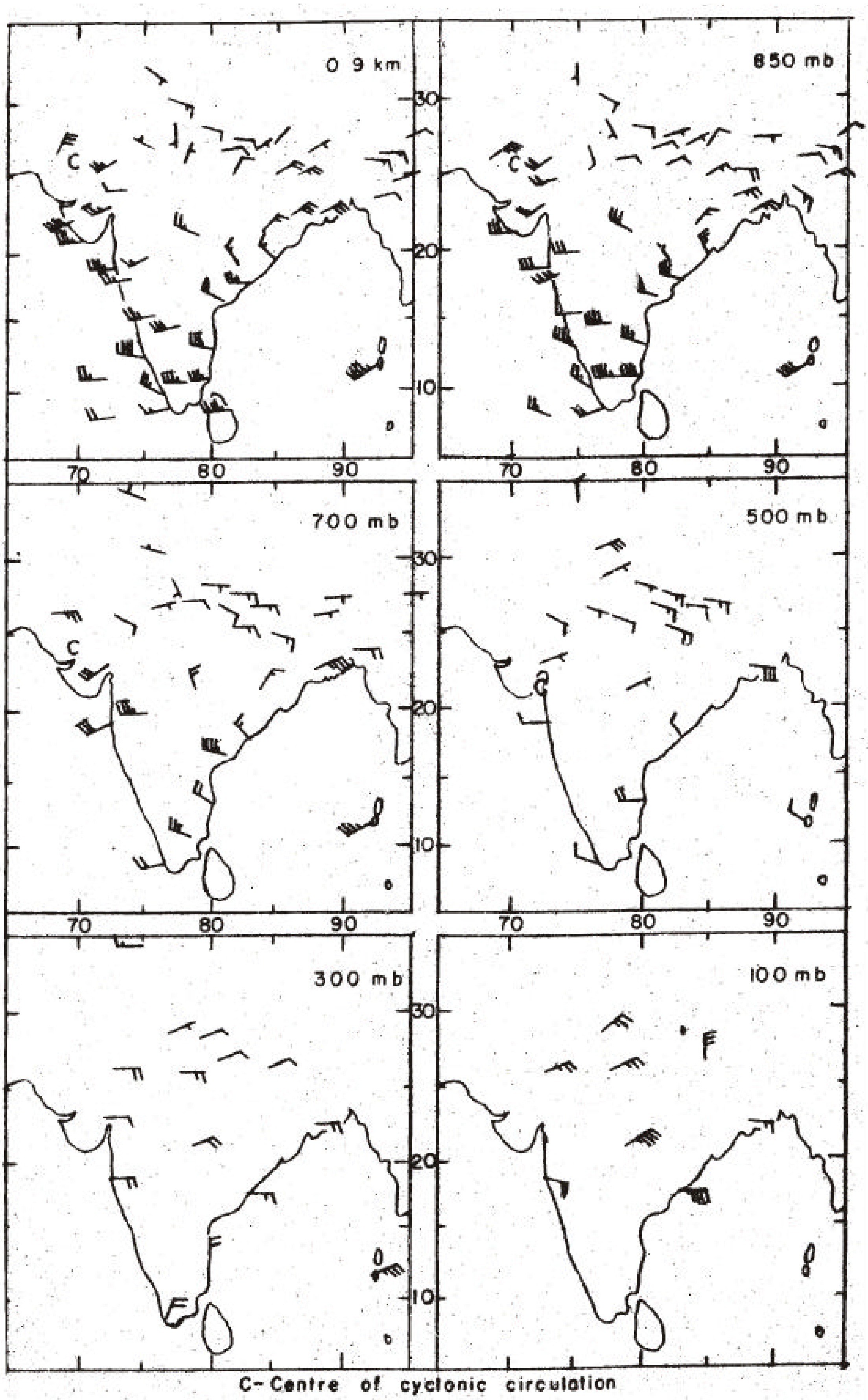


Fig. 7.24(b) Upper Winds 26 Jul. 1967 00 GMT

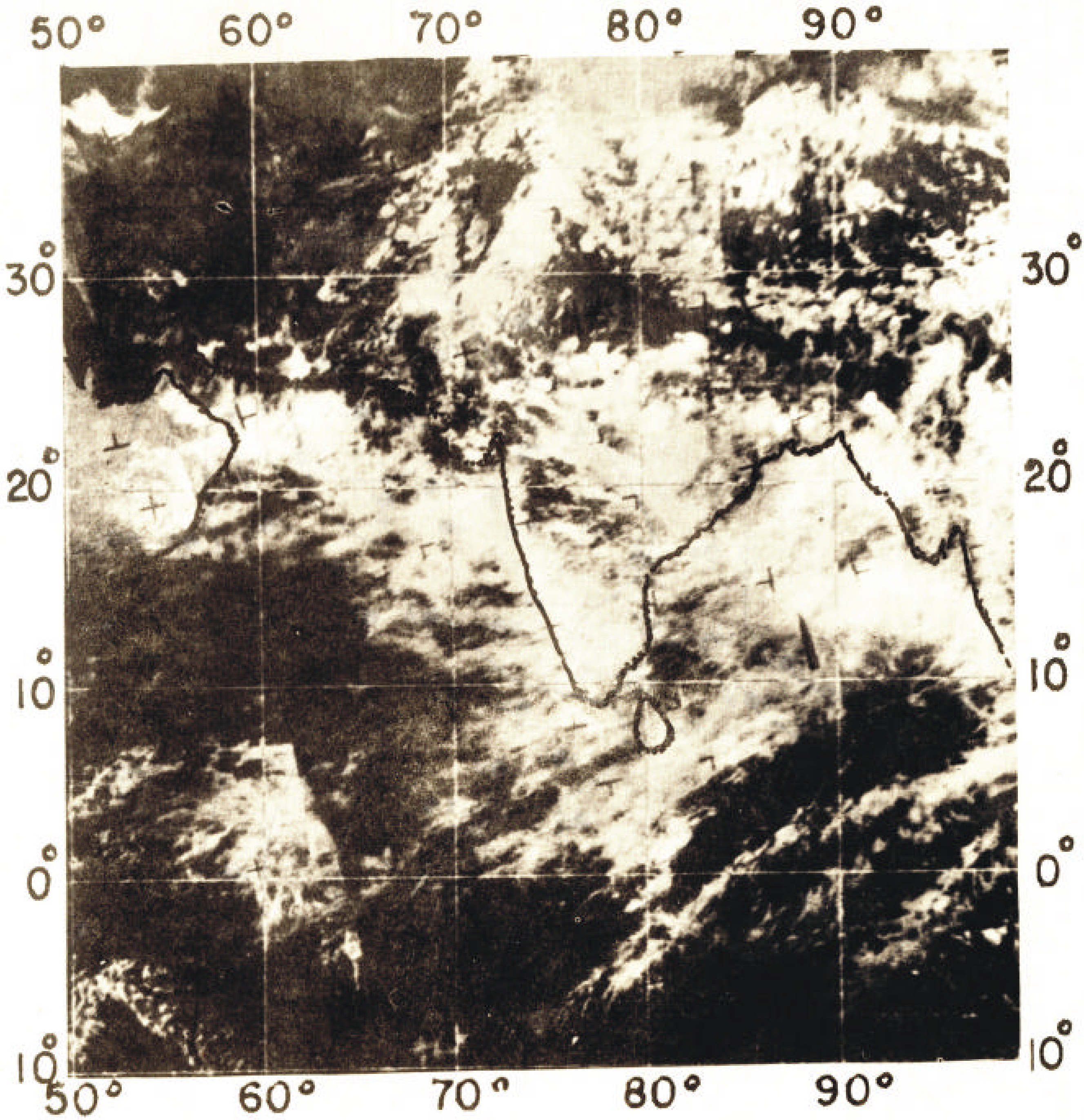


Fig. 7.24 (c) ESSA-5 26 July 1967

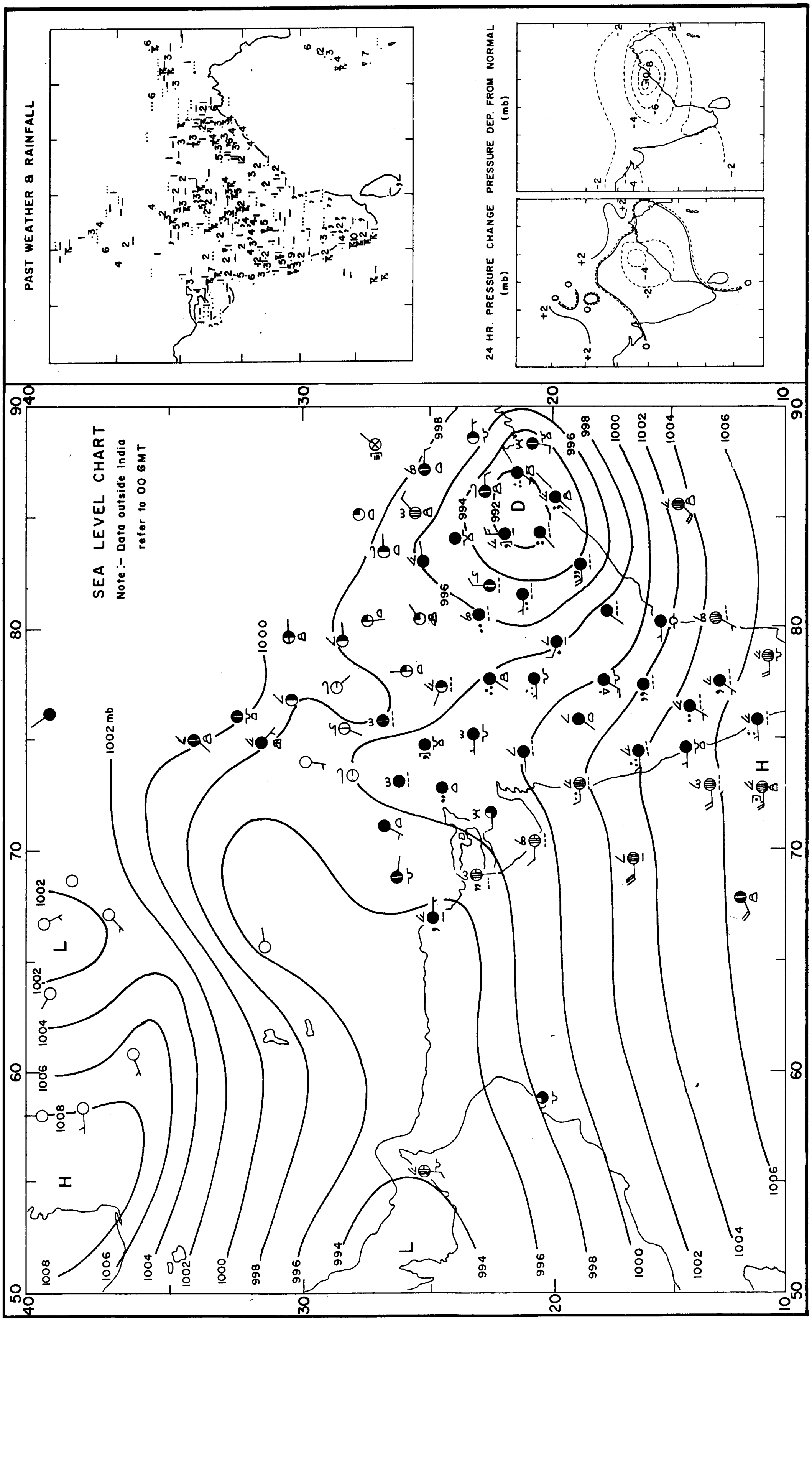


Fig. 7.24(d) Synoptic charts 0300 GMT 27 July 1967

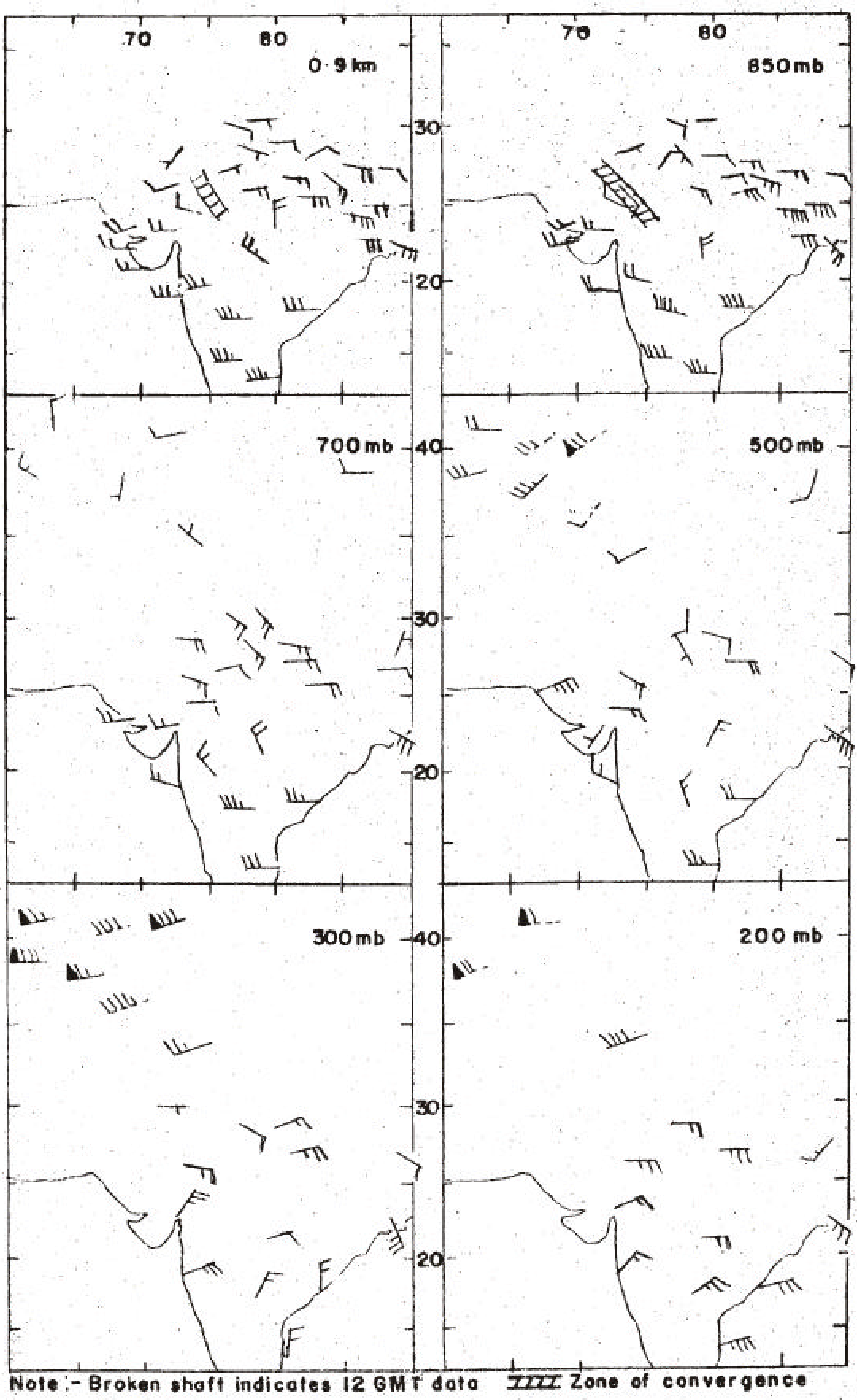


Fig. 7.24(e) Upper Winds 27 Jul. 1967 00 GMT

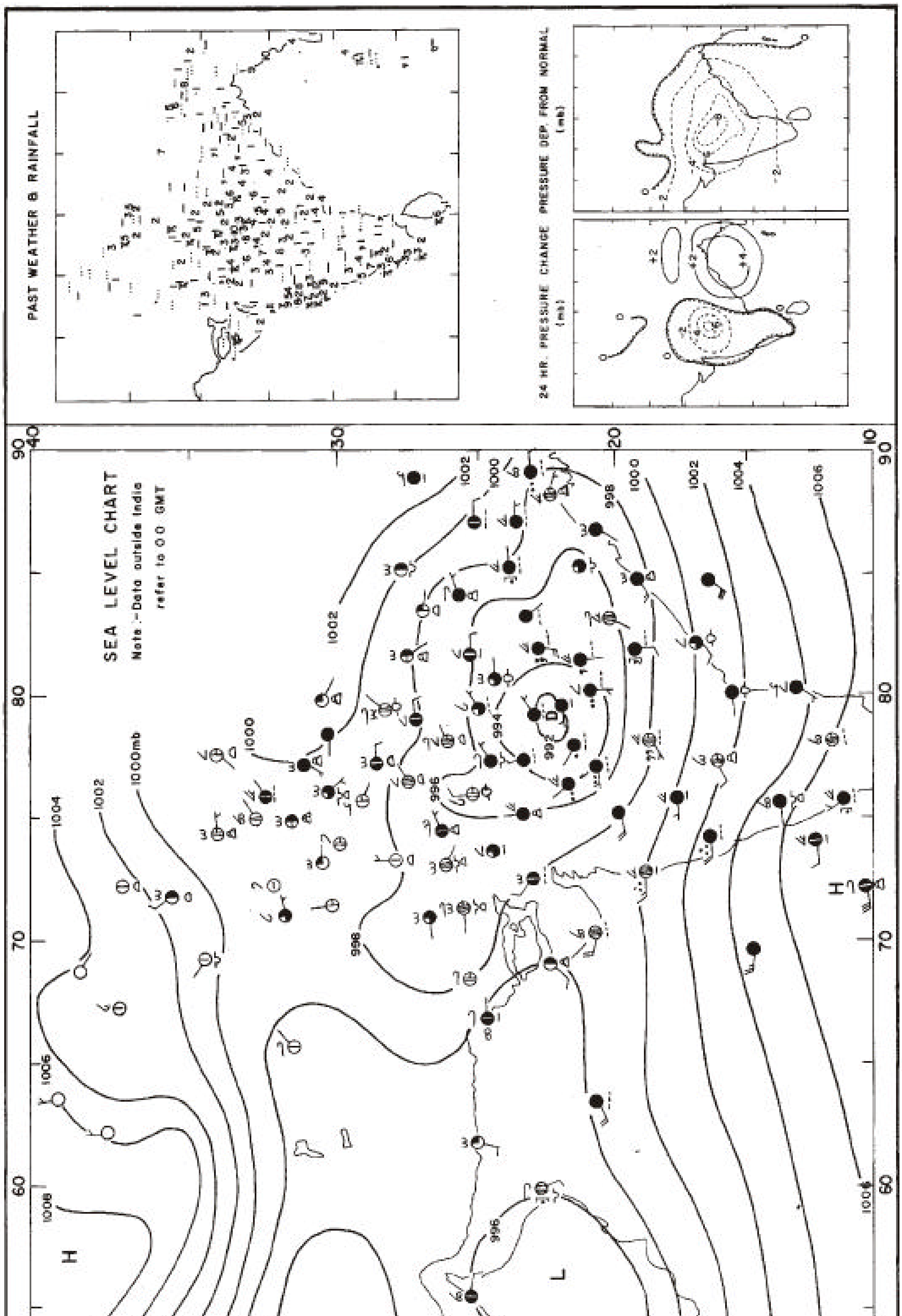


Fig. 7.24(f) Synoptic charts 0300 GMT 28 July 1967

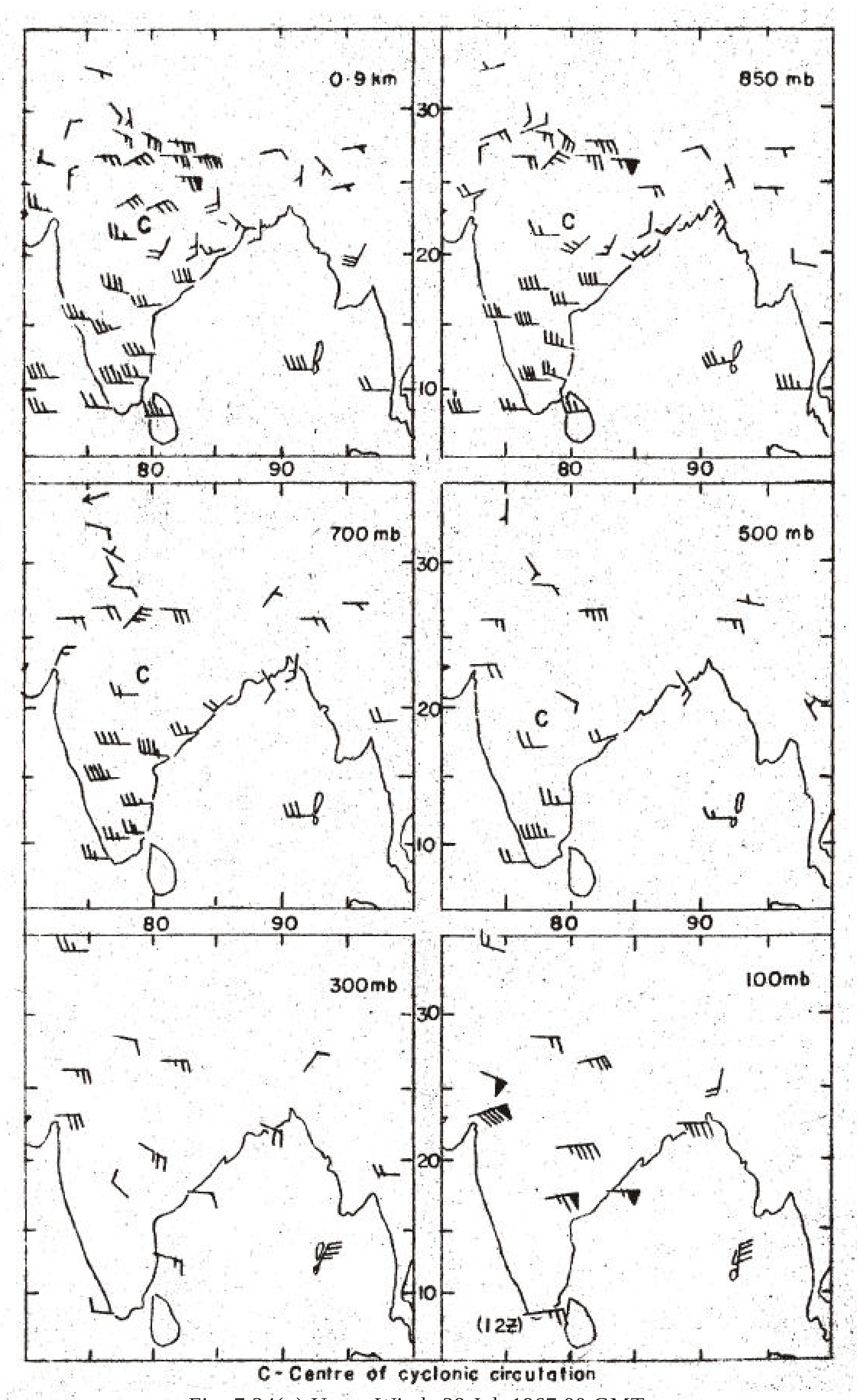


Fig. 7.24(g) Upper Winds 28 Jul. 1967 00 GMT

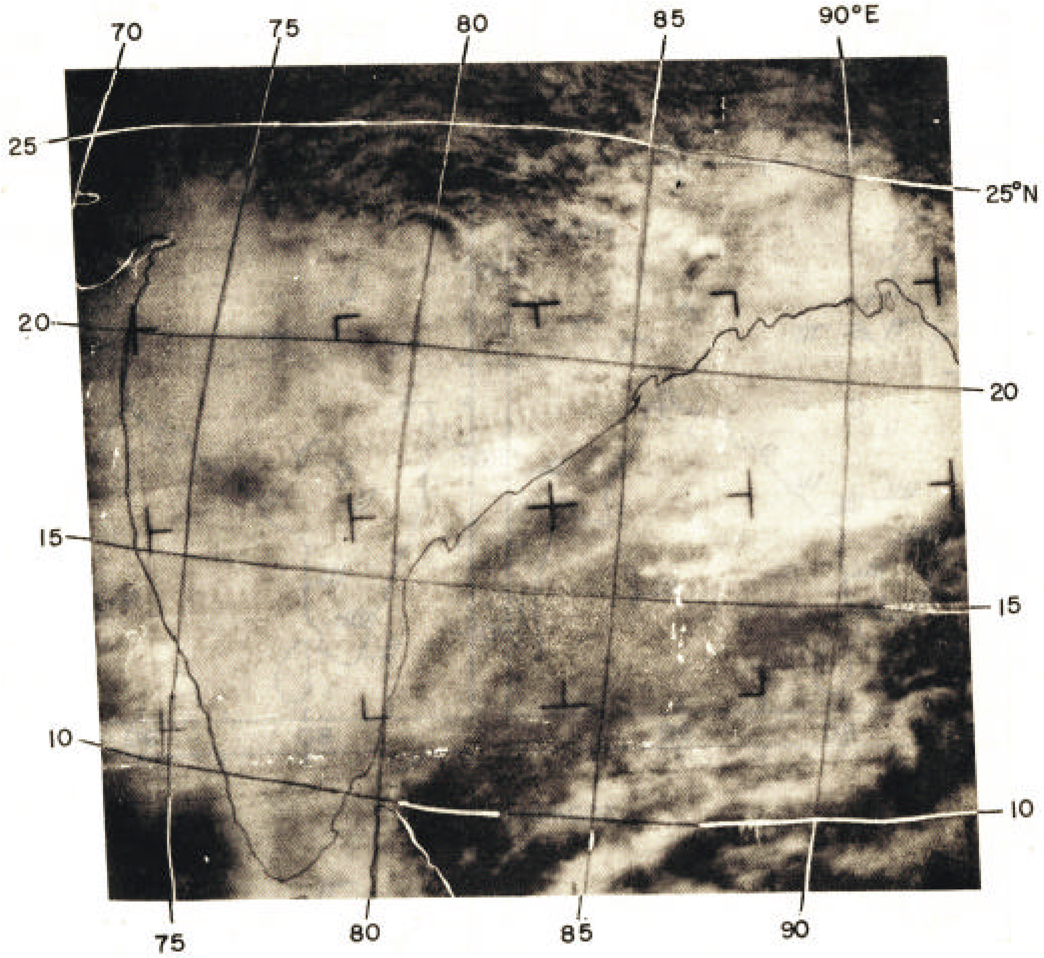


Fig. 7.24(h) NIMBUS-2 28 July 1967

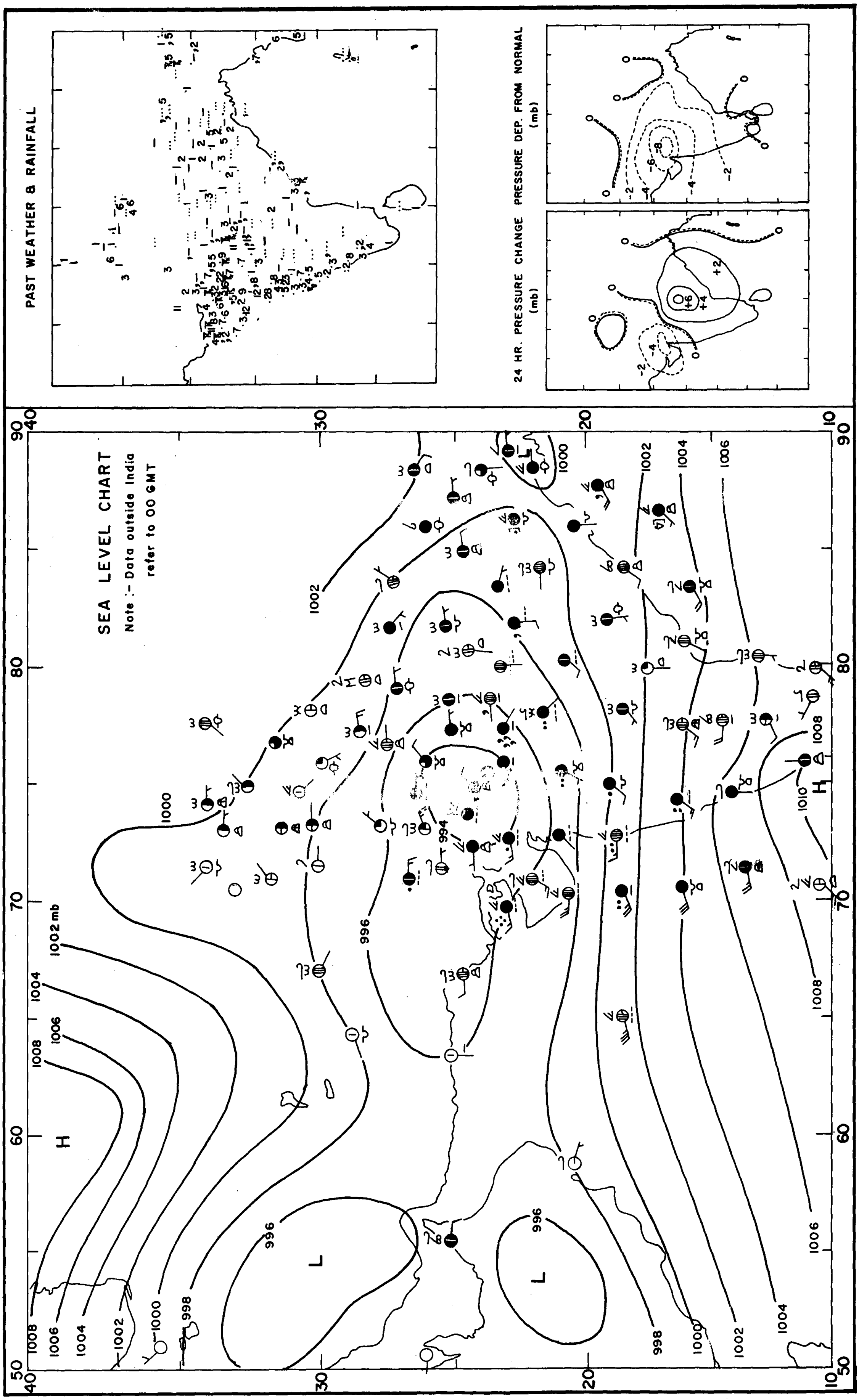
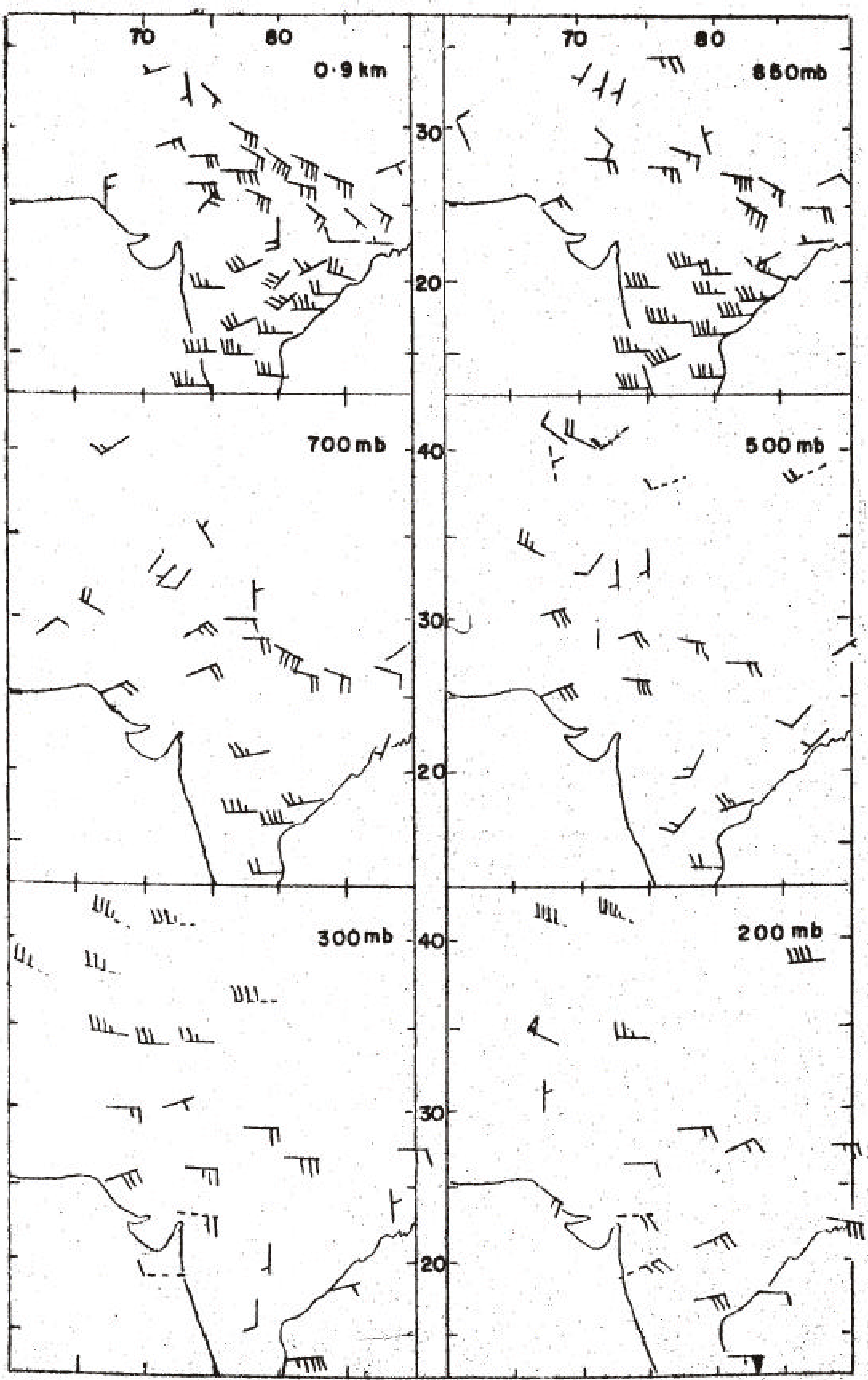


Fig. 7.24(i) Synoptic charts 0300 GMT 29 July 1967



Note :- Broken shaft indicates 12 GMT data

Fig. 7.24(j) Upper Winds 29 Jul. 1967 00 GMT

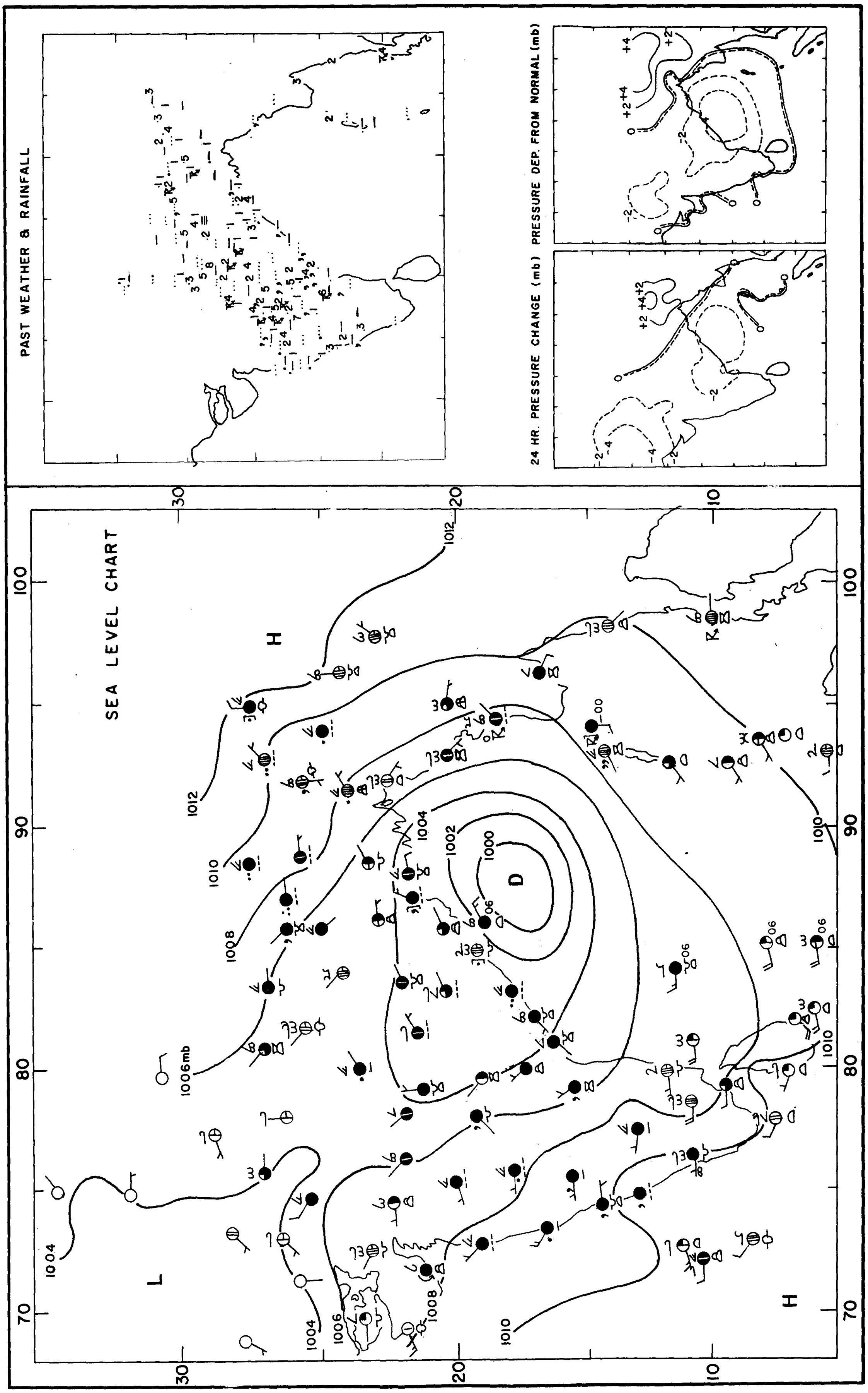


Fig. 7.25(a) Synoptic charts 0300 GMT 18 September 1969.

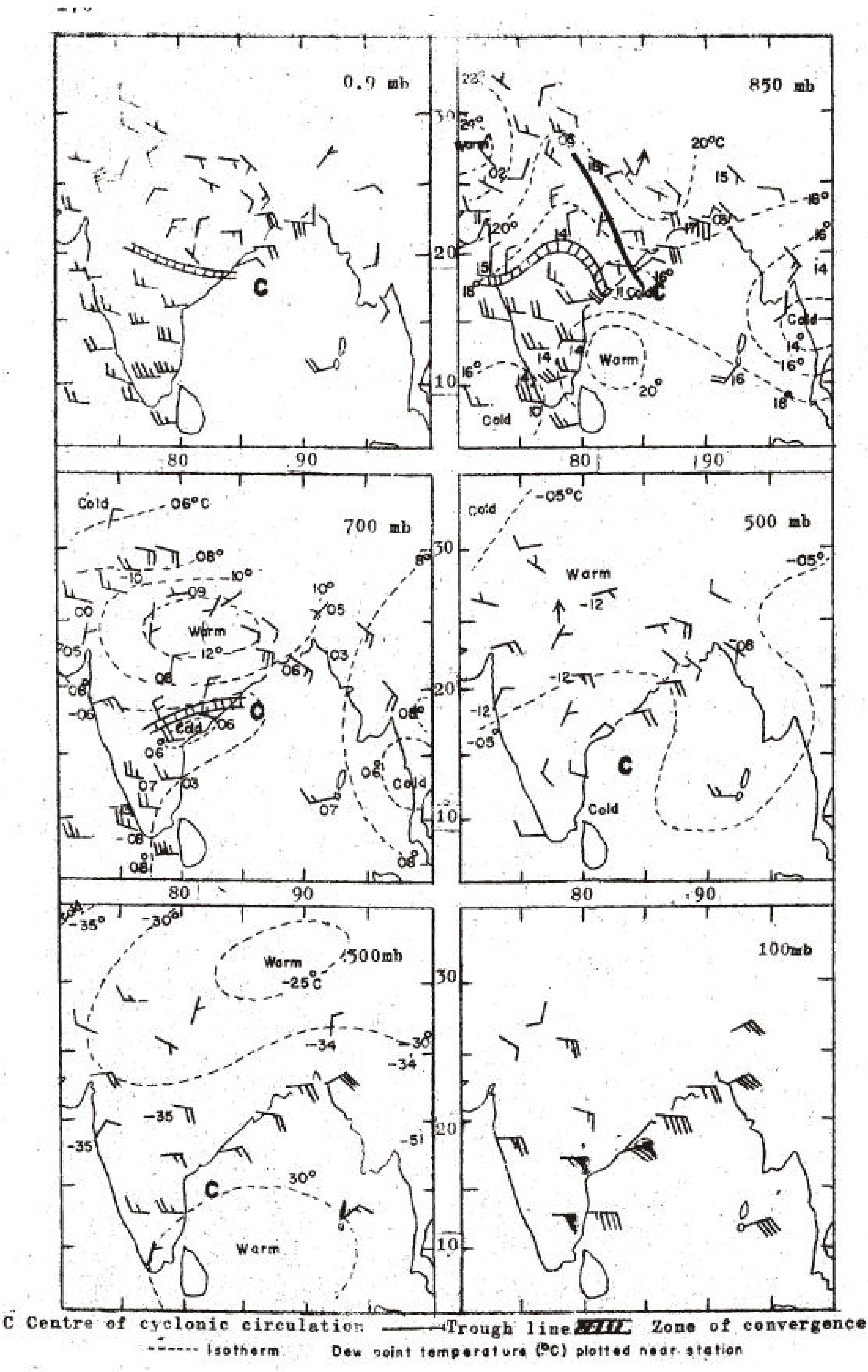


Fig. 7.25(b) Upper Winds 18 September 1969.

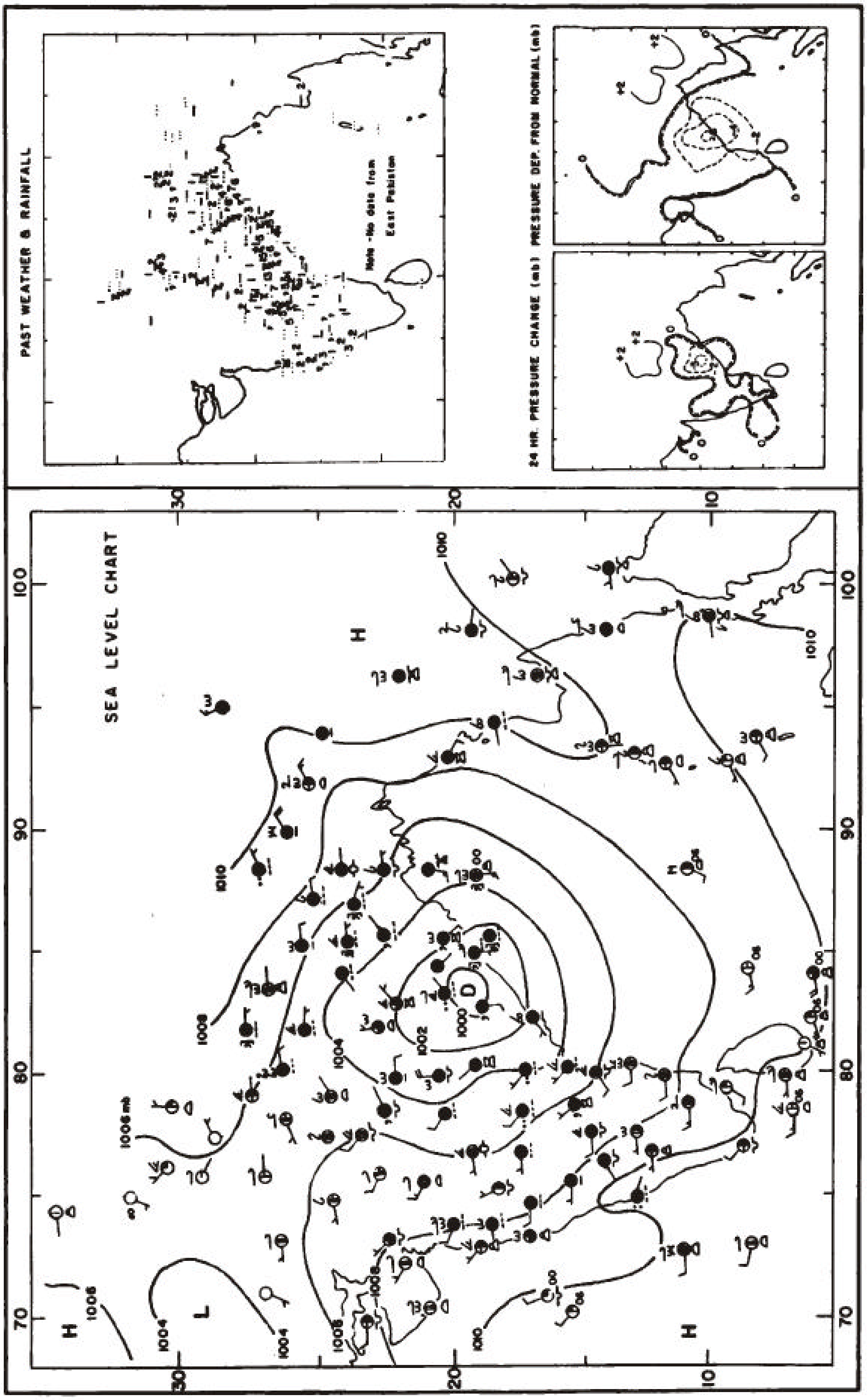


Fig. 7.25(c) Synoptic charts 0300 GMT 19 September 1969.

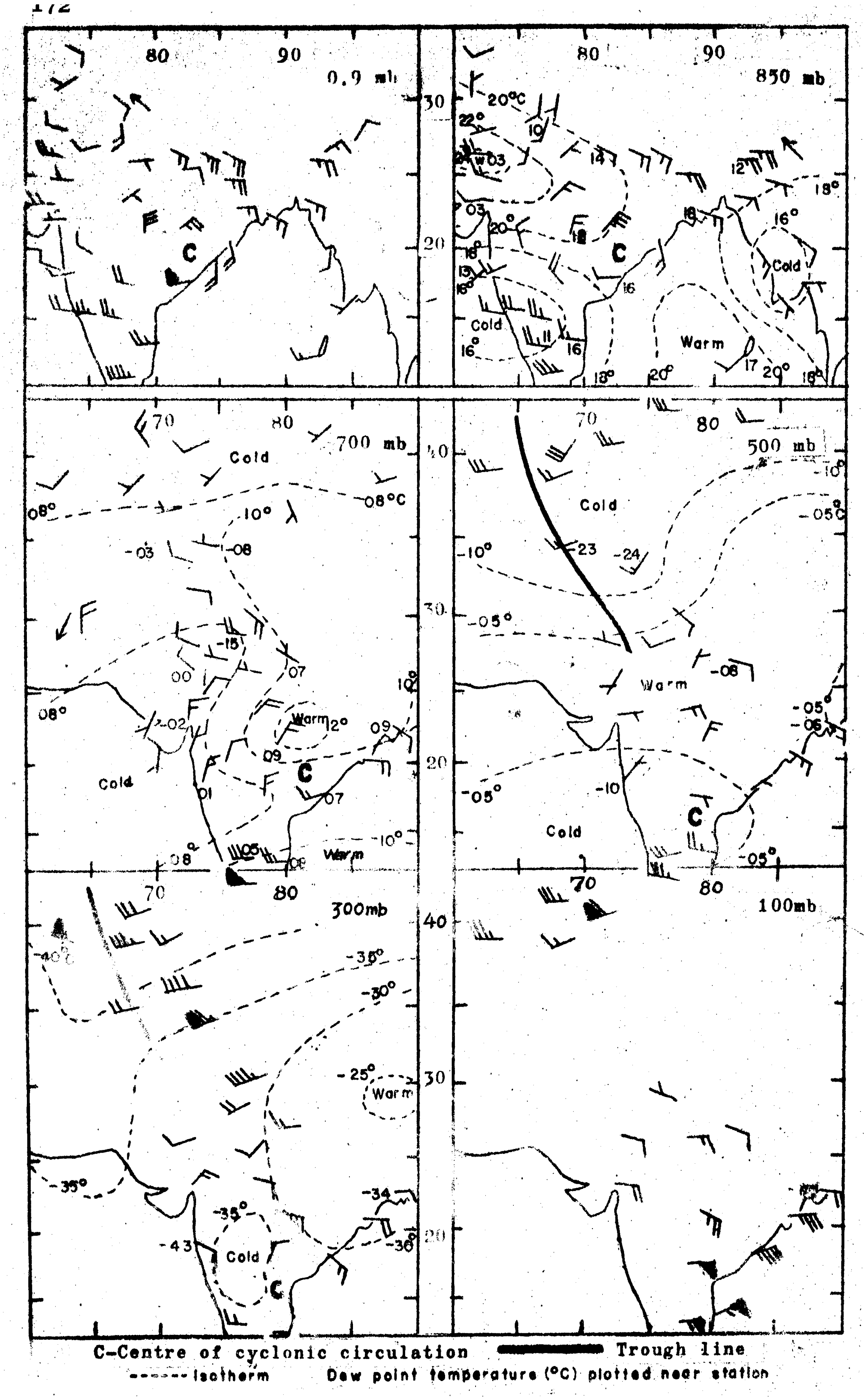


Fig. 7.25(d) Upper Winds 19 September 1969.

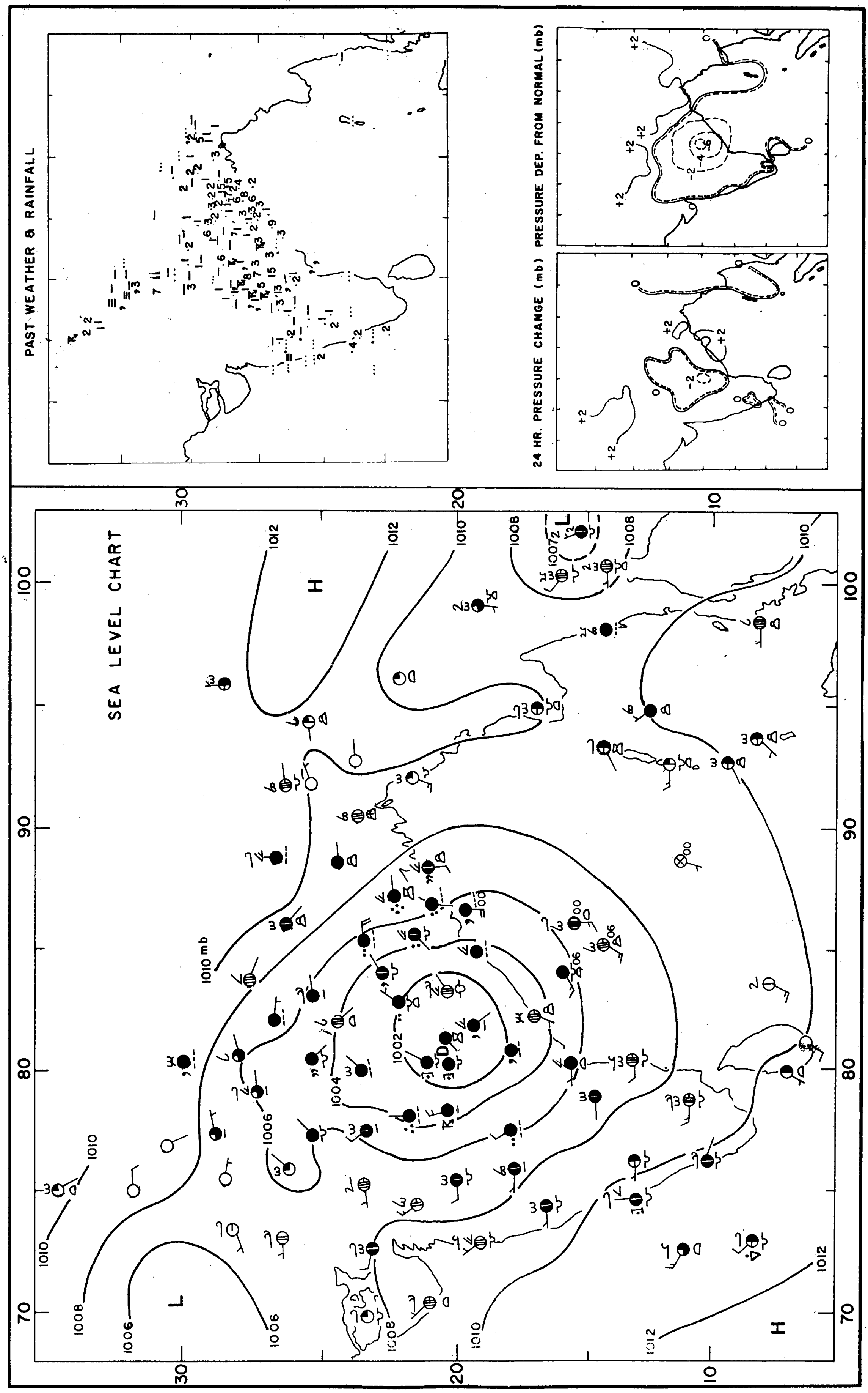


Fig. 7.25(e) Synoptic charts 0300 GMT 20 September 1969.

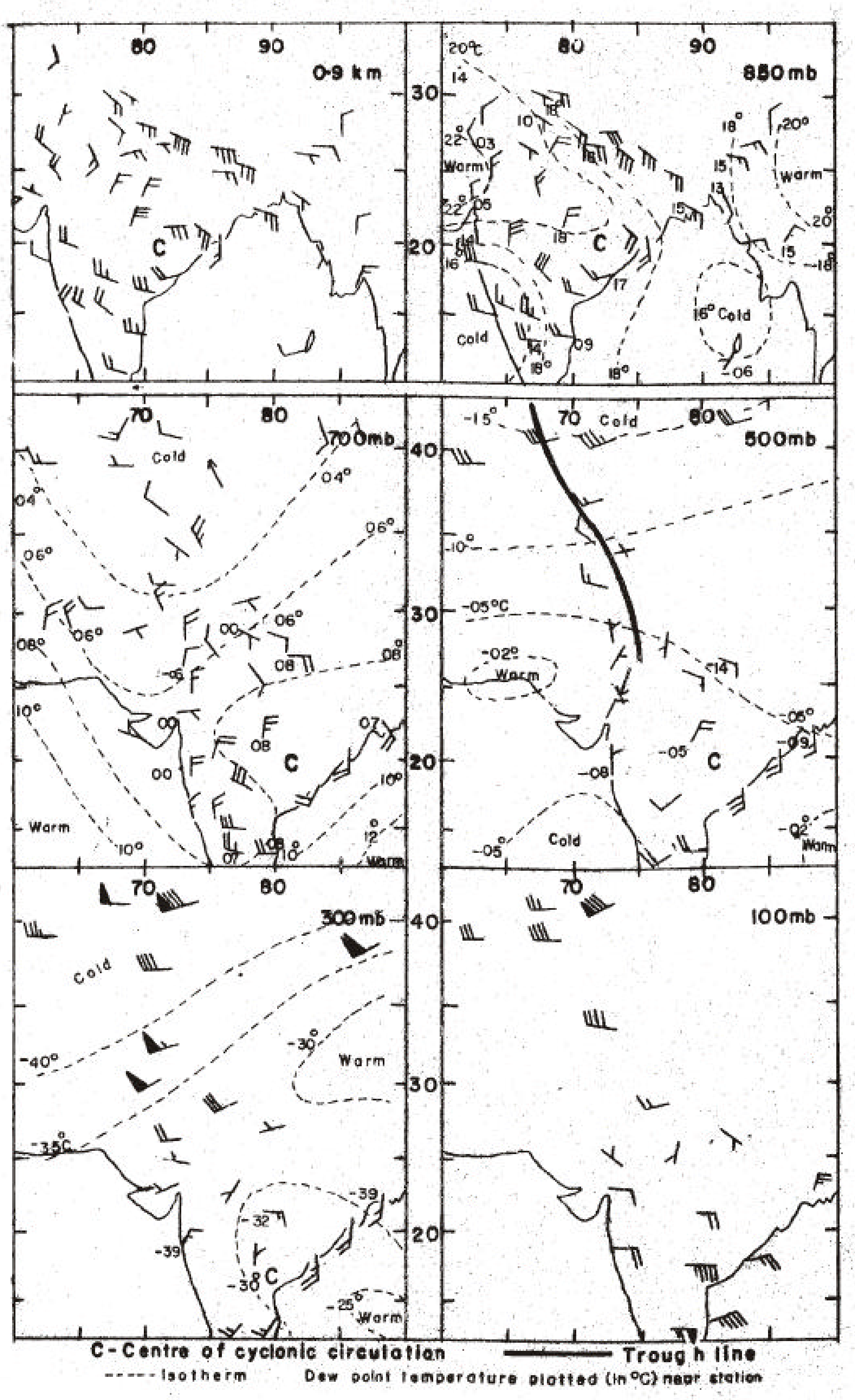


Fig. 7.25(f) Upper Winds 20 September 1969 00 GMT.

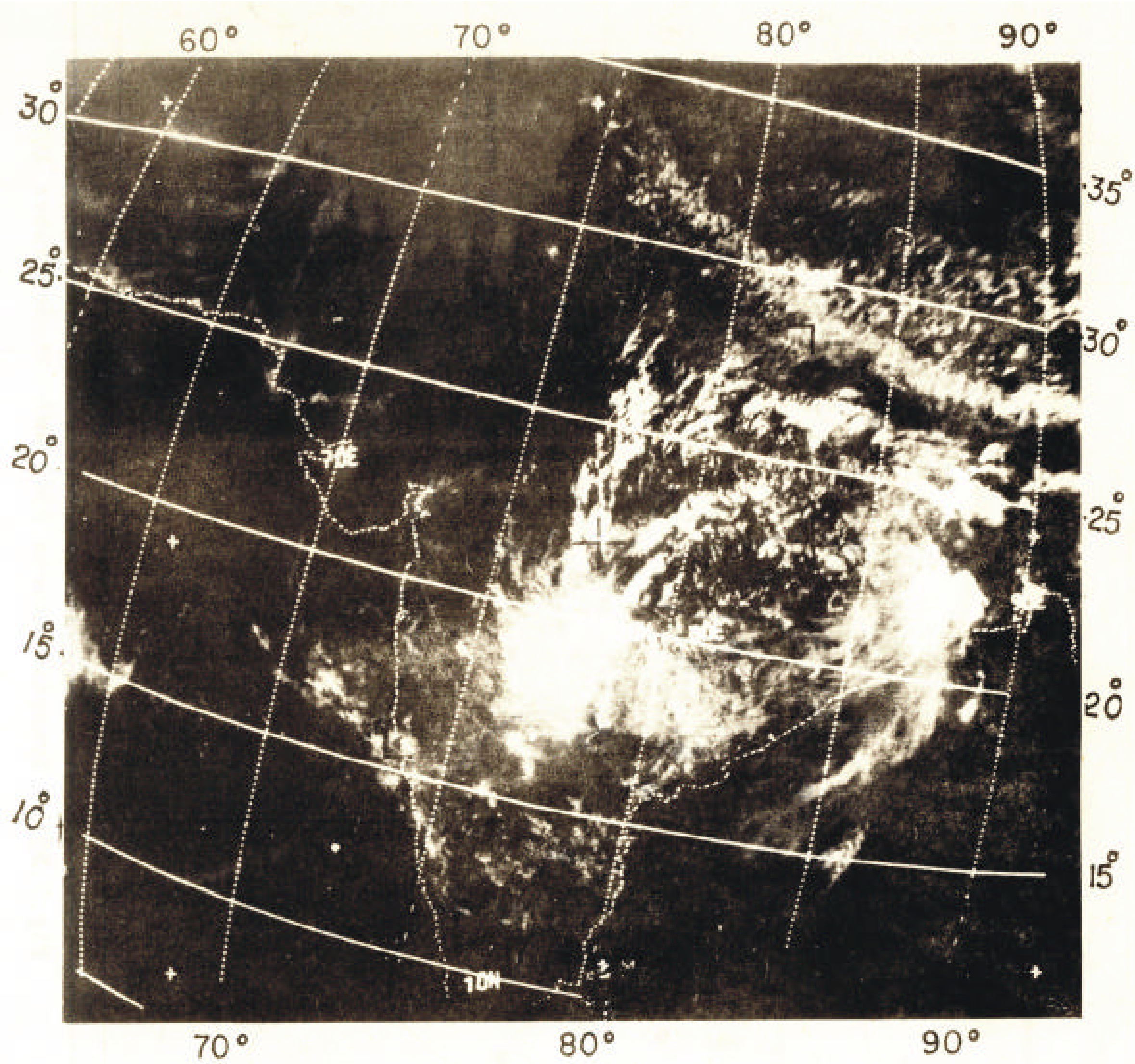


Fig. 7.25 (g) ESSA-9 20 September 1969

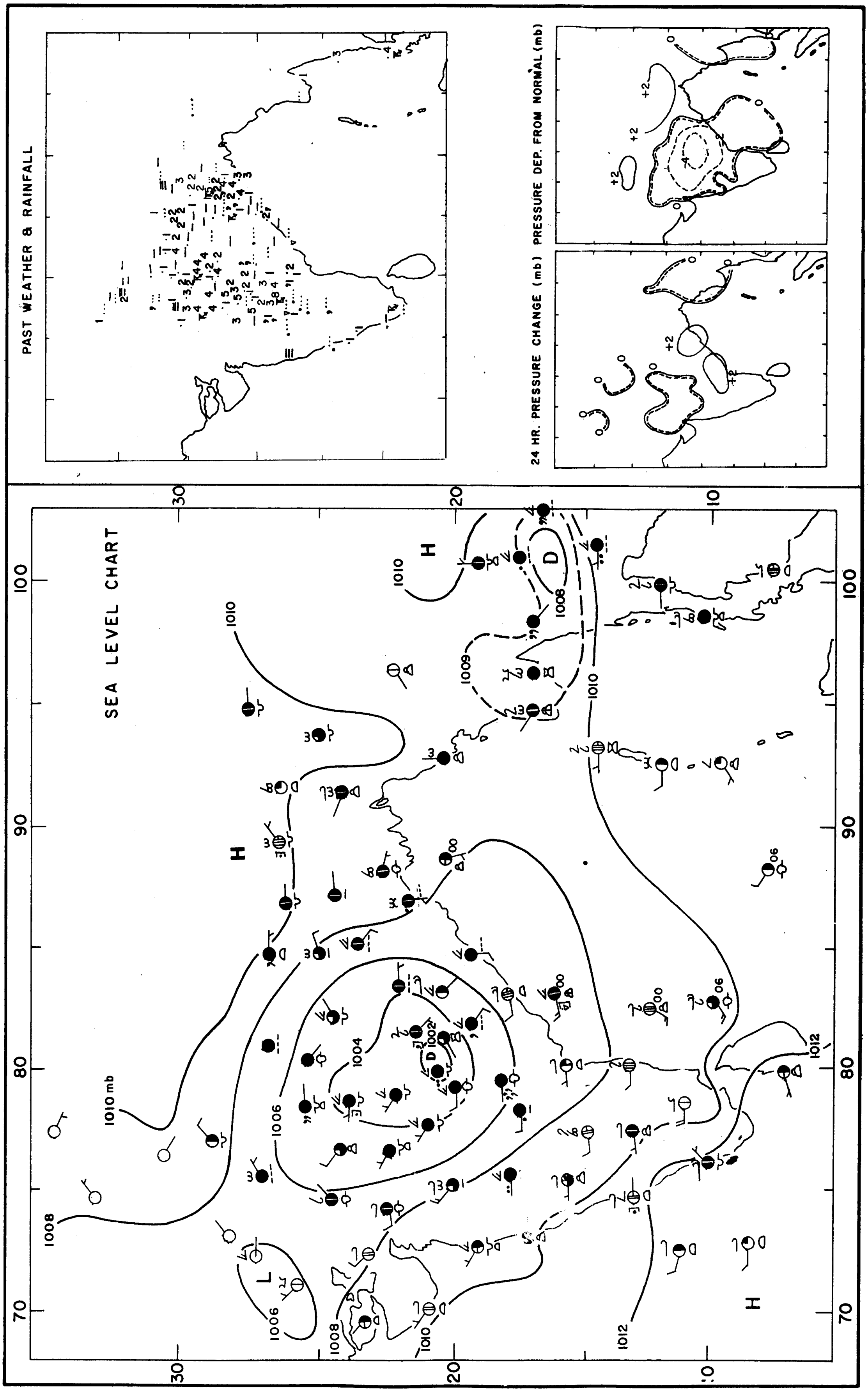


Fig. 7.25(h) Synoptic charts 0300 GMT 21 September 1969

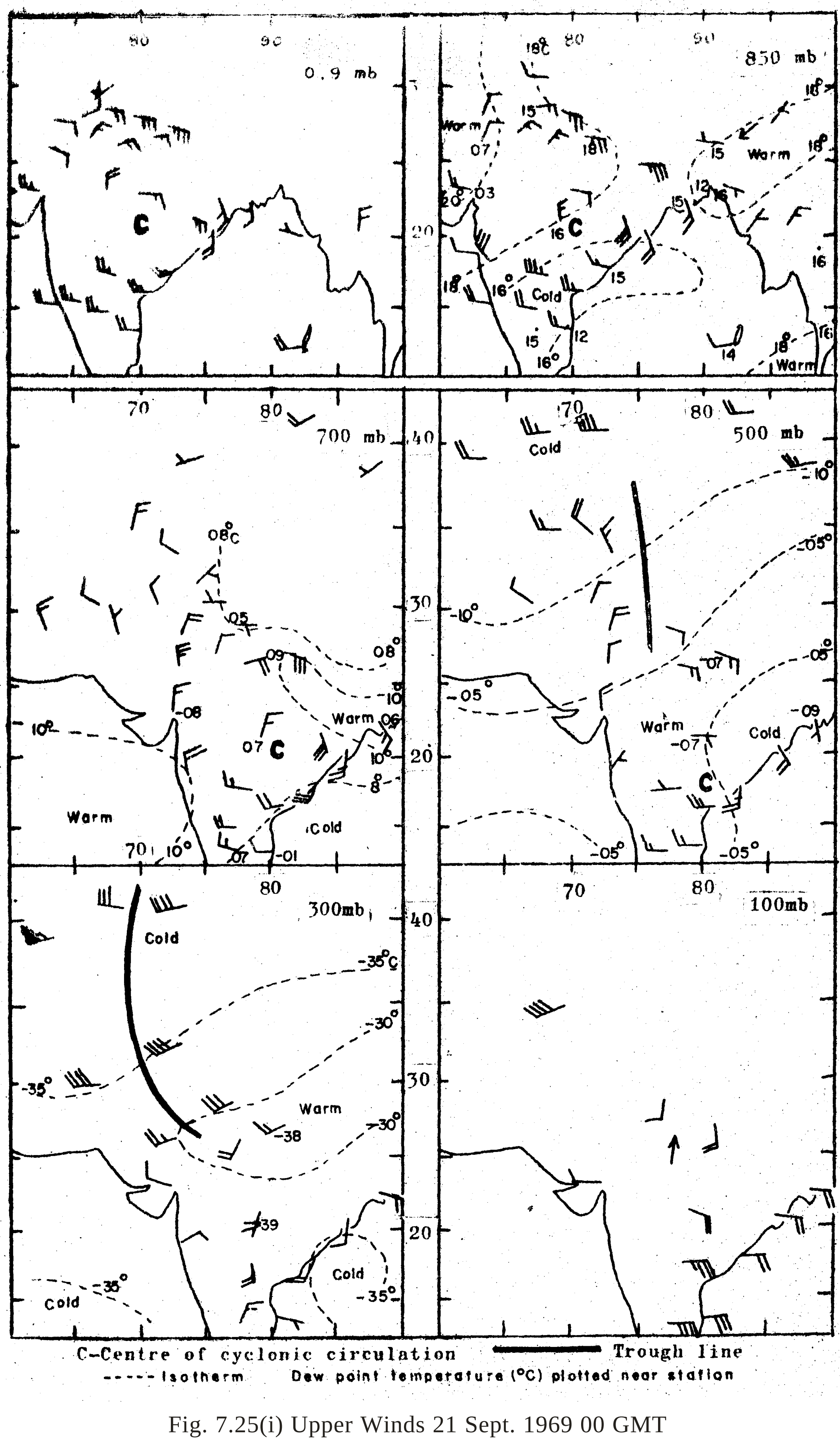


Fig. 7.25(i) Upper Winds 21 Sept. 1969 00 GMT

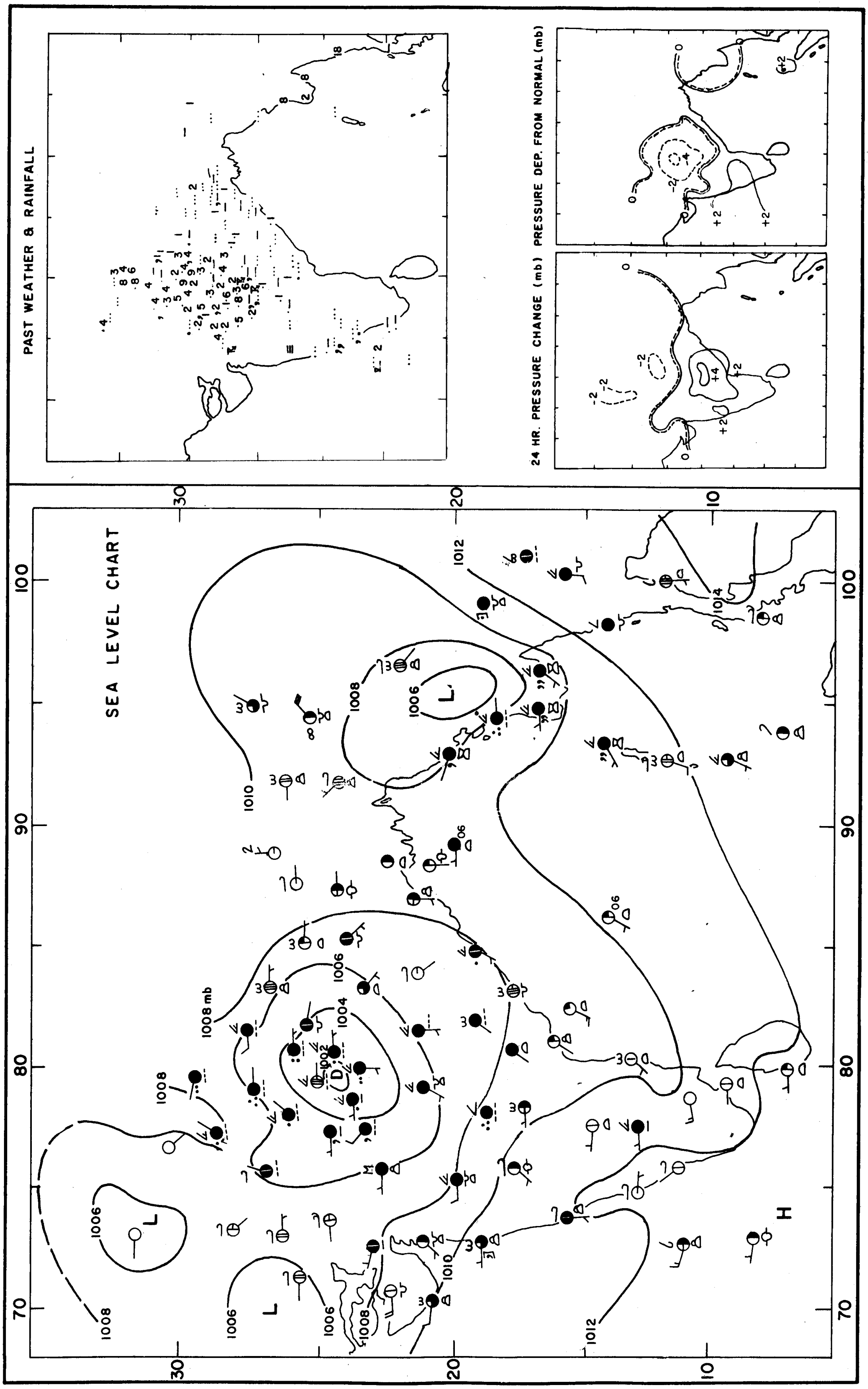


Fig. 7.25(j) Synoptic charts 0300 GMT 22 September 1969

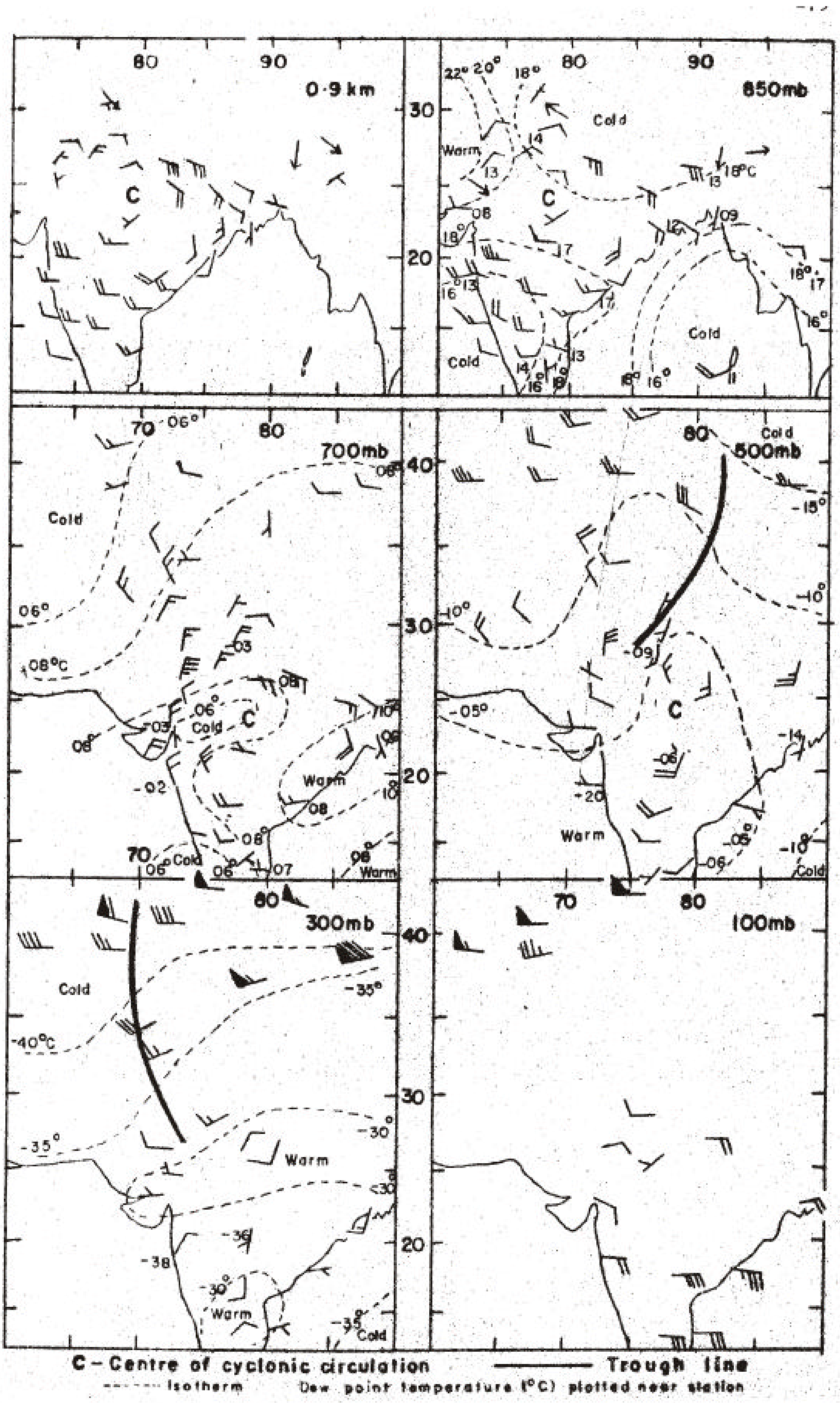


Fig. 7.25(k) Upper Winds 22 Sept. 1969 00 GMT.

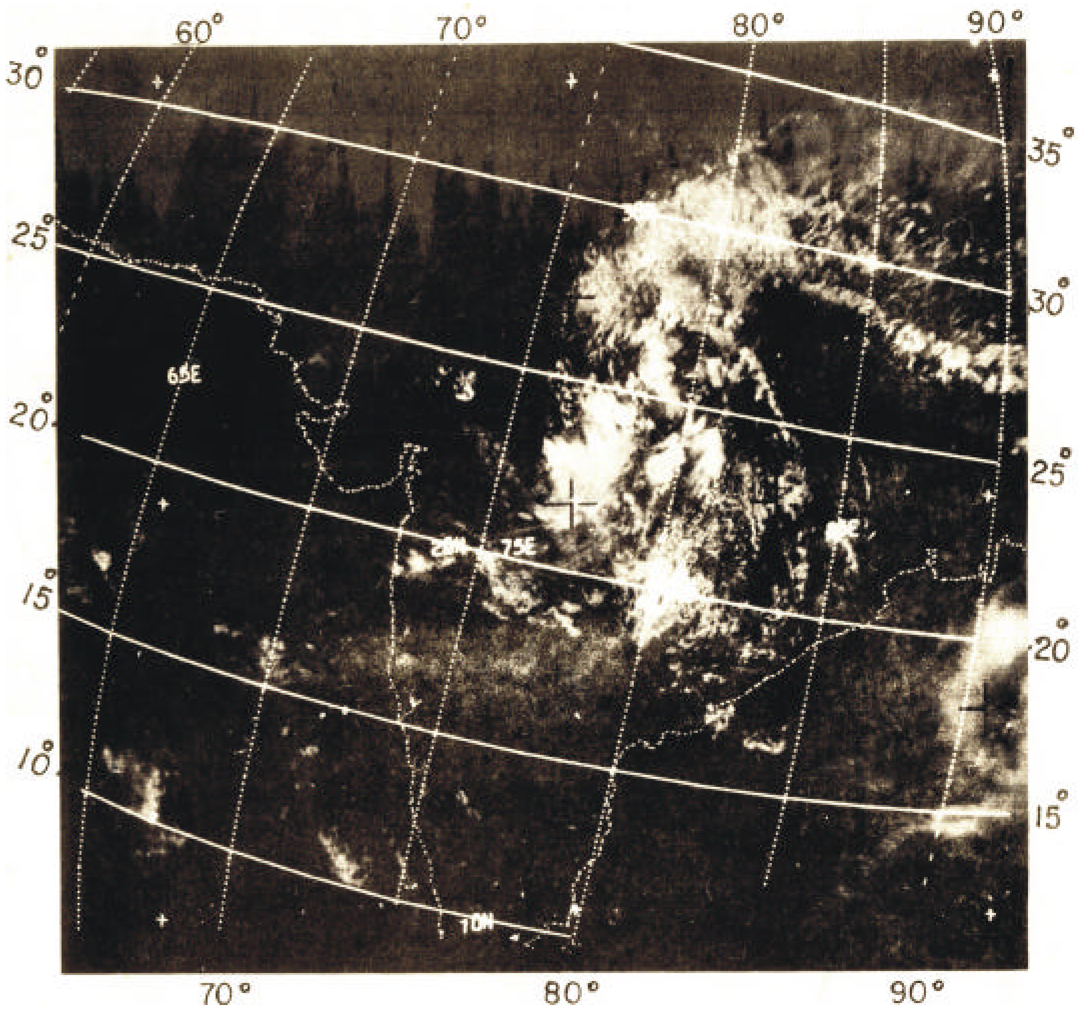


Fig. 72.5 (I) ESSA-9 22 September 1969

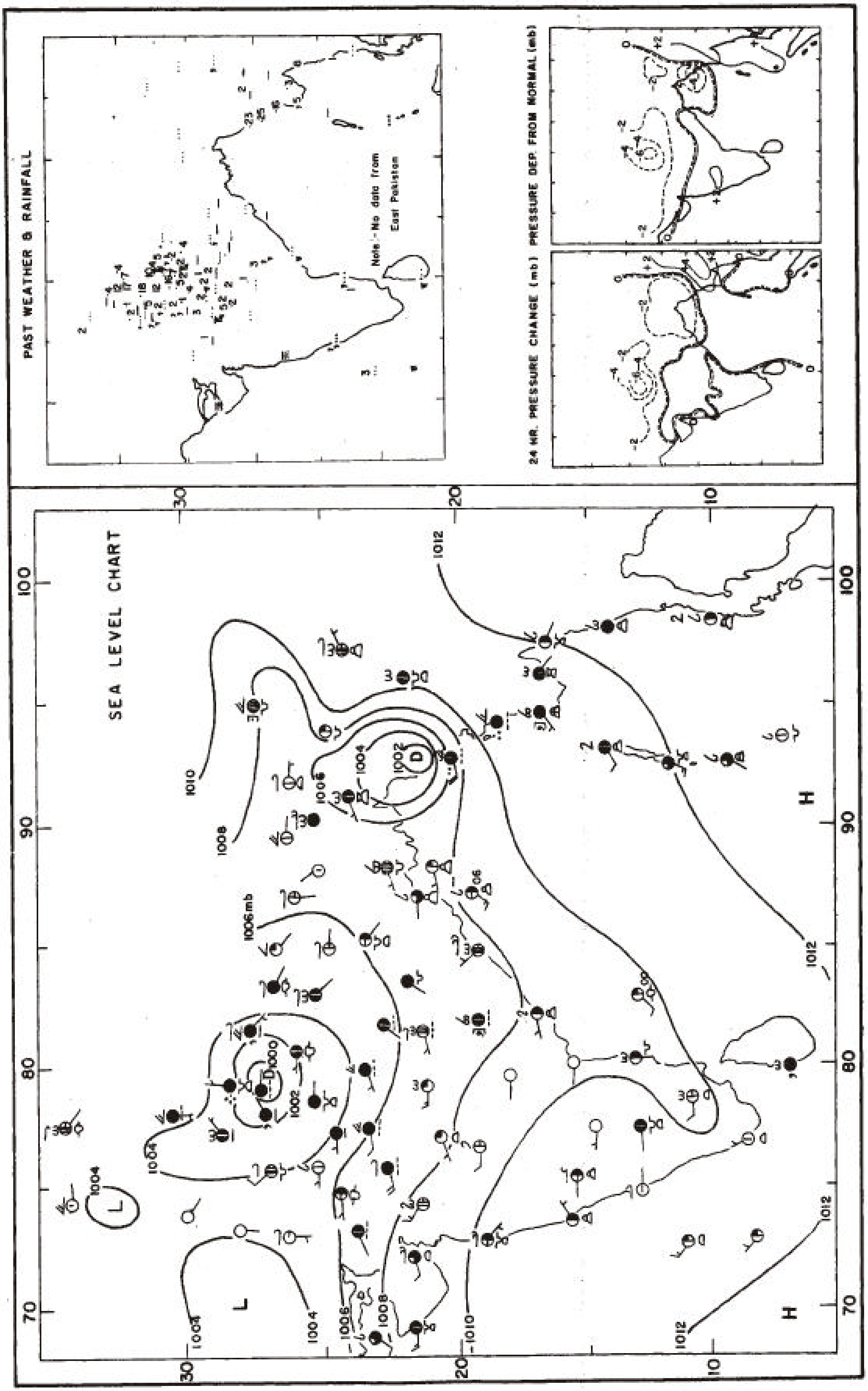


Fig. 7.25(m) Synoptic charts 0300 GMT 23 September 1969

including northern and adjoining central Bay needs no stress. In the mean–time, available data have to be studied more closely for structural variations of monsoon depressions at different levels and associated behavioural patterns.

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CHAPTER 8

BREAKS IN MONSOON

8.1 There are periods when the monsoon trough is located close to the foot of the Himalayas which leads to striking decrease of rainfall over most of the country but increase along the Himalayas and parts of northeast India and southern Peninsula. Such a synoptic situation is referred to as 'break in monsoon'.

8.2 Ramamurthy (1969) has catalogued the 'breaks' in July and August from 1888 to 1967. He adopted the criteria of the monsoon trough not being seen on sea–level chart as well as upto 850 mb and this synoptic pattern persisting for more than two days to define a 'break'. His statistics about 'breaks' are summarised in the Table 8.1

Table 8.1

Statistics of breaks in monsoon

Month	No. of breaks	No. of break days	Average duration (days)	Longest break (days)	Most frequent duration (days)	No. of break days in		
						First ten days	Second ten days	Remaining period
July	53	306	5.8	17	4	81	117	108
August	55	356	6.5	20	3	115	159	82
Commencing in July and ending in August	5	47		21				

August is slightly more susceptible to 'break' days and longer 'breaks', particularly the middle of the month. The beginning of July and end of August have less number of break–days. In these eighty years, there was no break in twelve years.

8.3 In a 'break' situation, the monsoon trough is at the foot of the Himalayas or not noticed at all, the surface winds all becoming westerlies. Similar conditions may prevail in upper air. This is in contrast to the mean position of the trough from Ganganagar to Calcutta at sea level and further south aloft. Even when the trough is present at sea level close to the foot–hills in some 'break' situations, at a little higher level it may be effaced. The usual slope southwards with height, due to temperature decreasing southward is not present. Perhaps the warmer air spreading across northern India from the west and the cooling associated with rainfall along the Himalayas may cause temperature increasing towards south in 'break' situations, at least in the vicinity of the residuary trough near the foot hill. The pressure gradient at surface over the Peninsula weakens while it increases over the Gangetic plain. This is well brought out in departure of pressure from normal becoming negative from Punjab eastward along the foot of the Himalayas, but positive from Gujarat to north Bay (2 to 4 mb and even 8 mb). There can be a negative pressure departure over the extreme south of the Peninsula also. Westerlies may become very strong north of 20^{o} N in the first 2 km. All these features may, sometimes, not develop along the whole length of the trough at the same time. The above features are illustrated in the case of 7th August, 1965 in Figs. 8.1 (a) to (d). Winds reached 50 kt at Bareilly (in north India) at 1 km, while over the Peninsula upper winds were weaker than normal.

8.4 In contrast to depressions giving extensive rainfall, during 'breaks' rainfall almost ceases over most parts. Heavy falls occur in and near Himalayas but not simultaneously along the

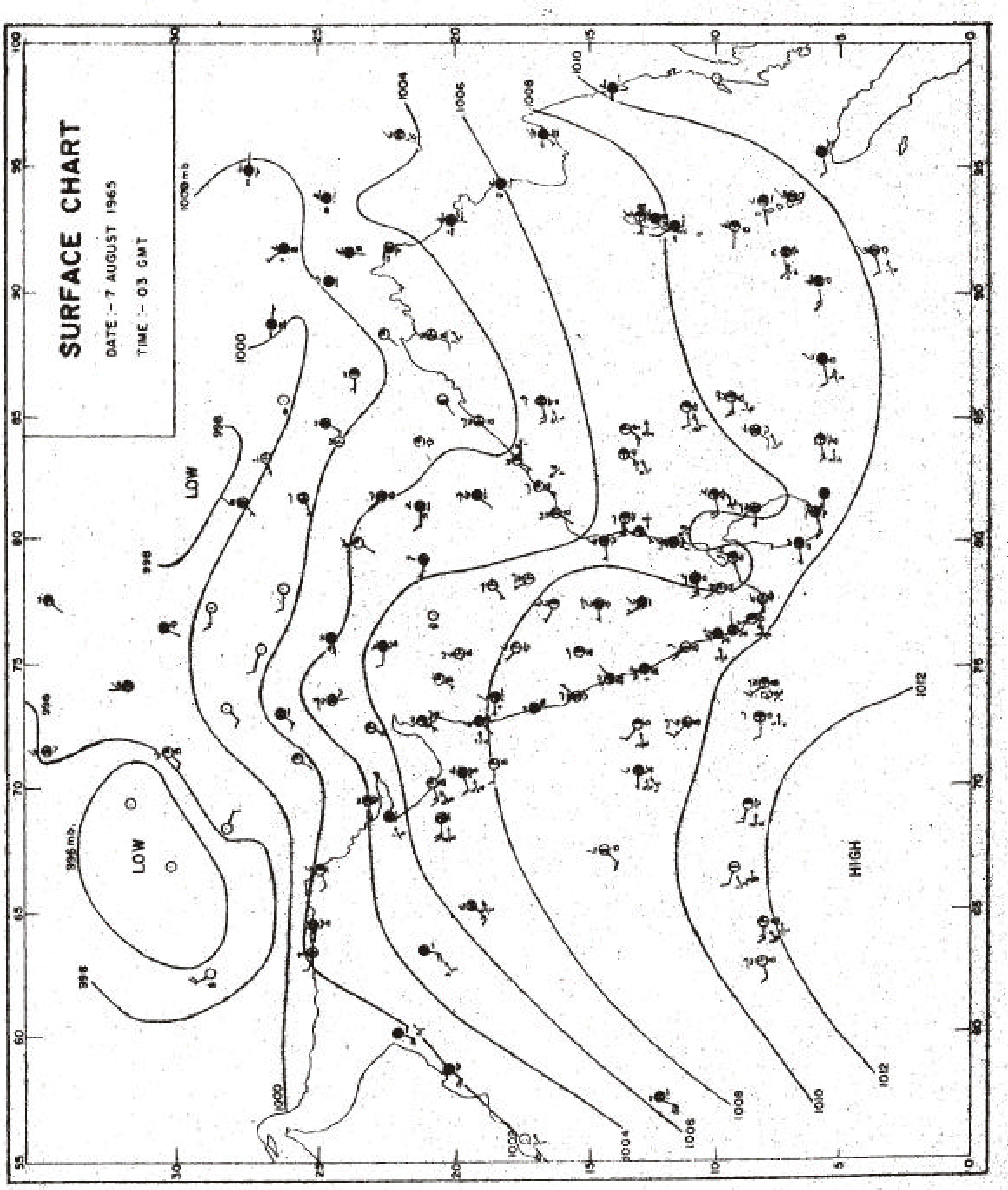
SURFACE CHART
DATE :- 7 AUGUST 1965
TIME :- 03 GMT
LOW
LOW
HIGH
996 mb.
998
1000
1002
1004
1006
1008
1010
1012

Fig. 8.1 (a)

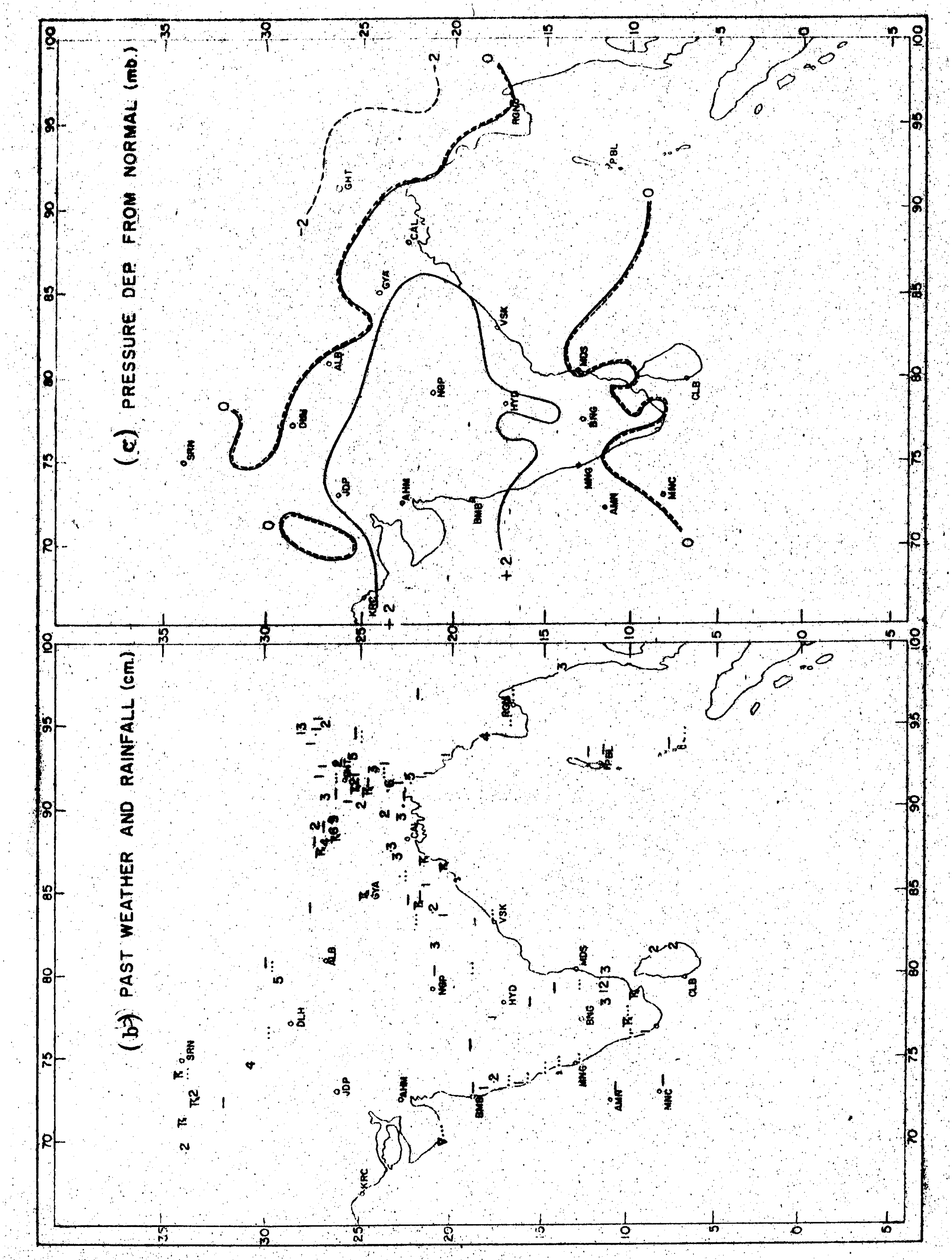

Fig. 8.1 (b) & (c) 7 August 1965 0300 GMT

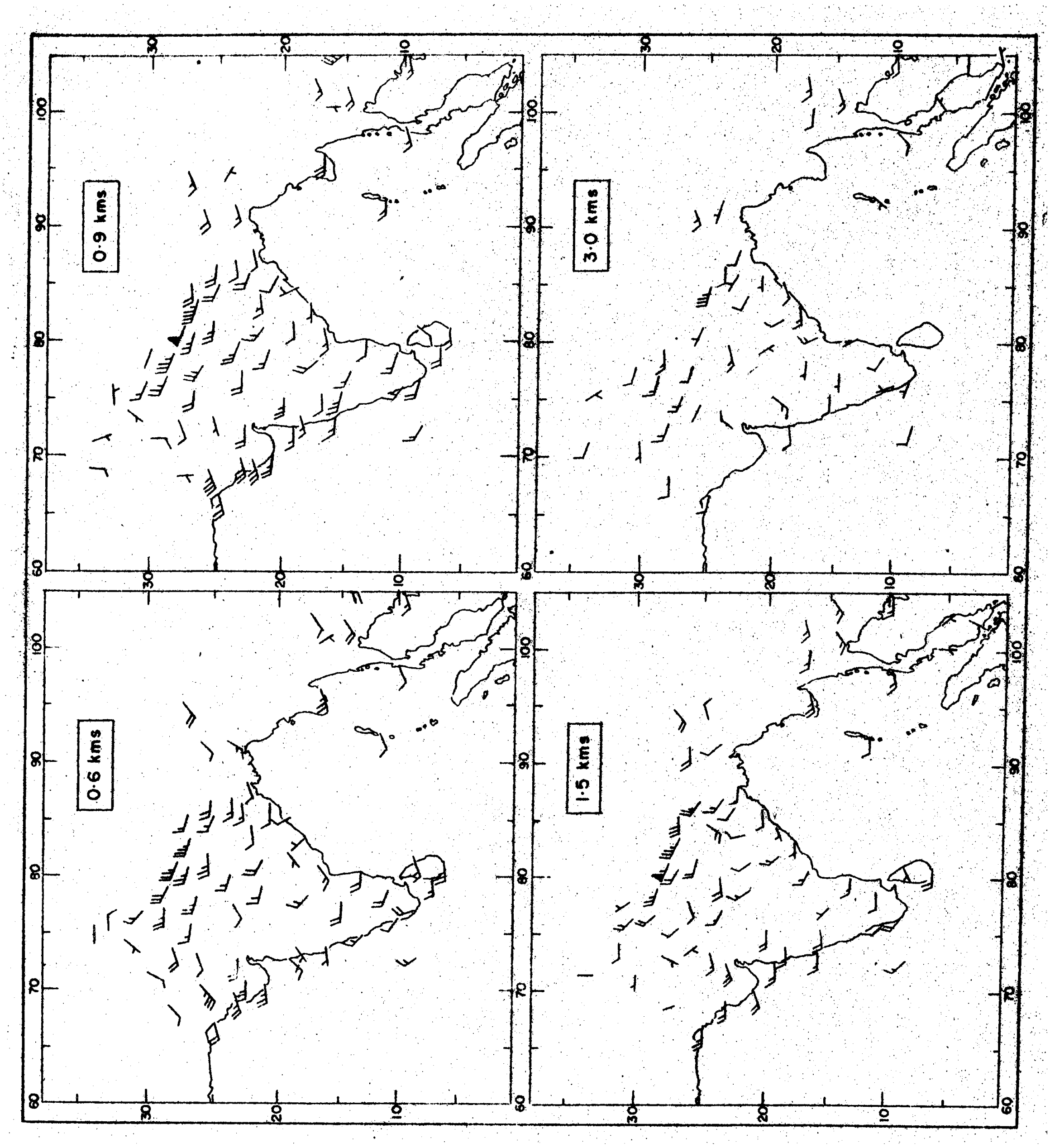

Fig. 8.1 (d) Upper Winds 7 August 1965 00 GMT

whole length. Influence of eastward moving troughs in westerlies further north seems to cause these patches of heavy rains. Himalayas to the east of 85° E are susceptible to much heavier falls than to the west. In northeast India, this above normal rainfall during 'breaks' extends to plains, well into Bihar, West Bengal and Assam. In the south, Tamil Nadu and Rayalaseema get more rains, as thundershowers. Figure 8.2 shows the average percentage departure of rainfall from normal in 12 'break' cases covering 83 days as presented by Ramamurthy (1969).

Areas not affected by high rainfall deficiency (>50 %*)* during 'breaks' are Jammu and Kashmir, Rajasthan, Gujarat State, Vidarbha, Madhya Maharashtra, Madhya Pradesh, Telangana and Orissa. Himachal Pradesh shows generally below normal rains in 'breaks', even though the western Himalayas are known to get heavy rains. In prolonged 'breaks', the air mass even over western Himalayas becomes dry and rains decrease. Perhaps it is the effect of this that gave sub–normal rainfall in this study, as the cases studied had a duration between 4 and 12 days. The percentage increase of rainy days in Assam and Tamil Nadu is only 20 % to 30 %, while the rainfall almost doubles. This is possible as the intensity of rains in favoured areas gets enhanced. Still there is considerable difference in the spatial distribution of rainfall in areas of increased precipitation. Assam and sub–Himalayan West Bengal have fairly widespread rains on about 90 % 'break' days, while in Bihar Plains it is only 46 %, Tamil Nadu 11 % and Rayalaseema 17 %. Dry days are still 42 % in Rayalaseema but only 8 % in Tamil Nadu, Heavy falls (>7 cm) common in the northern areas, occur only on 5 % of days in these southern parts and in isolated fashion.

8.5 Ananthakrishnan (l97l) finds a distinct minimum of rainfall around 14–18 August at several stations over north and central India, fitting with the higher frequency of 'breaks' about the middle of the August. Ramaswamy (1973) has compared the normal rainfall for each day in August with the average per day for the whole month. He finds a period between 7 and 15 August when daily rainfall is less than the month's daily mean in central and western parts, but more in Himalayan belt, northern plains, northeast India and extreme south Peninsula. He calls this a normal period of large scale 'break' in southwest monsoon. The type of variation in rainfall on which these conclusions are based is shown in Fig. 8.3. The excesses and deficiencies of average rainfall in different areas during 7–15 August over the mean daily rainfall for the month are shown in Fig. 8.4. The 'break' begins and ends relatively earlier over the Punjab plains, upper and lower Gangetic plains and Brahmaputra basin than over the rest of the country. 'Break' commences and ends later in the extreme southeast of Peninsular India than elsewhere over the country. Almost all stations experience 'breaks' which last between 6 and 11 days.

Ramaswamy also shows Indo–Gangetic plain west of 82° E and Western Himalayas as getting more rain during break while Ramamurthy finds them as getting less rainfall.

8.6 Fig. 8.5 shows ten year running means of 'breaks' and depressions which have opposite effects on the rainfall of the country.

8.7 The sub–tropical anticyclone normally over northwest India above 850 mb swings southwards and becomes more marked, the maximum displacement being at 500 mb, from about 30° N to the south of the 20° N. Weak troughs appear over the Bay of Bengal and south Peninsula in lower troposphere. The easterly jet is perhaps a little stronger. Ananthakrishnan and Ramakrishnan (1963) find that the area of strongest easterlies shifting to 19° N during 'break', from the normal position of 10^{o} N. The meridional flow of southerlies in lower troposphere and northerlies aloft, usually present south of 25° N and called the monsoon cell, contracts during 'break' upto 18^{o} N only and the Hadley Cell to north expands and covers the rest of the country (Fig. 8.6).

8.8 Raghavan (1973) has pointed out that as the 'break monsoon' intensifies, the troposphere as a whole in the central parts gets warmer. Warming is more in upper troposphere as can be seen from Fig. 8.7. This cannot be accounted for by advection and has to be due to

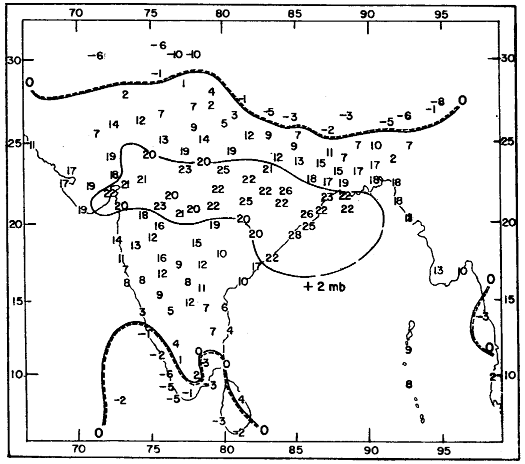

Fig. 8.2 (i) Mean Pressure departure (0300 GMT) during 'break'

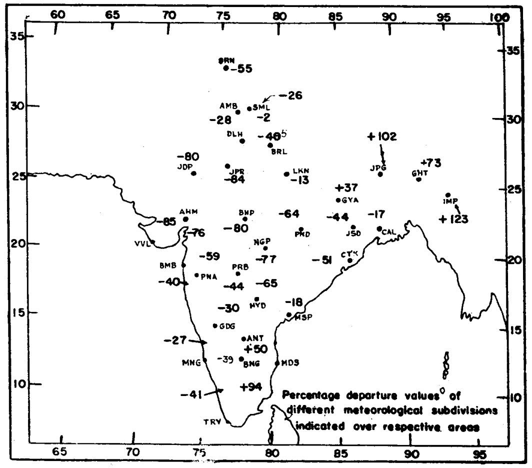

Fig. 8.2 (ii) Mean departure (%) of rainall during 'breaks'

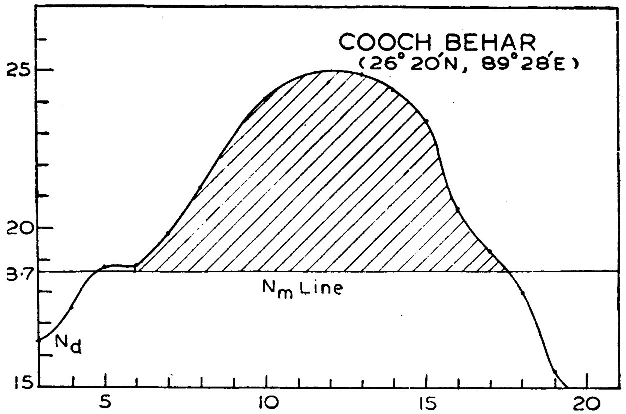

Fig. 8.3 (a) The smooth curve intersecting the Nm lines is the Nd curve. The hatched area is the Primary Crest and represents the period of break in the monsoon at the station associated with an excess of rainfall (Ramaswamy, 1973)

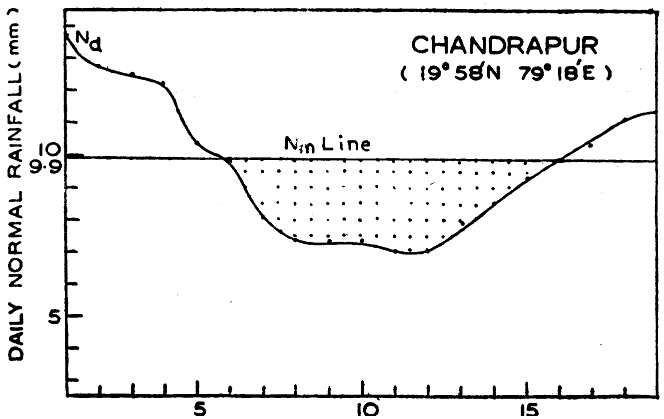

Fig. 8.3 (b) The stippled area is the Primary trough and represents the period of break in the monsoon at the station, associated with a dificieny in rainfall (Ramswamy, 1973)

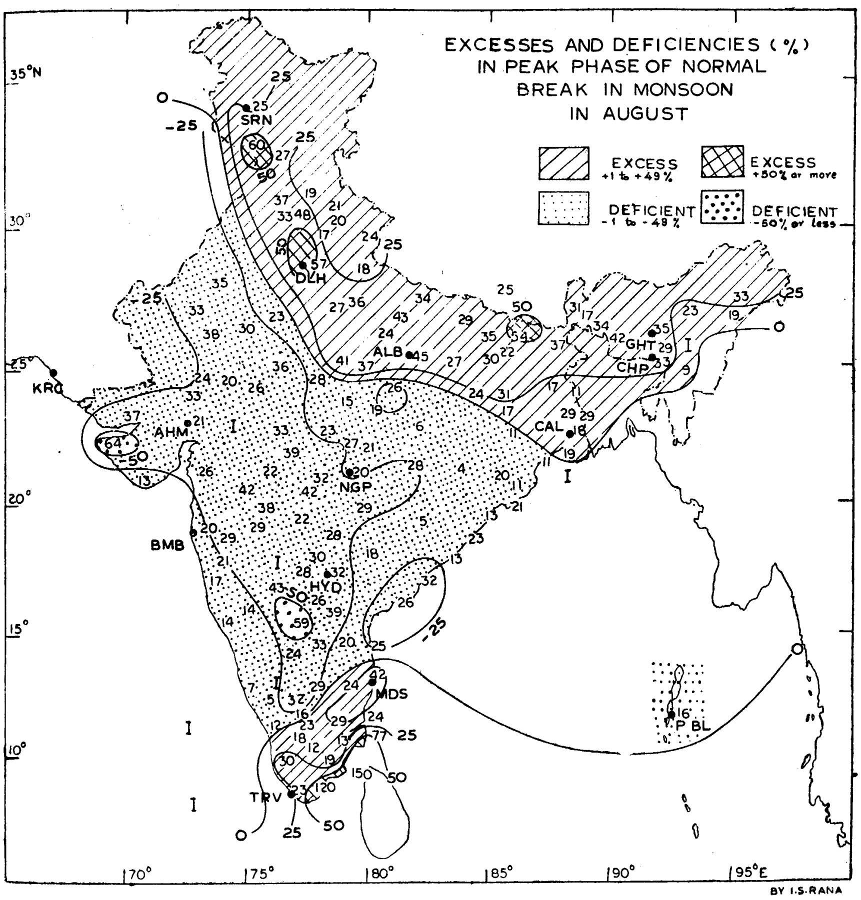

Fig. 8.4 Letter I - plotted at some stations stands for 'Indeterminate' (Ramaswamy, 1973)

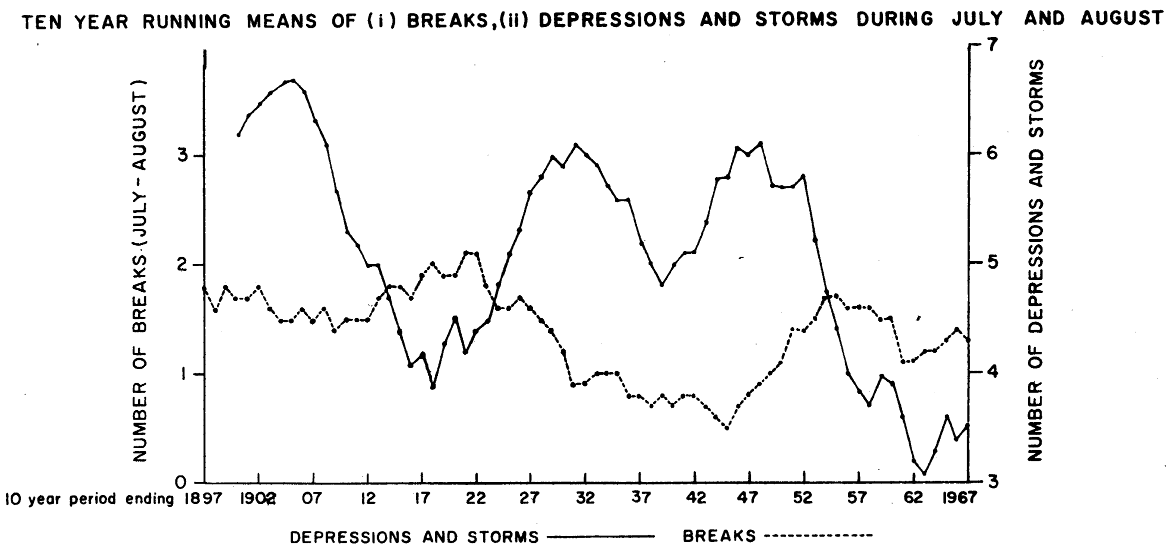

Fig. 8.5

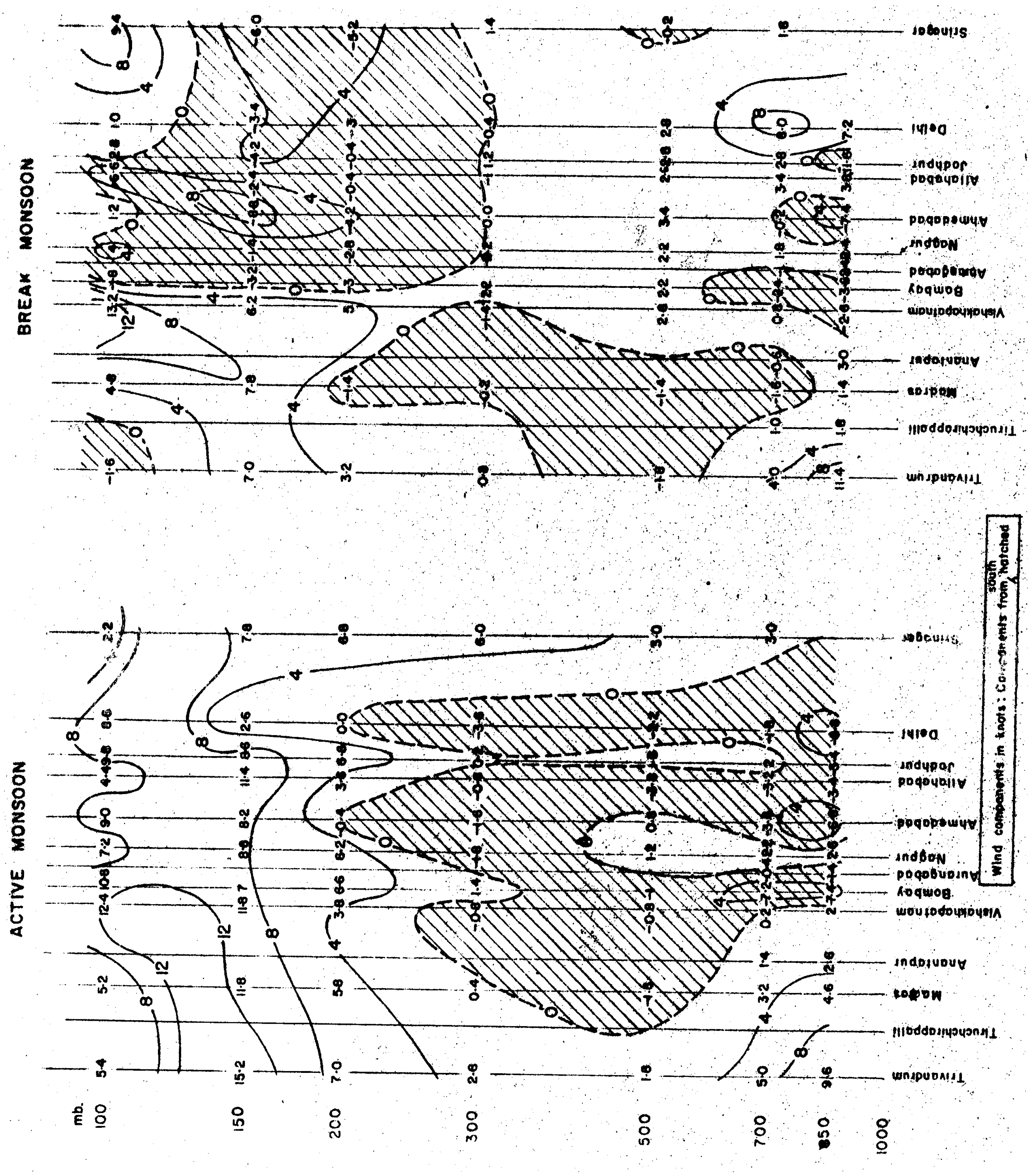

Fig. 8.6 Mean Meridional Circulation over India

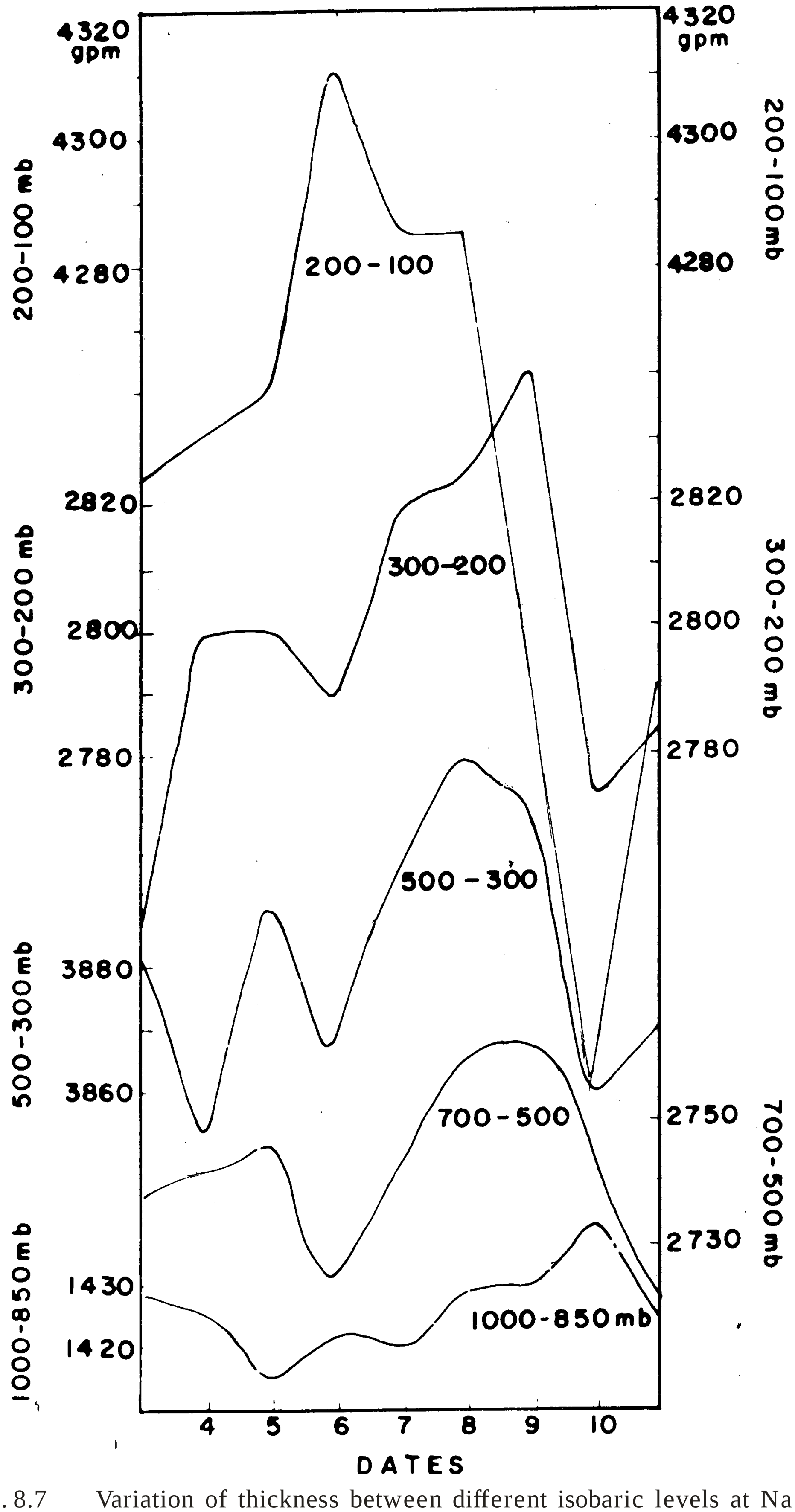

Fig. 8.7 Variation of thickness between different isobaric levels at Nagpur during the 'break' monsoon of August 1965 (Raghavan, 1973)

subsidence, perhaps from the air ascending in such a situation near the Himalayas and extreme south Peninsula. The development of a ridge coincides with this area of subsidence and gives an impression as if the anticyclone from Iran side has extended southeastwards.

8.9 The results of both Ramamurthy (1969) and Ramaswamy (1973) show that middle of August is most susceptible to 'breaks'. Discussing the position of the Equatorial Trough, Riehl (1954) remarks that it moves farthest from the equator in August and even over the continent of Africa it does not reach its highest latitude until late July and early August. The monsoon trough also seems to reach the northernmost position about the middle of August as indicated by the frequency of 'breaks'.

8.10 The sequence of development of 'break' and subsequent revival of monsoon has been studied by different workers. According to Malurkar (1950), the monsoon trough shifts towards the foot–hills of the Himalayas when a depression moves to these mountains, which he associated with more southerly travel of extra–tropical systems. Such a movement of depression always takes the monsoon trough to the foot–hills but the problem is to foresee how long it will remain there, Pisharoty and Desai (1956) found that passage of westerly waves across the Tibetan Plateau and adjoining Himalayas in quick succession leads to 'breaks'. Ramaswamy (1962 & 1965) finds that during 'break' pronounced low index circulation prevails in the middle latitude westerlies north of the Himalayas. Large amplitude troughs protrude into Indo–Pakistan area at 500 mb and aloft. They get retarded and elongated during movement across Tibetan Plateau which is a region of weak basic current. The troughs are highly diffluent. The associated upper divergence causes heavy rainfall over and near the Himalayas. The ridge in the rear of the trough causes the extension of the anticyclone over northwest and central India and northern parts of the Peninsula. When the 'break' lasts for a week or more the large amplitude trough remains quasi–stationary over north India. This type of case is illustrated in Fig.8.8 relating to a 'break' from 7 to 10 July, 1967. Ramaswamy (1971) has presented the 500 mb chart (Fig.8.9) averaged for 4–13 August 1965, for a break period showing the prevalence of a mid–latitude trough of large amplitude. Whether all 'breaks' conform to this sequence does not seem to have been verified.

Raman (1955) associates 'breaks' with a typhoon or its remnant in the western Pacific moving to the north of Lat. 30^{o} N, provided there is no depression or typhoon in the China Seas south of 30^{o} N nor unsettled conditions or a depression already present in the Bay.

Koteswaram (1950) connects 'breaks' with westward moving lows at low latitudes (10^{o} N) in the Bay, prominent at 700 mb. Out of 19 years with 'breaks', in eleven cases, 'break' commenced within two days (earlier or later) of the appearance of the 'low'. Movement of such lows weakens the north–south pressure gradient over the Peninsula. After spanning the Bay, some lows may move along the east coast to central and adjoining north Bay or after crossing into the Arabian Sea move along the west coast. When it moves to a sufficiently high latitude, the normal pressure gradient is reestablished and monsoon revives. Fig. 8.10 illustrates such a case. The displacement southward of the eastern cell of the sub–tropical high pressure belt in mid–troposphere, to the positions of premonsoon months, favours movement of weak low pressure systems at its southern periphery across Tennasserim and south Bay instead of upper Burma north of 15^{o} as when monsoon is strong (Desai, 1970).

8.11 Monsoon most often revives after a 'break', by the formation of a low or depression in the north Bay. Gradual formation of the monsoon trough in the normal position also takes place sometimes.

8.12 Not all 'breaks' seem to develop in the same manner and any relationship between the three sequences described in para 8.10 viz. the system in the westerlies, the low pressure areas in low latitudes and the typhoons of the China Seas, has to be further studied. It has not been possible to link all the shifts in the position of the monsoon trough satisfactorily with other synoptic

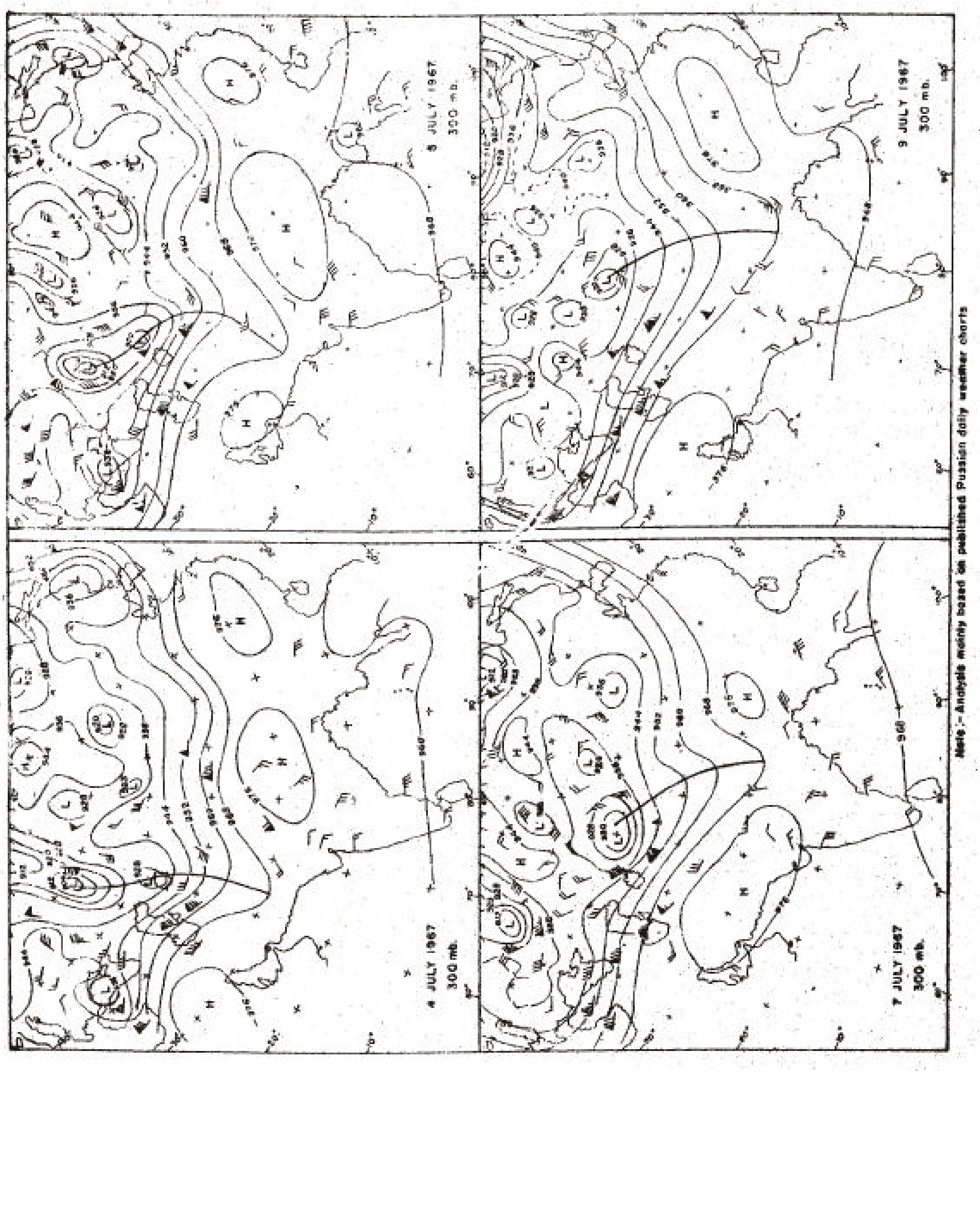

Fig. 8.8 Passage of trough in westerlies during 'break'

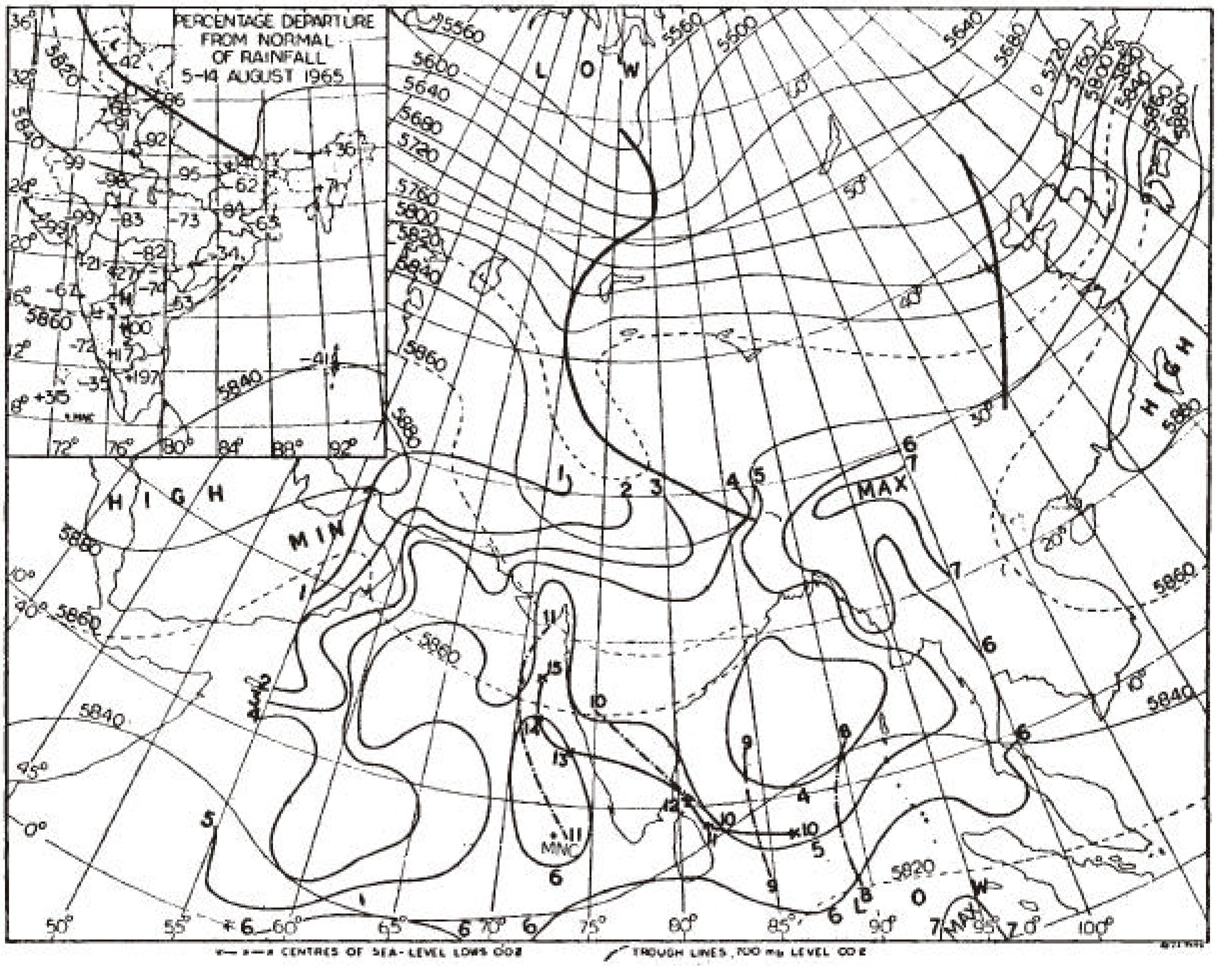

Fig. 8.9 - Mean contours (500 mb) 4 - 13 August 1965.
Mean cloudiness (Satellite) in oktas - 5-14 August 1965
(The figures by the side of the crosses indicate the dates and the figures at the two ends of the trough lines at 700 mb. level over south Bay of Bengal and south Peninsula indicate the corresponding dates.)

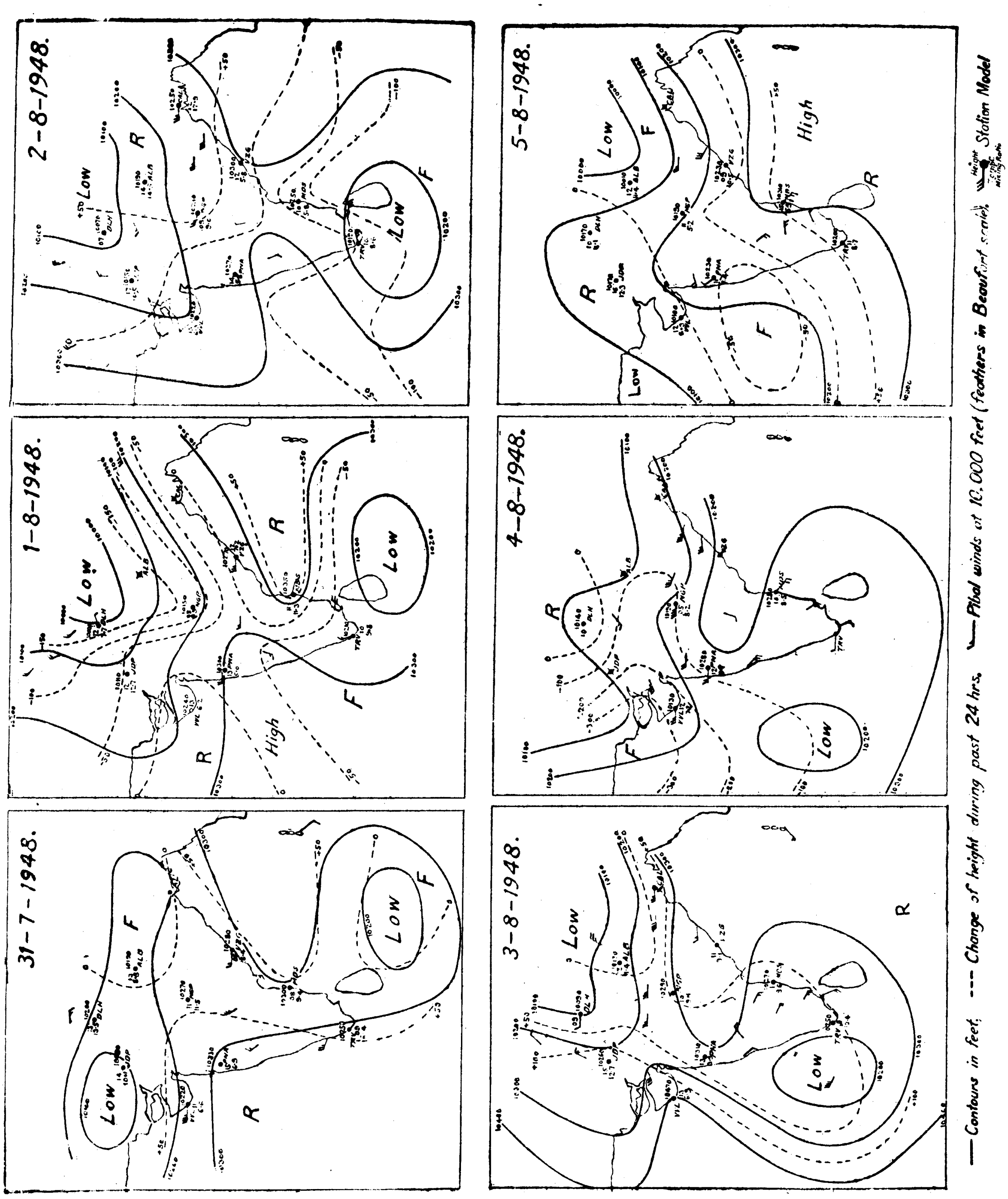

Fig. 8.10 Constant pressure charts for 700 mb

conditions to provide a basis for forecasting and extrapolation from short–period trends has, therefore, often to be resorted to.

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CHAPTER 9

OTHER SYNOPTIC SYSTEMS

9.1 Weather during the monsoon period varies from one area to another and from one day to another over the same area. These variations are connected with synoptic patterns in surface and upper air. The most important system of monsoon depressions has been dealt within an earlier chapter and so also the monsoon trough which is of a semi–permanent nature. Other synoptic systems will be covered in this chapter. Most systems extend upto mid–troposphere.

9.2 Lows

9.2.1 Low pressure areas, less intense than monsoon depressions, also form quite frequently during the monsoon and are responsible for causing substantial rainfall. Excluding the residuary lows out of weakening monsoon depressions, the lows fall into two distinct categories – (i) which form over northwest and adjoining central Bay just like depressions but do not develop upto depression intensity and (ii) which develop in situ over northern India. The following table (from Srinivasan, Raman and Mukherjee 1972) gives statistics about these lows for a 20–year period (1950–1969) :

Table 9.1

	Jul.	**Aug.**	**Sep.**
No. of lows developed over Bay	14	8	14
Average life (days)	5.5	4.0	6.5
No. of lows developed over land	22	28	1
Average life (days)	4.6	4.2	5.0

On the whole. the lows forming over land are at a more northerly latitude than the mean track of monsoon depressions.

9.2.2 In about a quarter of the cases, the precursory signs of depressions over the north Bay and adjoining central Bay do not develop sufficiently to form two closed isobars at 2 mb interval. With only one closed isobar, systems move inland as a low. Precursory signs and later behaviour of these weaker systems also are very much like those of depressions. Although movement is less regular and slower, life period and rainfall pattern (amounts being less, though upto 20 cm) are alike, In vertical extent also they are comparable to monsoon depressions. The diffuseness of the pressure field may cause variable extent and movement may not be so systematic

The low of 14th to l6th July 1962 discussed by Rao et al (1970) illustrates all the above features, The remnant of a depression which crossed Viet Nam coast emerged as a low into northeast Bay on 14th (Fig. 9.1a). Figs. 9.1b & c clearly show the movement, though all system influencing formation of lows in north Bay may not be as intense. By l6th morning (Figs. 9.1 d & e) this low had moved to Bihar Plateau and caused rainfall, like a depression. The low reached southwest Uttar Pradesh and adjoining east Rajasthan by 18th (Fig. 9.1 f).

9.2.3 The other type of lows forms over land north of Lat. 22° N, mostly over Bihar, north Madhya Pradesh and south Uttar Pradesh, Their mode of formation seems rather similar to

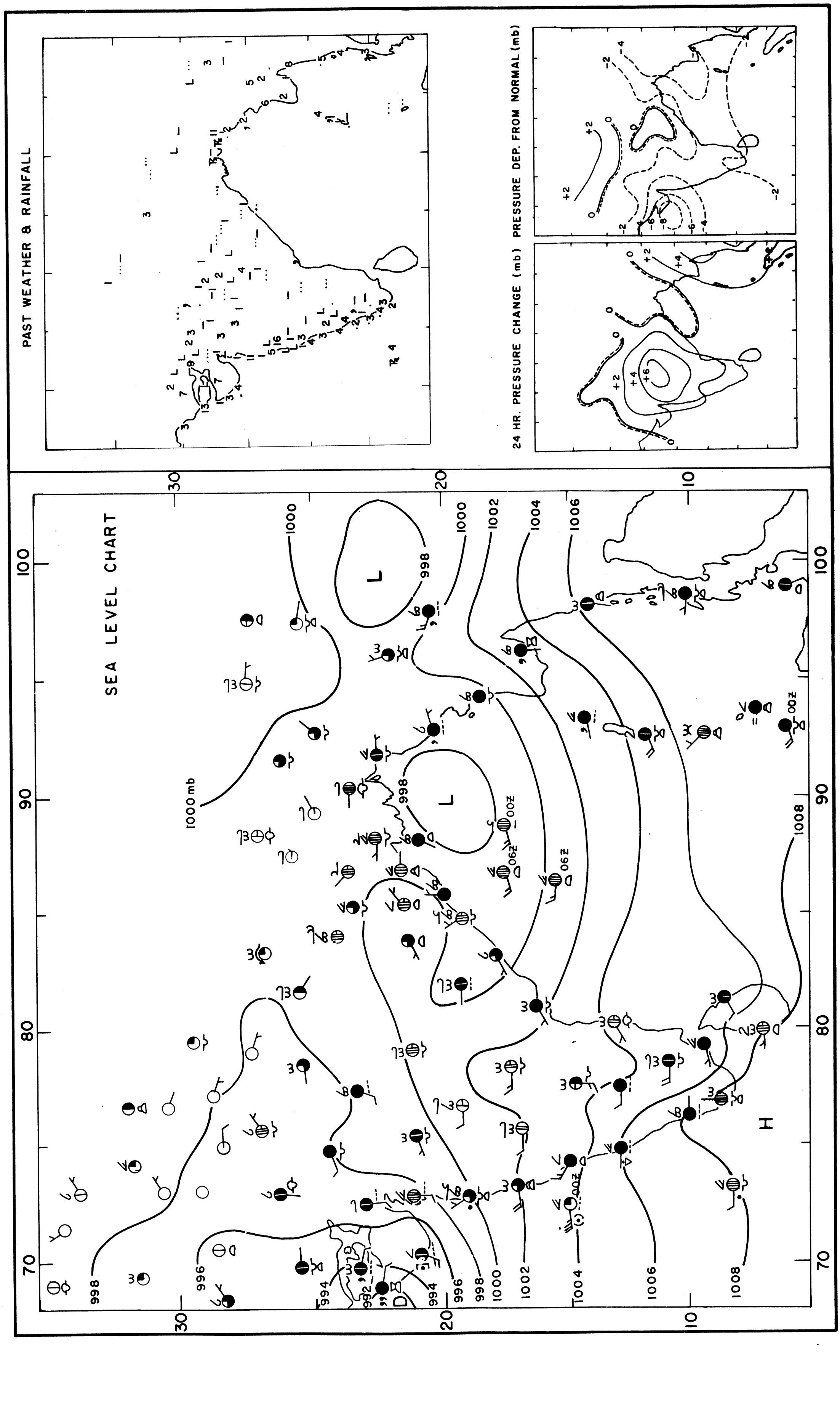
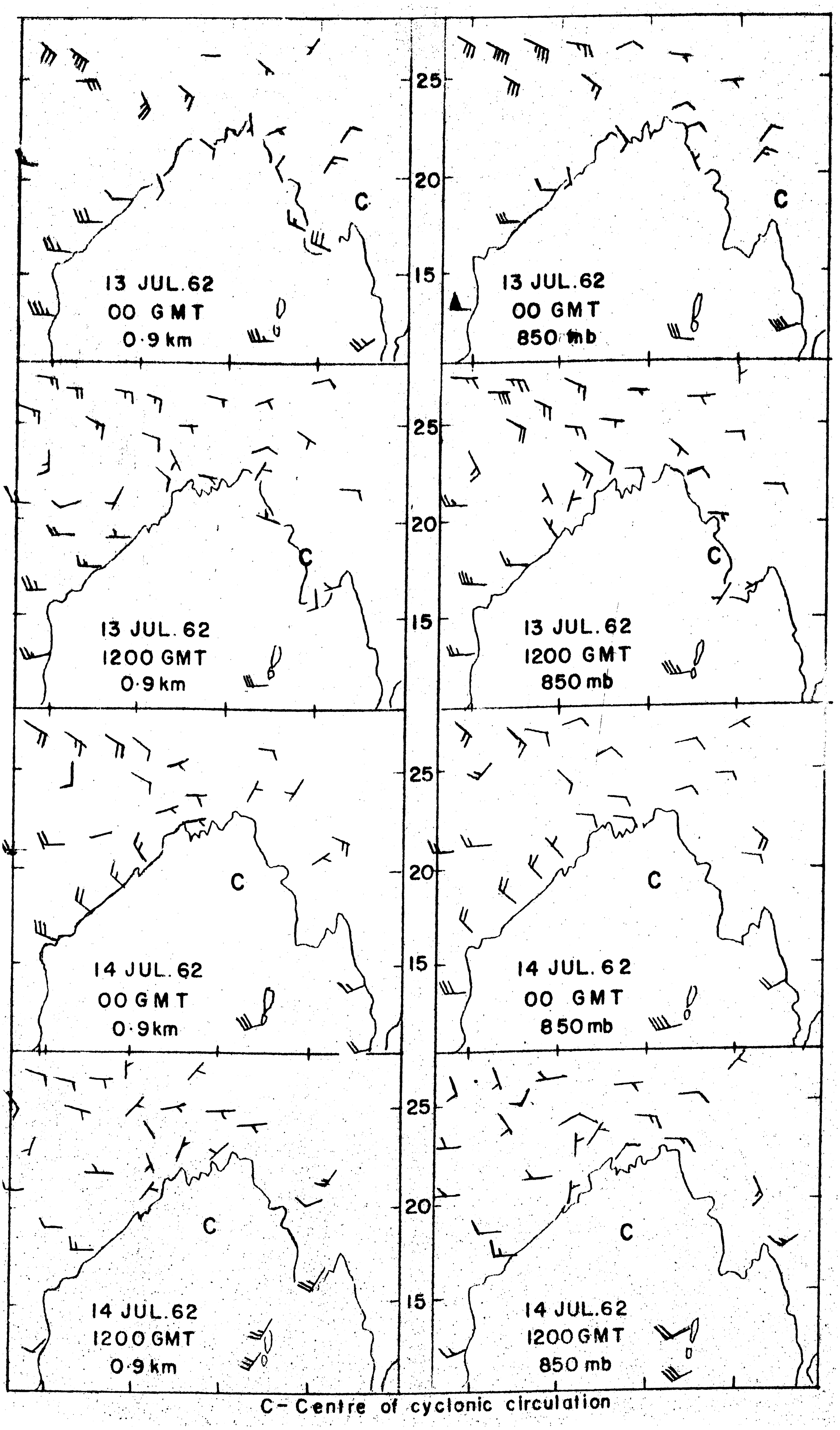


Fig. 9.1 (a) Synoptic charts 0300 GMT 14 July 1962



C – Centre of cyclonic circulation

Fig. 9.1 (b) Upper Winds

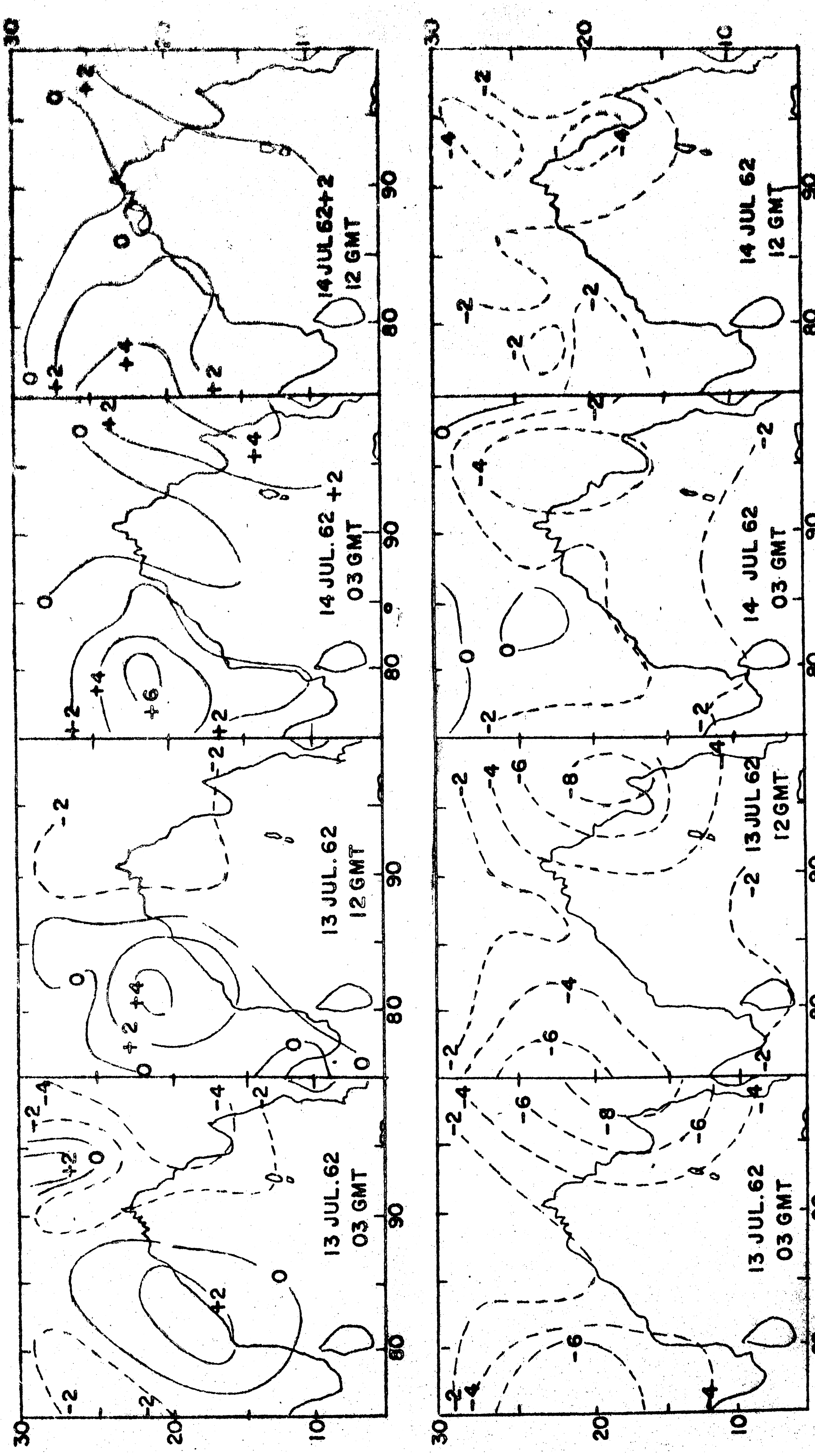


Fig. 9.1 (c) 24 hours pressure change (mb) (upper), Pressure Departure (Lower)

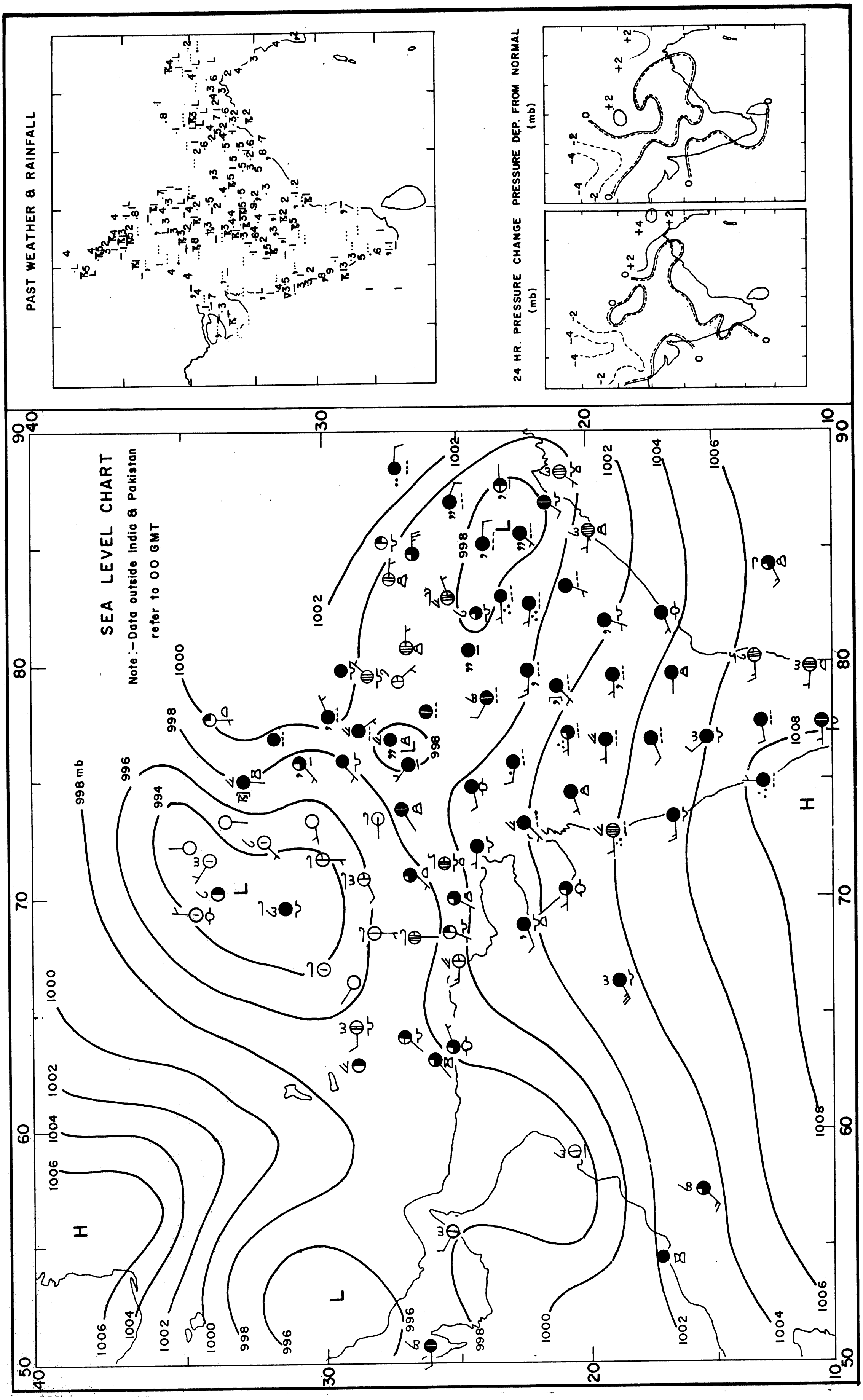


Fig. 9.1 (d) Synoptic charts 0300 GMT 16 July 1962

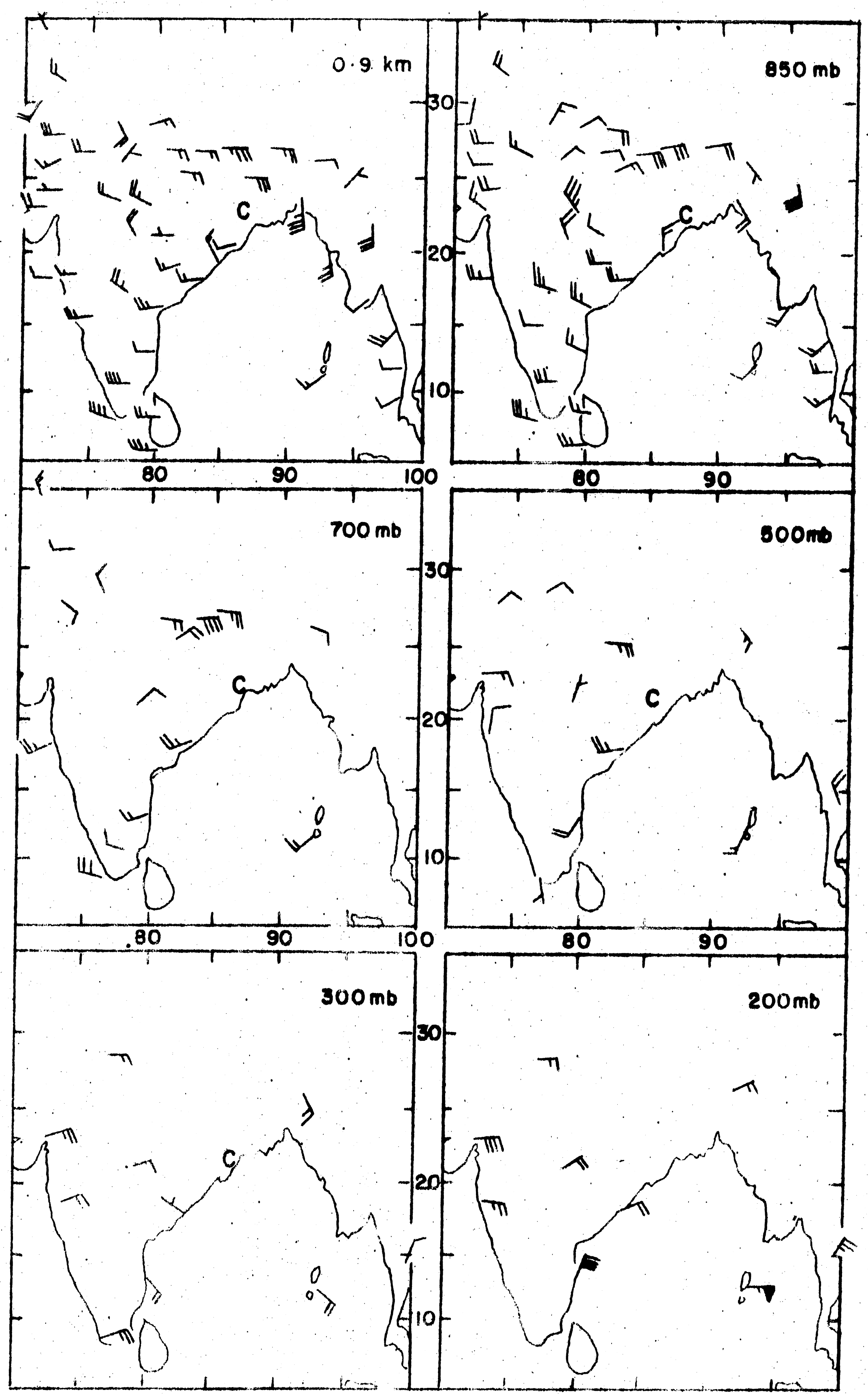


Fig. 9.1 (e) Upper Winds 16 July 62 00 GMT

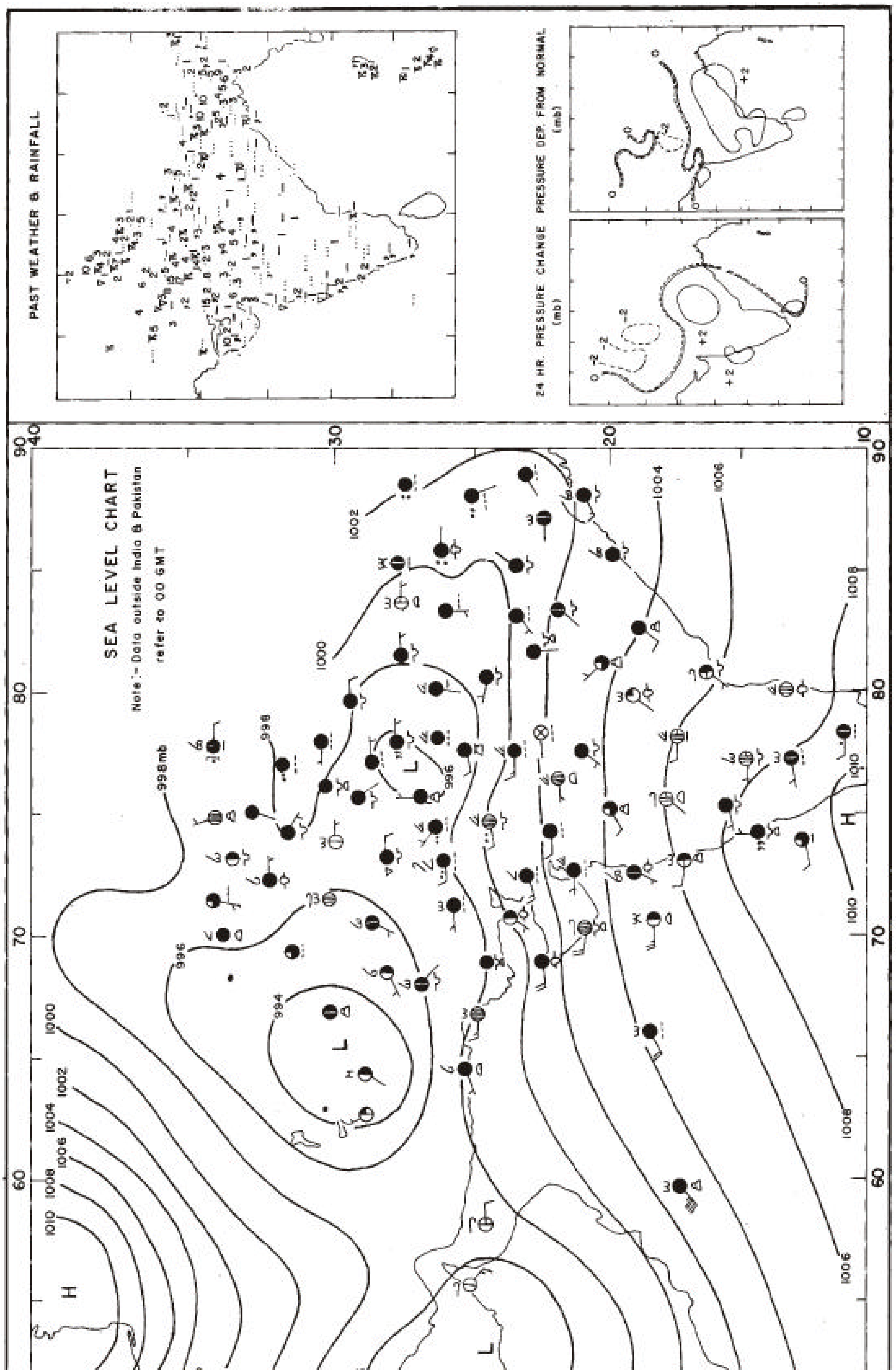


Fig.9.1 (f) Synotic Charts 0300 GMT 18 July 1962

induced lows' in the winter season, under the influence of troughs in middle latitude westerlies at mid–and upper tropospheric levels, though no formation without a westerly trough has not been established. With a pre–existing monsoon trough, a slight fall in pressure can lead to the appearance of a low for short periods. But in cases studied in some depth, the influence of a trough in westerlies seem to have prevailed. Such lows also dissipate when the westerly system of higher latitudes moves away eastwards. Surface lows develop about 3 to 5 degrees longitude to the east of 500 mb trough. As the former circulation appears at surface and in the lower troposphere, it is very difficult to link up with the higher latitude system on account of the Himalayan massif. Movement of the southern periphery of the westerly troughs across Pakistan and northwest India is considerably masked by the prevailing sub–tropical ridge in mid–troposphere and aloft. Thus the success in linking up the development of such 'lows' with the extra–tropical westerly troughs is better in diagnostic studies rather than in operational weather forecasting.

These land lows are generally smaller in extent than the lows arriving from the east. Formation is mainly to the west of Long. 85° E. Life period is about two to three days. They give concentrated rainfall when the system underlies the forward portion of the mid–tropospheric trough. Preferred sectors of rainfall have not been noticed. The system shows slow movement eastward. Pressure changes and departures from normal have to be carefully watched to identify areas of development. In rare cases such systems have deepened upto depression stage.

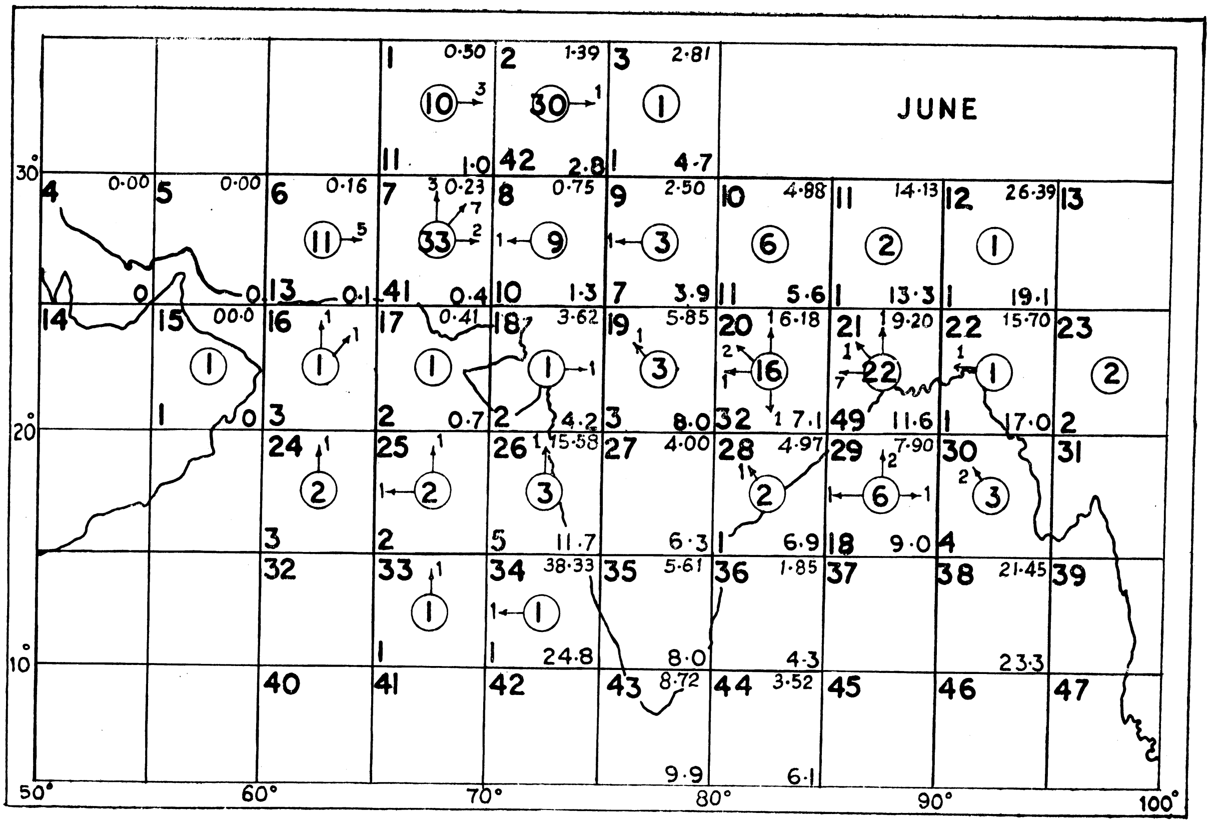
9.2.4 Tripathi (1963) has prepared diagrams showing the relation between cyclonicity and rainfall over the Indian region. ' Cyclonicity' denotes a synoptic situation in which at least one closed isobar exists over a 5 degree square with associated cyclonic circulation upto 1.5 km or more. Fig, 9.2 (a to d) show (i) number of cyclonic systems in 5 degree square (within circle at centre), (ii) numbers which moved in different directions (indicated by arrows), (iii) total number of days systems lay in that square (figure at lower left hand corner), (iv) average monthly rainfall per station in inches (figure at upper right–hand corner) and (v) average monthly number of rainy days (figure at lower right–hand corner). The data used are for 1944 to 1953. Excluding areas where orography has dominating influence on rainfall, there is a correlation of 0.55 (n = 56) between number of rainy days and number of days of cyclonicity. Cyclonic flow at surface causes frictional convergence which is important in building up vertical motion. It is also possible that low level convergence causing rainfall is itself responsible for developing cyclonic vorticity and low.

9.3 <u>Trough in Monsoon Westerlies</u>

9.3.1 During weak monsoon conditions, in the westerly sweep across northern India, weak troughs are noticed to develop in the lower troposphere over the plains, mainly to the east of the longitude 80° E. They cause increased rains for about 2° longitude on either side and farther to the east. Rainfall is not widespread and heavy as in the field of depressions and lows. But the increase of rainfall from a 'break' situation is quite marked, scattered heavy falls also occurring. These troughs are more numerous, in July than in other monsoon months. The possible influence of westerly troughs of higher latitudes has been traced in some cases. Occasionally extension of the lower tropospheric trough in monsoon westerlies into the Bay has been the starting point for depression formation.

9.3.2 A case of a well–marked north–south trough in monsoon westerlies during a 'break' situation, from 23 to 27th August 1966 reported by Rao, Srinivasan, Ramakrishnan and Raman (1970) and Srinivasan, Raman and Mukherjee (1972) is shown in Fig. 9.3 (a to j). The trough appeared on 22nd over the central parts of the country along long. 80°/83° E, upto 500 mb, moved to 89° E by 24th and 93° E by 26th when it was weakening. A middle latitude westerly trough also seems to have been moving from 22nd to 25th, north of 30° N in upper troposphere in nearly the same longitudes. Rainfall increased on 24th and 25th near the trough line, particularly

(a)



(b)

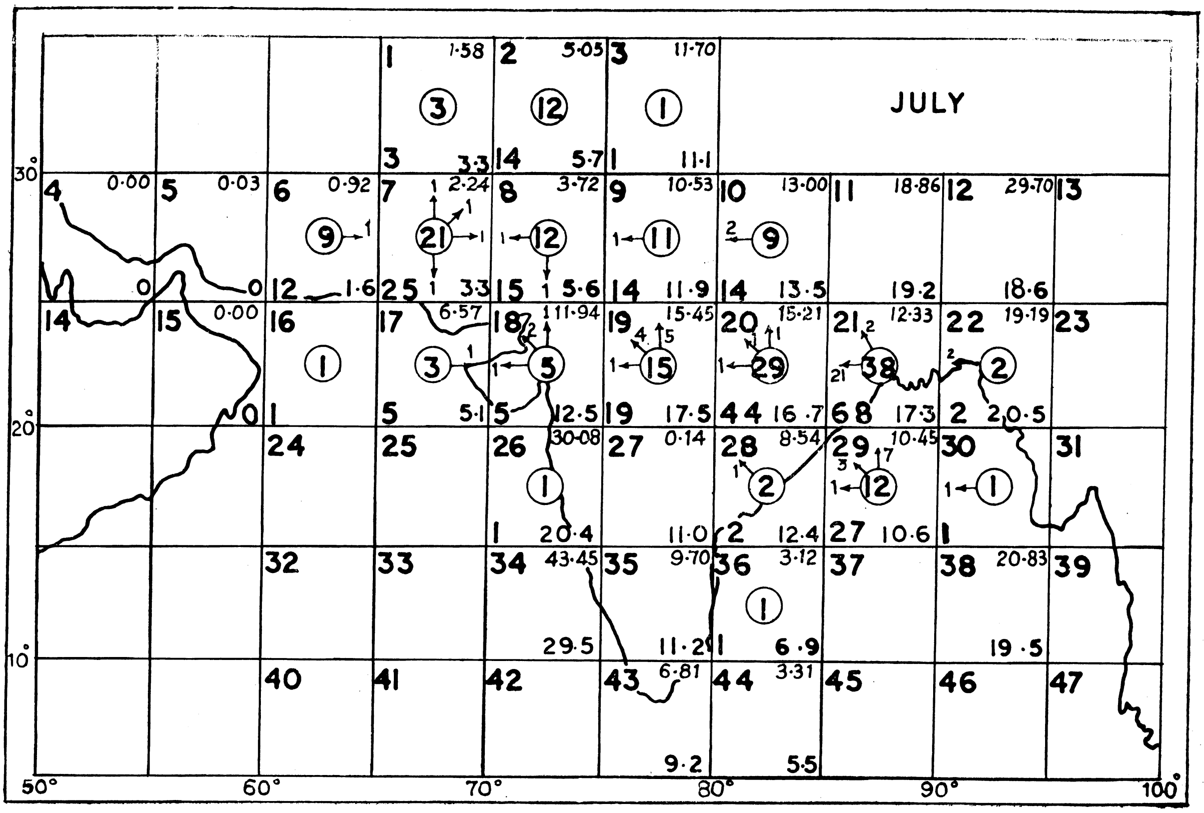
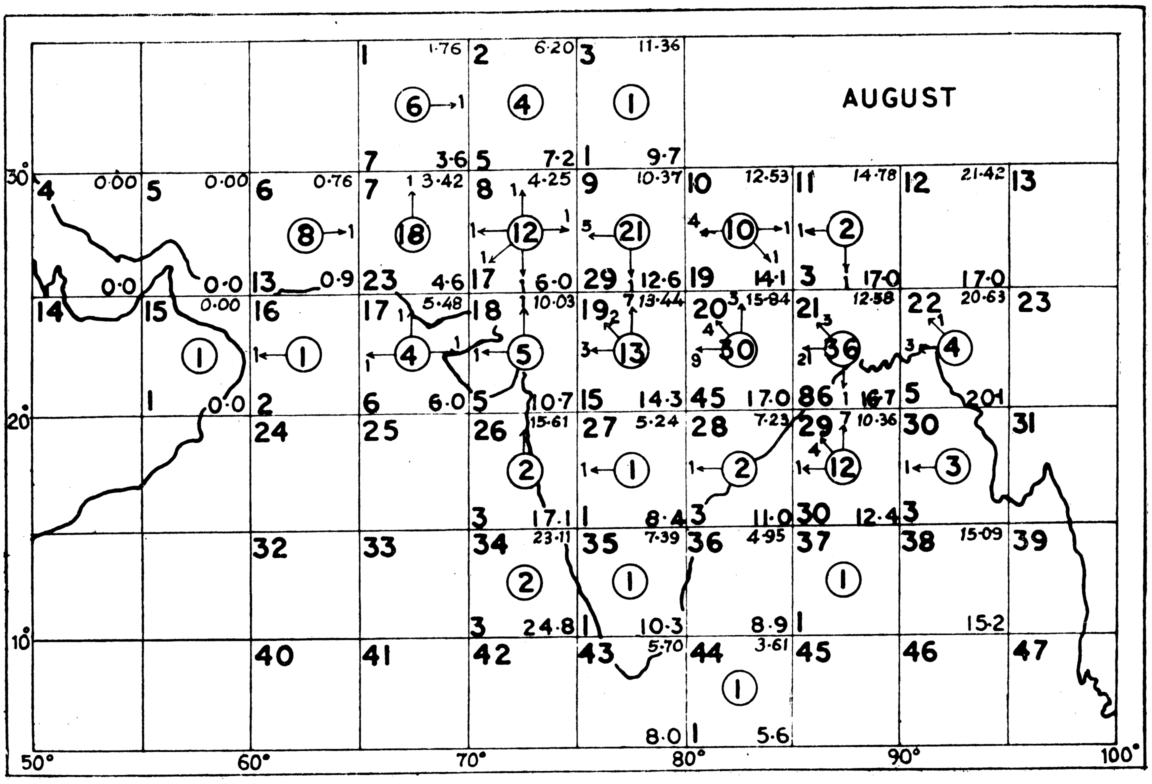


Fig.9.2

(c)


AUGUST
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10°
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70°
80°
90°
100°

(d)

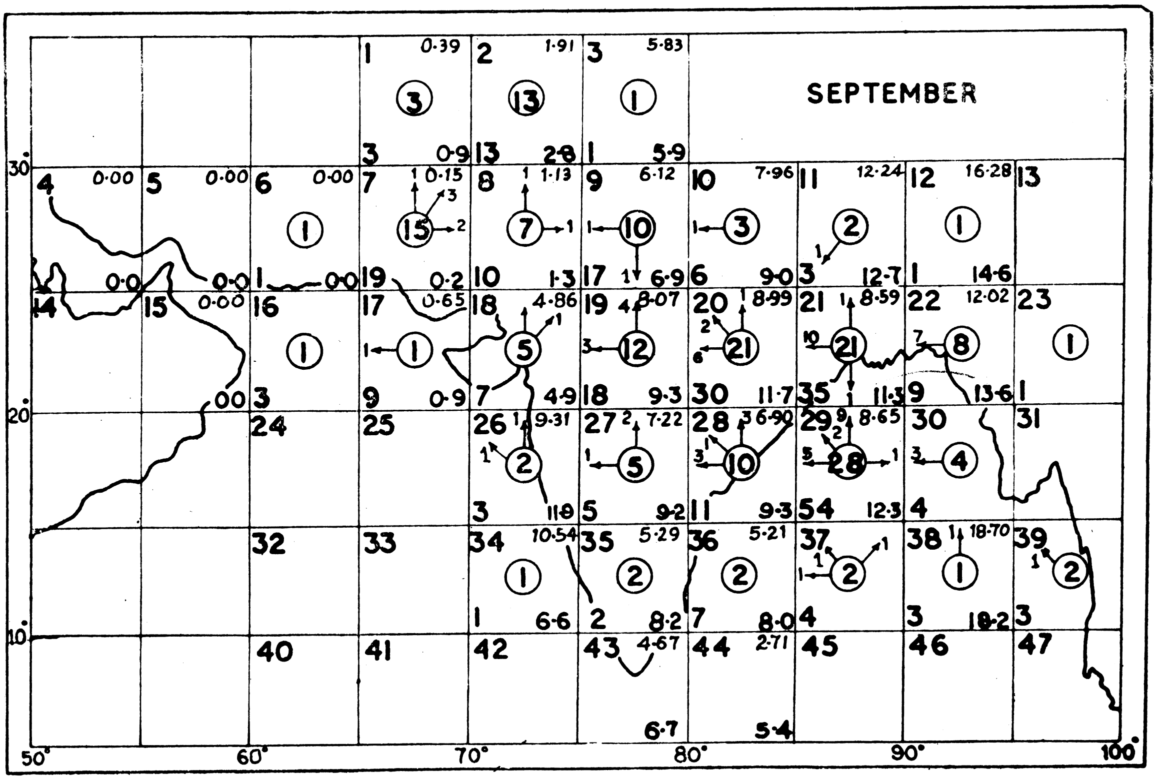

SEPTEMBER
30°
20°
10°
50°
60°
70°
80°
90°
100°

Fig. 9.2

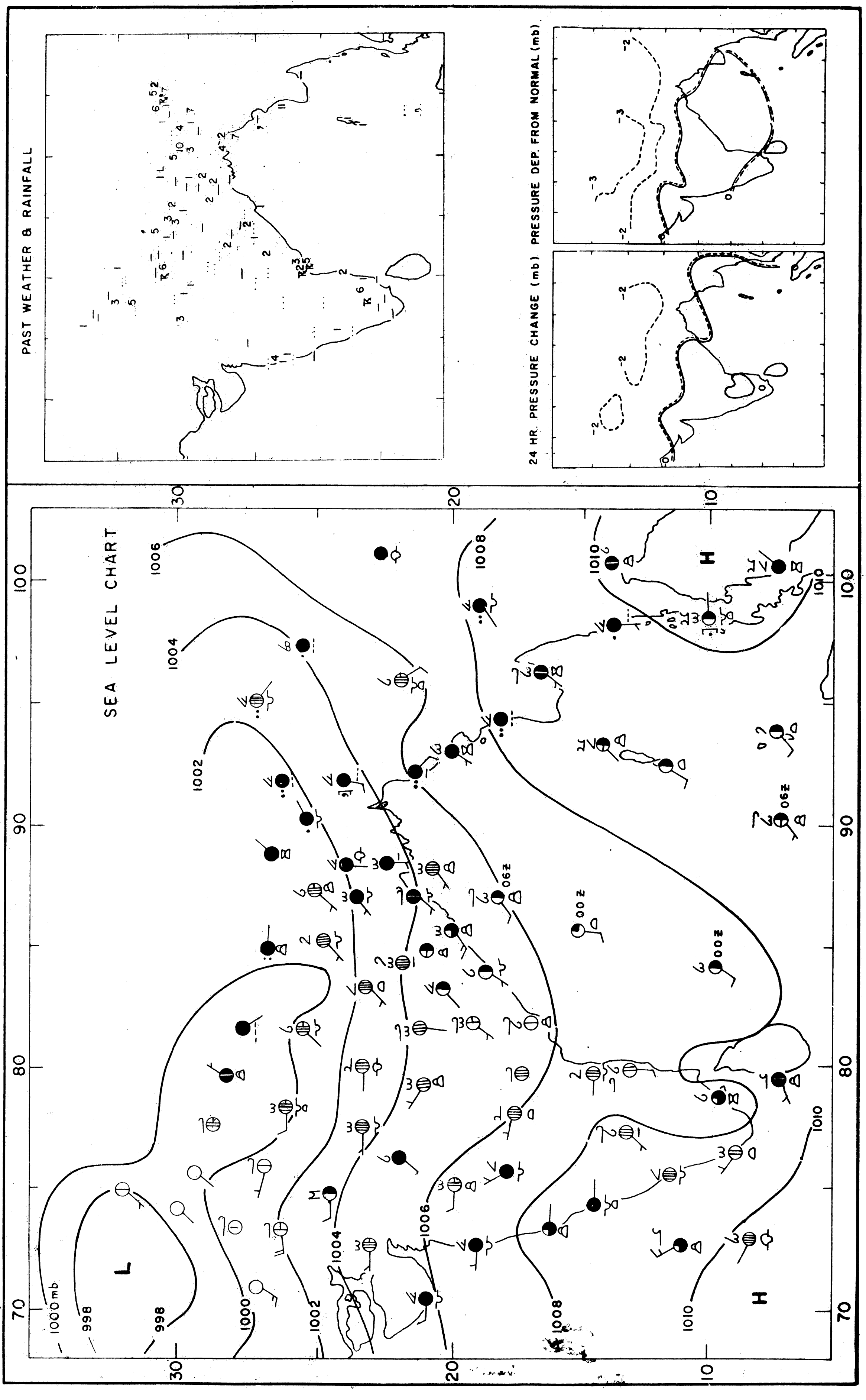


Fig.9.3 (a) Synoptic Charts 0300 GMT 22 August 1966

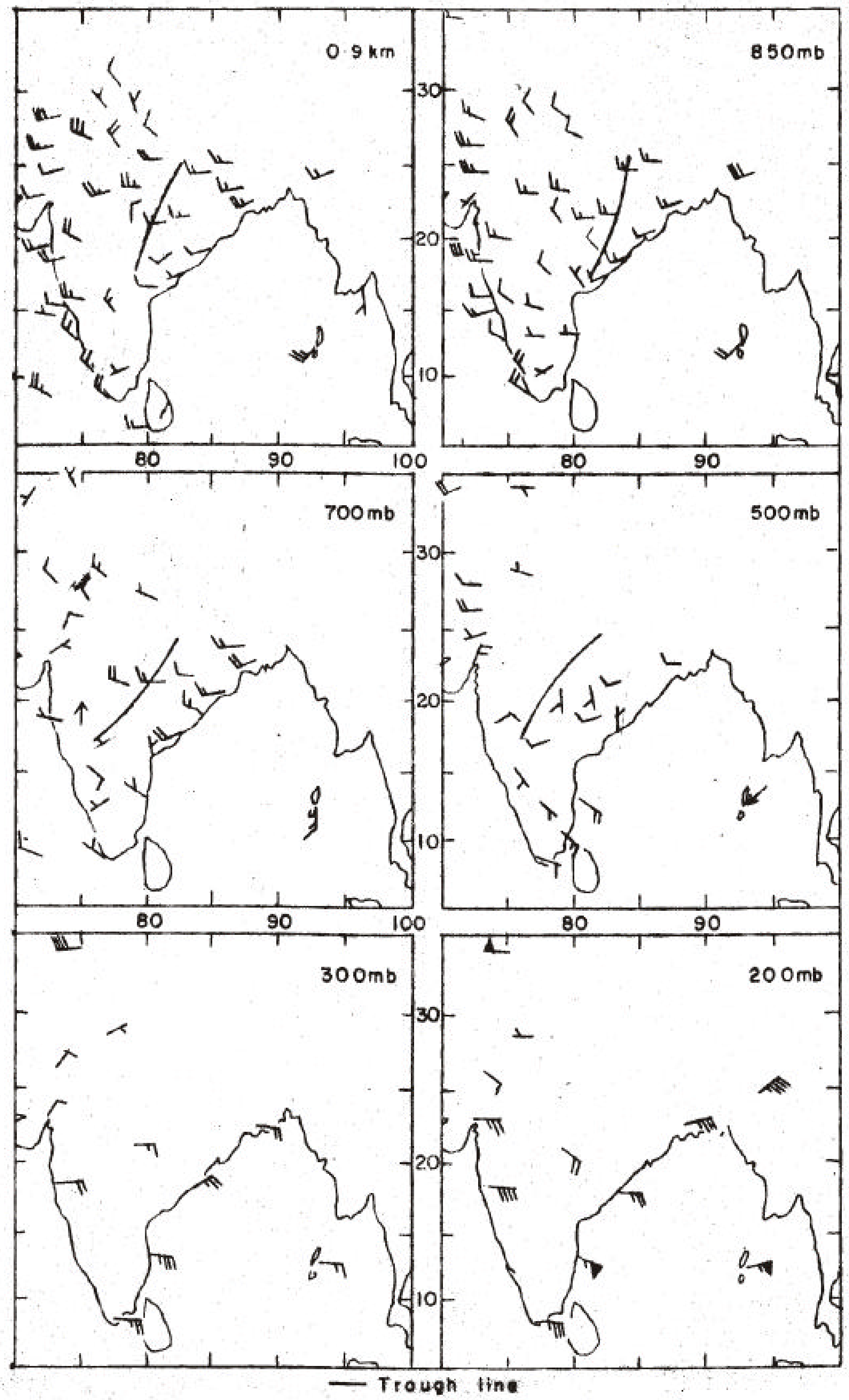


Fig. 9.3(b) Upper Winds 22 Aug. 1966 00 GMT

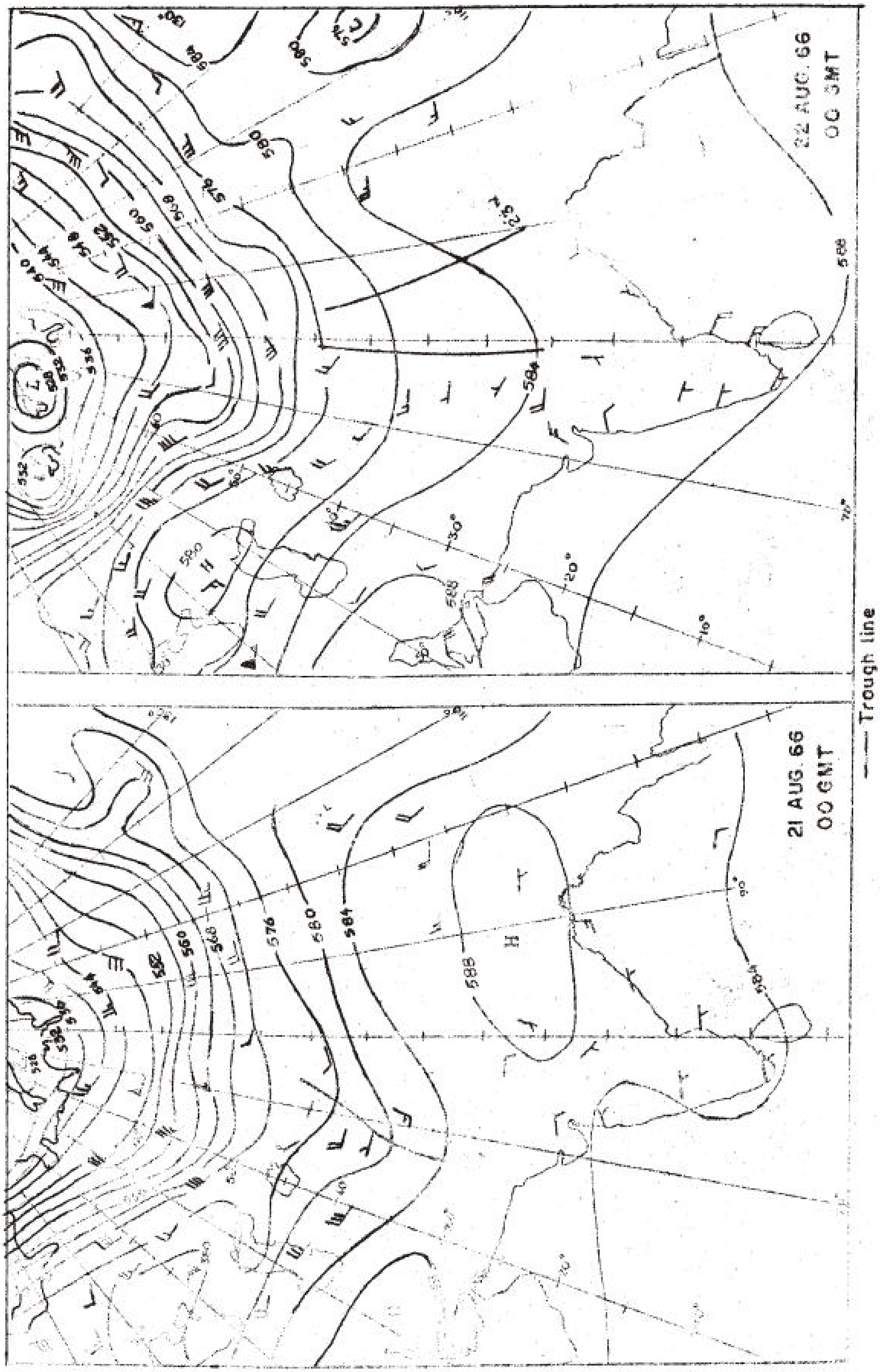


Fig. 9.3(c) Upper air chart (500 mb)

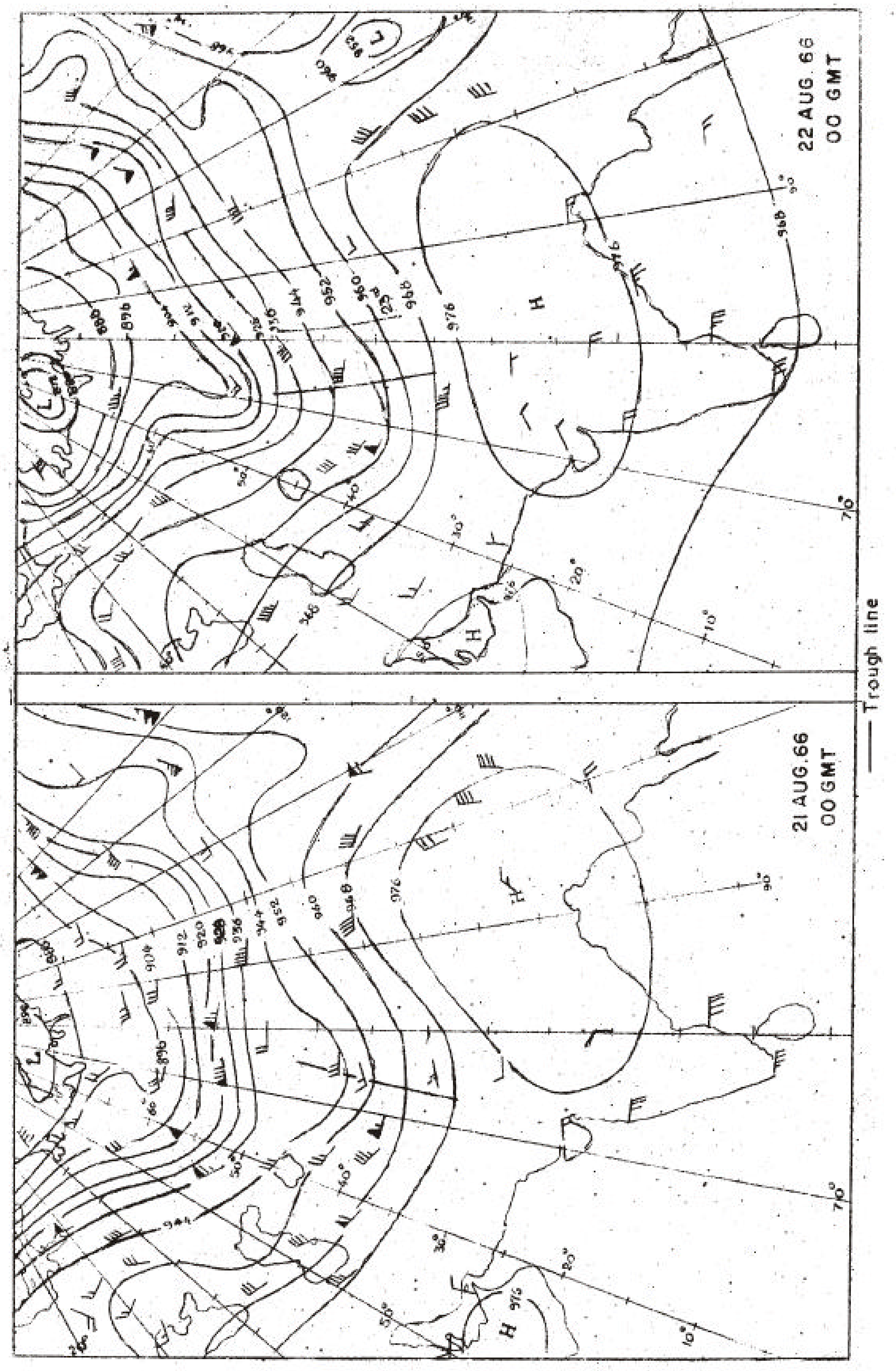


Fig. 9.3(d) Upper air chart (300 mb)

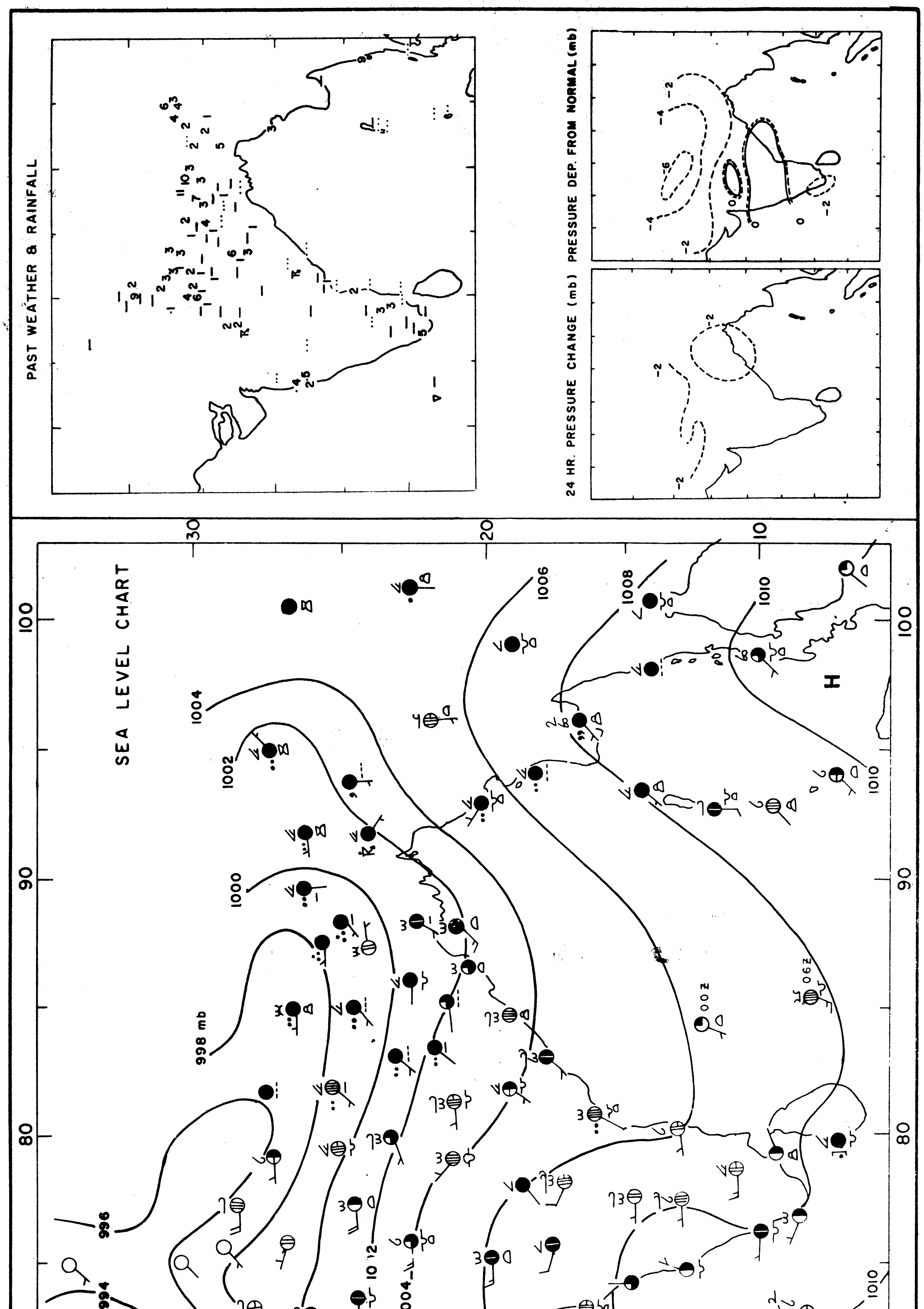


Fig. 9.3(e) Synoptic Charts 0300 GMT 23 August 1966

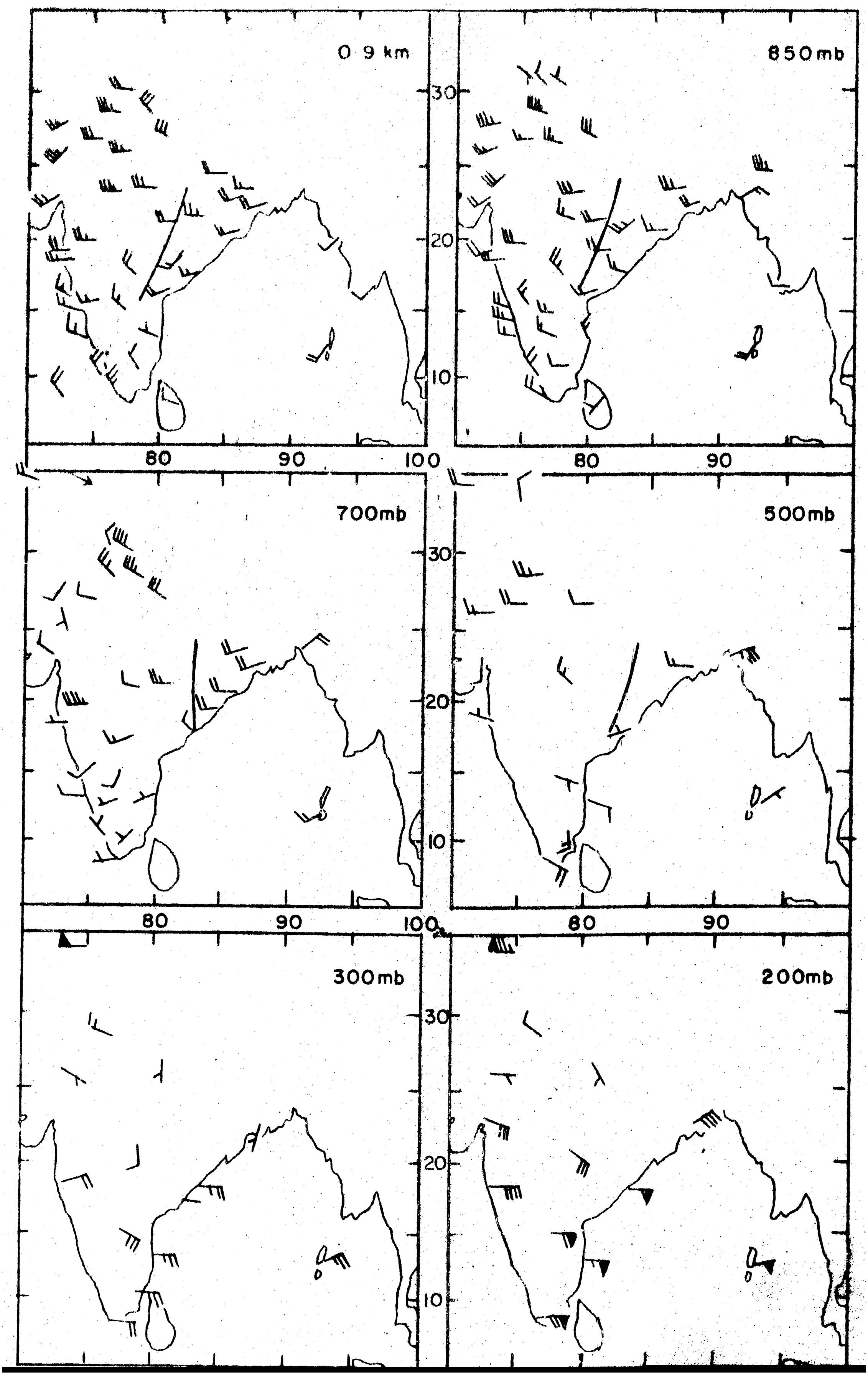


Fig. 9.3(f) Upper Winds 23 Aug. 1966 00 GMT

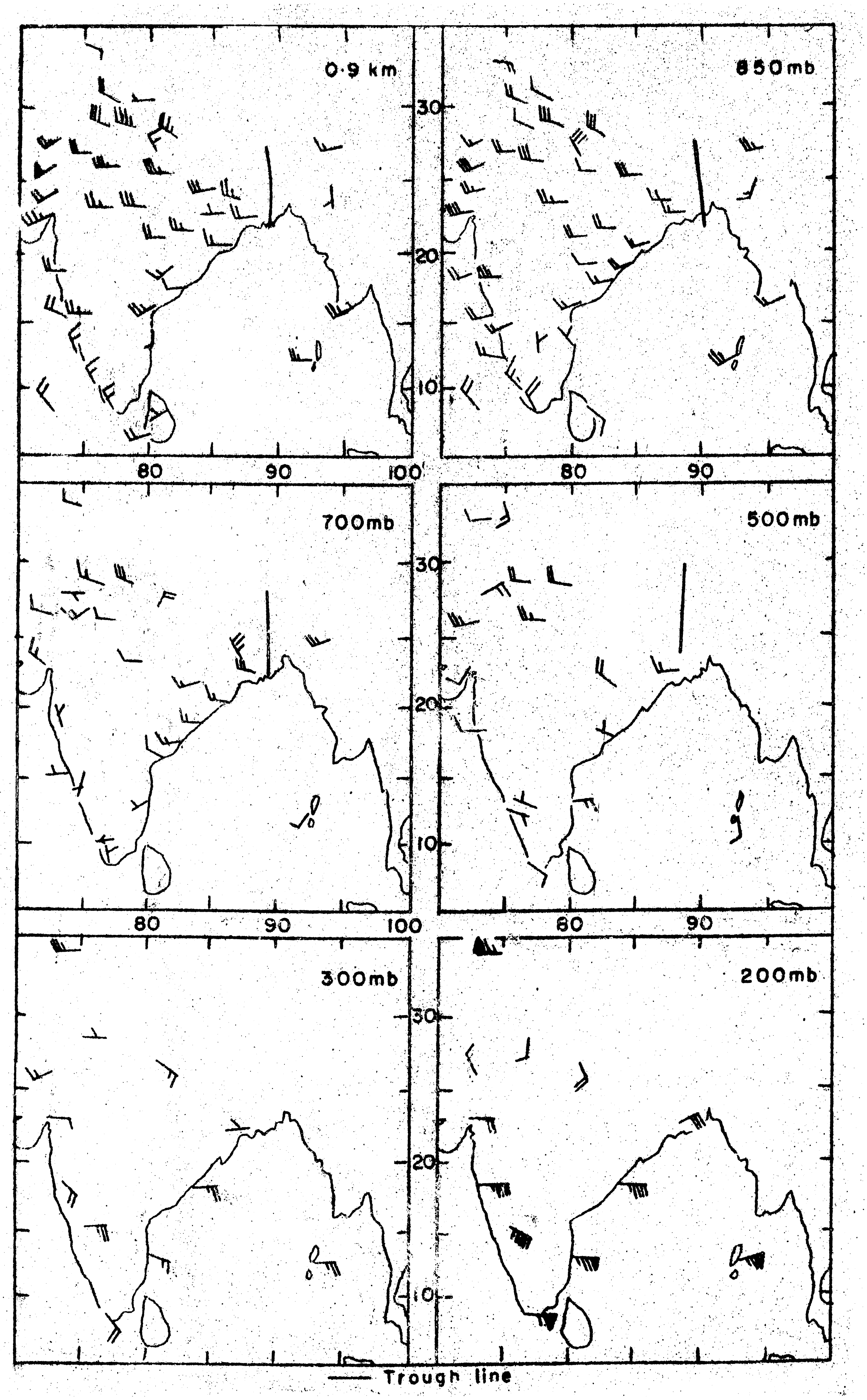


Fig. 9.3(g) Upper Winds 24 Aug. 1966 00 GMT

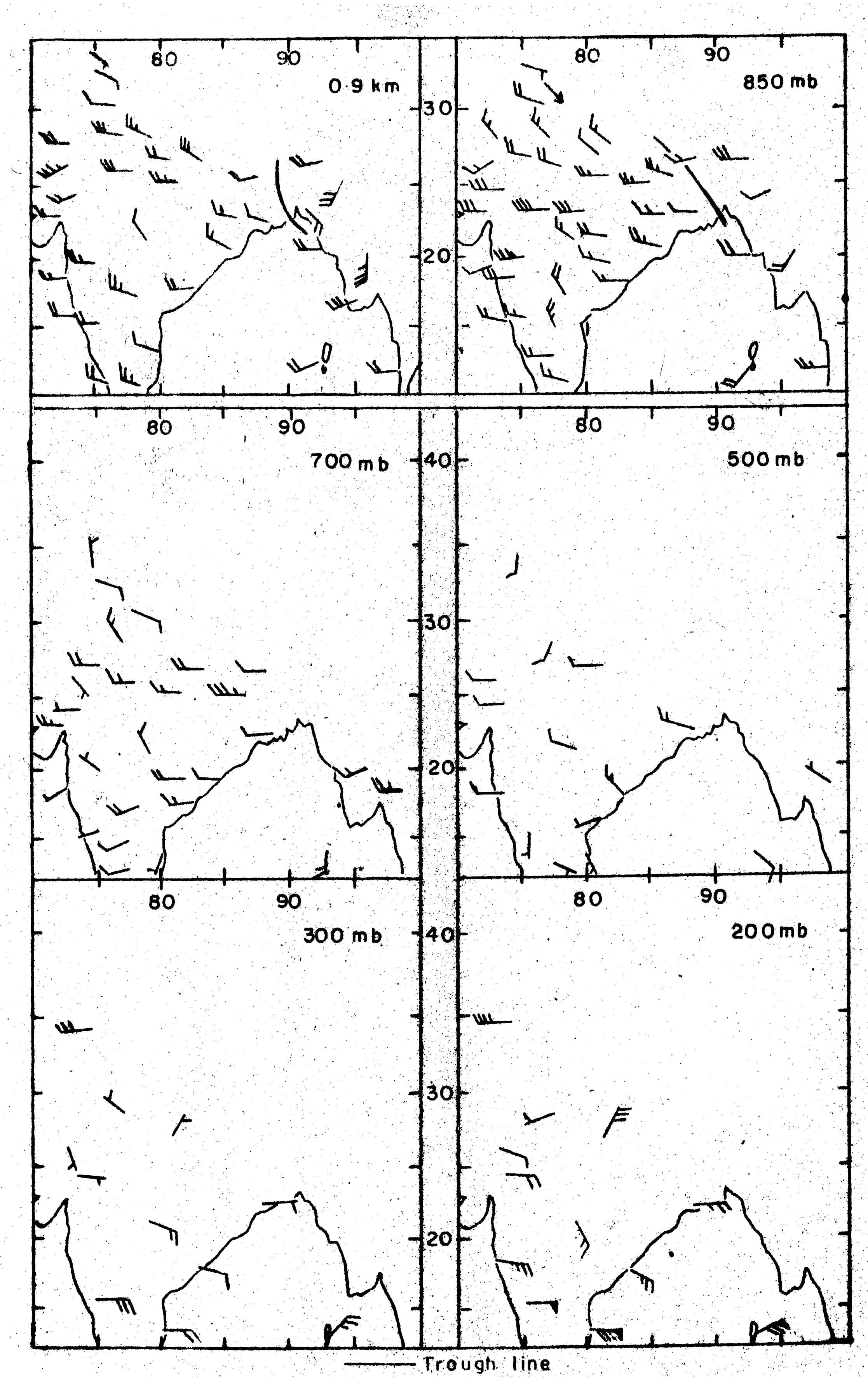


Fig. 9.3(h) Upper Winds 25 Aug. 1966 00 GMT

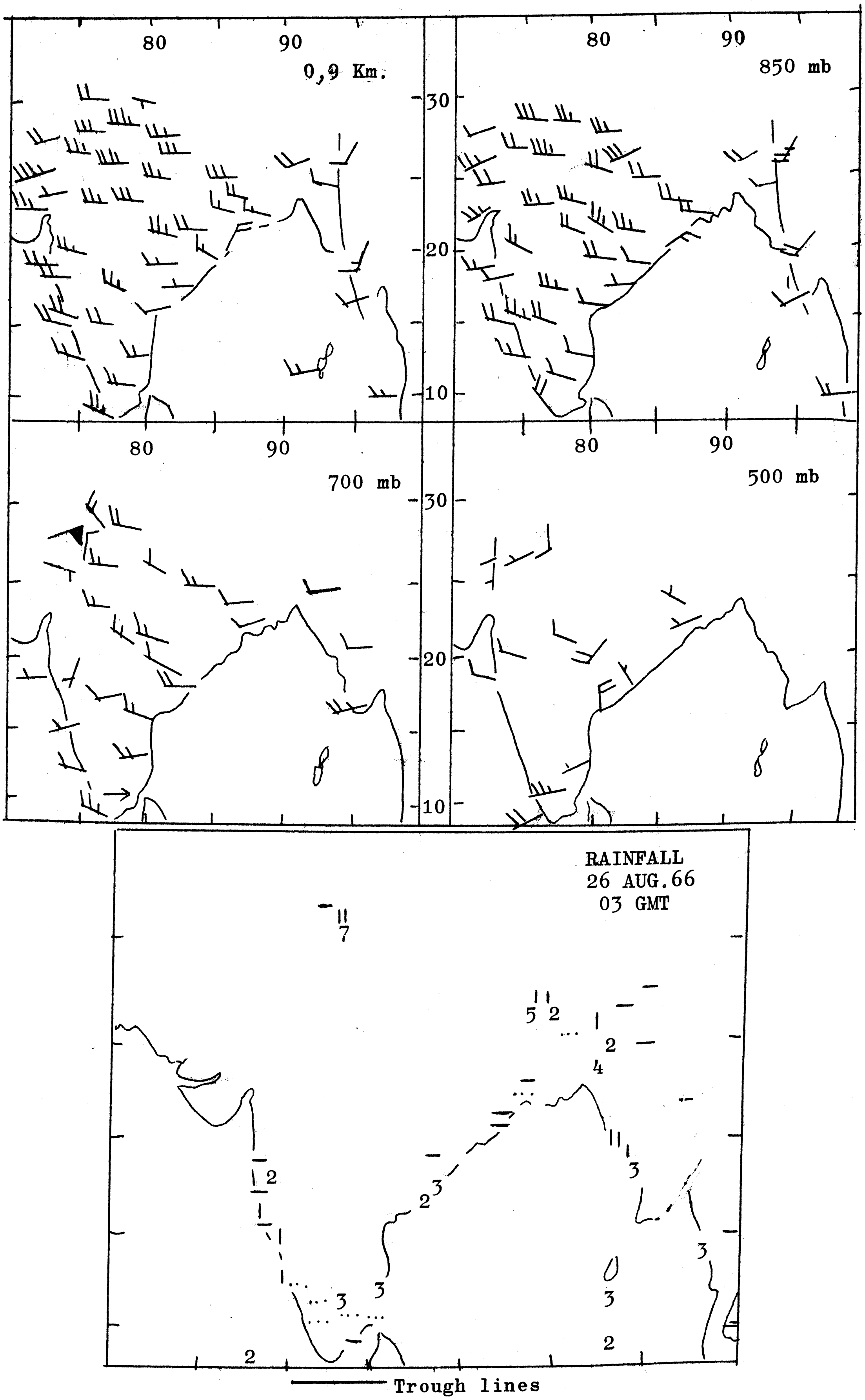


Fig. 9.3(i) Upper Winds 26 Aug. 1966 00 GMT

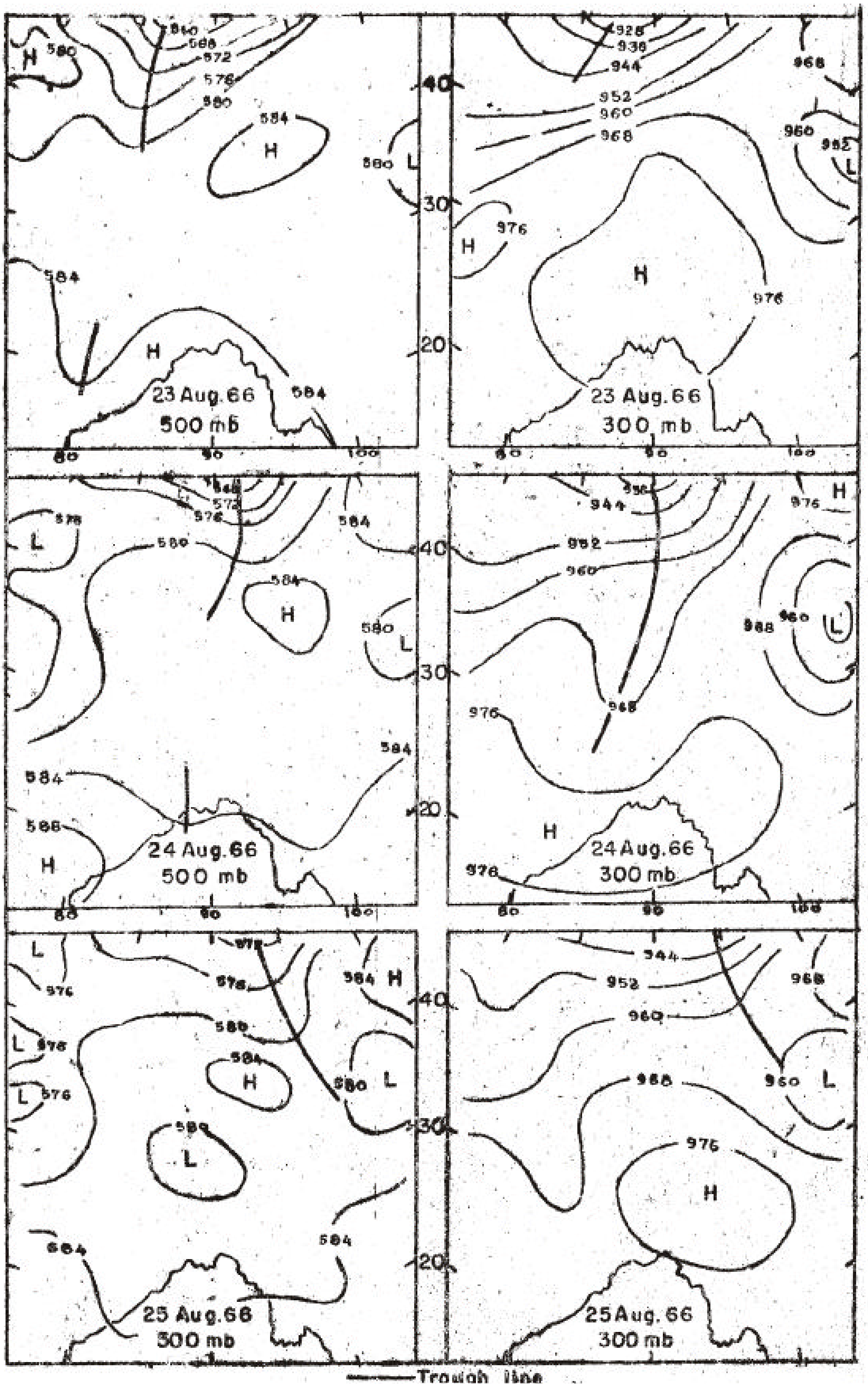


Fig. 9.3(j) Upper Air Chart 00 GMT

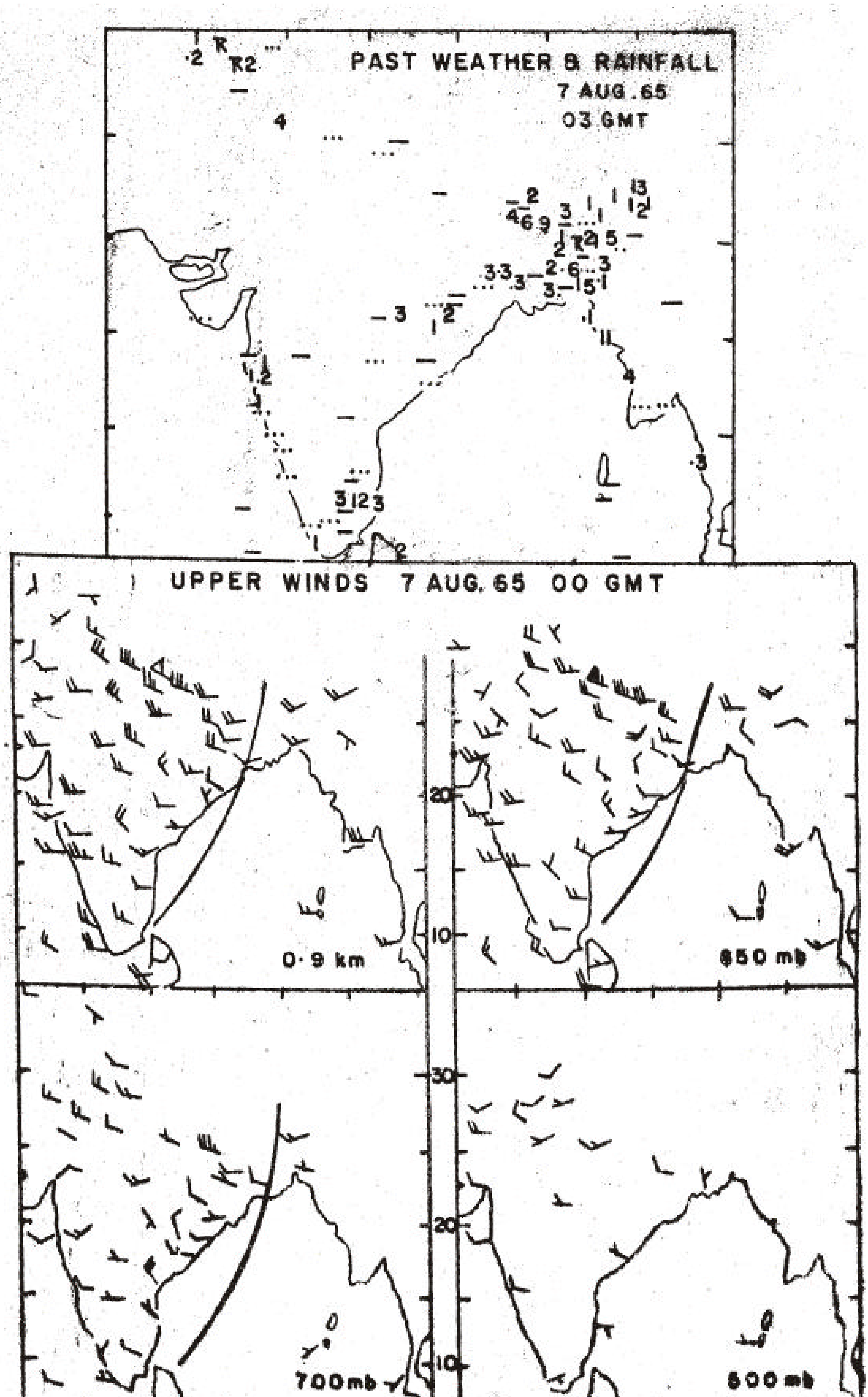

PAST WEATHER & RAINFALL
7 AUG. 65
03 GMT
UPPER WINDS 7 AUG. 65 00 GMT
0·9 km
850 mb
700mb
500 mb
Trough line

Fig. 9.4(a)

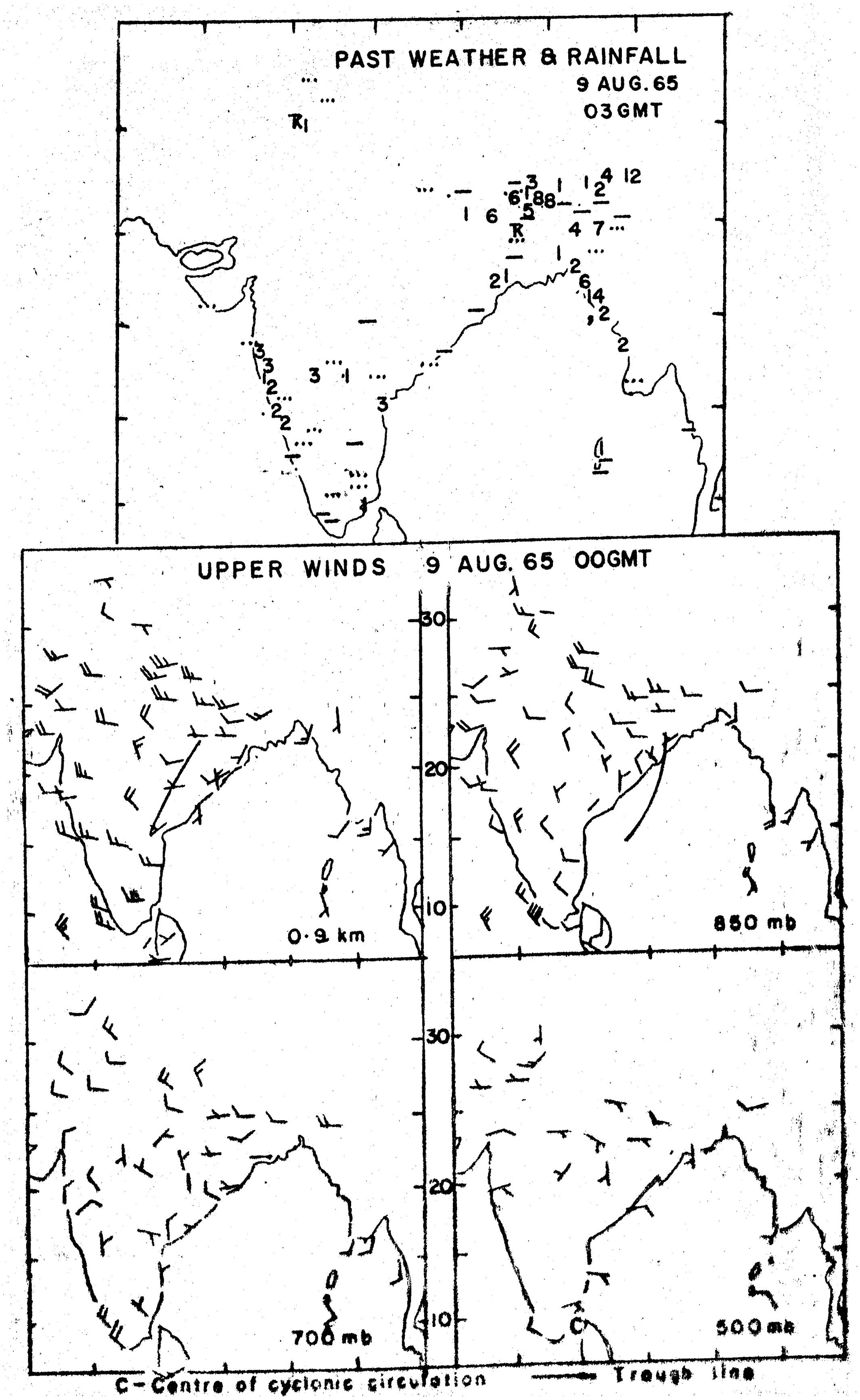

PAST WEATHER & RAINFALL
9 AUG. 65
03 GMT
UPPER WINDS 9 AUG. 65 00GMT
0·9 km
850 mb
700 mb
500 mb
30
20
10
C
C – Centre of cyclonic circulation
Trough line

Fig. 9.4(b)

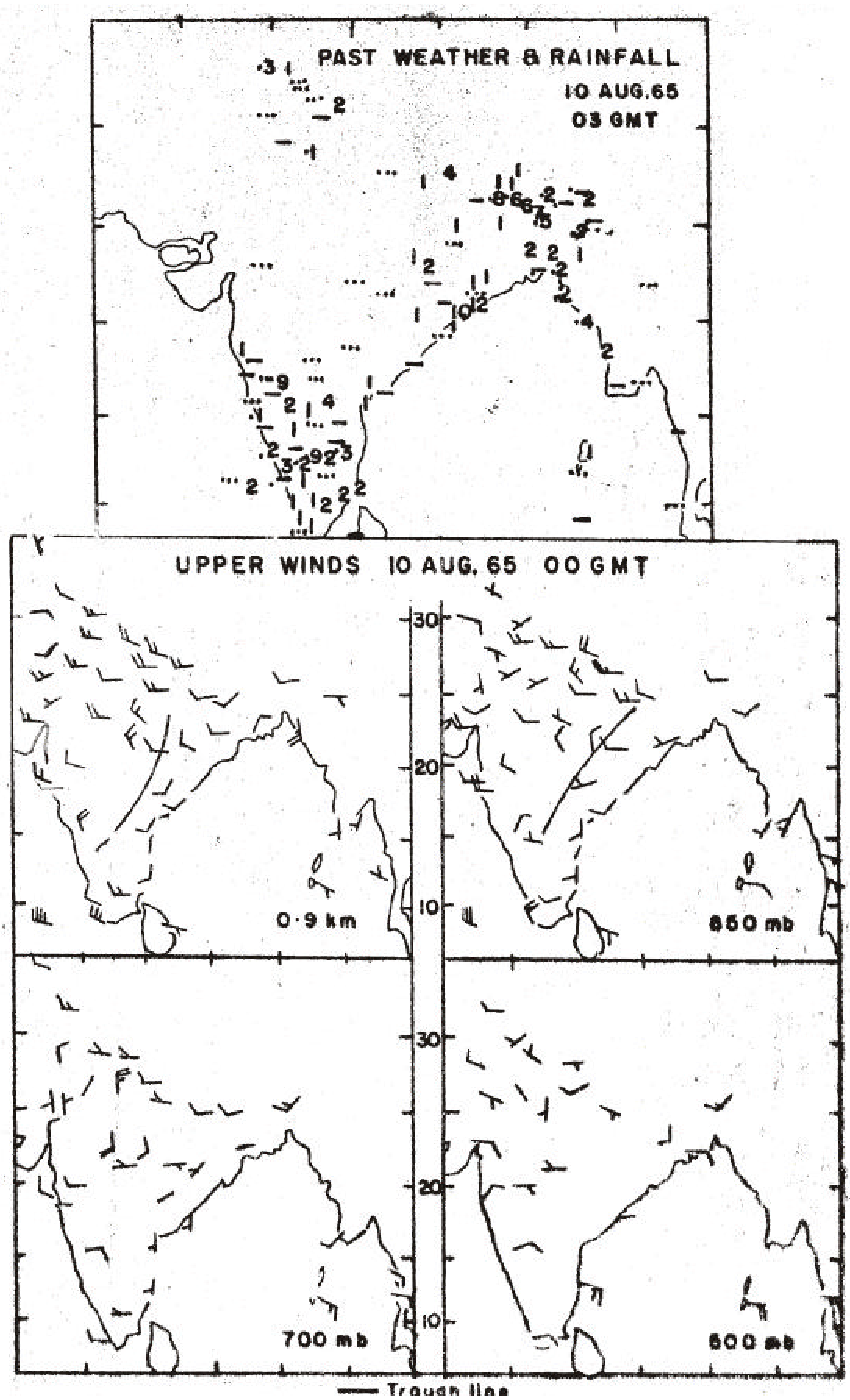

PAST WEATHER & RAINFALL
10 AUG.65
03 GMT
UPPER WINDS 10 AUG. 65 00 GMT
0·9 km
850 mb
700 mb
500 mb
Trough line

Fig. 9.4(c)

ahead of it, perhaps under the additional influence of the system in higher latitudes.

These troughs also appear to move westwards if they come under the influence of westward moving troughs in mid–troposphere in low latitudes (Figs. 9.4 a to c).

9.4. Effect of Middle Latitude Westerly Systems on Monsoon

9.4.1 When the southwest monsoon is fully established over India, middle latitude westerlies prevail to the north of Lat. 30° N only. Still systems in these westerlies seem to exercise considerable influence on the monsoon weather over northern India. The extra–tropical systems themselves consist of (i) troughs and (ii) ridges in rear. It may be presumed that the ridges in rear can sometimes develop into a warm high, Northern India is just at the periphery of these systems and it is difficult to trace their trailing end across the sub–tropical ridge over northern India in mid–and upper troposphere. Their affect is of four kinds (Rao and Srinivasan, 1970).

i) intensifying or developing lower tropospheric lows or troughs,

ii) enhancing rainfall in pre–existing systems,

iii) causing recurvature of depressions and lows and

iv) leading to onset of break conditions.

Mooley (1957) has pointed out cases of enhancement of activity of monsoon over Punjab and west Uttar Pradesh with the passage of western disturbances across extreme north India, On some occasions in the absence of lows and depressions from the east, if there is a western disturbance moving across extreme north of the country and eastern Himalayas, the monsoon trough shifts to the foot of the Himalayas and break situation develops. Chakravorty and Basu (1957) have also put forward a similar view.

9.4.2 The first three effects are opposite to the last and specific differences between the conditions leading to the two types of developments have to be elucidated. Even 'in break' cases traced to westerly troughs, rainfall is enhanced by the mid–latitude troughs before the 'break' sets in. 'Breaks' are characterised by the sub–tropical ridge becoming more marked and swinging southwards. Hence whether a 'break' sets in or not after the passage of a westerly trough may depend upon the intensity of the ridge in the rear rather than the passage of the trough itself. Even after a 'break' has set in, bursts of rain along the Himalayas travelling eastwards seem to be associated with the passage of subsequent troughs in westerlies, which however are perhaps not able to break up the warm high built up earlier. A warm high building up and strengthening the seasonal sub–tropical ridge may explain 'breaks' persisting for one to two weeks sometimes. The passage of a trough by itself cannot account for the subsequent prolonged 'break'. On the other hand, it tallies with the behaviour of warm highs.

9.4.3 The effect of westerly troughs on rainfall seems to be more over north west India. An examination of the daily rainfall over stations in western Himalayas and the adjoining plains shows, on the average, 8 to 10 peaks in rainfall in July and August. This rainfall is quite often moderate to heavy and forms a distinct zone separate from the rainfall belts further south. The rainfall peaks are found to be associated with passage of low pressure systems to the north of this country in the middle latitude westerlies. This is brought out in Fig. 9.5 which gives the daily 700 mb heights at Kuche (41° 43' N, 83° 04' E) in Sinkiang and rainfall over Western Himalayas in July and August 1968. This type of influence of westerly troughs on rainfall can occur without the monsoon trough shifting towards Himalayas. Monsoon can be even active over the central parts of the country, sometimes with a depression moving along the usual track. Pressure changes at sea–level give indication of the movement of the lows/troughs in westerlies, as fall of

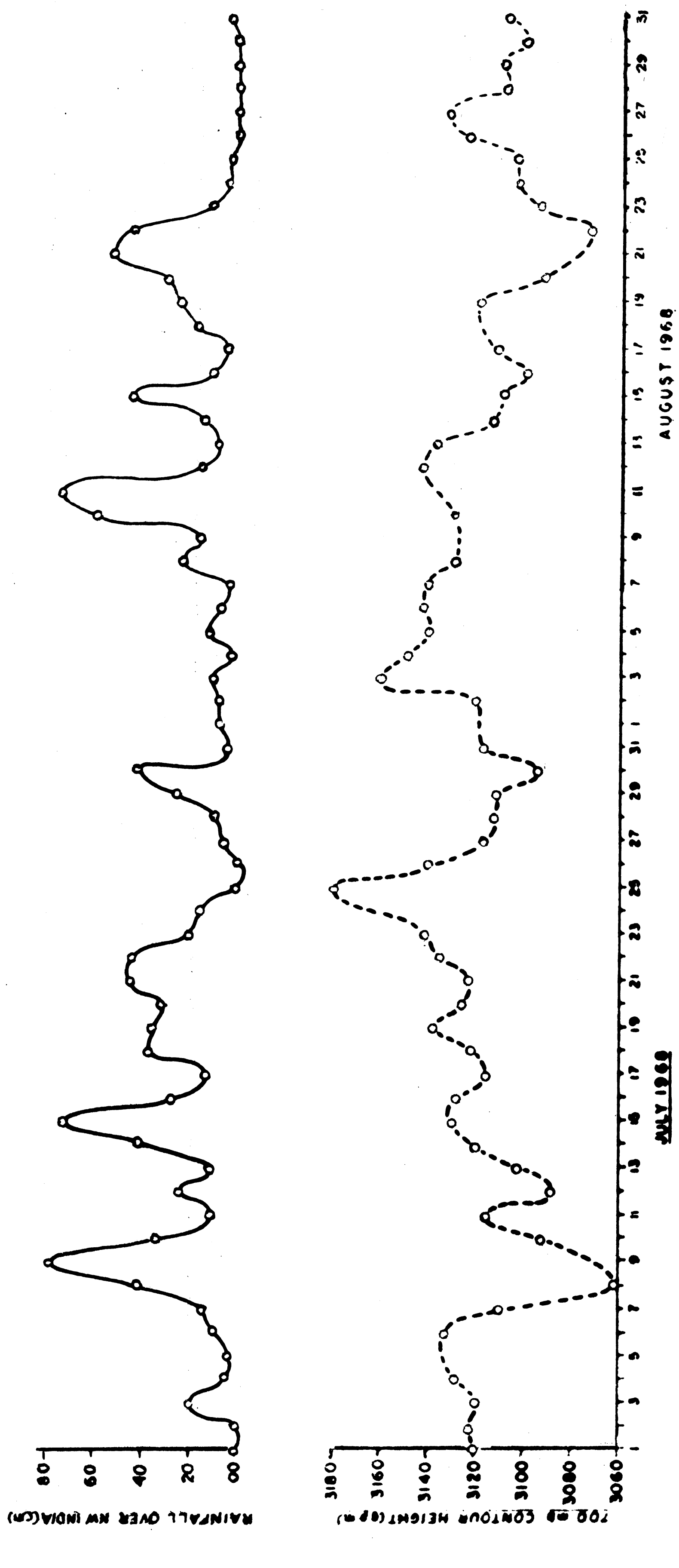


Fig. 9.5 700 mb contour at Kuche (Lat. 40 ° 45' N 83° 04' E) and daily rainfall over western Himalayas. July - August 1968 (Rao et al, 1970)

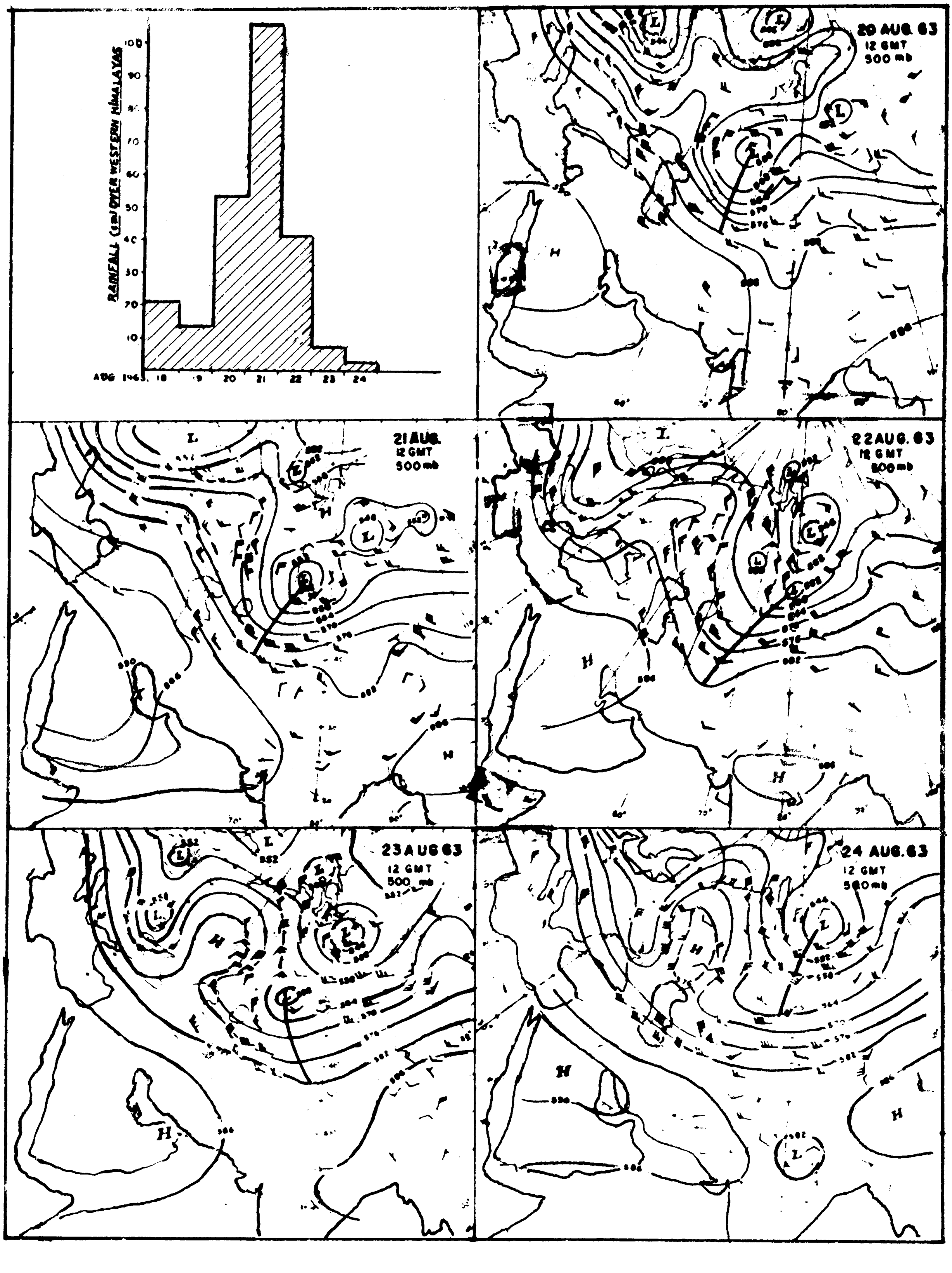


Fig. 9.6 Movement of a westerly trough and rainfall over western Himalayas (Rao et al, 1970)

pressure in advance and rise in rear. The lower tropospheric winds sometimes show southwest/south flow over west Uttar Pradesh, Haryana, Punjab and adjoining areas. A case of 20–22nd August 1963 is presented in Fig. 9.6 as a sequence of 500 mb charts and rainfall over Western Himalayas.

This kind of effect may extend sometimes upto Gujarat. During 7 to 10 July 1968, a well marked extra–tropical low moved across central Asia between 45 ^{o} and 55^{o} N with its frontal system extending southward upto 40^{o} N. Associated upper trough while moving across Russian Turkistan and Sinkiang increased in amplitude, Under their influence, lows developed over northern parts of Pakistan and also over southeast Baluchistan, Sind and neighbourhood between 7th and 8th. They activated monsoon over northwestern parts of the country as far south as Gujarat. As this extra–tropical system moved further eastwards, it apparently triggered off a depression over southwest Uttar Pradesh in the pre–existing monsoon trough on 10th. This sequence is illustrated in Fig, 9.7 (a to k) (Rao, Srinivasan, Raman and Ramakrishnan, 1970) .

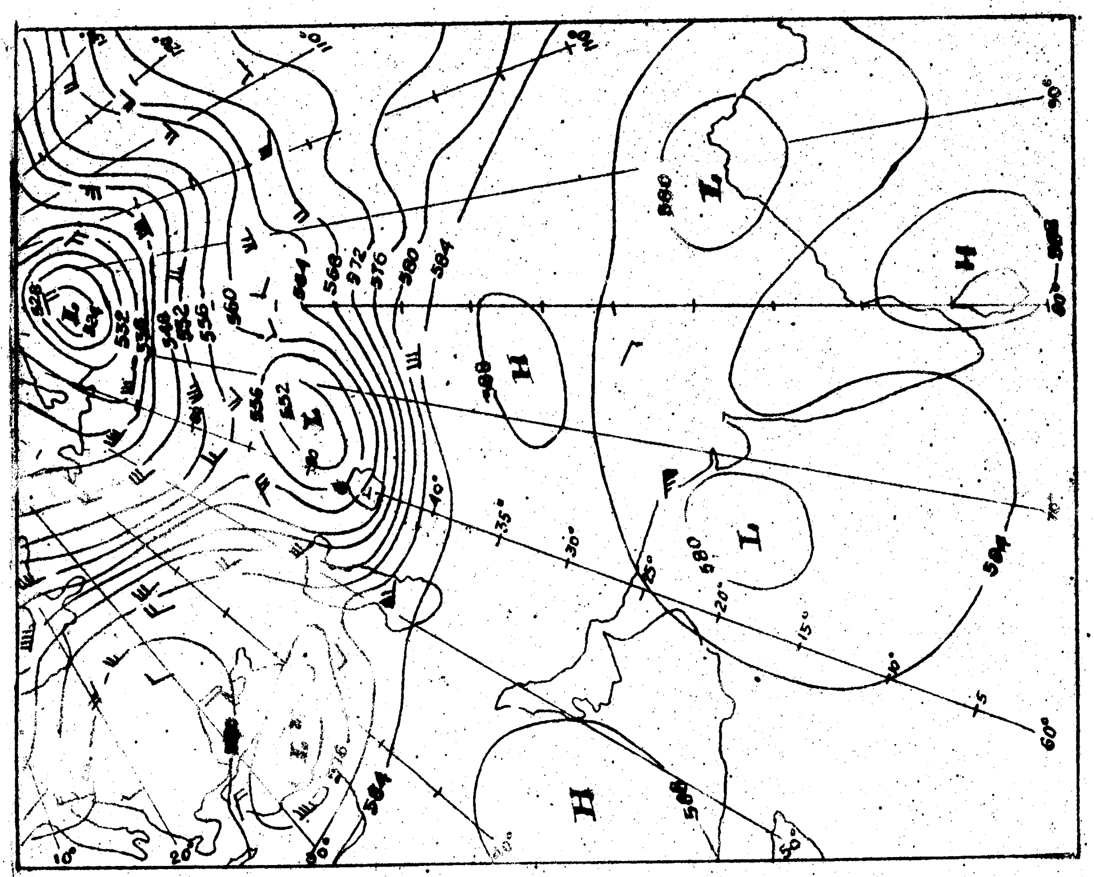
9.4.4 In another instance, a monsoon depression formed over northwest Bay on 2nd September 1966, and moved westnorthwestwards to 20.5^{o} N, 80.5^{o} E by 5th. Later it moved northwest till 7th when it began to recurve northwards. Weakening on 8th, it moved away northeastwards. In this case there was a well–marked low at 42^{o} N, 60^{o} E and the associated upper trough reached upto 25^{o} N. The sequence of satellite pictures in Fig. 9.8 shows the southward extension of middle latitude frontal system upto Kashmir on 3–4 September, formation of an open vortex to the west of it on 5th and the joining up of the middle latitude cloud system and the monsoon depression on 7th.

9.4.5 Even when the monsoon trough is along the Himalayas and favourable for monsoon activity in the neighbourhood, rainfall over north Assam and sub–Himalayan West Bengal is more when a middle latitude westerly trough is extending upto Tibet, The case of 7th to 11th July 1967 (Figs, 9.9 a to e)shows the passage of a middle latitude westerly trough on a break–type of situation (Srinivasan, Raman and Mukherjee 1972). Between 5 and 6 July, the monsoon trough shifted progressively northwards and the eastern half lay close to the foot of the Himalayas. A feeble north–south oriented trough lay over sub–Himalayan West Bengal on 7th upto 850 mb. At 500 mb and above, a trough in mid–latitude westerlies lay along 81^{o} E but the movement of the monsoon trough to the foot of the Himalayas had taken place at least a day earlier. As this upper trough lay near Long, 86^{o} E on 8th and 9th, rainfall in northeast India increased right upto Head Bay, This is not a case of break situation being caused by a westerly trough but rainfall being enhanced in a break situation by a westerly trough extending right into the plains.

9.4.6 Malurkar (1958) recognised that the weather produced by monsoon depressions could be accentuated by extra–tropical eastward moving disturbances at a higher latitude. The weather on the southern side of the high latitude disturbance may be further increased by orography near Himalayas.

9.4.7 A shortcoming in the studies of the effect of middle latitude systems on the Indian southwest monsoon has been that, only out–of–the–way behaviour of the monsoon systems is sought to be explained by tracing middle latitude systems. No comprehensive analysis of all systems in middle latitude westerlies moving at the periphery of the Indian area has been undertaken to unravel their varying interaction with the monsoon synoptic patterns.

9.4.8 In American longitudes, Cressman (1948) regards that the interaction between high and low latitudes depends upon the arrangement of sub–tropical cells, the two types being called zonal and meridional. In meridional flow in summer, anticyclonic cells are poleward of 35^{o} N and large extended troughs, between the anticyclonic cells, may reach from arctic to equator. In zonal flow which predominates when the sub–tropical high pressure cells are centred equatorward of 35^{o} N with their major axes oriented east–west, waves in westerlies and tropical systems do not



Upper Air Chart 7 July 1968 00 GMT

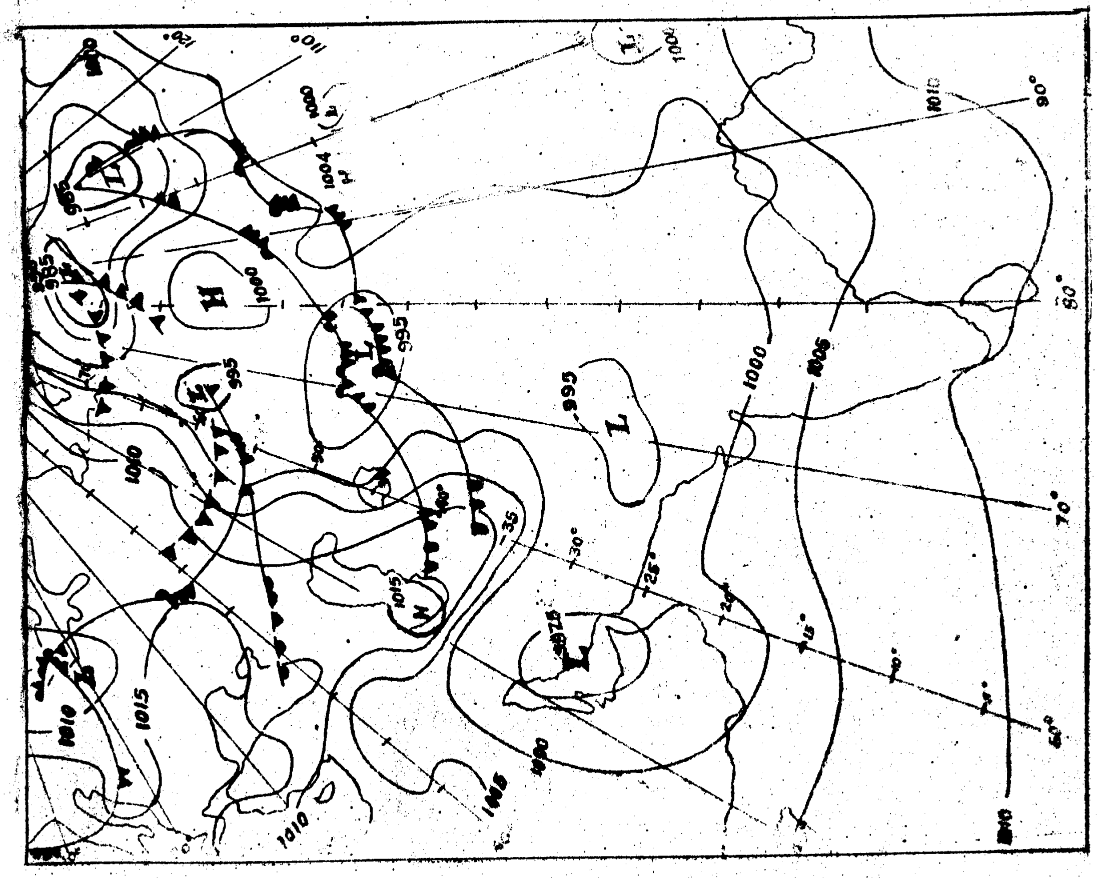


Fig. 9.7(a) Surface Chart 7 July 1968 00 GMT

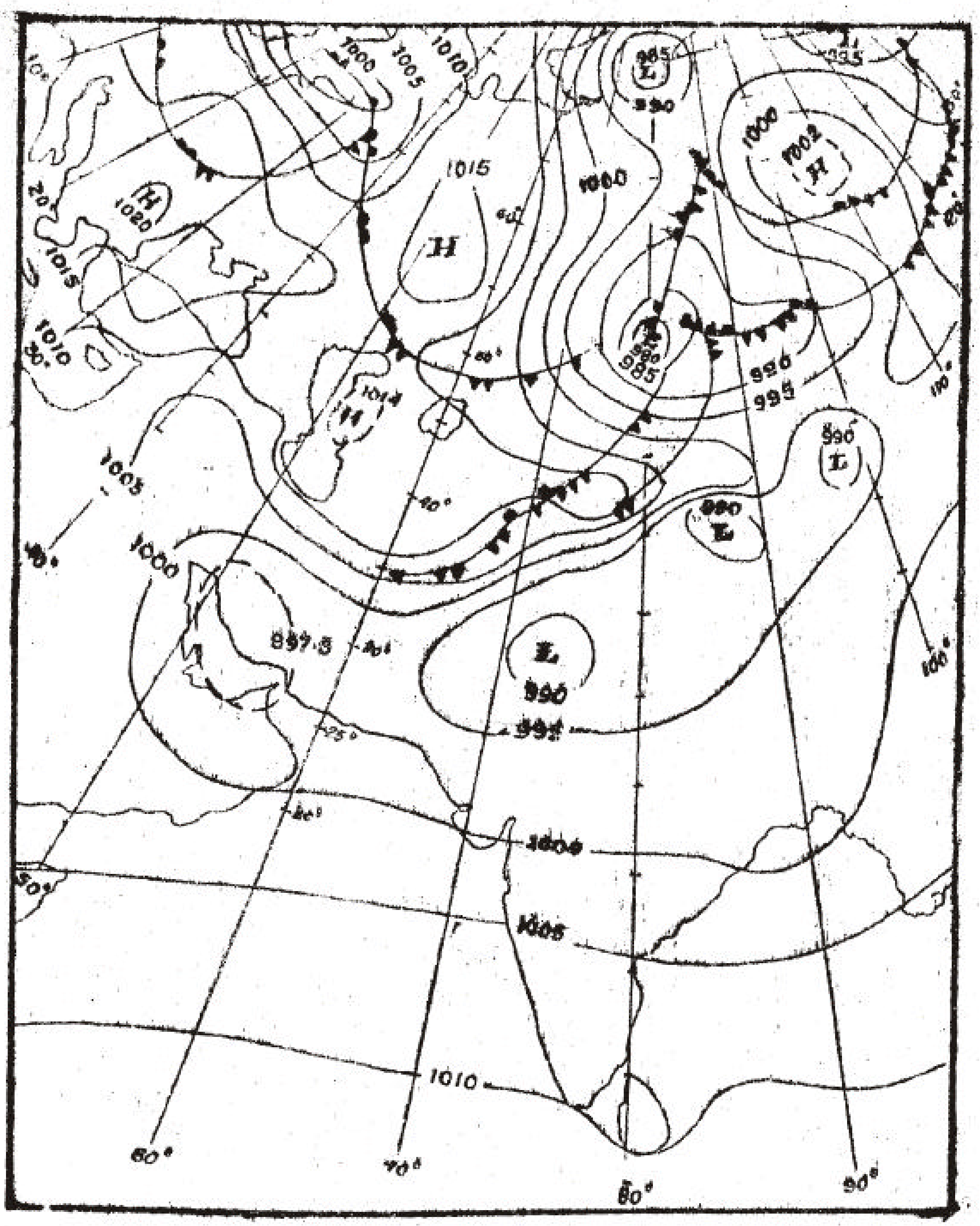


Fig. 9.7(b) Suraface Chart 8 July 1968 00 GMT

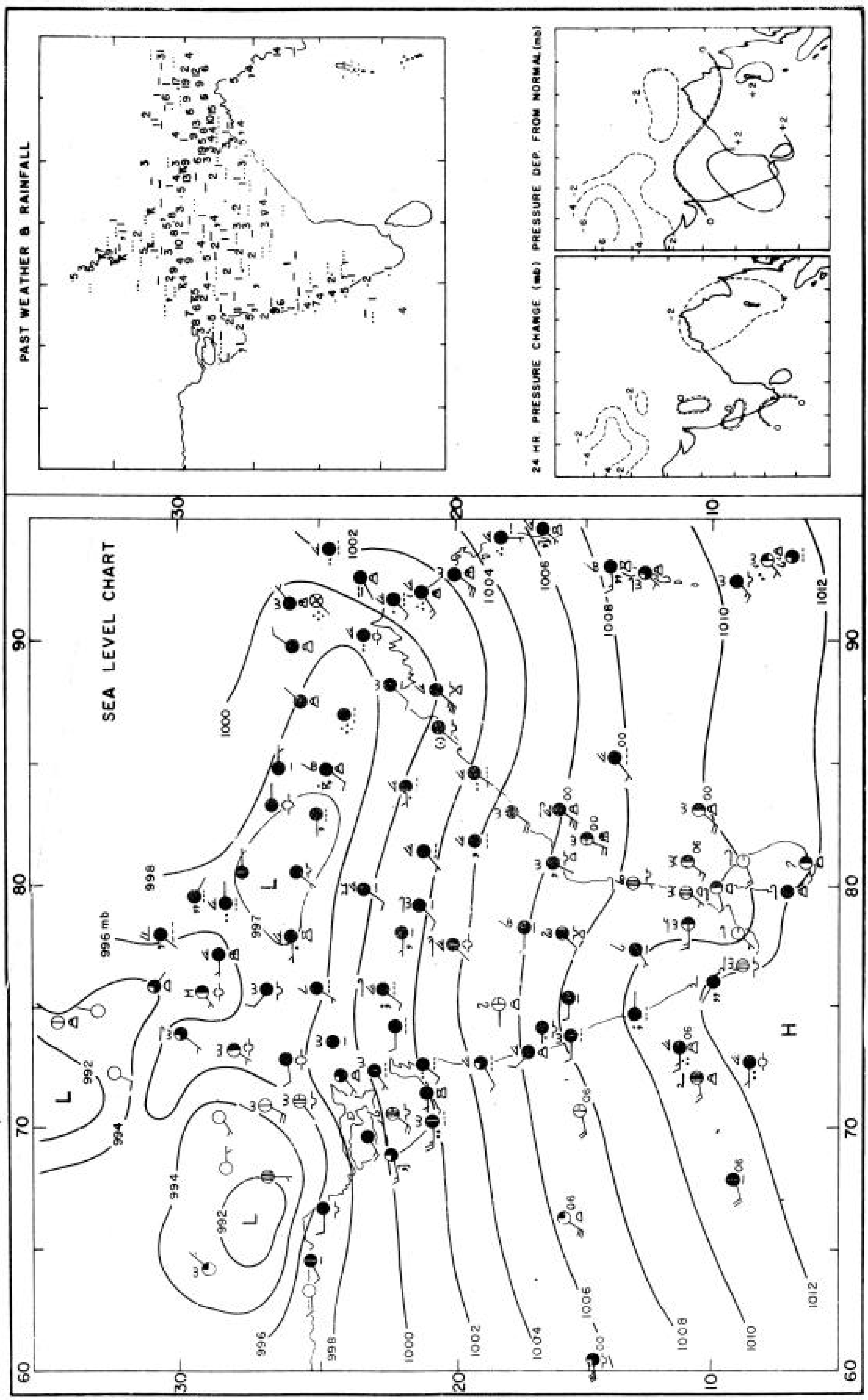


Fig. 9.7 (c) Synoptic charts 0300 GMT 8 July 1968.

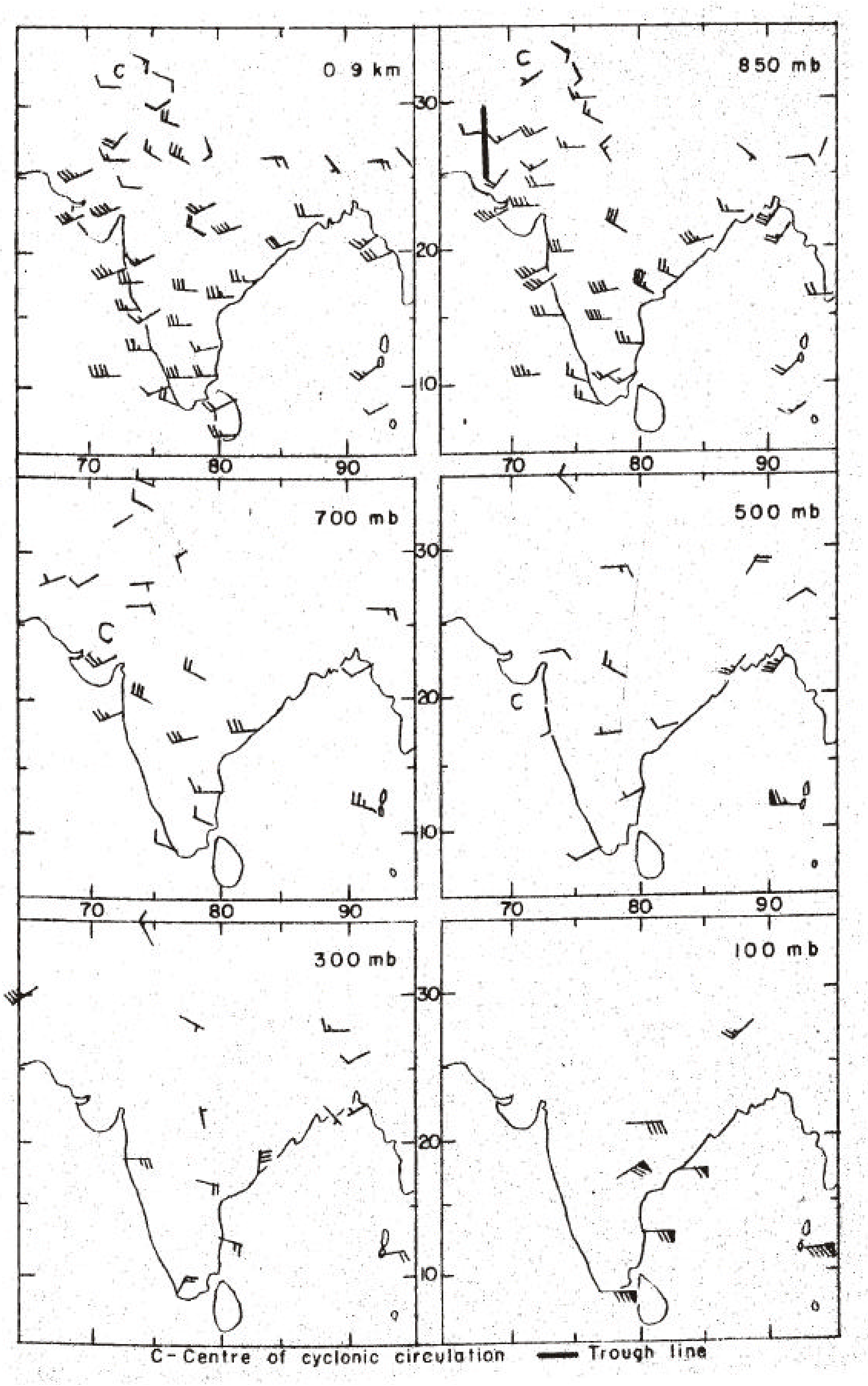


Fig. 9.7(d) Upper Wind 8 Jul 1968 00 GMT

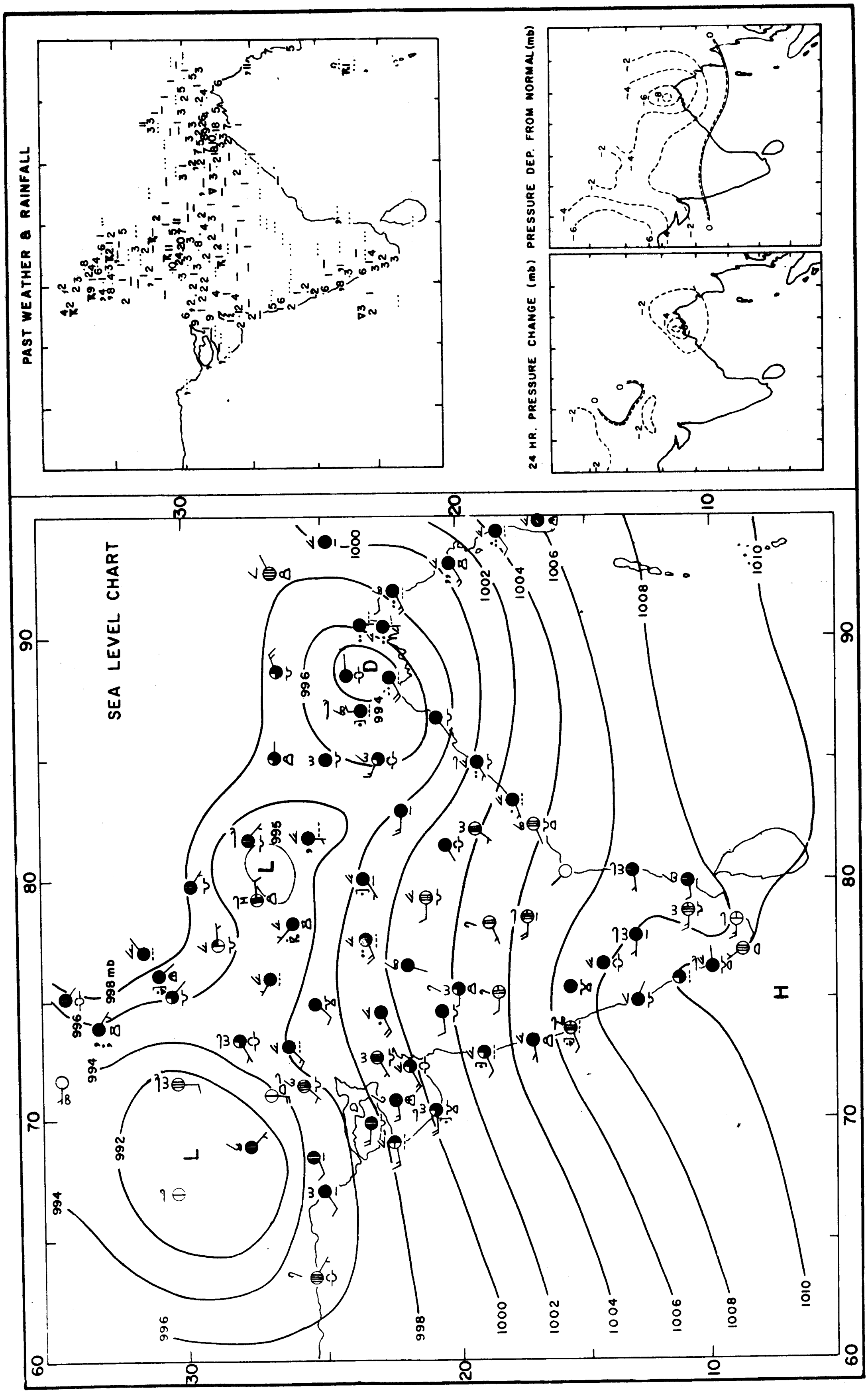


Fig. 9.7 (e) Synoptic charts 0300 GMT 9 July 1968.

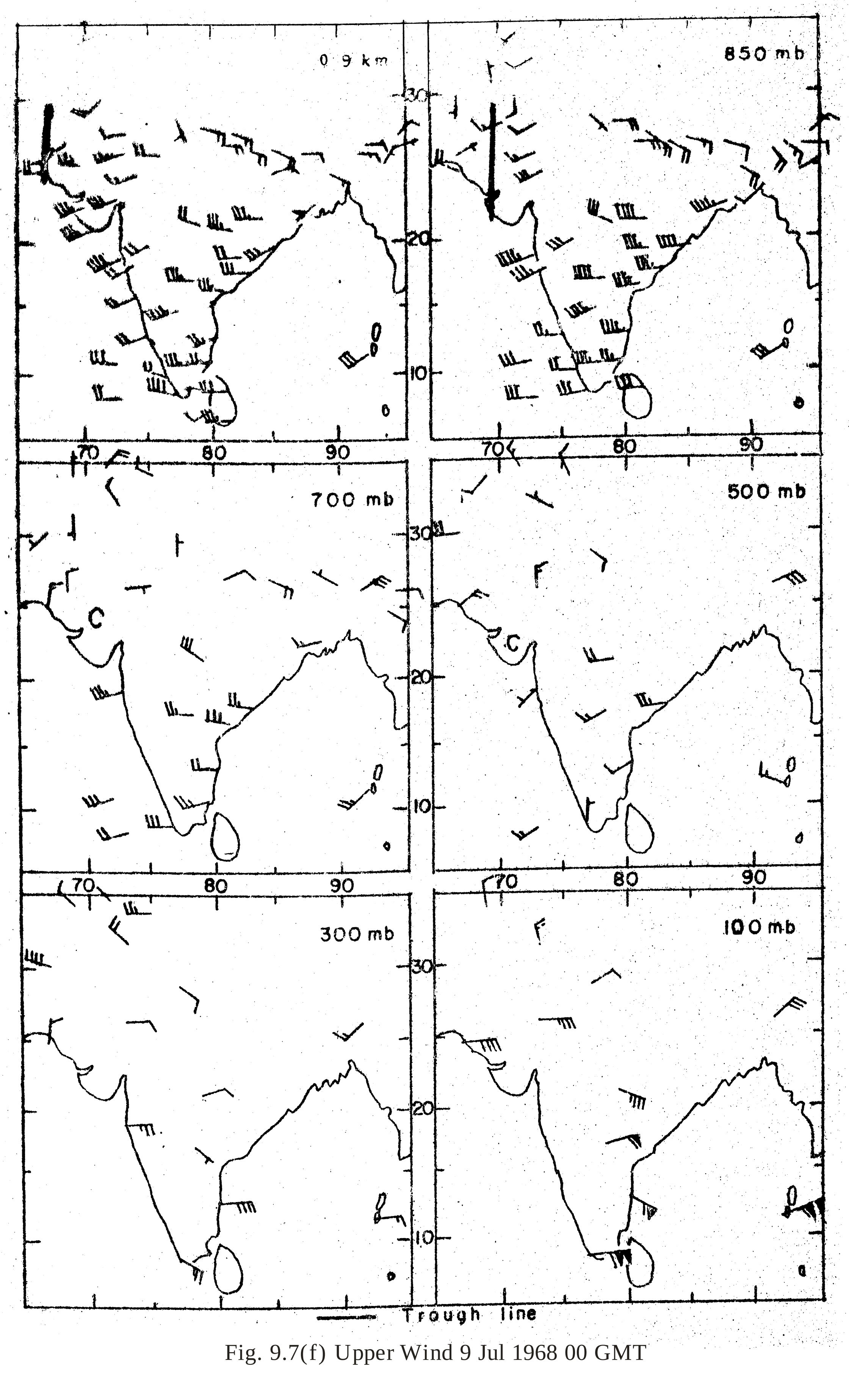


Fig. 9.7(f) Upper Wind 9 Jul 1968 00 GMT

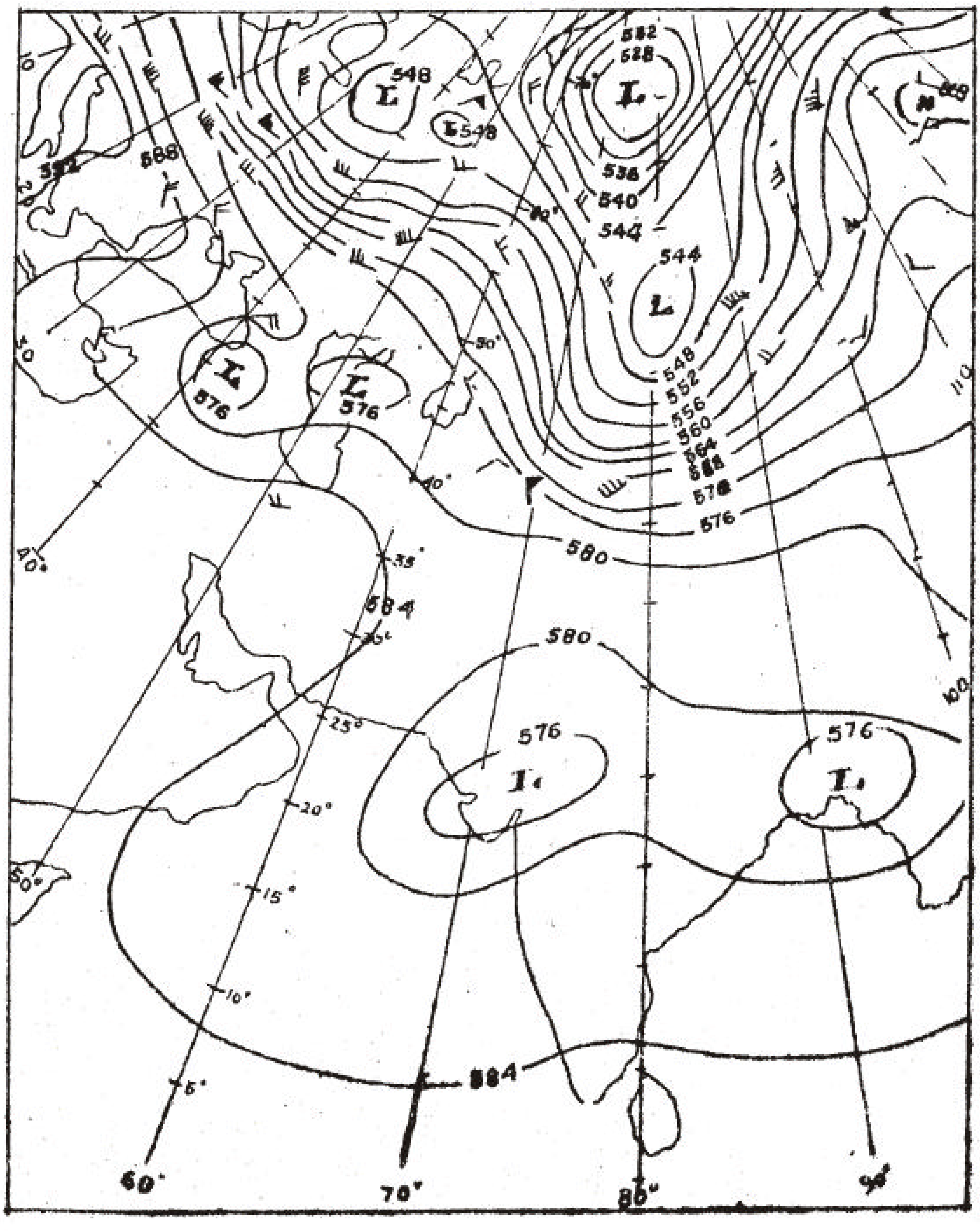


Fig. 9.7(g) Upper Air Chart 9 July 1968 00 GMT (500mb)

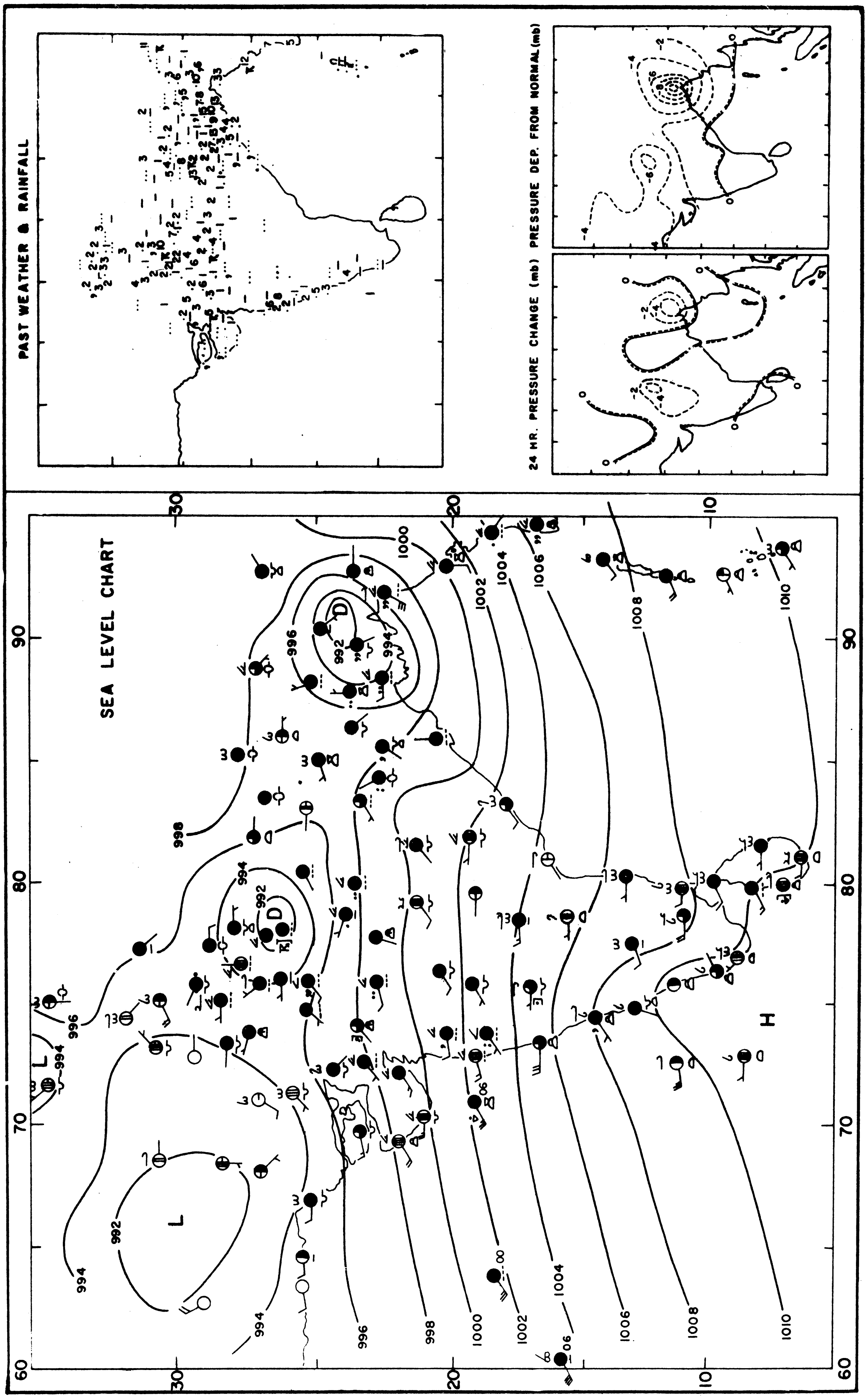


Fig. 9.7(h) Synoptic Charts 0300 GMT 10 July 1968

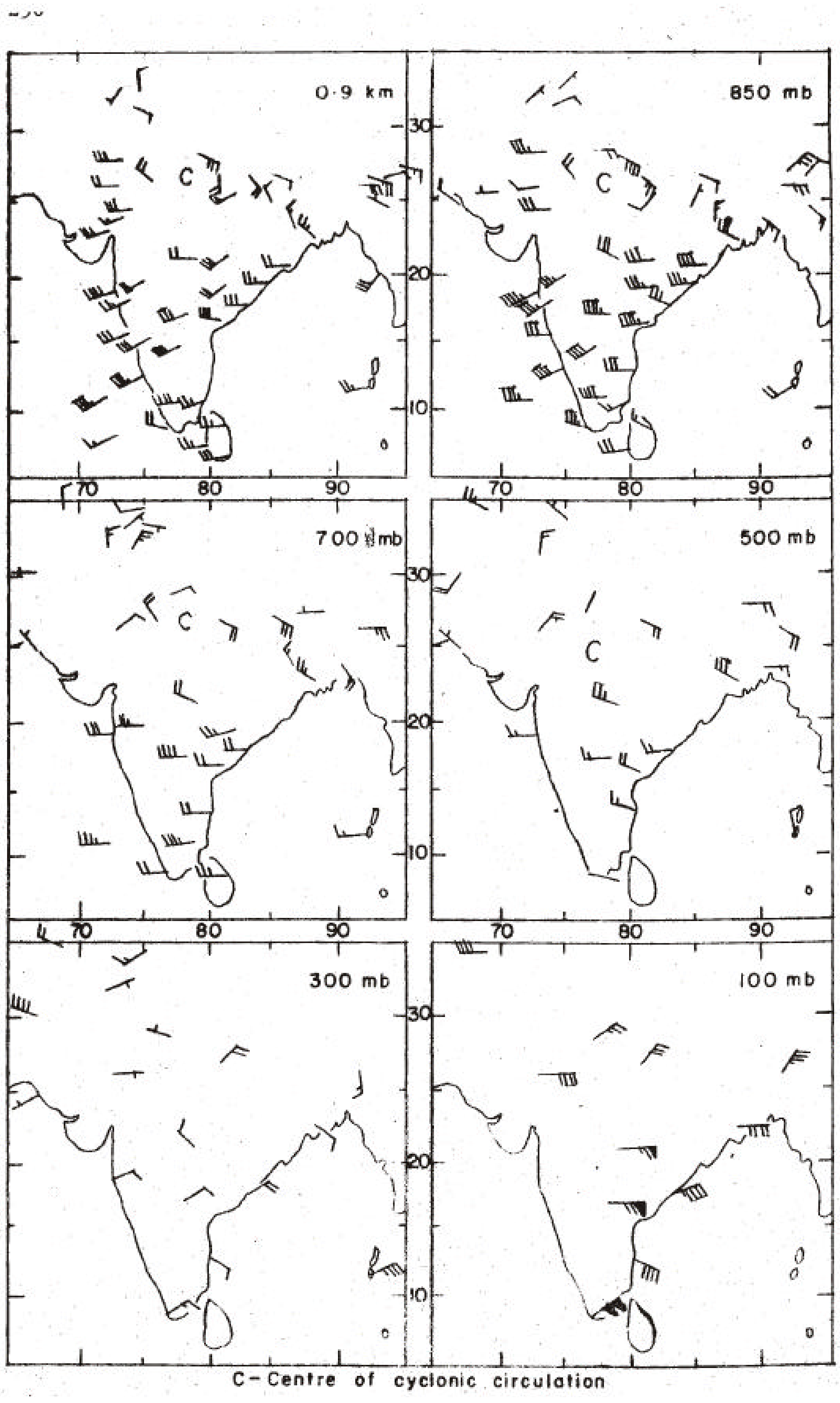


Fig. 9.7(i) Upper Winds 10 Jul. 1968 00 GMT

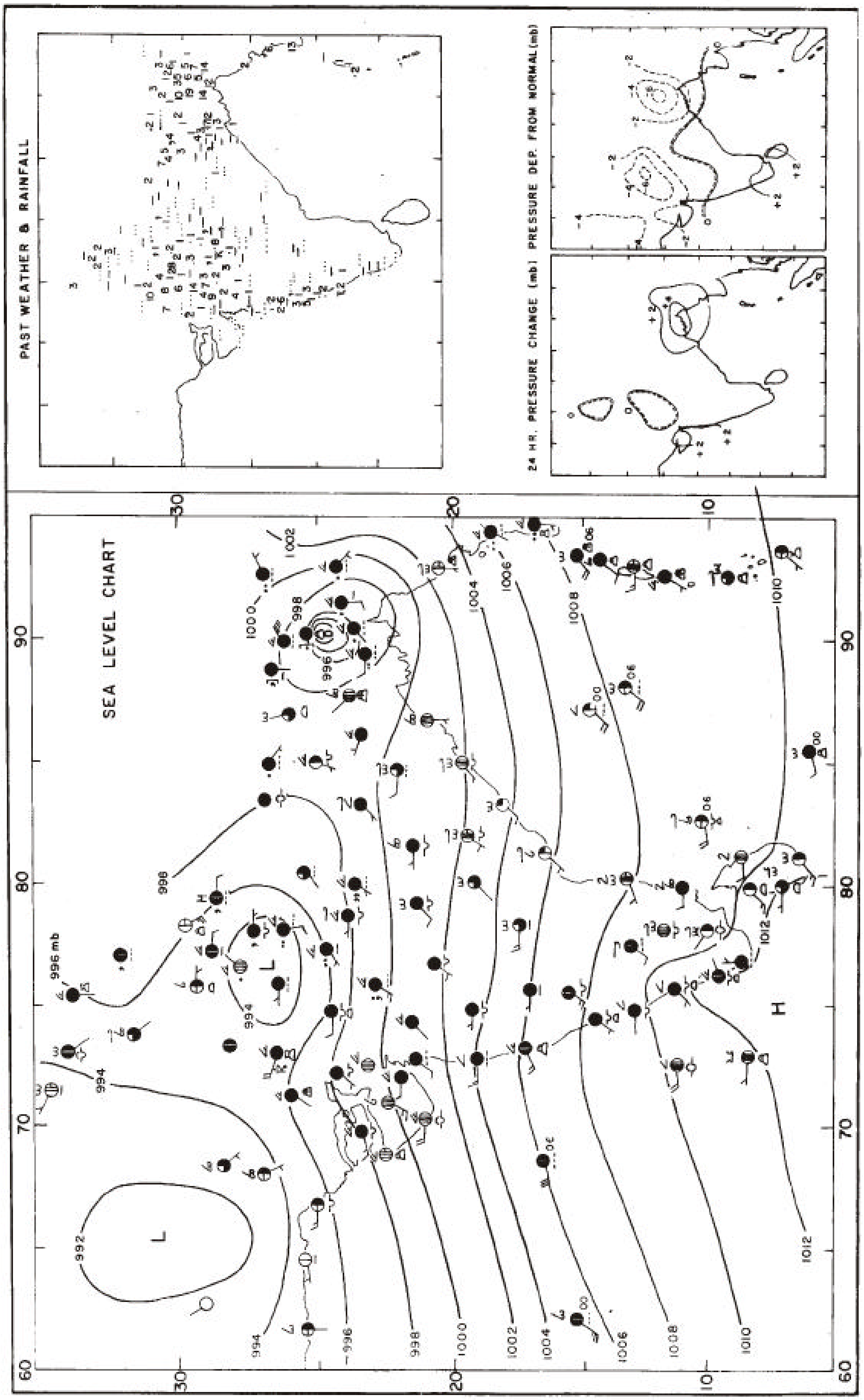


Fig. 9.7(j) Synoptic Charts 0300 GMT 11 July 1968

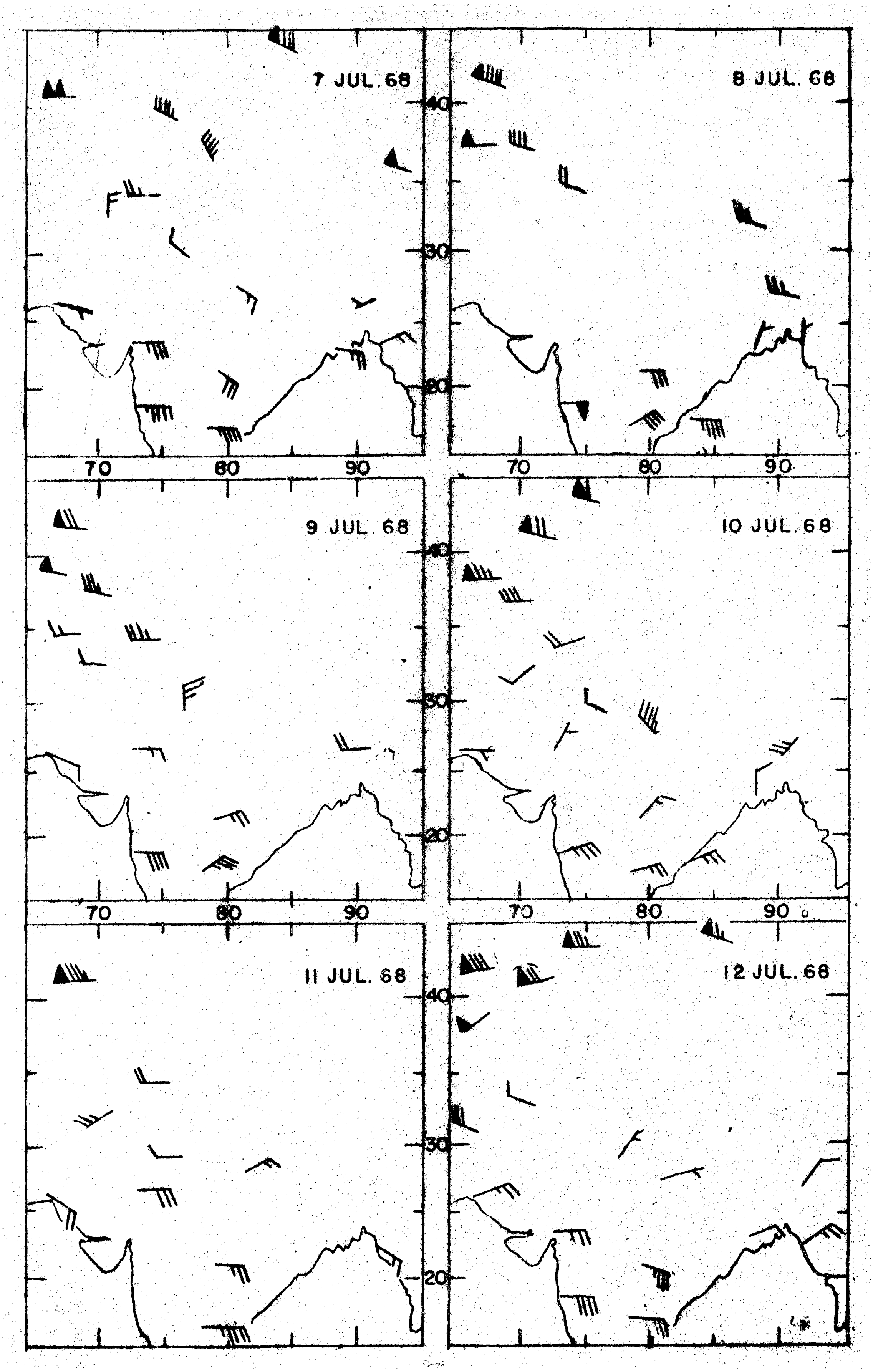


Fig. 9.7(k) Upper Winds 200 mb 7-12 July 1968 00 GMT

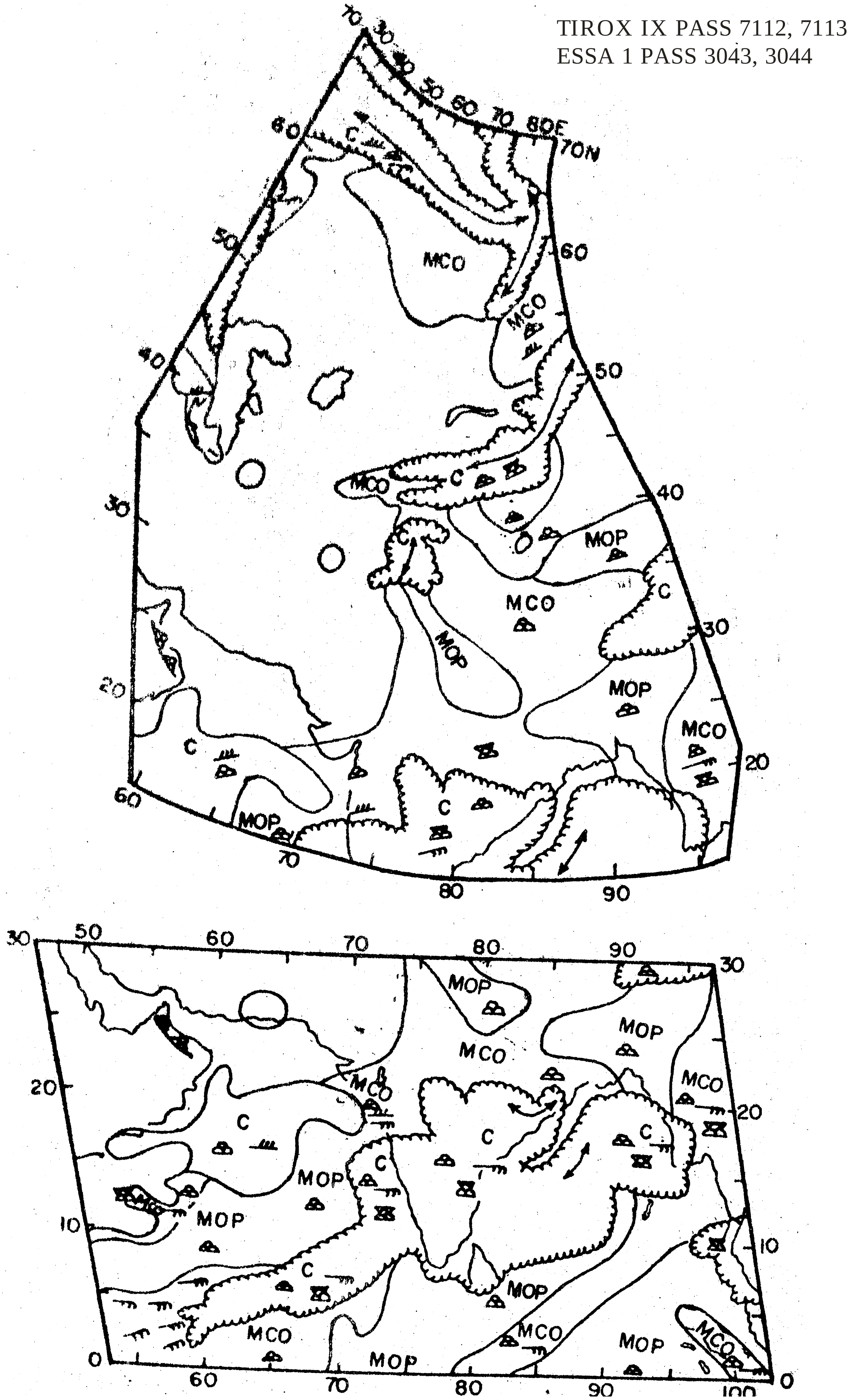


Fig. 9.8(a) Satellite Nephanalyses 3 September 1966

TIROX IX PASS 7125
ESSA 1 PASS 3058, 3059

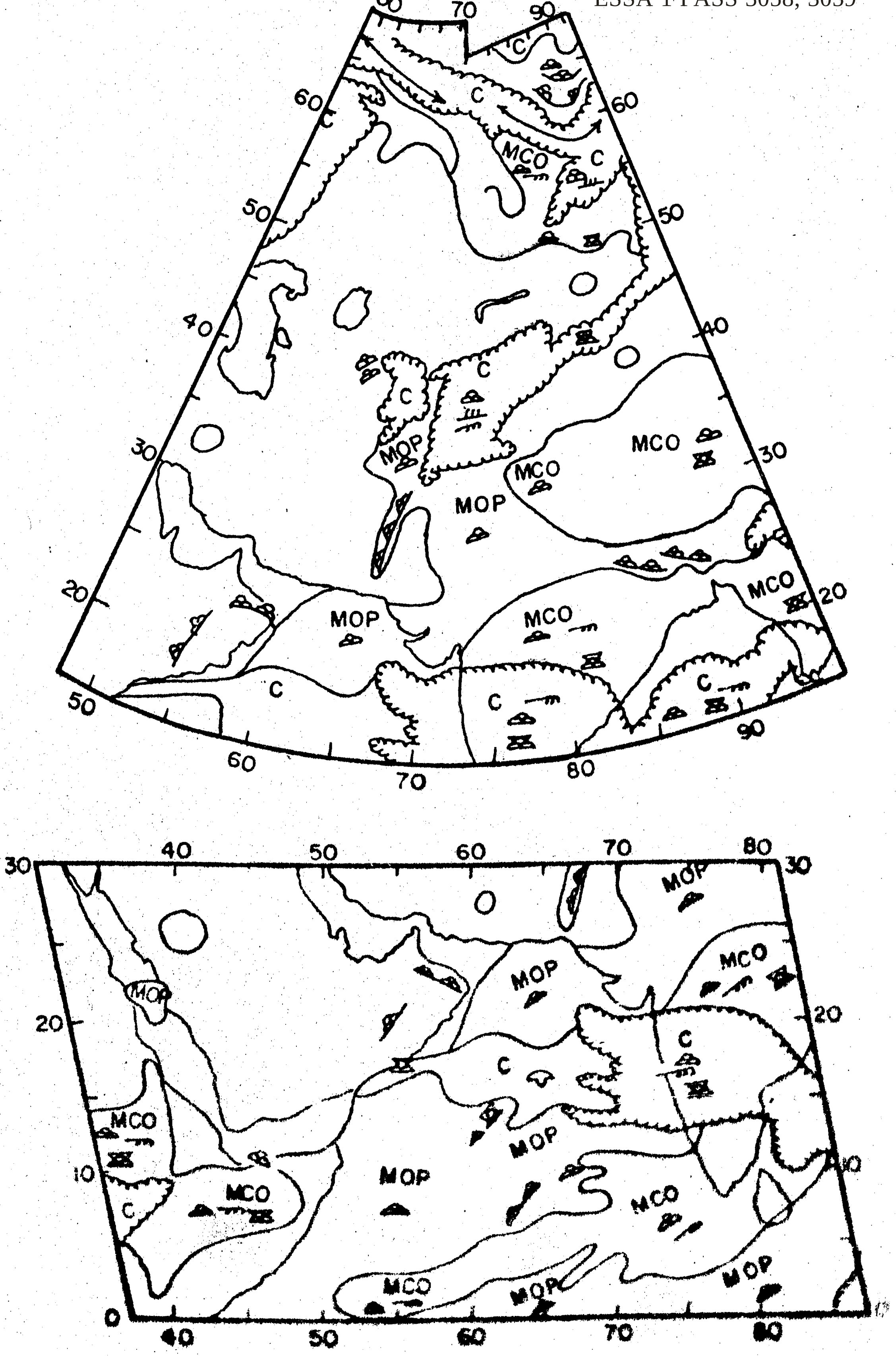


Fig. 9.8(b) Satellite Nephanalyses 4 September 1966

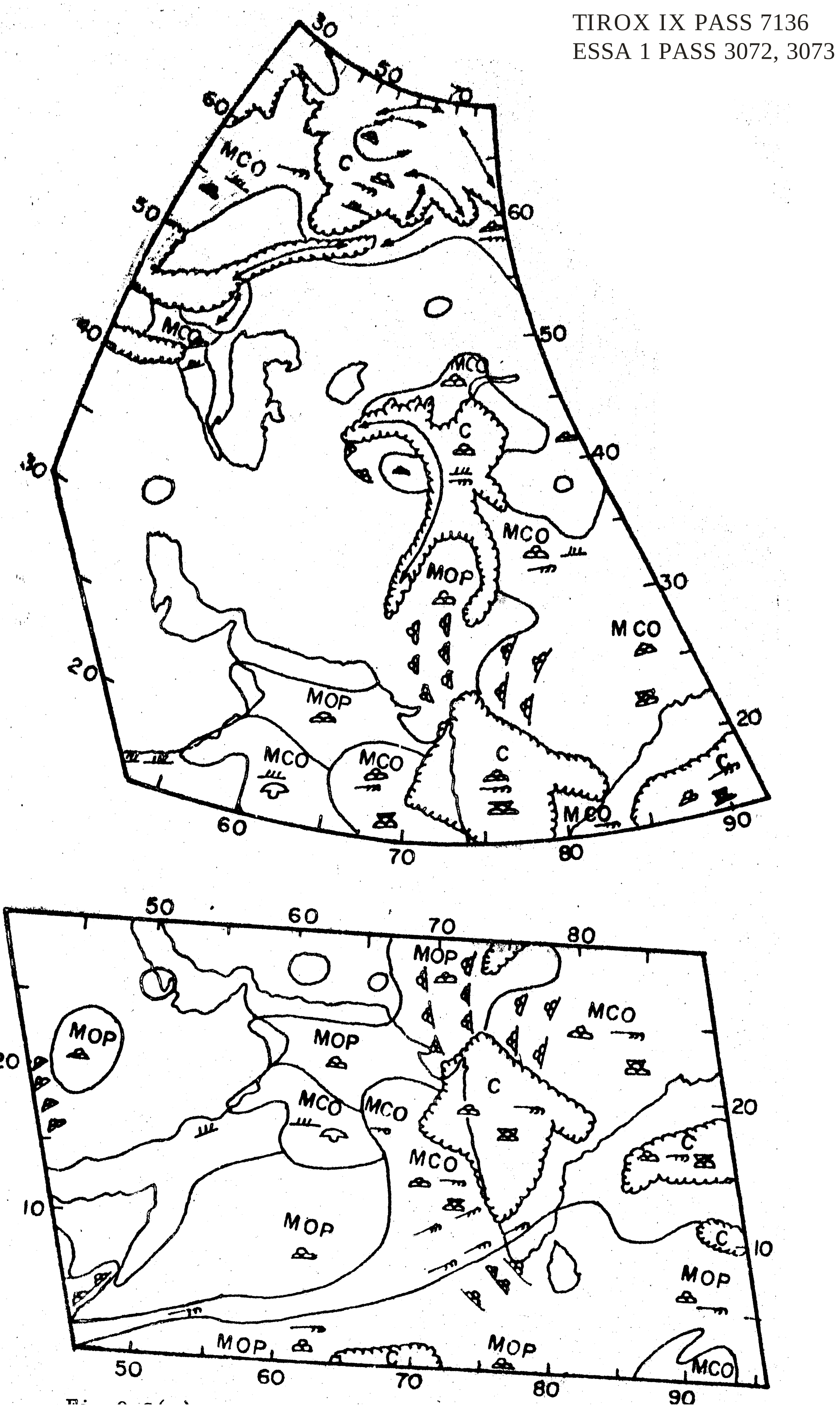


Fig. 9.8(c) Satellite Nephanalyses 5 September 1966

TIROX IX PASS 7160
ESSA 1 PASS 3102, 3103

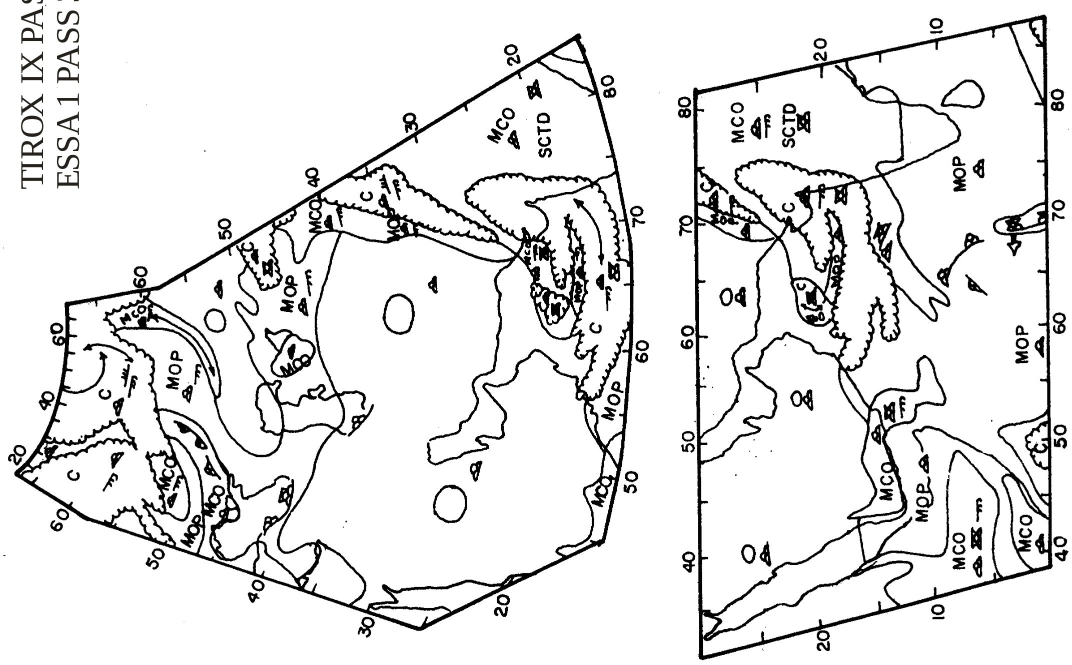


Fig. 9.8(e) Satellite Nephanalyses 7 September 1966

TIROX IX PASS 7148
ESSA 1 PASS 3086, 3087

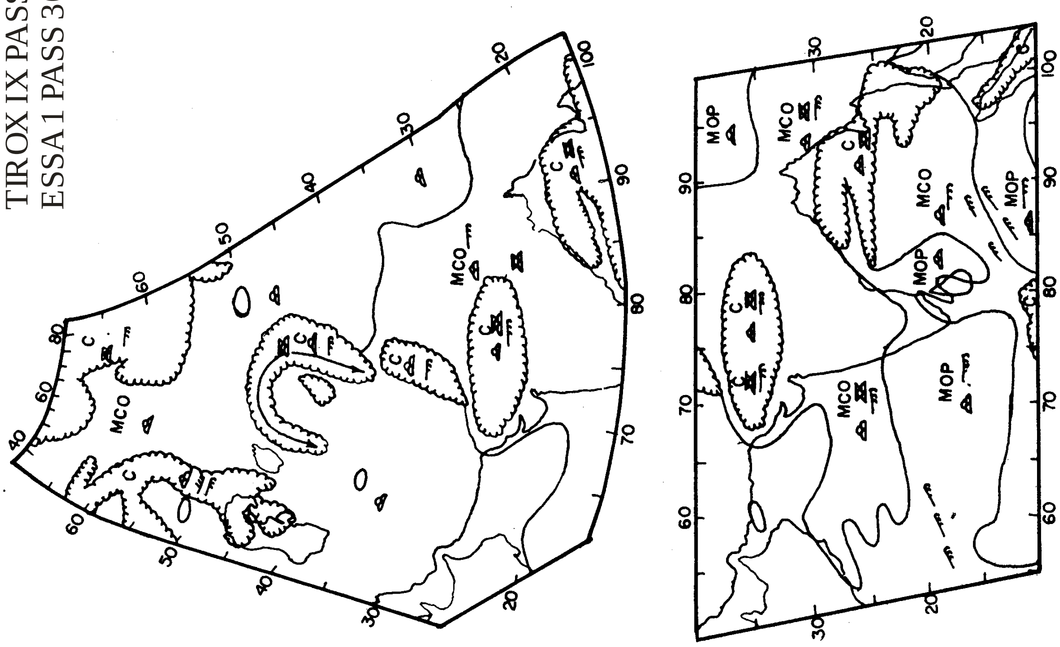


Fig. 9.8(d) Satellite Nephanalyses 6 September 1966

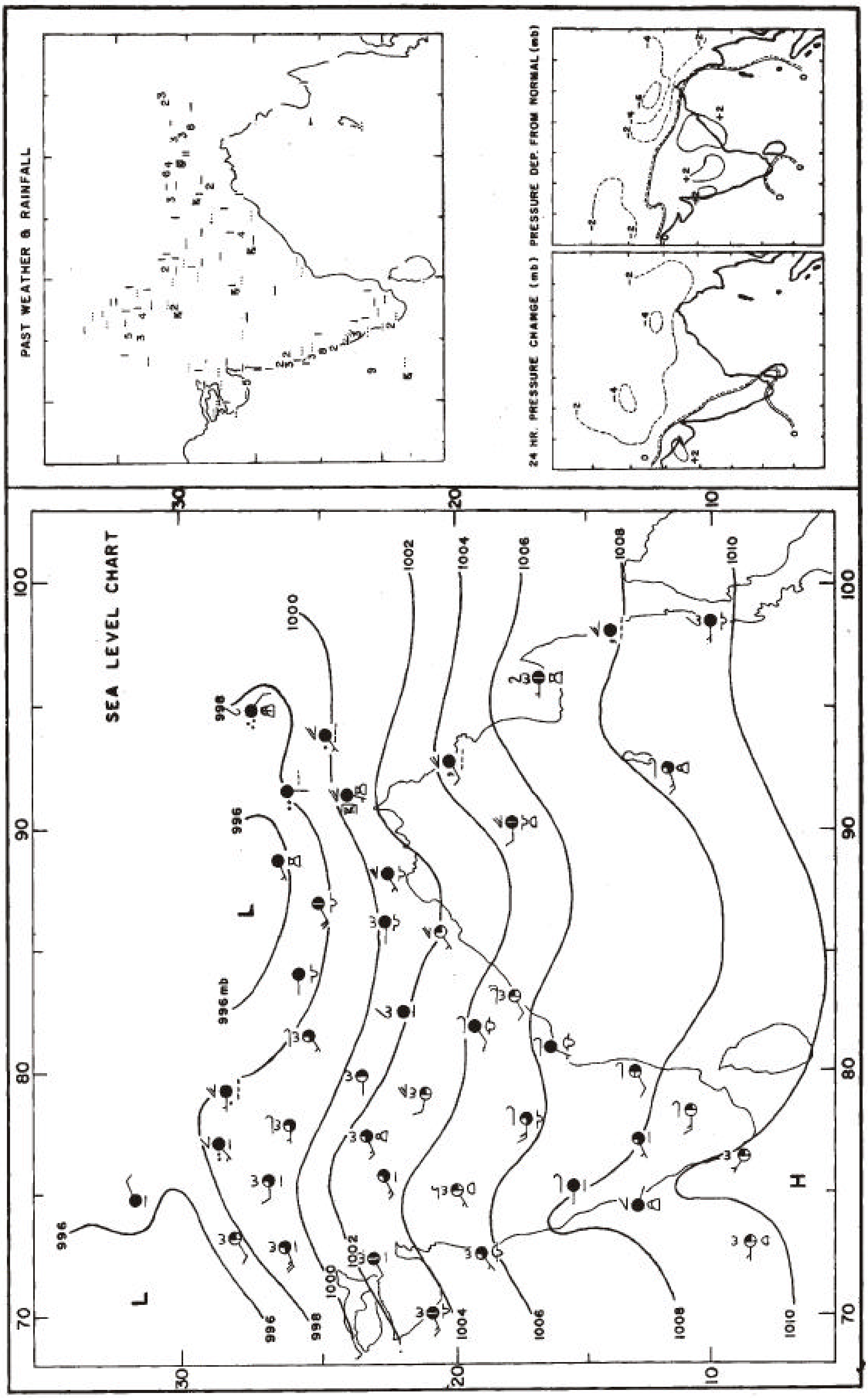


Fig. 9.9(a) Synoptic Charts 0300 GMT 7 July 1967

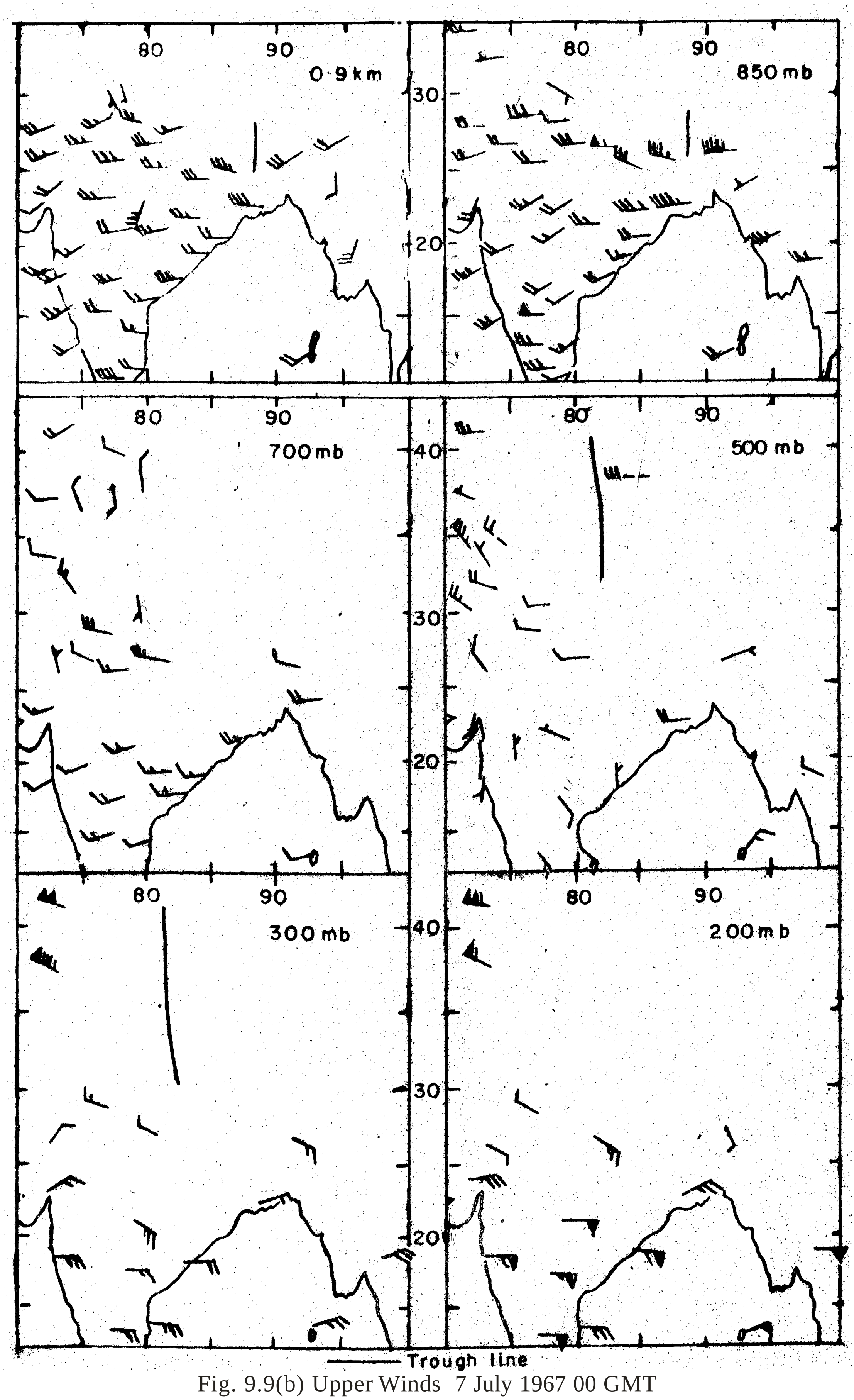


Fig. 9.9(b) Upper Winds 7 July 1967 00 GMT

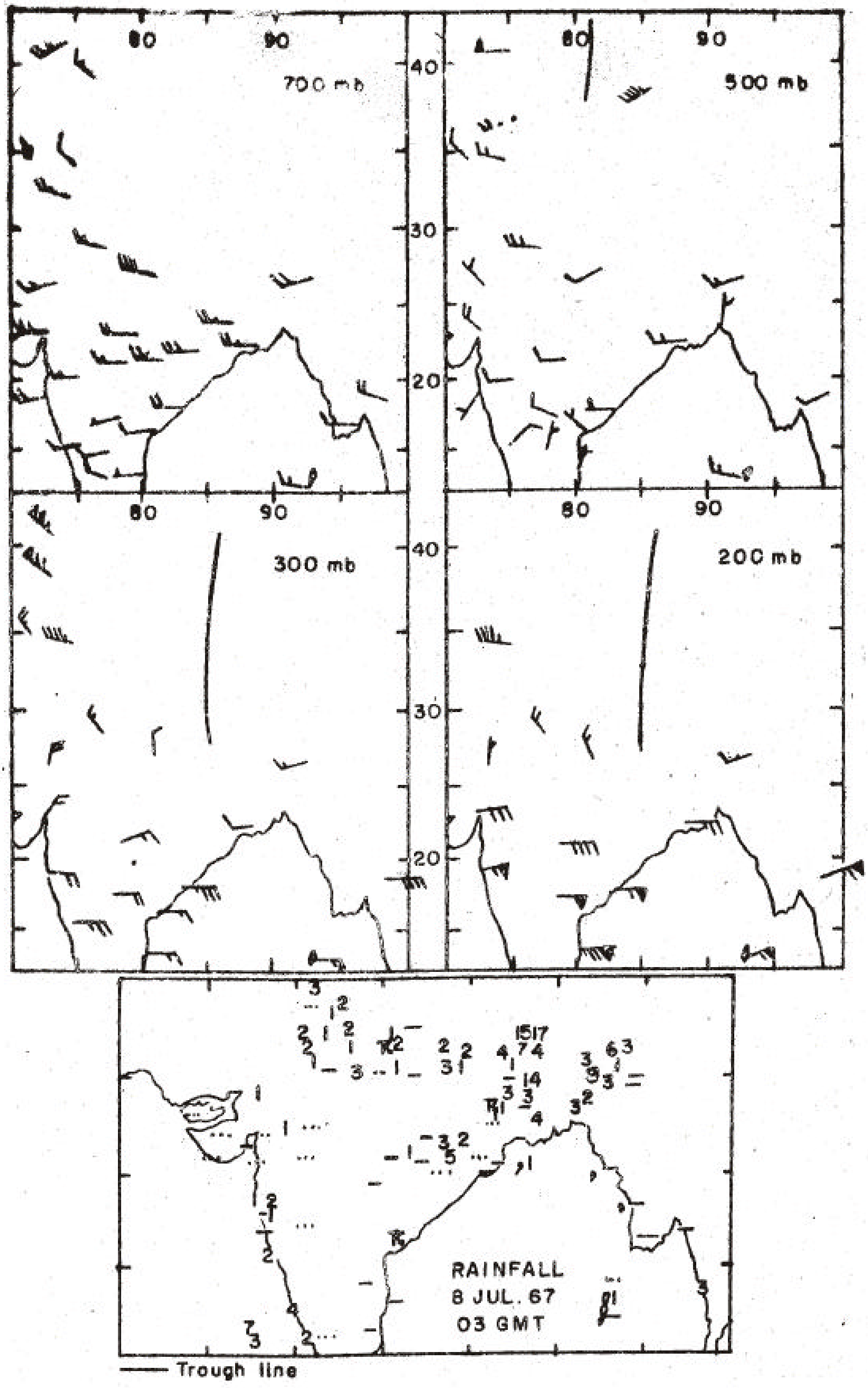


Fig. 9.9(c) Upper Winds (00 GMT) and Rainfall 8 July 1967

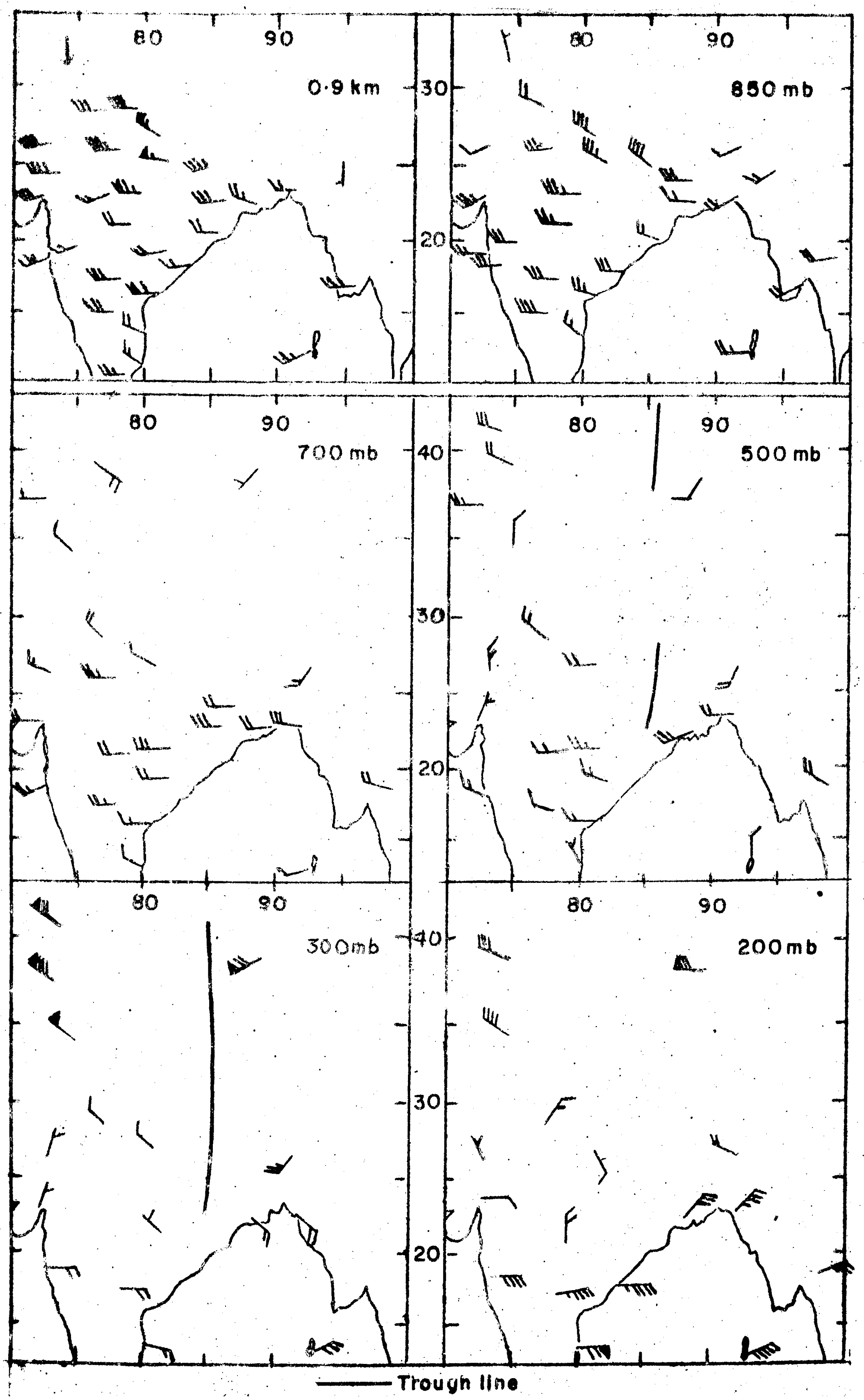


Fig. 9.9(d) Upper Winds 9 July 1967 00 GMT

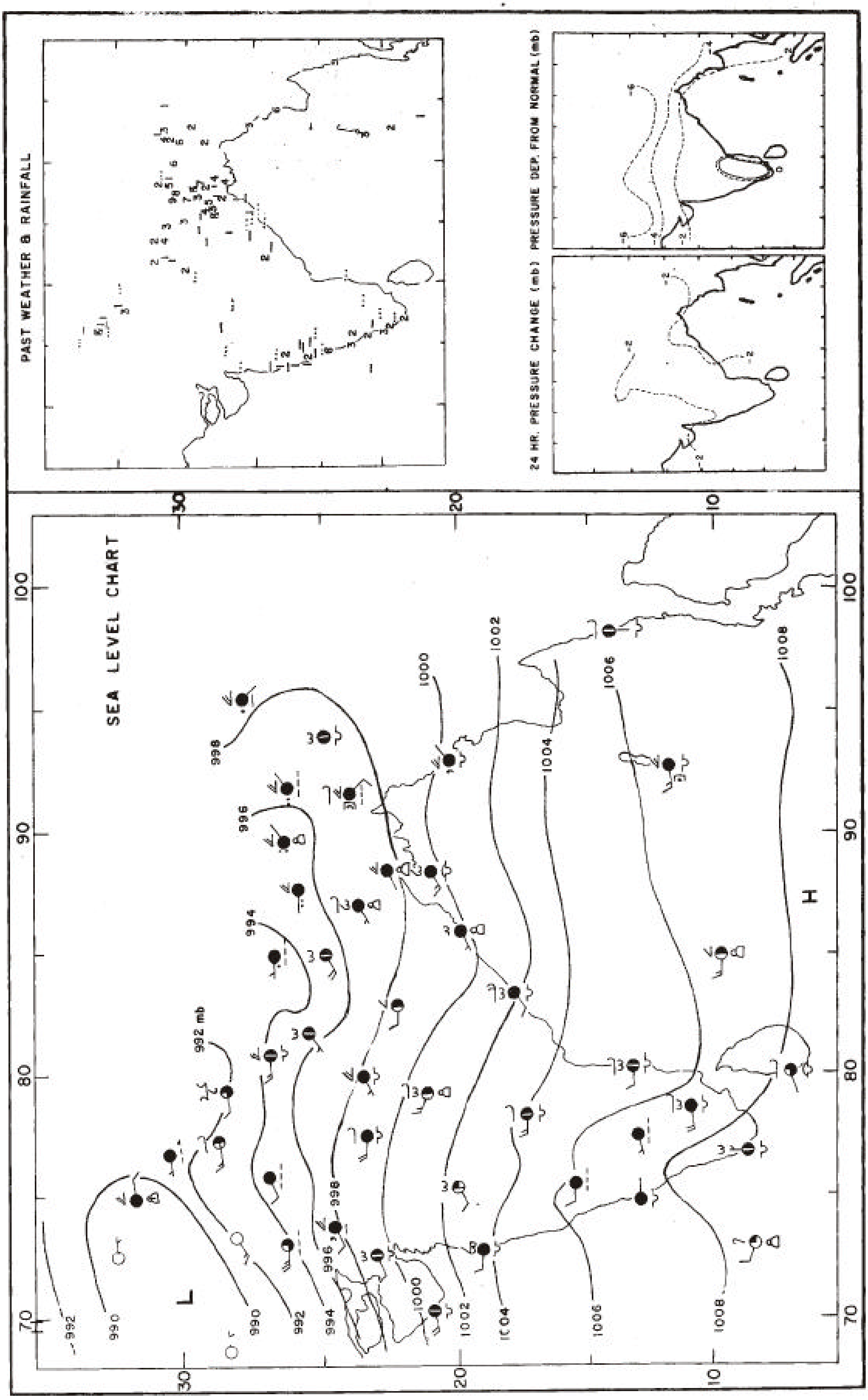


Fig. 9.9(e) Synoptic Charts 0300 GMT 9 July 1967

interact While whole hemisphere may exhibit one or the other of these characteristics, frequently one portion has a more meridional and another a more zonal circulation. The meridional pattern is similar to the instances cited by various authors, of trough in middle latitude westerlies interacting with monsoon systems.

9.5 Trough off the West Coast of Peninsula

9.5.l That the advance of the monsoon into Kerala is often associated with a weak trough along and off that coast has already been mentioned. This type of system quite frequently develops off the west coast of India, anywhere from north Kerala to south Gujarat, during the period of southwest monsoon, and is responsible for the strengthening of the monsoon in terms of rainfall, in the adjacent parts of the coastal belt. The trough is in contrast to the normal weak ridge over the Arabian Sea Islands, Kerala and adjoining Coastal Karnataka from June to August, which extends further north in weak monsoon conditions. The development of the trough is seen as a weakening of the usual pressure gradient and southerly/easterly surface winds in some belt as against the normal westerlies. This contrast is also in comparison with neighbouring areas to north or south. 24–hour pressure changes and departures from normal may also indicate the trough. A small closed circulation may be embedded in some troughs. Morning charts give better indication than the evening charts. The troughs form more often near Coastal Karnataka and slowly shift 2^{o} latitude per day northward, though they may also appear and disappear in situ over any area. Pressure gradient increases to the south of the trough particularly as it shifts to north. Upper winds are affected only in the lower troposphere and that too, often below 1 km. Cyclonic vorticity may still be present above the surface trough, as wind shear. As the troughs move to Maharashtra Coast, upper winds are drawn into the circulation to a greater depth.

9.5.2 In this type of synoptic situation, the variation between sea and land of surface drag on the winds in the boundary layer seems to be important. Due to the greater drag over land than over sea, air flow in the boundary layer will have a larger component towards low pressure. When the run of the isobars near the coast has lower pressure towards sea, this results in convergence along the coast. Bryson and Kuhn (l96l) have derived the horizontal divergence of the volume transport within the atmospheric friction layer as

$$\left[\Delta C_D/(f\Delta y)\right]\left[(sSin\,a)^2-(sCos\,a)^2\right]$$

where ΔC_D is the drag coefficient difference across the coast, the x coordinate is the coast, a is the angle between the surface wind and the coast and s is the surface wind speed.

9.5.3 Over Kerala, 90 % of the active monsoon conditions are with such troughs but often with additional features as indicated in Table 9.2.

Table 9.2
Synoptic features causing Active Monsoon over Kerala

Synoptic feature	Percentage of occasions of active monsoon over Kerala
a) Surface trough alone	8
b) Surface trough and strong lower tropospheric winds, with backing to W to W S W direction	42
c) Surface trough and mid–tropospheric trough/low	25
All three features	15

Activity of monsoon over the Arabian Sea Islands is most frequently associated with type (c). Usually the southern portion of a trough off Coastal Karnataka extends to north Kerala, but not further south. Nearly half the number of active to vigorous monsoon situations in Konkan and three quarters of such occasions in Coastal Karnataka are with troughs off the west coast. Heavy rainfall is generally in the southern portion of the trough. But the intensity and distribution of rainfall along the coast may vary widely in association with troughs.

9.5.4 Figs. 9.10(a to k) show a trough which appeared off Kerala and Coastal Karnataka on the 11th July 1969 and moved to Saurashtra and off Konkan by the 14th (Srinivasan, Raman, Mukherji and Ramamurthy 1972). On that day (l4th), the trough linked with the low over northwest India and adjoining Gujarat. Persisting for another day, the system weakened on the l6th. Easterly surface winds along the coast indicated the trough. Upper winds were affected as only a weakening of westerlies upto 12th in the lower troposphere and strengthening over Laccadives from 13th onward. As the system came near Konkan, lower tropospheric winds were showing a cyclonic circulation. A trough line at 600 mb along 15^{o} N on 11th over the Peninsula moved to 17^{o} by 13th. In the lower troposphere moderate to strong westerlies spread upto 20^{o} N by 16th. The belt of active monsoon rains along Kerala and Karnataka coasts on 11th progressively moved northward.

9.5.5 Jayaram (1965) has illustrated (Fig.9.1l) that heavy rainfall associated with a trough off the Konkan Coast is to the south of the apex of the trough and this belt moves progressively northward with the shift of the trough.

9.6 Sub–tropical Cyclone

9.6.1 'Cut–off' cold upper level cyclones developing equatorward of main polar westerly stream in the Pacific north of Hawaii, were called by Simpson (1952), 'sub–tropical cyclones'. Many of these are associated with surface cyclones. Their formation is always preceded by the injection of cold air aloft through the agency of large–amplitude troughs in polar westerlies. Ramage (1962) brought out that in these systems, 600 mb to 400 mb is the layer of largest pressure gradients, strongest winds and greatest convergence, increasing from periphery inwards, over a distance of 500 km. Within a radius of about 100 km from the centre of the sub–tropical cyclone, only scattered clouds are noticed and the area is compared to the eye of regular cyclones. Between 100 and 500 km, vigorous upward motion and deep precipitating clouds are present. Their base is, however, at 700 mb. Another layer of clouds is between surface and 800 mb, subsidence intervening between the two cloud layers. This last feature distinguished sub–tropical cyclones from other tropical lows. A sub–tropical cyclone is also distinguished from a large amplitude trough in polar westerlies associated with a surface low, in that bad weather in the latter system is generally concentrated east of the trough axis. Although the genesis of sub–tropical cyclone is out of such a trough, weather is more symmetrically developed in the former. Fig. 9.12 gives a schematic representation of the sub–tropical cyclone (Ramage 1971).

9.6.2 Miller and Keshavamurthy (1968) have presented their analysis of composited data for the period 2 to 10 July 1963 of the area between 60^{o} and 81° E and 8^{o} and 28^{o} N as evidence of existence of sub–tropical cyclones accounting for most of the monsoon rains of western India and northeast Arabian Sea. Extensive aircraft reconnaissance reports of wind, weather and temperature (dropsondes) in East Arabian Sea were available during the period. Their charts for surface and 700, 600 and 500 mb are given in Figs. 9.13 (a to d). A trough extends at surface from Saurashtra and Kutch to off north Konkan coast, while closed circulations exist at 700 mb at 20.5° N, 73° E, at 600 mb at 20° N, 72.5° E and at 500 mb at 19° N, 72° E. Southward slope with height is quite a common feature in many monsoon lows, including depressions, which can be explained by the thermal field. In their composited temperature field, the centre of the cyclone at 700 mb was colder than its environment, at 600 mb neither warmer nor colder and at 500 mb warmer. The lapse rate was less within the circulation than on its periphery.

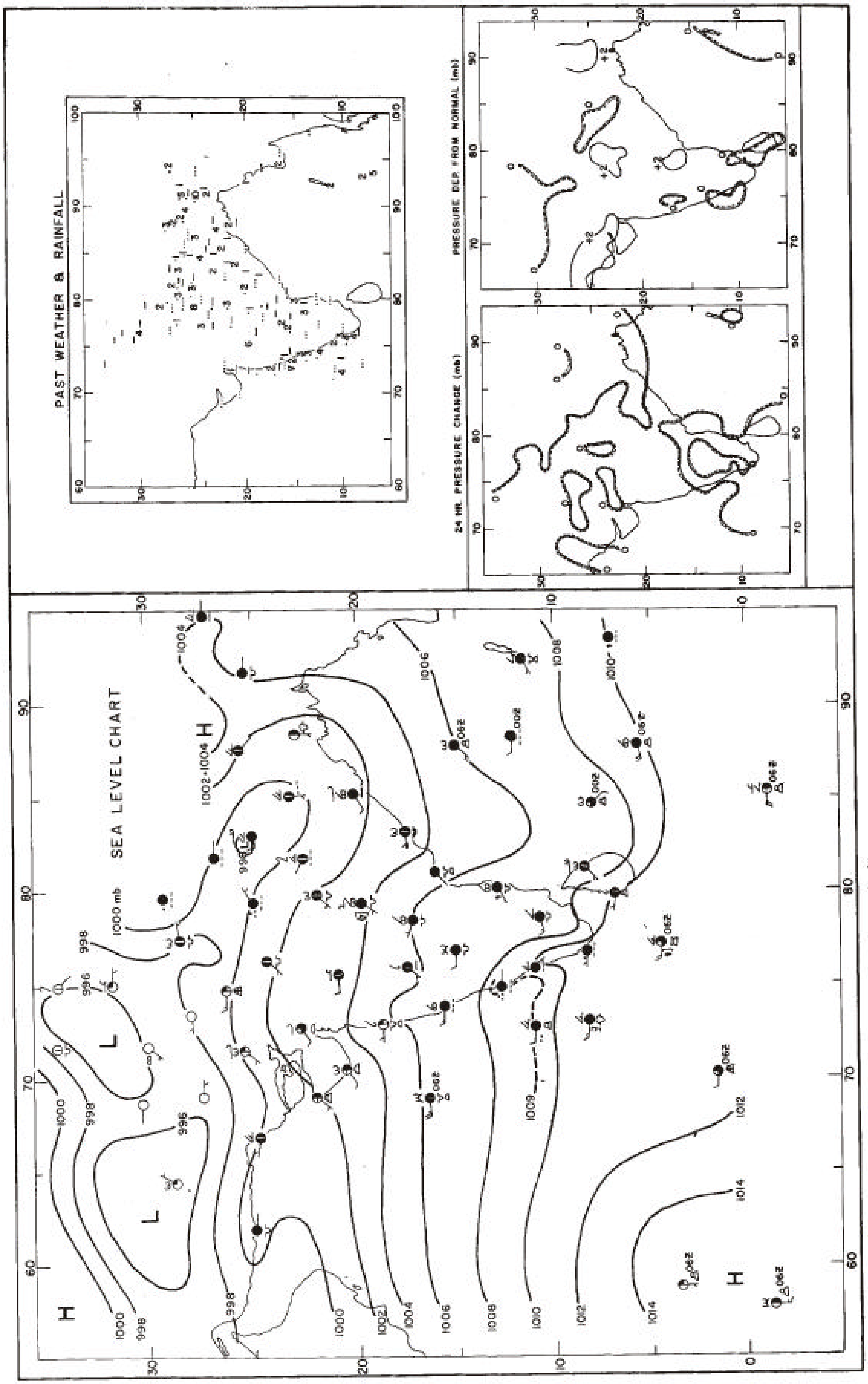


Fig. 9.10(a) Synoptic Charts 0300 GMT 11 July 1969

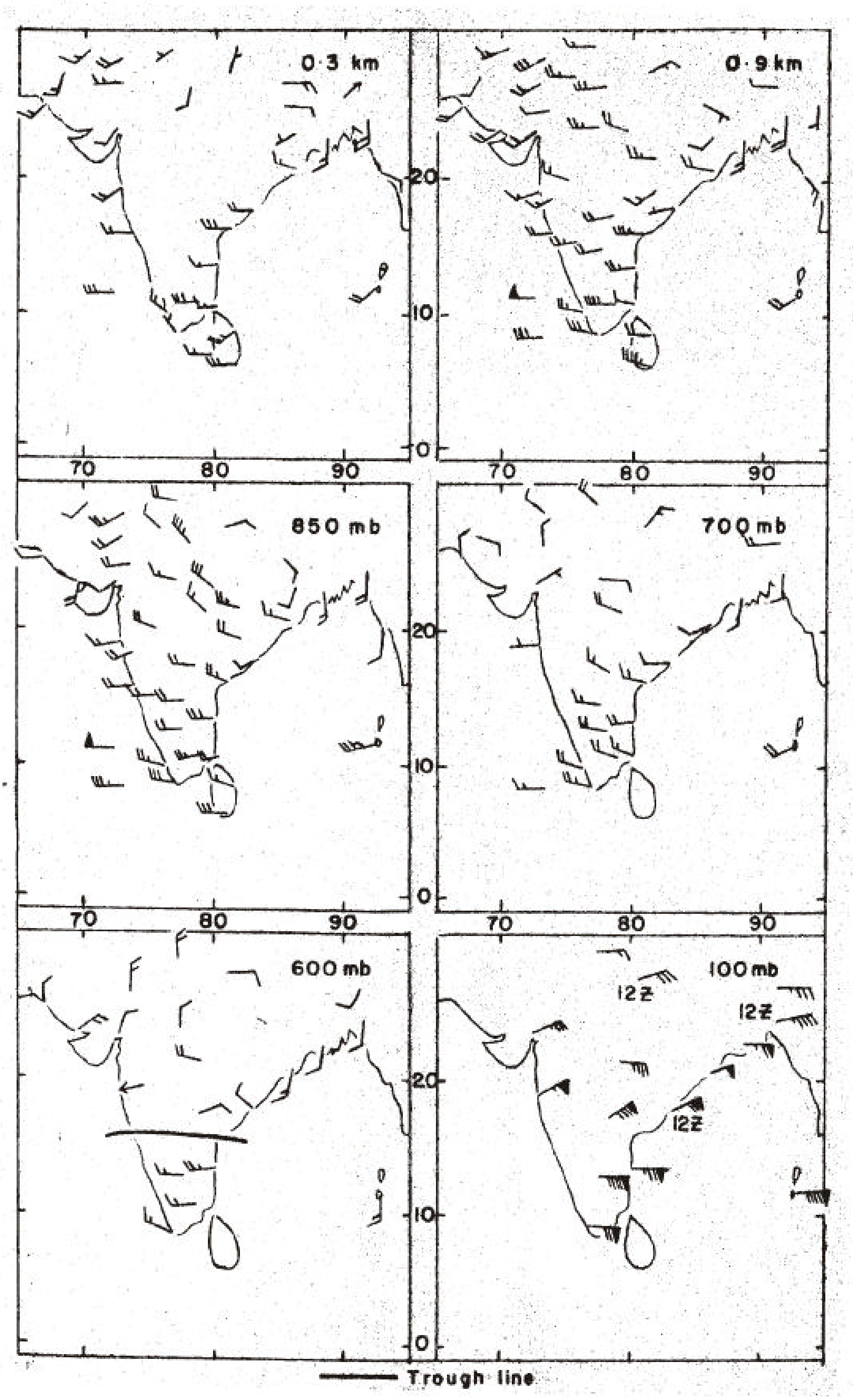


Fig. 9.10(b) Upper Winds 11 July 1969 00 GMT

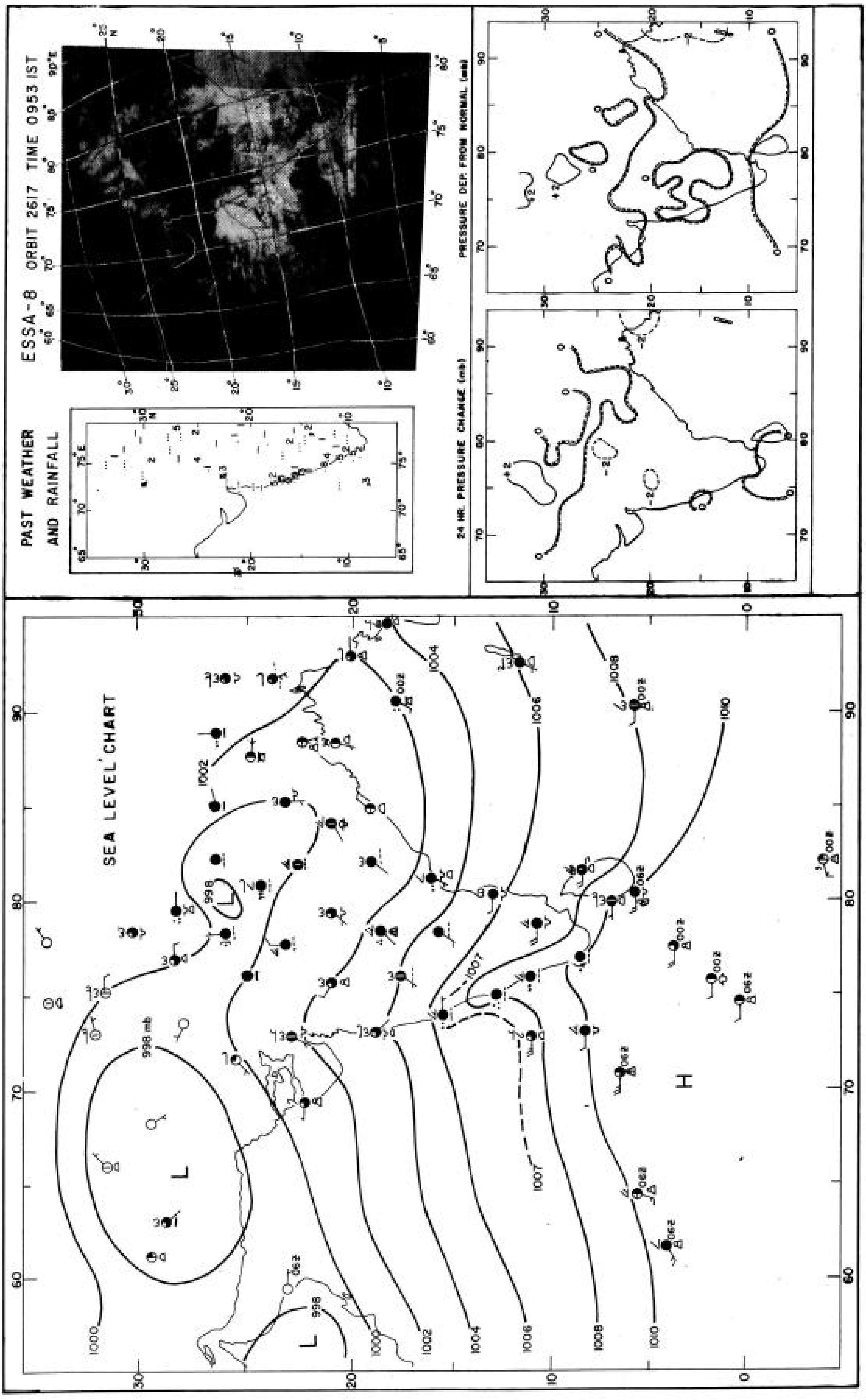


Fig. 9.10(c) Synoptic Charts 0300 GMT 12 July 1969.

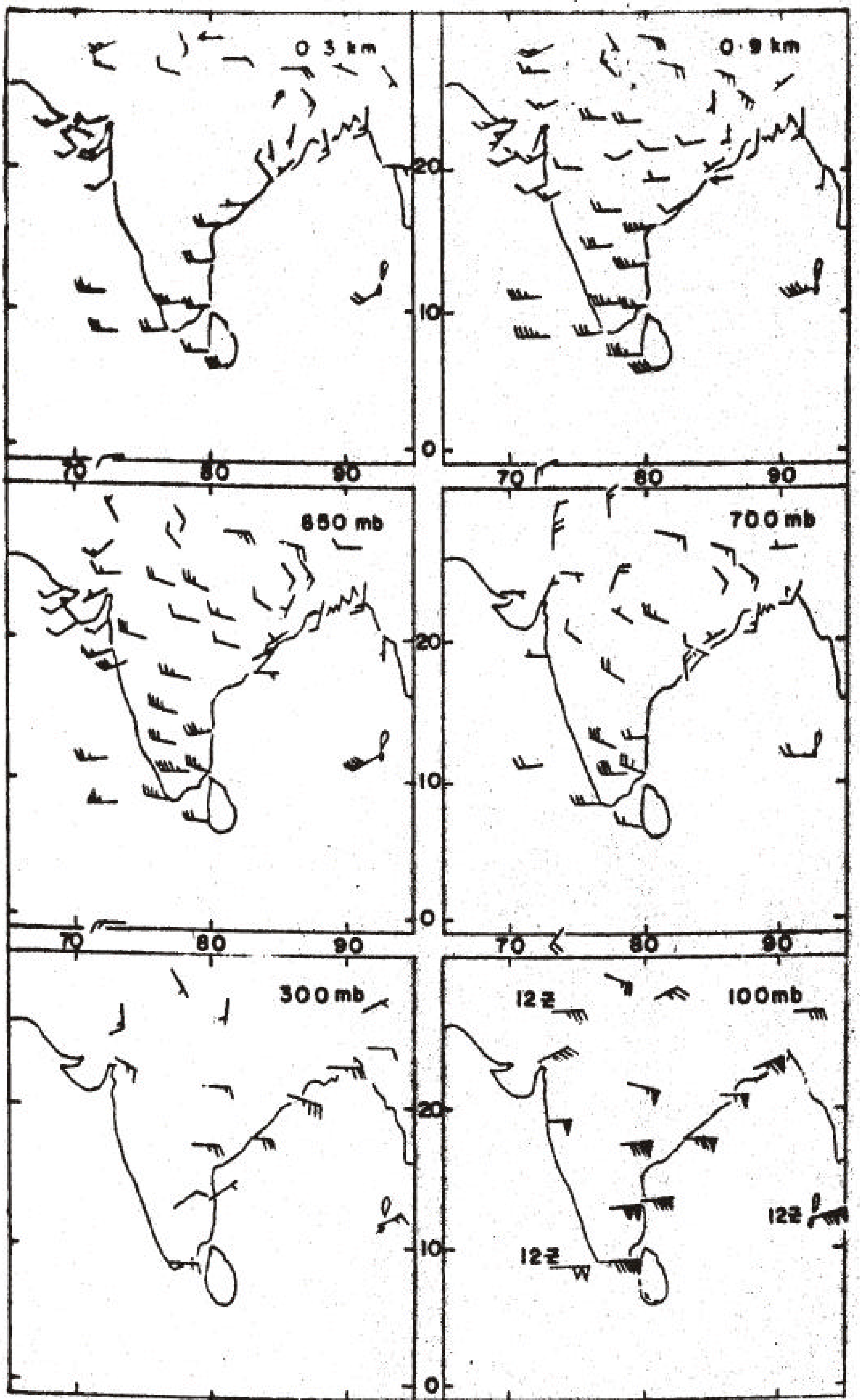


Fig. 9.10(d) Upper Winds 12 July 1969 00 GMT

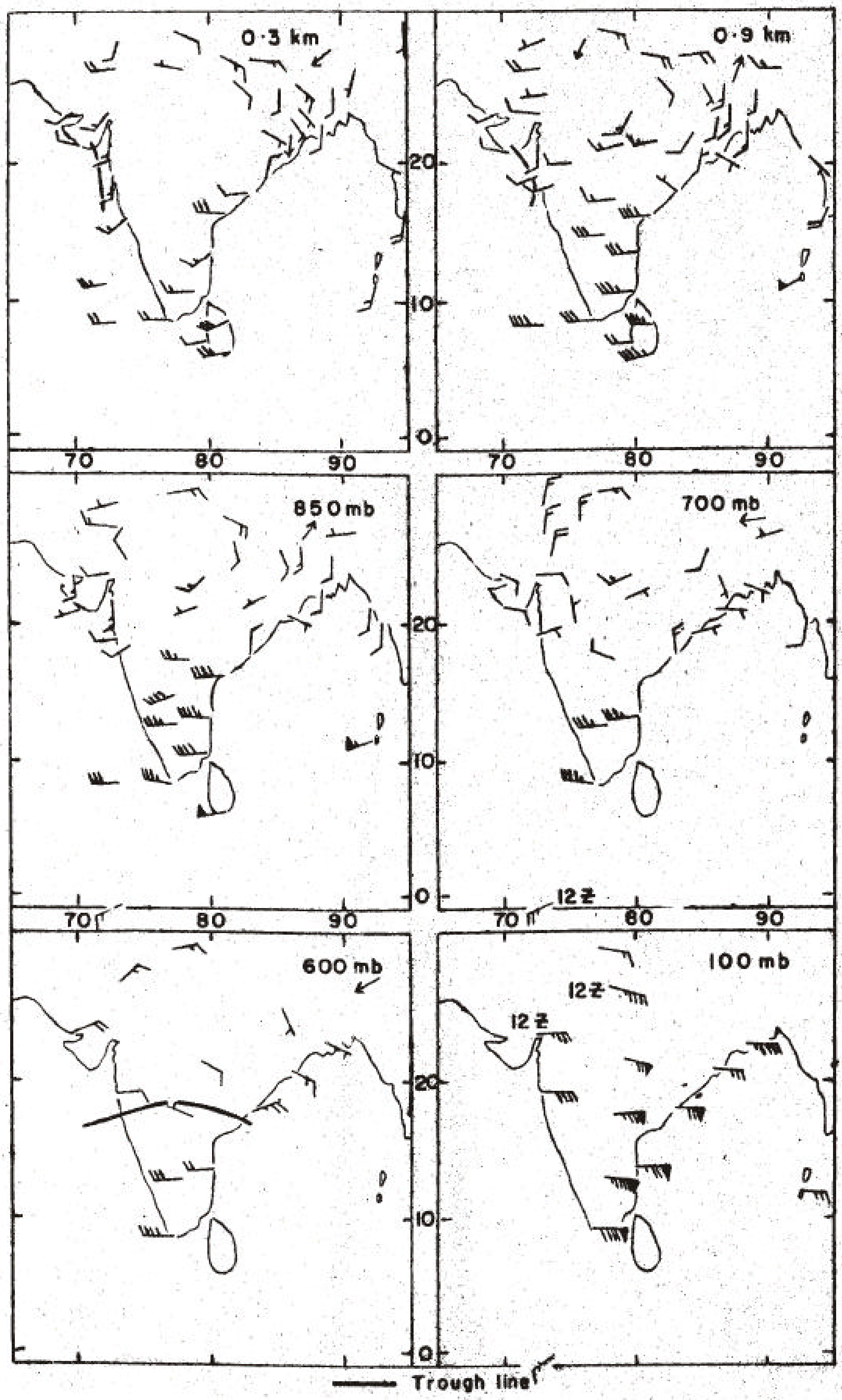


Fig. 9.10(e) Upper Winds 13 July 1969 00 GMT

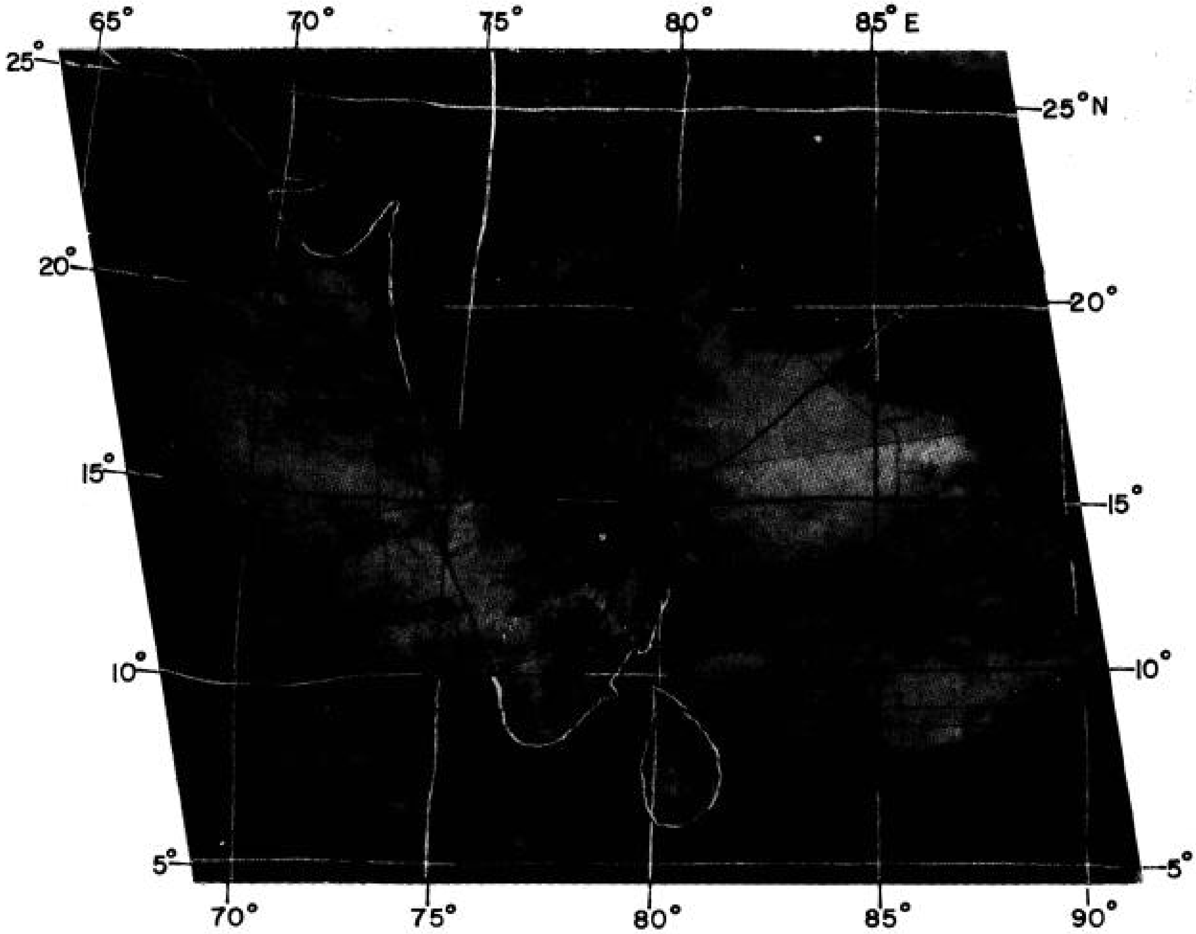


Fig. 9.10(f) NIMBUS-3 13 July 1969

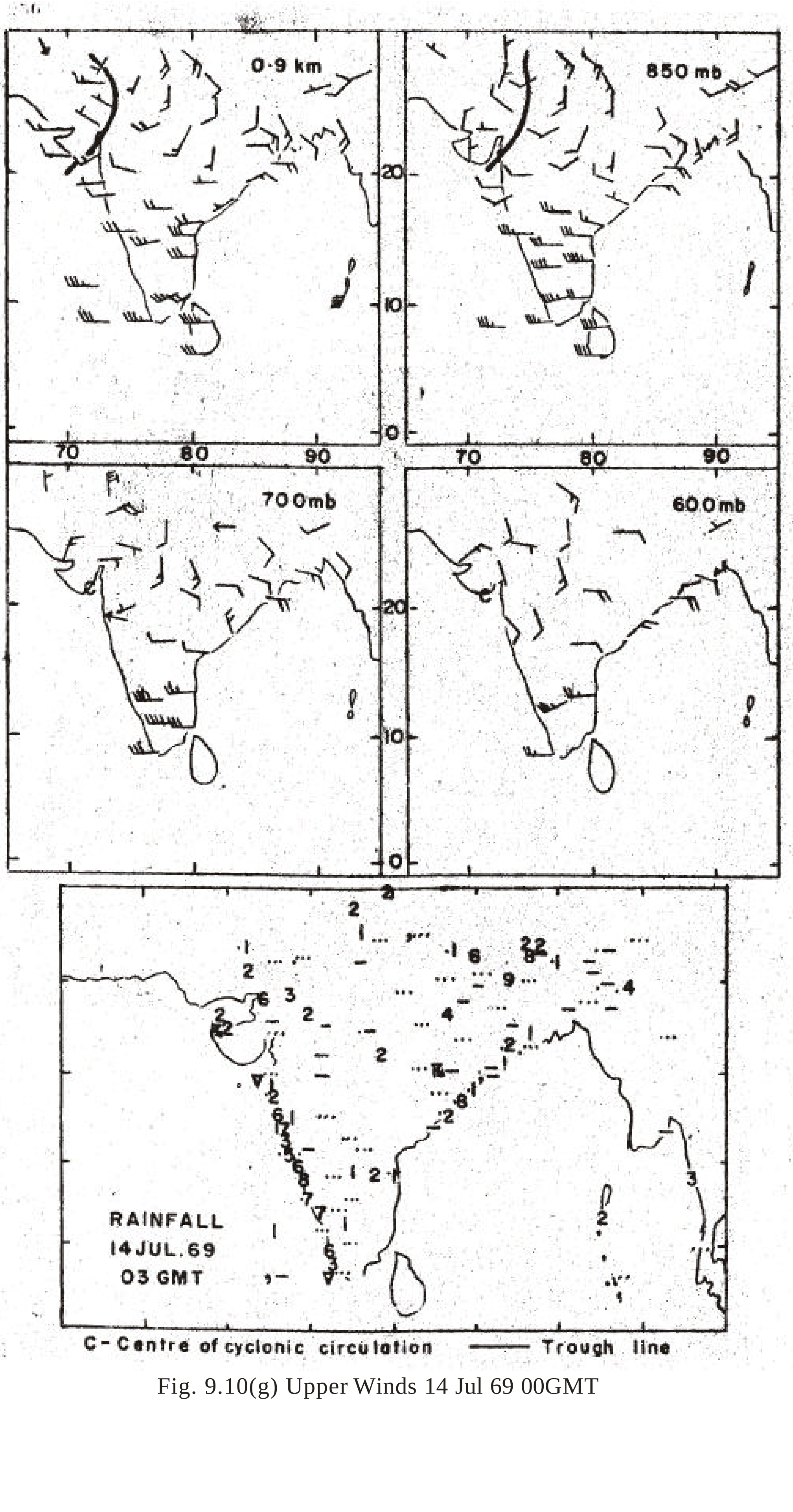


Fig. 9.10(g) Upper Winds 14 Jul 69 00GMT

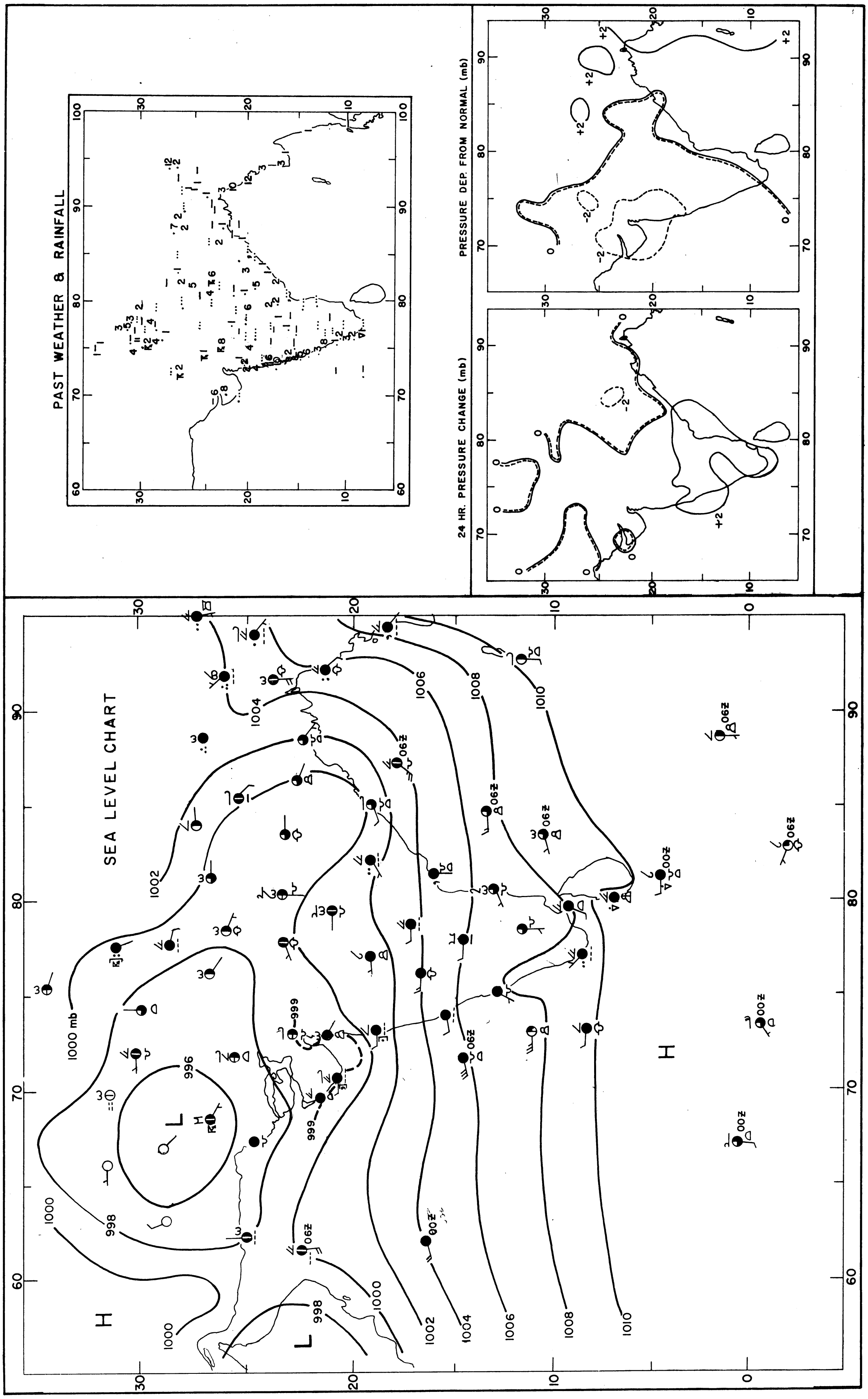


Fig. 9.10(h) Synoptic charts 0300 GMT 15 July 1969

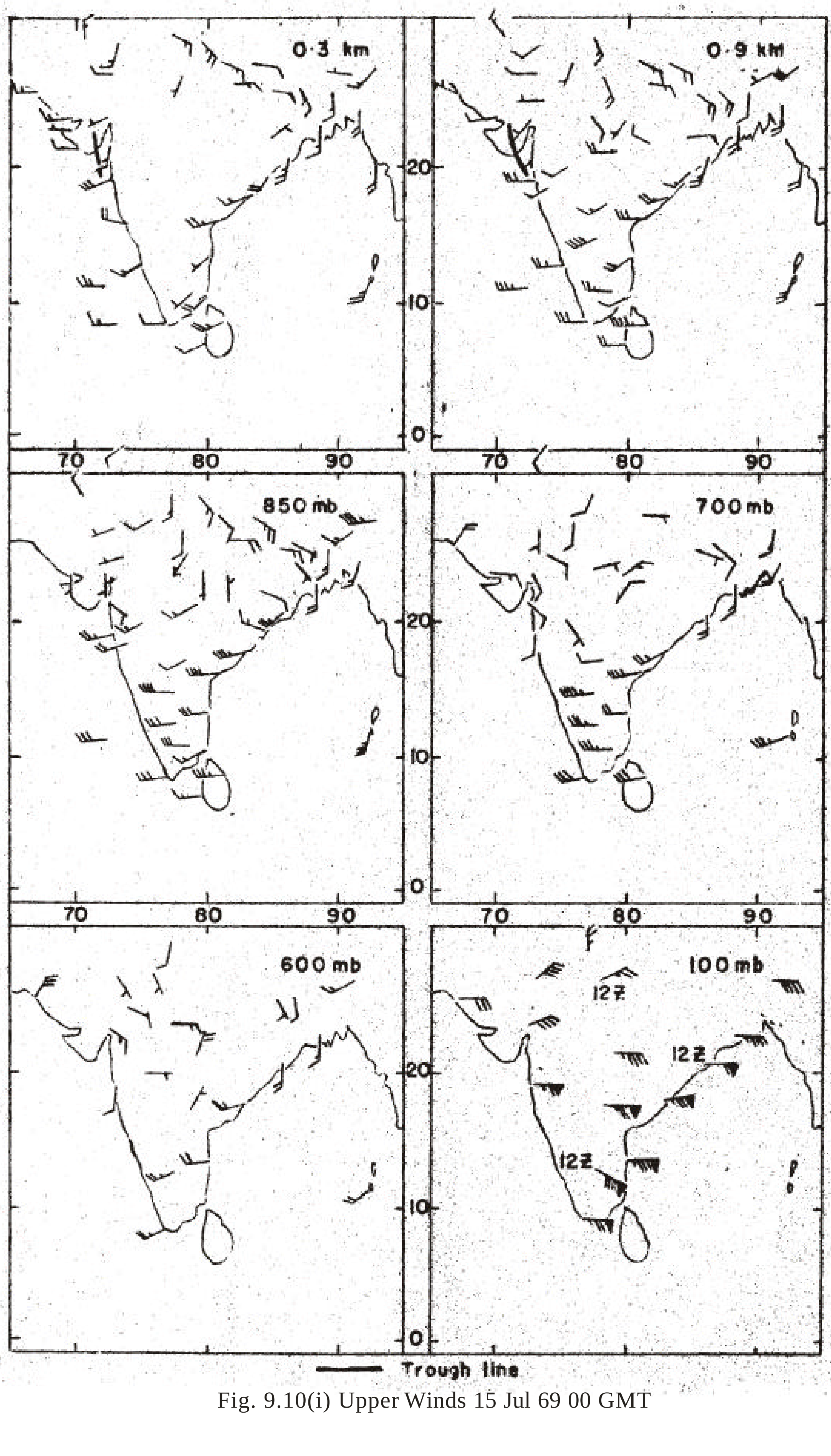


Fig. 9.10(i) Upper Winds 15 Jul 69 00 GMT

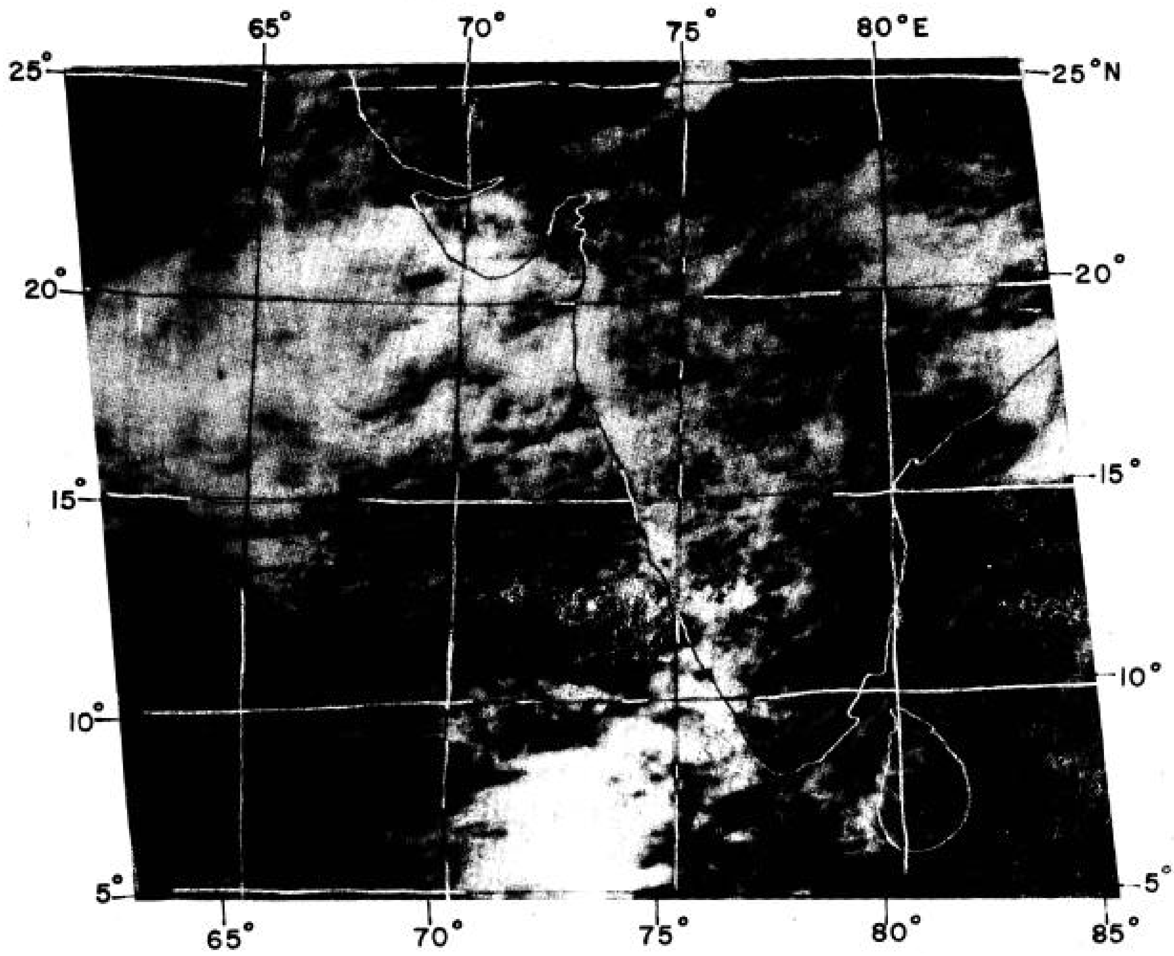


Fig. 9.10(j) NIMBUS-3 15 July 1969

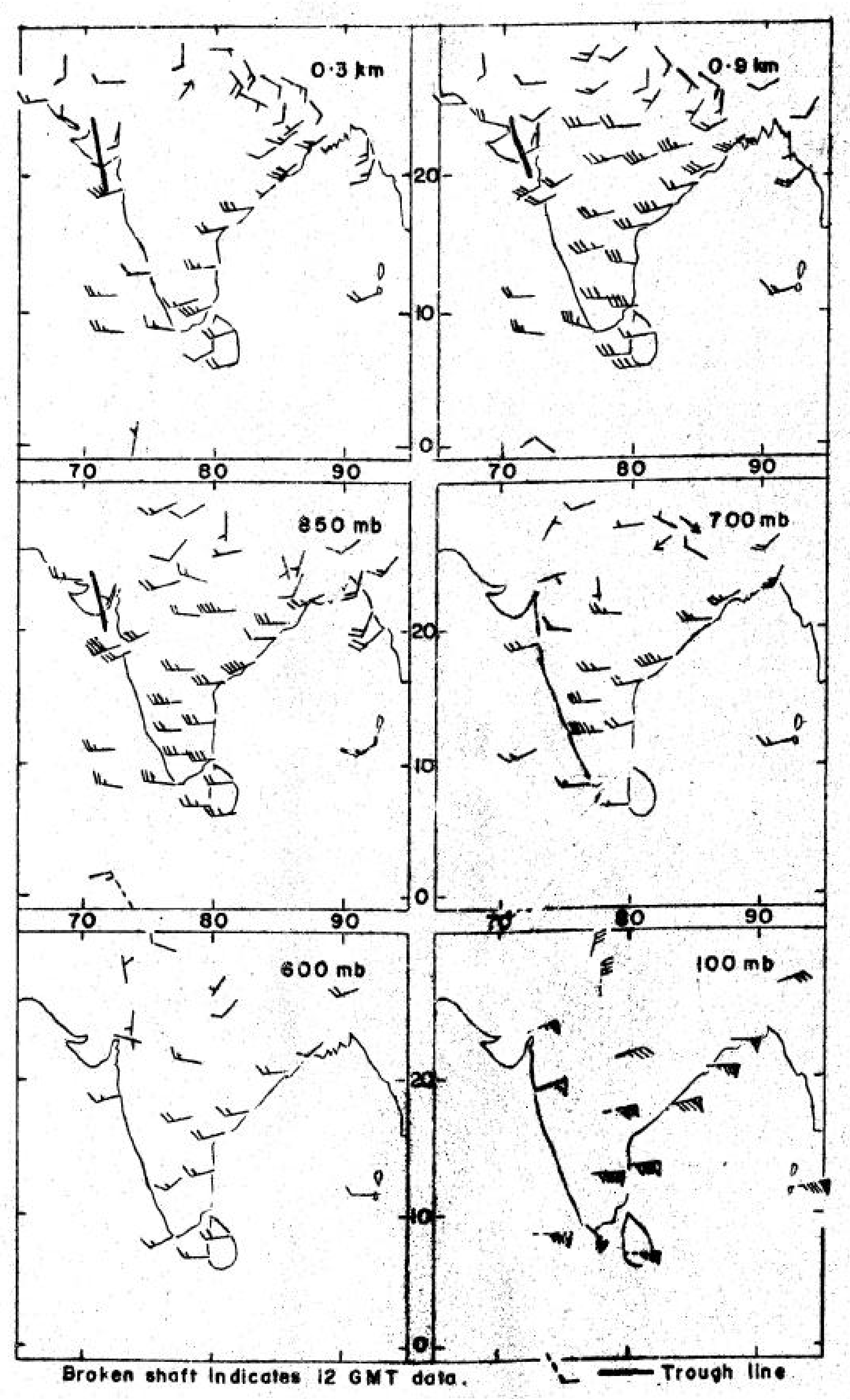
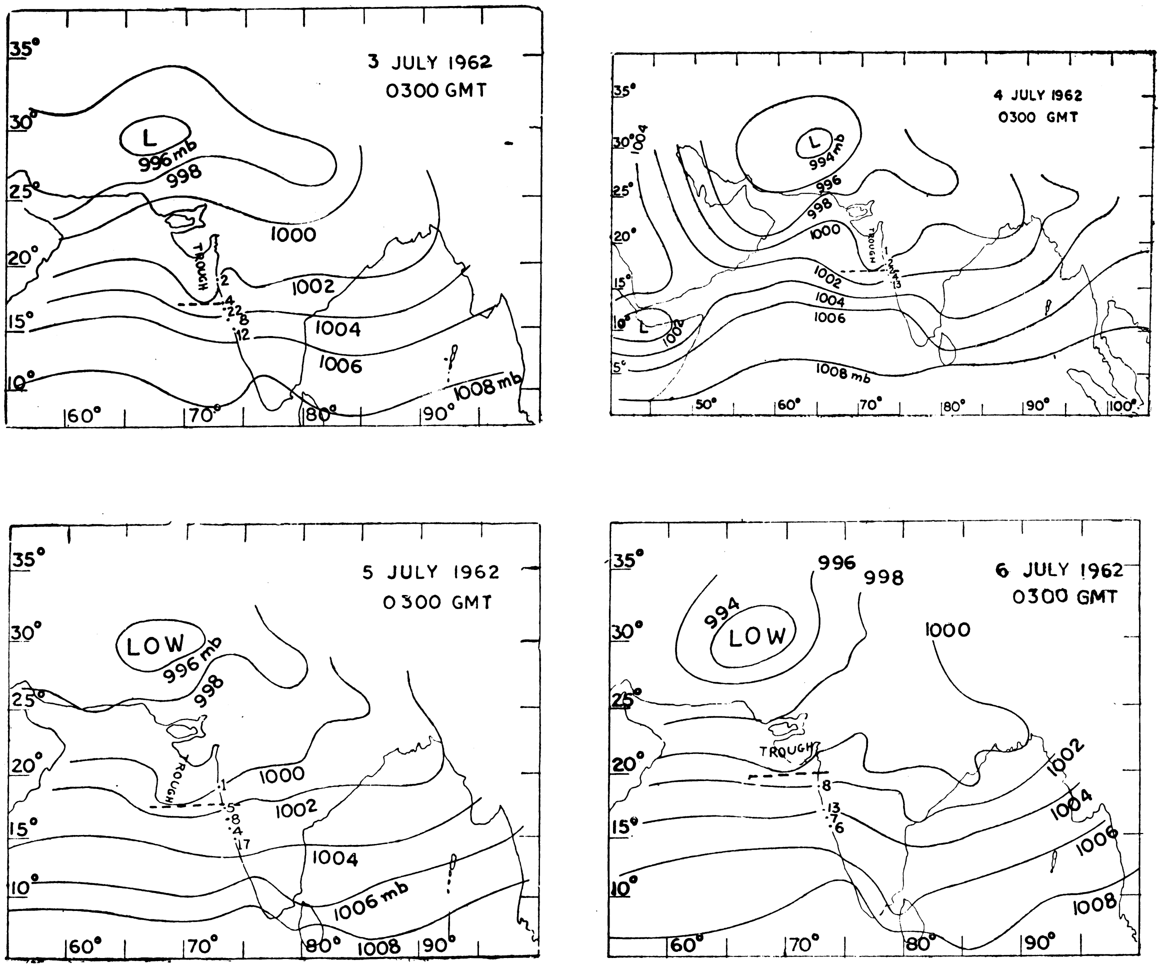


Fig. 9.10(k) Upper Winds 16 Jul 69 00GMT


3 JULY 1962
0300 GMT
L
996 mb
998
1000
1002
1004
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4 JULY 1962
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1008
6 JULY 1962
0300 GMT
994
LOW
TROUGH

Fig. 9.11

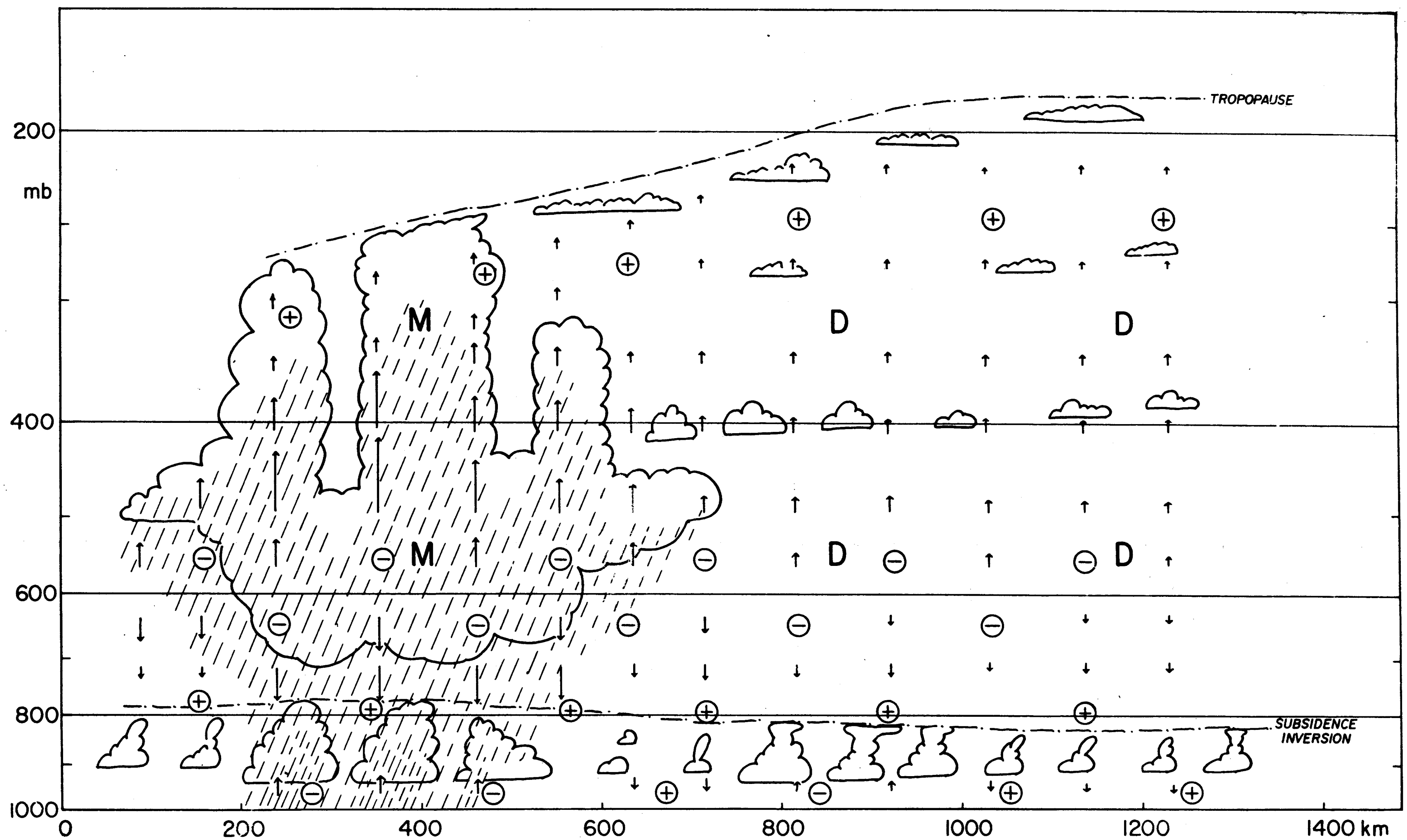


Fig.9.12 Schematic radial cross section of a sub-tropical cyclone. Severe constriction of the horizontal scale permits only gross features of the vertical motion and cloud systems to be shown. Divergence is indicated by plus signs, convergence by minus signs. Regions of vertically moving air undergoing dry adiabatic temperature changes are denoted by D and regions undergoing moist adiabatic temperature changes by M. (Ramage, 1971).

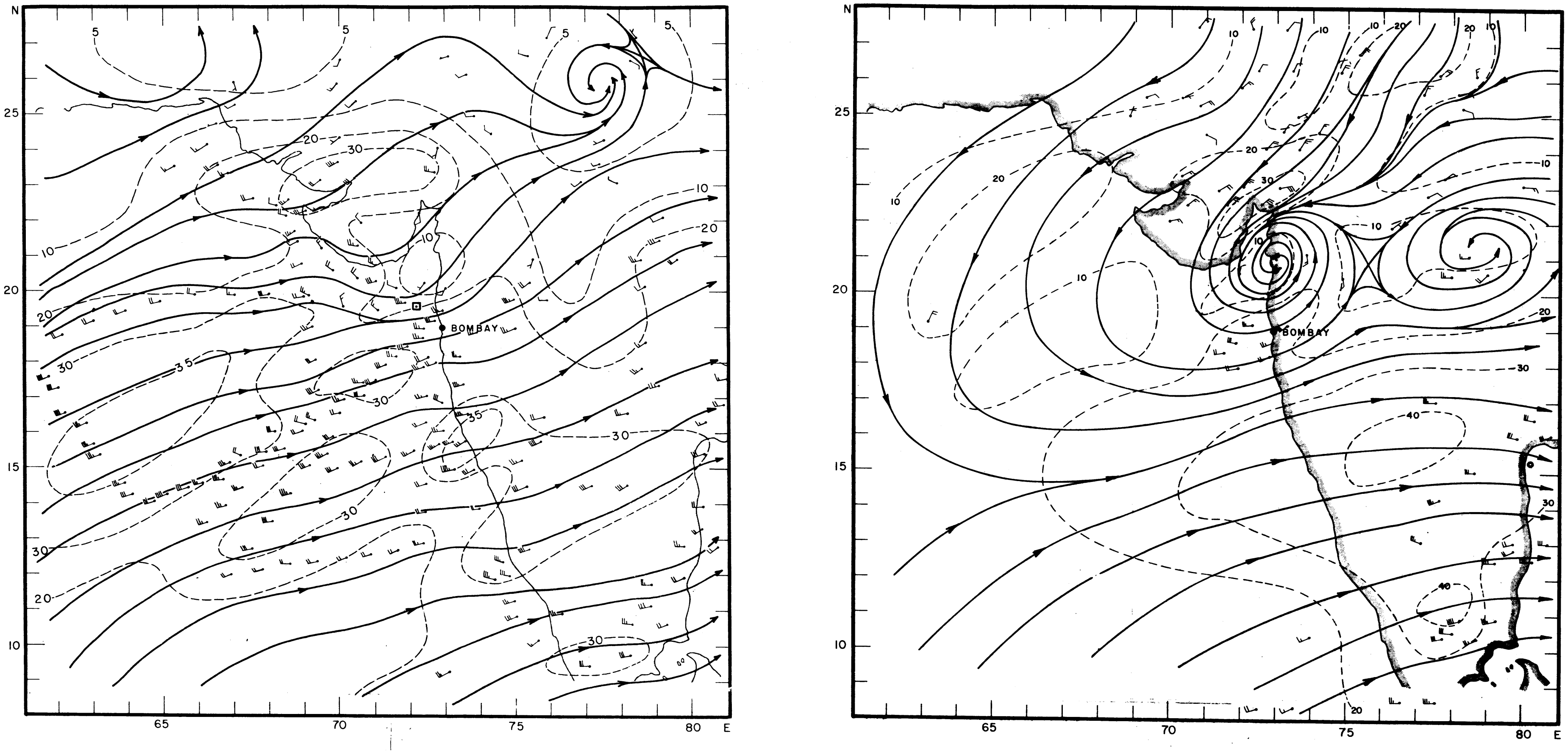
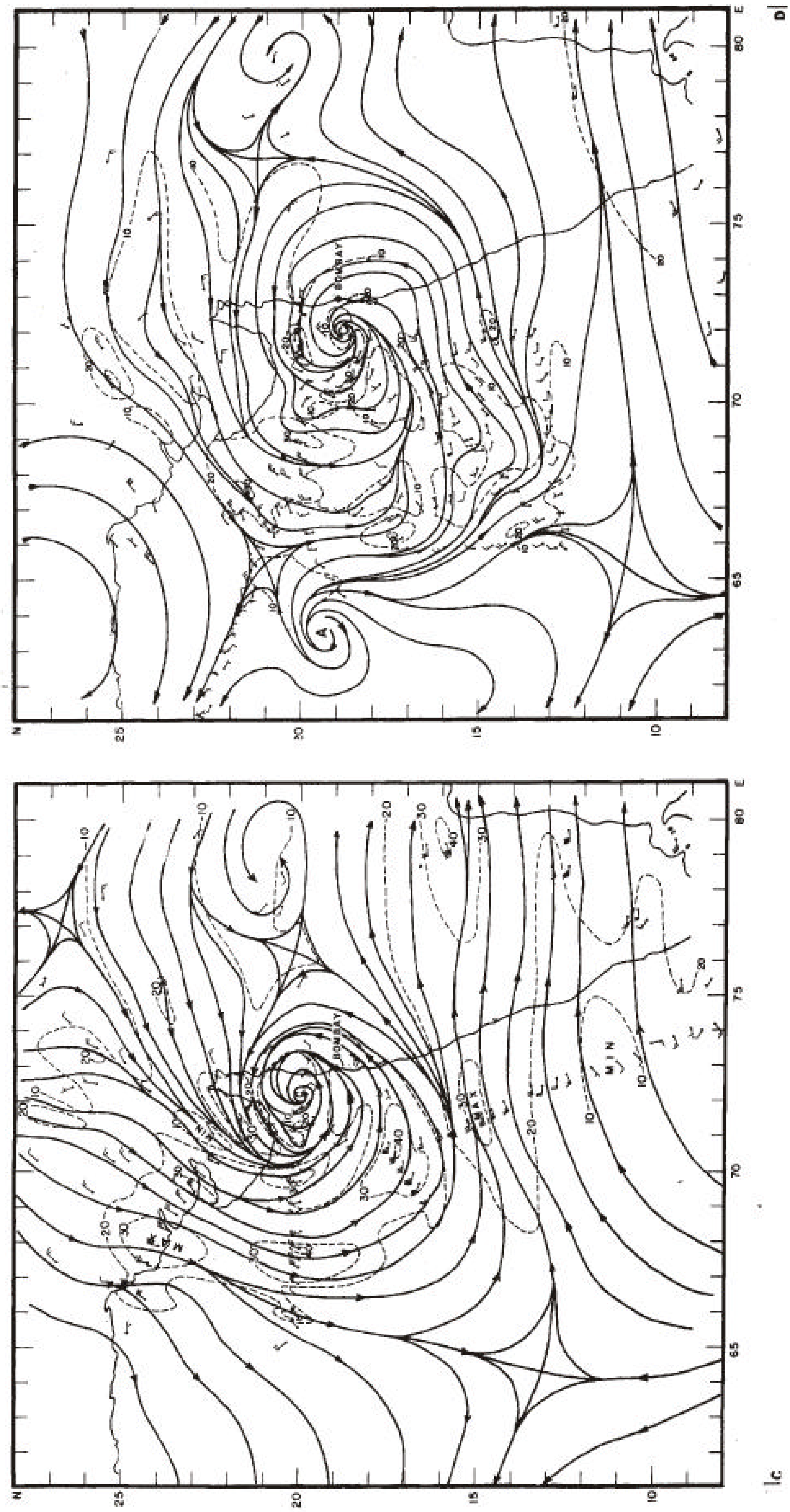


Fig.9.13 Composite kinematic analyses for
(a) the low level (500 m to 900 m)
(b) 700 mb.
Solid lines are streamlines; dashed lines are isotachs. The letter 'C' over the Konkan coast denotes the centre of the wind field for the Arabian Sea cyclone (Miller and Keshavamurthy, 1968).



9.13 Composite kinematic analyses for (c) 600 mb. (d) 500 mb.
Solid lines are streamlines; dashed lines are isotachs. The letter C' over the Konkan coast denotes the centre of the wind field for the Arabian Sea cyclone. The letter 'A in chart (D) denotes the centre of the anticyclone over the Arabian Sea (Miller and Keshavamurthy, 1968)

9.6.3 Before presenting their computations of derived parameters in the area from the unsurpassed wealth of data, though spread over a few days, the reason for identifying this system as a sub–tropical cyclone and regarding such systems as frequent and important for Indian monsoon dynamics is worth examining. While a trough off the Konkan coast is frequent, it is only occasionally associated with closed low at 700 mb. Compositing has smoothened out the surface feature much more, as the shift in position of the surface low was greater. For example, closed low was present at surface on 3rd and 4th which is not seen in their composite charts. On 3rd, a ship was reporting wind of 50 kt from northwest within 1° of centre. A trough at 600 mb or so in the area is a climatological feature which by itself is innocuous in causing rainfall. The climatological mean is not due to frequent presence of sub–tropical cyclones. No evidence has been presented to show that cold air with a large amplitude trough in westerlies has reached 20° N and caused the formation or intensification of the trough in mid–troposphere. The differences in temperature field cited as between centre and periphery is of the same magnitude of the changes noticed in areas of intense convection in monsoon area brought about by vertical uplift. The mean monsoon ascent at all places has a lapse rate slightly in excess of saturated adiabatic lapse rate, more so over Gujarat and Rajasthan where continental influence is greater. With intense convection, lapse rate becomes nearer saturated adiabatic through a slightly lower surface temperature and it is not unusual for the ascent in the rain area to be colder than other regions in lower troposphere and slightly warmer aloft. This is more so when the drier northwestern parts of the Indian sub–continent are also considered. Not infrequently systems are somewhat more intense in mid–troposphere than at surface, as in this case. That alone does not seem to be a reason for designation as sub–tropical cyclones, as long as their formation is not as per the pattern noted by Simpson (1952). Systems existent in upper air only are being described by Indian meteorologists as low or trough in upper air. As mentioned earlier, some lows form to the north of 20° N, under the influence of large amplitude westerly troughs. Whether such lows have the characteristics of sub–tropical cyclones deserves further study.

9.6.4 Miller and Keshavamurthy (1968) have computed divergence, vorticity and vertical motion from streamline analysis and their results are presented in Figs. 9.14, 9.15 and 9.16, using the cyclone centre at 500 mb as the origin. Increase in relative vorticity with height upto 500 mb and slope of the line of maximum vorticity towards south and west are seen. As mentioned earlier, estimate of maximum vorticity at surface has been affected by compositing. The anticyclonic vorticity at the periphery is to be expected whenever cyclonic vorticity develops over a small area. The maximum anticyclonic vorticity is not more than a quarter of the cyclonic vorticity.

A comparison of streamlines with contours showed that a nonsteady state existed within 250 km of centre. The streamlines spiralling into the centre of the low pressure cut the contours at a sharp angle, especially in the southern half between 700 mb and 500 mb. This is explained as due to constant state of flux of the thermal field, due to changing intensities of rainfall, horizontal wind speeds and vertical motions, as in tropical cyclones. The maximum convergence at the centre is at 500 mb and the level of non–divergence at 300mb.

North of 23° N, convergence is seen at 300 mb and 200 mb, marked divergence between 700 mb and 400 mb (almost two–thirds of the convergence at the centre of the cyclone) and again convergence below.

Vertical velocity (ascent) near the centre is maximum at 300 mb and continues well above 200 mb explaining the rainfall intensities. In the descent at the periphery, velocities reach 15 cm sec^{-1} in upper troposphere near 64° E, about 40% of the rate of ascent at centre.

9.6.5 Miller and Keshavamurthy (1968) found in three sub–tropical cyclone sequences, that middle tropospheric development was preceded by cyclonic activity over eastern India, by above normal anticyclonic activity, above the heat low and by above normal cyclonic vorticity between 18° N

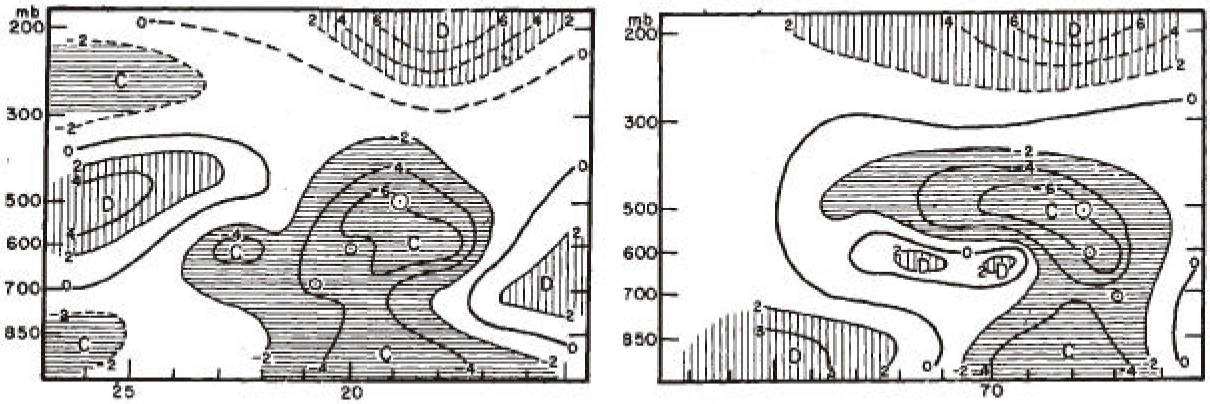


Fig.9.14 (a) Meridional cross section, (b) Latitudinal cross section of divergence through the 500 mb composite centre (l8.8°N, 72°E) of the upper level cyclone. Divergence is given in units of 10 $^{-5}$sec $^{-1}$ All values below 300 mb are based on the divergence computations at selected levels for 2,4 & 7July. Above 300 mb, the dashed lines represent divergence computed from 2nd July data only. The circled dots represent composite centres of the cyclone. The letter C and D denote centres of convergence and divergence respectively. (Miller and Keshavamurthy, 1968).

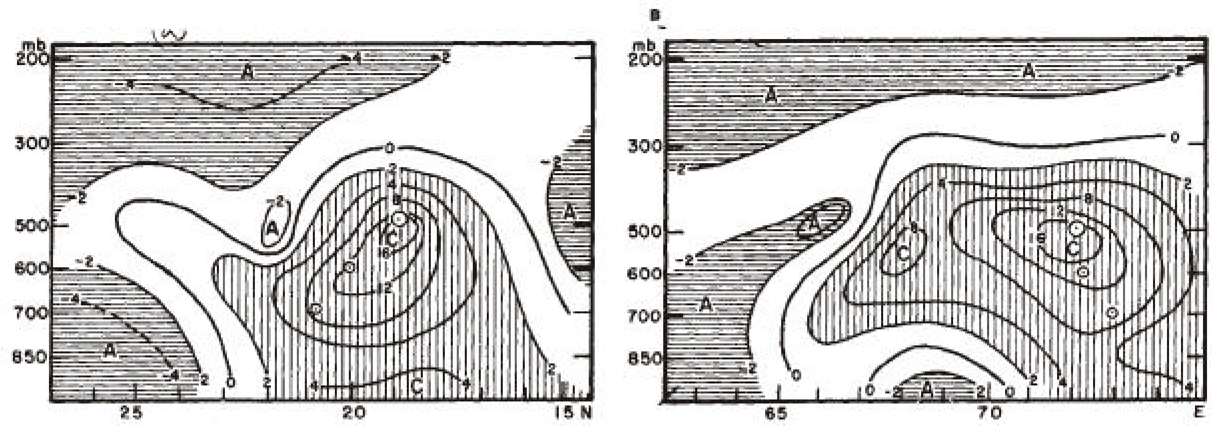


Fig.9.15 (a) Meridional cross section, (b) Latitudinal cross section of relative vorticity (in units of 10^{-5} sec^{-1}) through the 500 mb composite centre (l8.8°N, 72°E) of the upper level cyclone. Values above 300 mb are based on 2 July data. Those below,300 mb are based on data of 2, 4 and 7 July. The circled dots represent composite centres of the cyclone. The letters C and A denote centres of cyclonic and anticyclonic vorticity respectively. (Miller and Keshavamurthy, 1968)

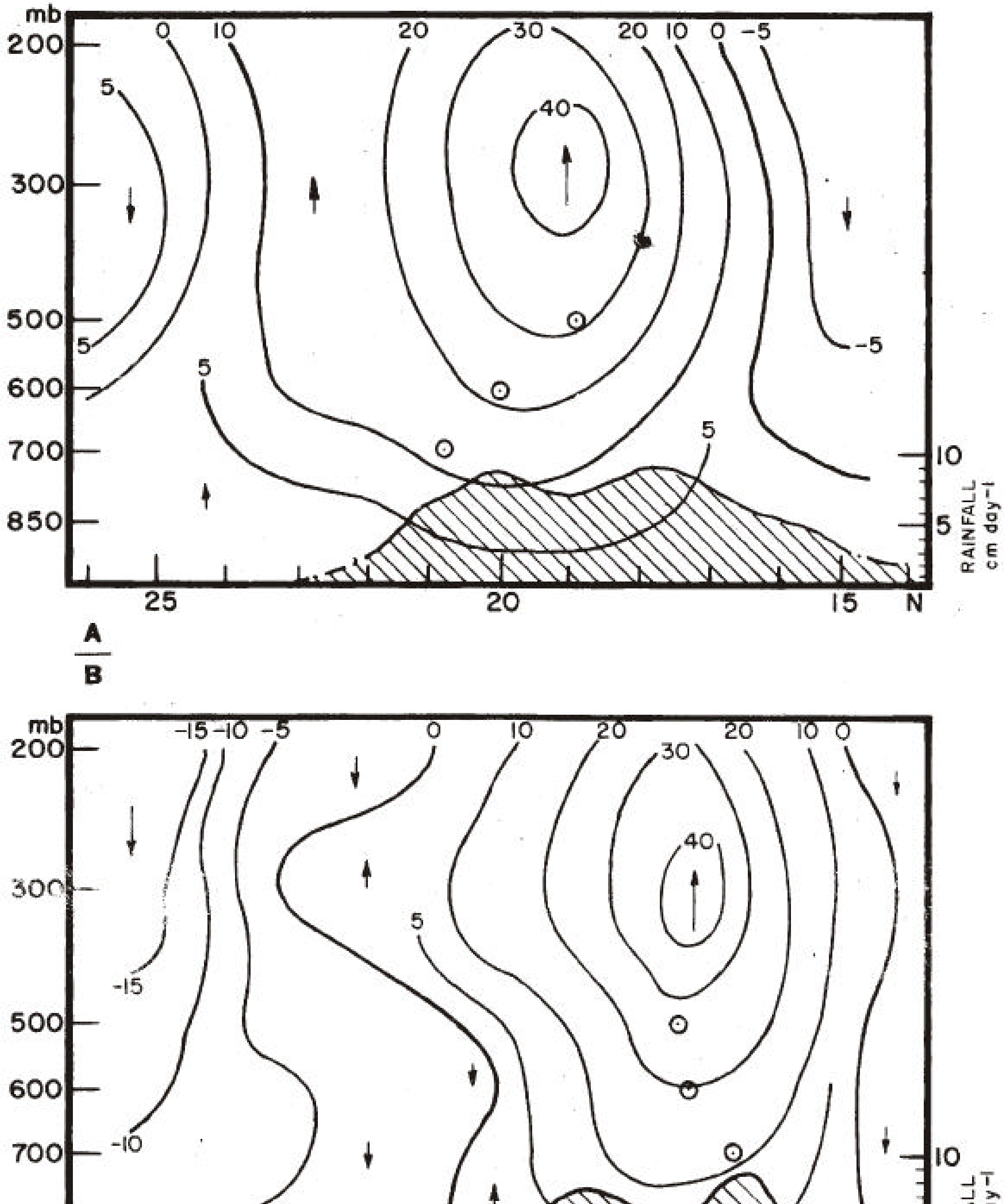


Fig.9.16 (a) Meridional cross section and (b) Latitudinal cross section, of vertical motion through the 300 mb composite centre (l8.8°N 72°E) of the upper level cyclone based on the composite divergence fields. Isotachs of vertical motion are in cm sec^{-1} with negative values downward. The hatched area below the vertical motion profiles shows the meridional and latitudinal variation in rainfall along and across the Konkan coast based on a composite analysis of all daily rainfall amounts from 2 to 8 July at coastal stations. (Miller and Keshavamurthy, 1968).

and 21° N. Then, as the sub–tropical cyclone intensified, anticyclonic vorticity above the heat low increased sharply. It has been known for a long time that with a depression or low forming in north Bay, monsoon strengthens off Konkan coast and one of the methods for this is for a trough to form off the coast at sea level which is described as sub–tropical cyclone by these authors.

9.6.6 Dissipation of sub–tropical cyclones in this area is ascribed to flow of drier air from northwest into the field at mid–tropospheric levels. This view is contrary to general experience in monsoon systems that as long as lower tropospheric moist air flow is maintained, moisture variations in mid–troposphere are controlled by vertical motion. Change in vertical motion causes both dissipation of lows and drying in mid–troposphere.

9.6.7 Ramage (1966) concludes that the quasi–steady heat low over the north west of the sub–continent is subject to vorticity divergence, to shed its vorticity (produced by the continuous radiational heating) into its neighbouring northeast Arabian Sea, to generate sub–tropical cyclones. He uses Pettersen's equation

$$\nabla.\left(z_a \vec{V}\right) = z_F - \left(\frac{\partial z_a}{\partial q}\right)\left(\frac{dq}{dt}\right)$$

for a surface of constant potential temperature, derived by assuming steady state and neglecting the contribution of vertical advection of the vorticity vector. The vorticity (z_F) of the heat source is positive, $\left(\frac{\partial z_a}{\partial q}\right)$ is negative due to the overlying ridge above the heat low and $\left(\frac{dq}{dt}\right)$ is positive with the convergence in the surface heat low. Hence positive divergence of vorticity has to take place from the heat low, But it is not established that this net export takes place into adjoining area of northeast Arabian Sea. The type of synoptic situation cited by Miller and Keshavamurthy as an example of sub–tropical cyclone, usually develops near 15 – 17^{o} N and progressively moves northwards into Saurashtra. If it is a process of export of vorticity from the heat low, the development should first take place in adjoining regions and move outward.

9.6.8 About the clouds and rain in such situations their (Miller and Keshavamurthy, 1965) conclusions are :

i) A large area of heavy Cb builds up mainly to the west (and southwest) of the mid–tropospheric cyclone centre.

ii) The heavy precipitation is generally near the coast but in the region of the mid–tropospheric trough it extends further out into the sea.

iii) The middle overcast is in the shape of a wedge extending from the coast towards the west upto 66^{o} E. It extends more on the southern side than to the north of the mid–tropospheric trough.

iv) All the heavy rain is not due to Cb clouds. The heavy rain in the southern portions of the mid–tropospheric trough, particularly along the coast, is due to stratiform clouds with embedded large cumulus.

These are illustrated in Fig. 9.17.

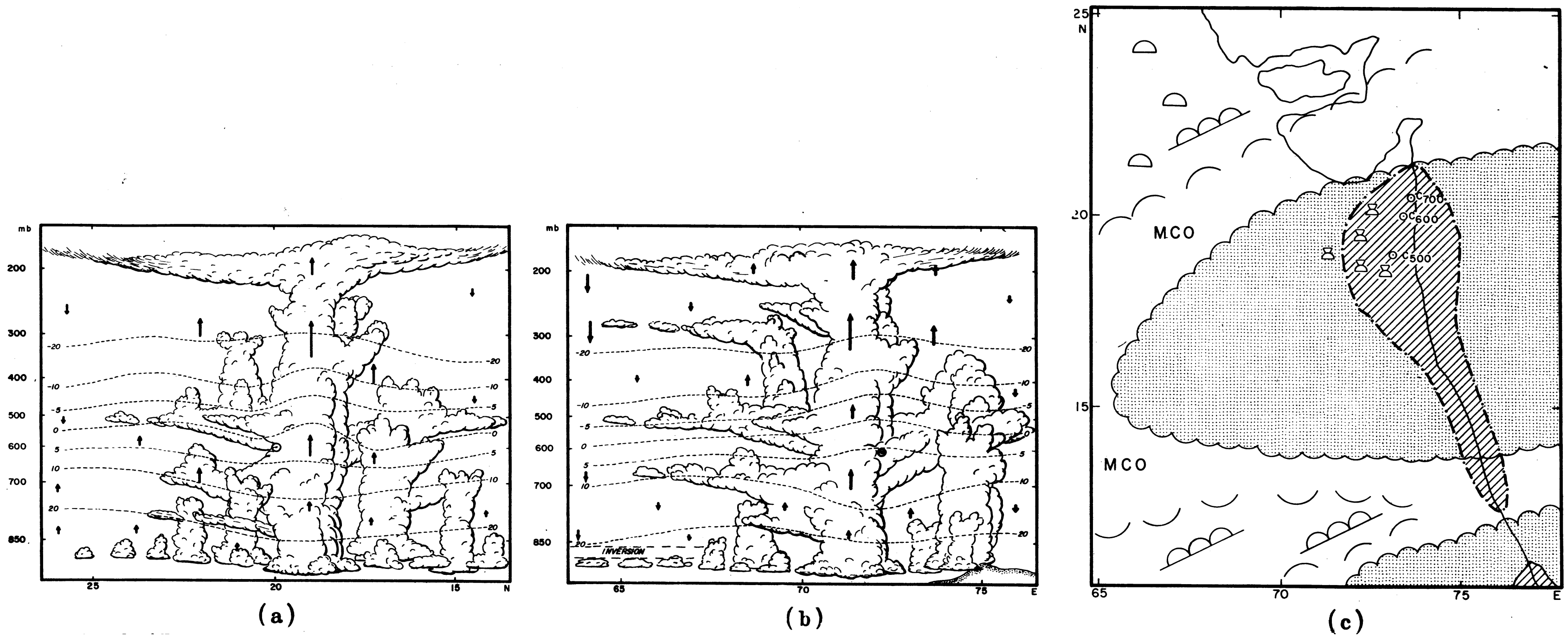


Fig.9.17 Composite Model showing the distribution of clouds, vertical motion and temperature through the cyclone centre with Chart (a) the longitudinal axis along 72°E Chart (b) the latitudinal axis along 18.8°N Chart (c) is a general plan of the weather based on weather distribution charts constructed for individual days. In charts (a) and (b) clouds are shown schematically on a greatly exaggerated vertical scale. The dashed lines are isotherms taken from the horizontal composite temperature charts. Solid arrows are vertical velocity vectors extracted from the cross sections in Fig. 9.16. The longest arrow is 40 cm sec^{-1} and the shortest 5 cm sec^{-1} The circled dot represents the 600 mb composite centre of the cyclone, the hatched area in Chart (c) shows the extent of rainfall exceeding 4 cm/day associated with the mid-tropospheric cyclone; and cumulonimbus symbols, surrounding the cyclone centres schematically show the region of greatest vertical cloud development. Stippled areas outline broken to overcast middle and high clouds; broken arcs show extent of broken to overcast low clouds and scattered high clouds (Miller and Keshavamurthy, 1968).

9.7 Upper Easterly Waves

9.7.1 In the upper troposphere, easterlies prevail during this period, increasing in speed with height, reaching jet speeds near 150/100 mb over some parts. Srinivasan (1960) finds this current dynamically unstable and is subject to perturbations like transverse waves (troughs) or wind maxima travelling from east to west. Wind speeds are greater than the rate of movement of these trough lines and there is convergence to the rear of the trough line and divergence in advance. Weather is caused by the upper divergence in advance. As the density of upper air stations is not close enough, vertical time–sections of winds and temperature or thickness between isobaric surfaces are found more useful to trace the movement of the systems. Isotherms drawn using thermal winds, which have a greater amplitude than the streamlines themselves, are also useful to locate the troughs.

9.7.2 These easterly troughs are most marked between 400 mb and 200 mb and have a wave length of 20 longitude and a speed of 10 to 15 kts. If the winds are strong, troughs do not extend to very high levels. Coldest air is near the trough–line with relatively warmer air ahead as well as in the rear. Srinivasan's estimate of ten troughs per month passing over Calcutta seems to be rather large, their number decreasing to north and south. According to his analysis over the year 1955 to 1957, 81 % of the rainy days over Gangetic West Bengal were one day before or on the day of passage of waves or troughs over Calcutta. Convergence in the lower levels is due to the seasonal sea level monsoon trough near about. In lower troposphere, synoptic patterns known to be favourable for development of weather occurred in these cases, but Srinivasan seems to lay more stress on the upper tropospheric troughs as the main mechanism. Thunderstorms coincided with the transit of the coldest part of the troughs. The time section of Calcutta in Fig.9.18 illustrates these features. Movement of a trough across the country is shown in Fig. 9.19 (a & b). General synoptic experience is that troughs in upper easterlies are rather weak systems and cannot be objectively followed, to be more useful in forecasting than lower tropospheric patterns.

9.7.3 Krishnamurthy (1971) has studied by spectral analysis percent variances as a function of wave numbers of 200 mb zonal (U) and meridional winds (V) and stream function for June, July and August of 1967 on a global scale in tropics. From 20^{o} N to 20^{o} S, most of the variance of zonal winds is in wave number 1. At 30^{o} N, wave number 2 has a larger variance than wave number 1. The spectra of meridional components are different and show that the variance is spread in the first ten wave numbers, suggesting significant disturbances in all scales larger than about 45° longitude.

He has studied the transient U waves as deviations from the mean for a 4–day period of several latitude circles. This was adopted as deviations from a long term mean were not meaningful. Motion of transient wave number 1 is generally eastward between 10^{o} N and 10^{o} S, although there are cases propagating westward. The motion of the transient wave number 2 was generally westward, although a few exceptions were noticed. Speeds are in the range of 120° to 45° longitude day^{-1}. The amplitude of wave number 2 is about one–third that of wave number 1. Such transient ultra long waves are an intermittent phenomenon, as a well defined wave for about 3 or 4 days is followed by a period with irregular features.

Wave numbers 7 and 8 of V were further investigated as their variance showed a peak in the northern hemisphere tropics. Amplitude of these waves is upto 12 kt which appears to fluctuate with a 3–5 day period. Westward propagation is intermittent, fast movement for short periods being 10–12° longitude day^{-1}. Large amplitude and rate of westward movement appear to be correlated. Large coherence of phase from day to day and that individual waves could be followed sometimes for 25 days lends weight to the reality of these waves. Such analysis is independent from one day to another.

A schematic drawing of a S W– N E tilted quasi–stationary long wave and S E– N W tilted short wave at 200 mb postulated by Krishnamurthy is shown in Fig. 9.20.

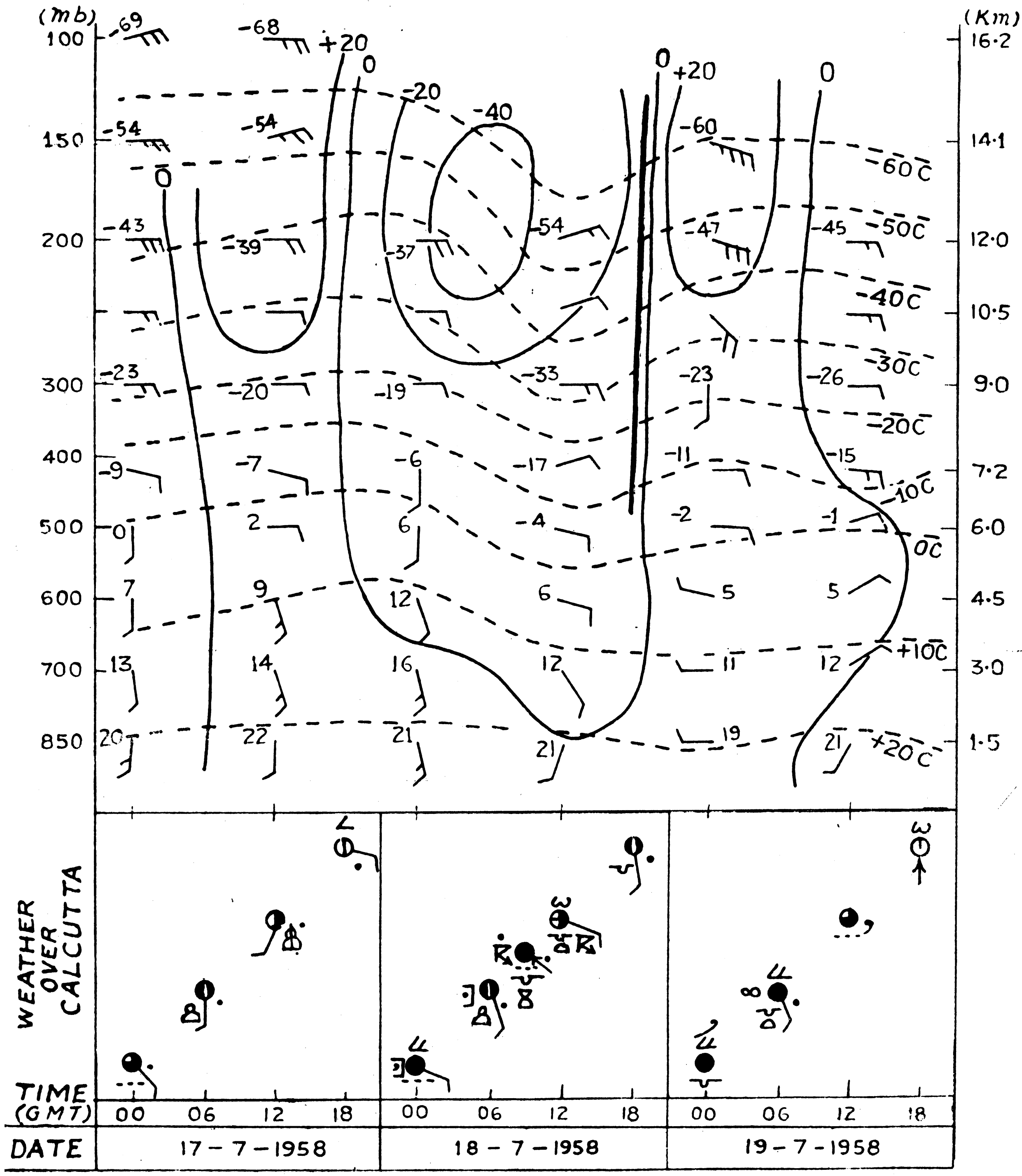
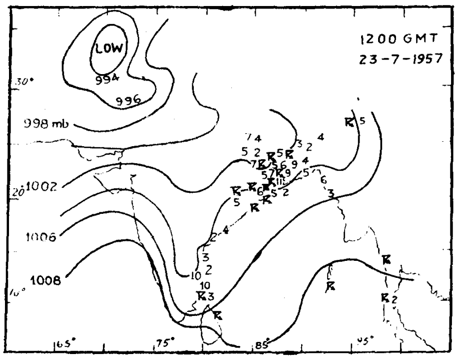
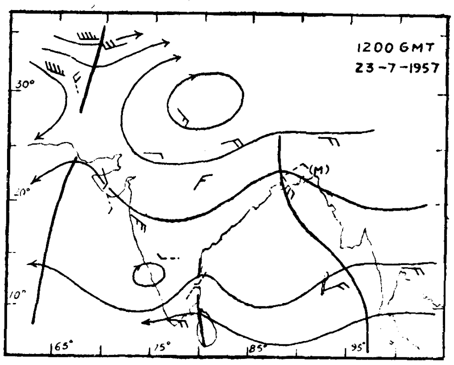


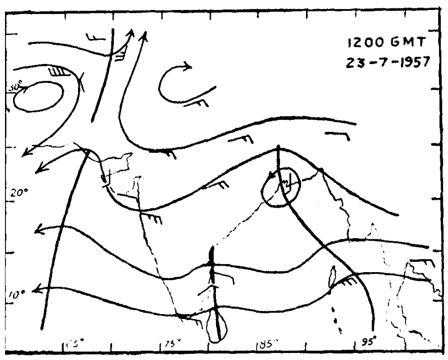
Fig.9.18 - Vertical time section for Calcutta 17-19 July 1958.
(Isopleths of 24 hr. height changes (in gpm) are shown by thin continuous lines and isotherms by broken lines. Thick continuous line shows the position of the upper trough. Temperatures (°C) at standard isobaric surfaces are also plotted).



(i) Sealevel isobars at 1200 GMT on 23 Jul.1957 and rainfall amounts (more than 1.5 cm) recorded at 0300 GMT on 24 Jul. 57 at stations in NE India, Bangladesh and east coast of Indian Peninsula. Thunderstorms wherever reported are shown.

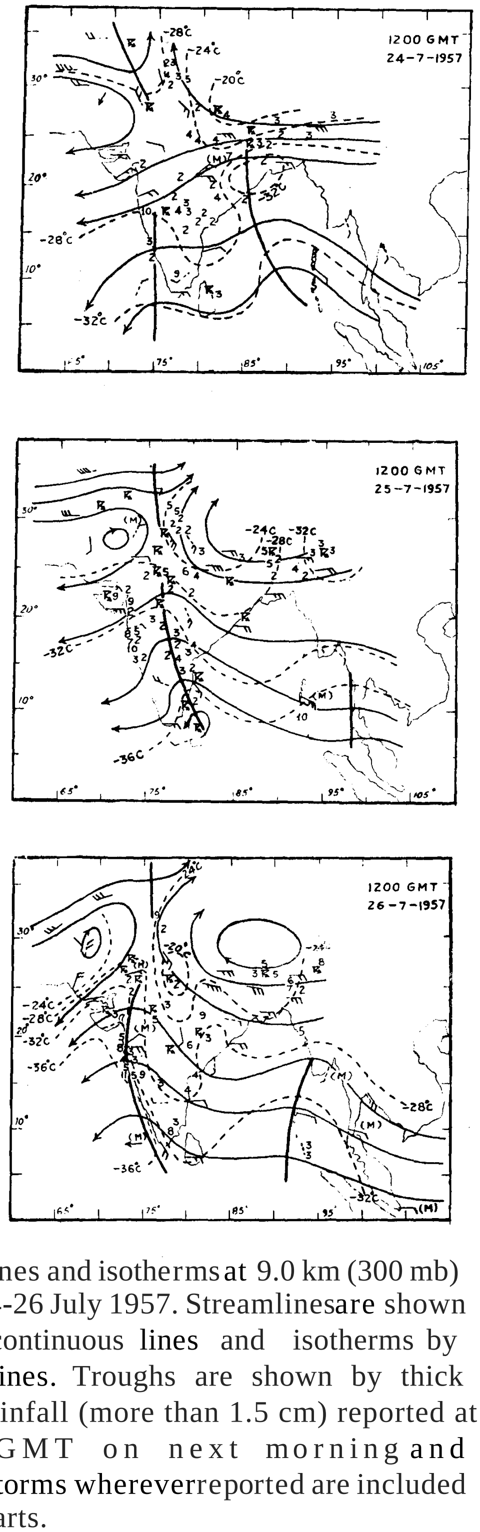


(ii) Stream lines at 9.0 km. (300 mb) Broken barbs indicate next lower level winds M - morning (00 GMT) data; continuous thick lines indicate the position of the troughs.



(iii) Stream lines of thermal winds 9.0 ~ 6.0 km (300 ~500 mb). Thermal winds (9.0 ~ 6.0 km) are also plotted. Trough lines are shown by thick continuous lines.

Fig. 9.19a



Streamlines and isotherms at 9.0 km (300 mb) level - 24-26 July 1957. Streamlines are shown by thin continuous lines and isotherms by broken lines. Troughs are shown by thick lines. Rainfall (more than 1.5 cm) reported at 0300 GMT on next morning and thunderstorms wherever reported are included in the charts.

Fig. 9.19b

9.8 Other Systems

9.8.l George (1956) adduces evidence of the formation of off–shore vortices along the west coast of India which cause concentrated heavy rainfall over the coast away from the Ghats. They have a diameter of 30–150 km (larger than Cb cells) and vertical extent of 0.3 to 1.5 km. The vortices move south to north along the coast, normal speed being 15 mph, though considerable variation is likely. Vortices are identified from the variations of directions of light winds at coastal stations. Nothing is known about the life period of such vortices and it is not possible to be sure that the features on the basis of which they are being identified are not circulations around cumulus cells. An illustration of the vortices is shown in Fig. 9.21. When there is a weak trough off the coast, there would be a tendency to place a number of short–lived vortices in the system.

Das Gupta (1967) reports mesoscale low pressure systems over northeast India influencing rainfall which can be sometimes located with the present synoptic network of surface stations.

9.8.2 Strong pressure gradient along west coast south of 20^o N, causes increase of monsoon rains, which may be as squally showers. Jayaram (1965) has brought out that pressure gradient of about 4 to 8 mb between 15^o and 20^o N is necessary for heavy rain in the Bombay belt. This combined with a trough off Konkan Coast, or a low near about Saurashtra or a depression forming in Bay or moving across west Madhya Pradesh, is more effective than pressure gradient alone. He has also brought out that a cyclonic shear of 20 kt or more per 500 km at 700 mb is very favourable for heavy rains near about Bombay.

Sarma (1969) studied the variation of heavy rainfall (>=5 cm) at Bombay between Santa Cruz and Colaba (29 km apart) in July to September. On days of large variation in rainfall, westerly winds are of the order of 30–40 kt and the depth 3 km or more, On days of little variation of rainfall, the speed is less than 20 kt and depth 1.5 km.

9.8.3 Rao and Hariharan (1959) report a case of widespread thunderstorm activity over the Peninsula which could be explained on the basis of convergence easily inferred from the wind flow in the lower troposphere as seen in Fig. 9.22 (a, b & c). In some areas, convergence was more than 0.l hr^{-1} and even 0.2 hr^{-1} .

9.8.4 The same authors (1957) reported thunderstorm activity over the Peninsula on 14 June 1950 (Fig. 9.23 a) during weak monsoon, with upper winds upto 500 mb being unusually northerly. This was associated with in–phase superposition a westward moving lower latitude trough and mid–tropospheric eastward moving trough in westerlies further north. Fig. 9.23(b) shows their location at 0200 GMT on 12th and 13th.

9.8.5 In the southern hemisphere trough, well defined vortices reaching the intensity of deep depression form and travel westwards, according to Srinivasan (1968). Frequently when such circulations come rather near the equator, they affect both the hemispheres, with clouding extending well into the north hemisphere. The rainfall of 6 cm on 31st August and 9 cm on 1st September 1964 at Minicoy are regarded as due to such a system at 4° S, 68° E (Fig. 9.24).

9.8.6 George and Narayanan (1973) have presented (Fig. 9.25) the variations in stratospheric and mesospheric winds in three monsoon seasons and the fluctuations in rainfall over Kerala and Lakshadweep. They are of the view that appearance of very strong easterlies over Trivendrum in the upper stratosphere and lower mesosphere is in association with vigorous monsoon over Kerala and south Peninsula. During weakening of easterlies or appearance of westerlies at those levels, monsoon activity decreases over the Peninsula, leading even to 'break monsoon' in the country.

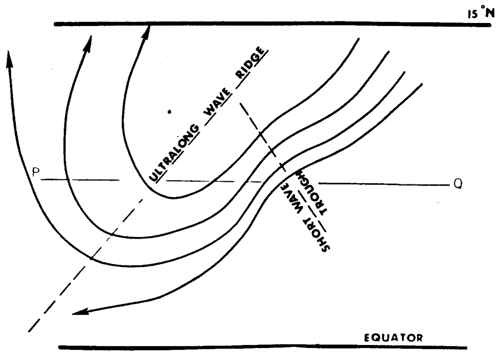


Fig.9.20 A schematic drawing of tilted quasi-stationary ultra long waves that contribute to poleward flux (across line PQ) of momentum and equatorward flux of kinetic energy; and a tilted short wave that contributes to an equatorward flux of both momentum and kinetic energy (Krishnamurti, 1971).

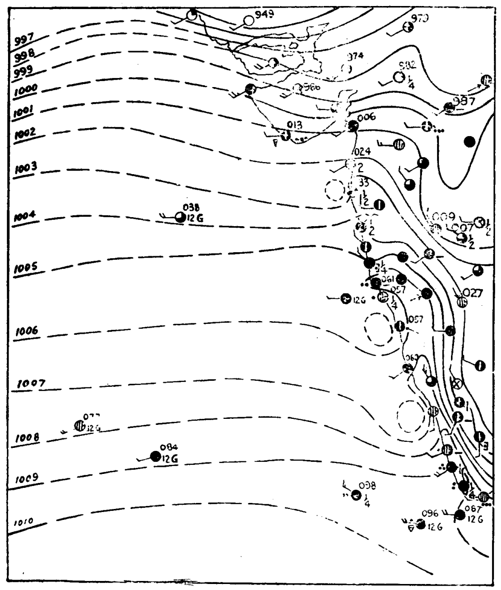


Fig. 9.21 - 12 July 1951 (1730 IST)

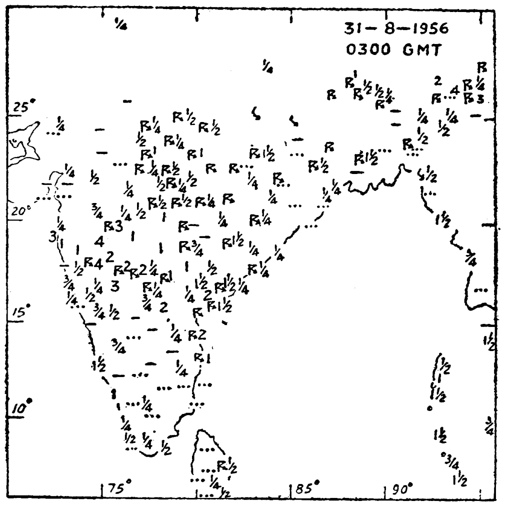
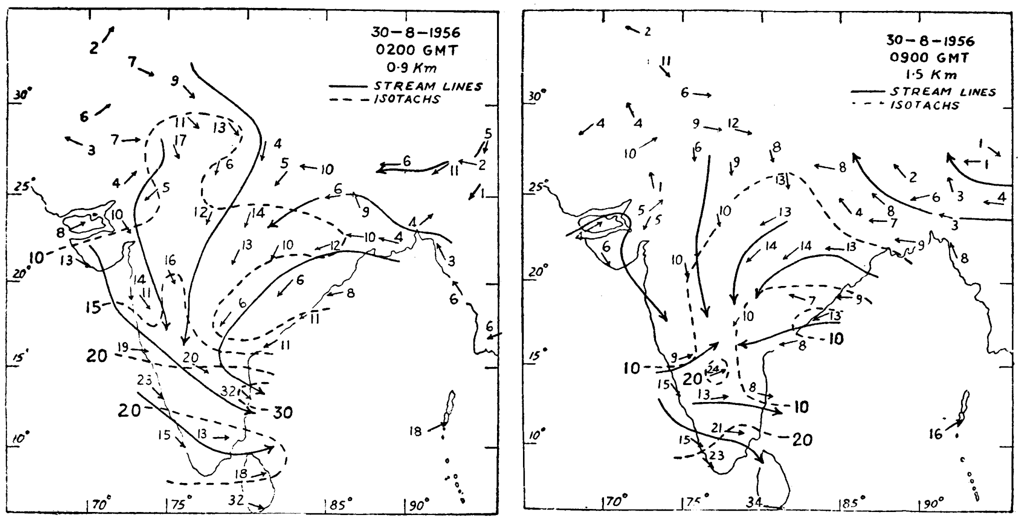
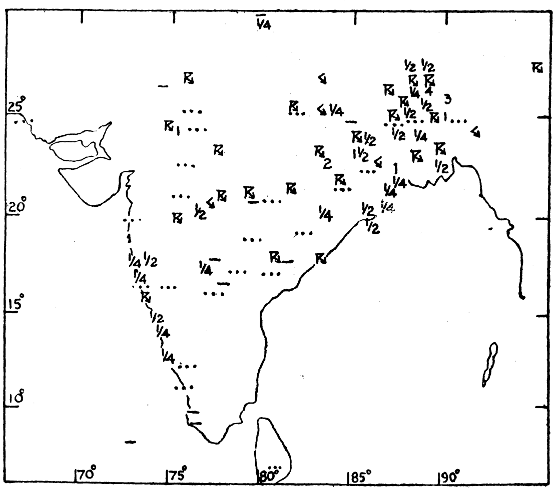


Fig.9.22 (a) Rainfall and thunderstorms during the 24 hrs. ending at 0300 GMT on 31 August 1956.

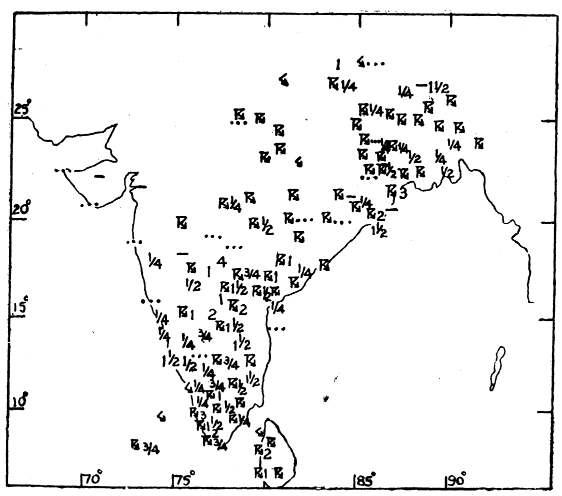


(b) (c)

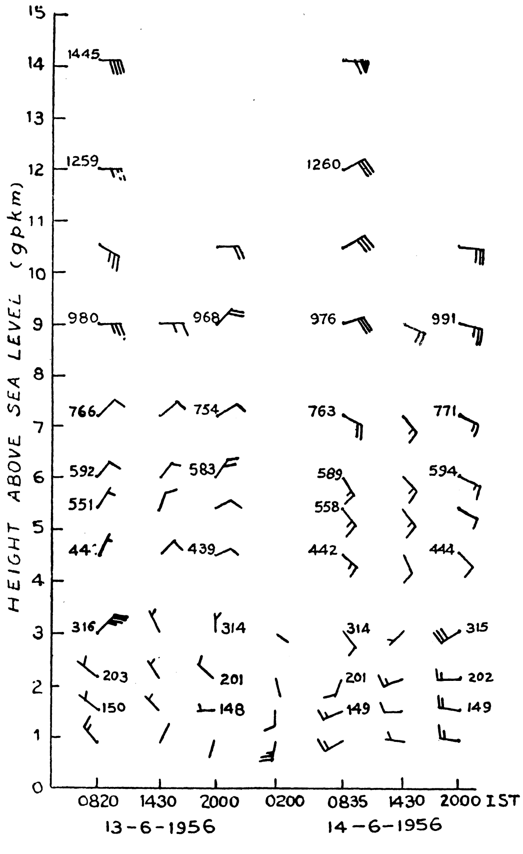
Winds, stream lines and isotachs.



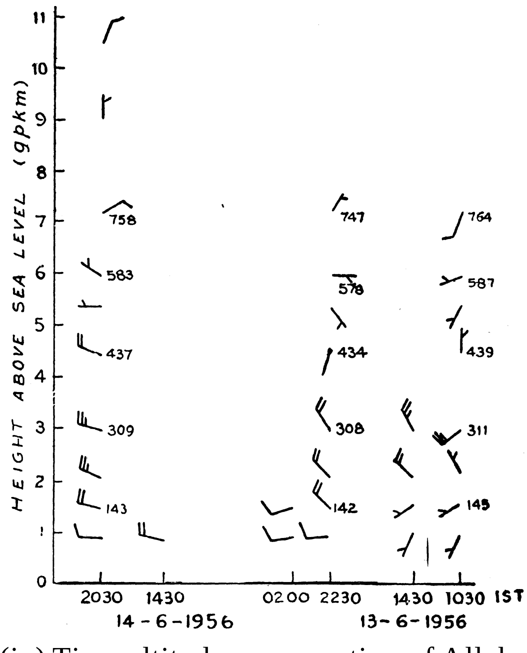
(i) Past weather and rainfall - 13 June 1956.



(ii) Past weather and rainfall -14 June 1956.



(iii) Time-altitude cross section of Madras.



(iv) Time altitude cross section of Allahabad

(Altitudes in tens of geopotential metres)

Fig.9.23(a)

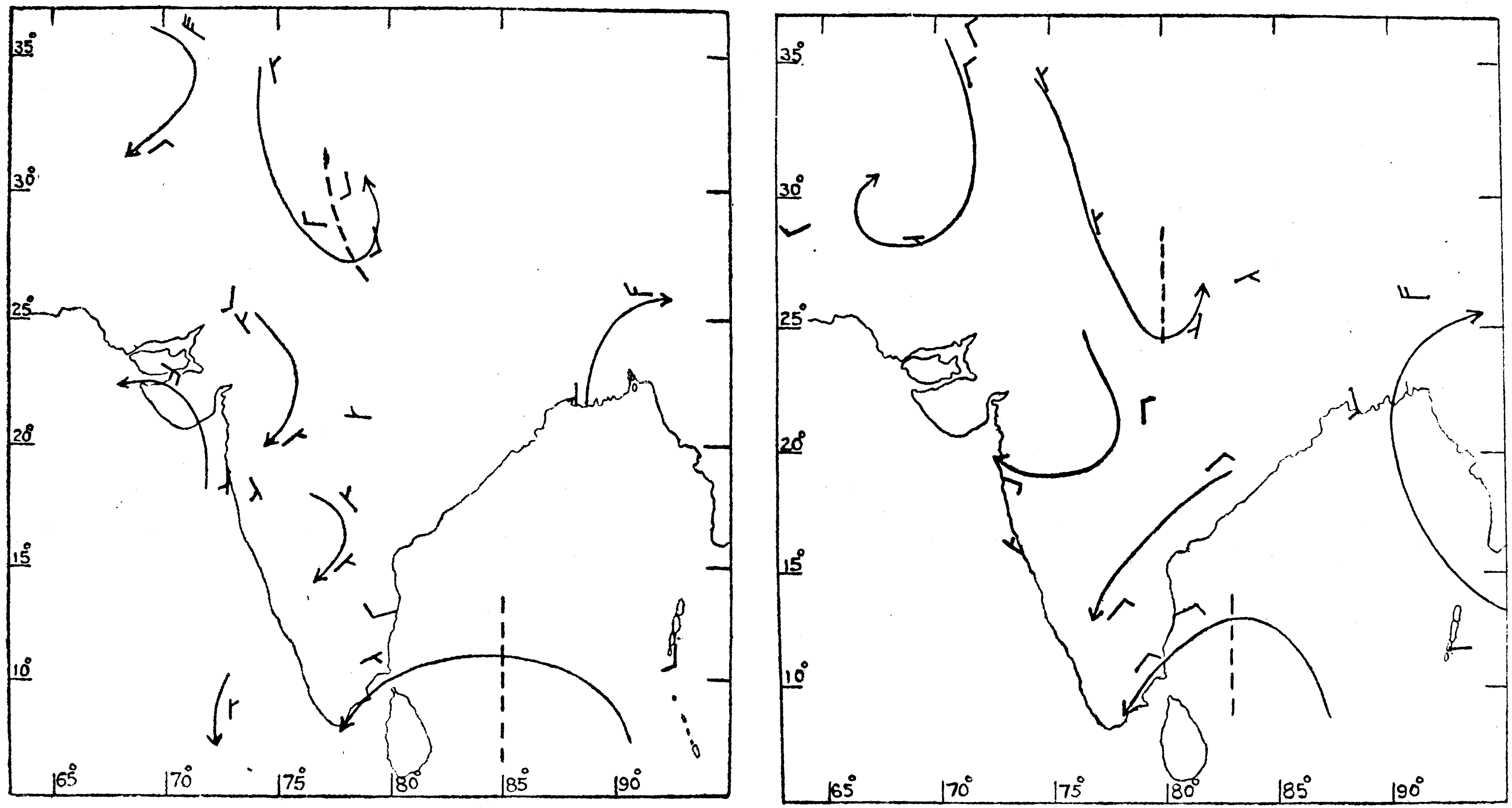
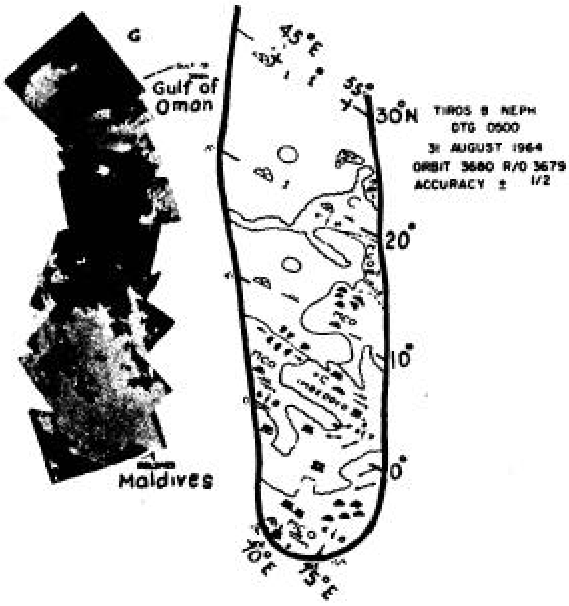


Fig.9.23(b) Upper winds at 5.4 km. (i) 0200 GMT 12 June 1950 (ii) 02 GMT 13 June 1950. Altitudes in tens of geopotential metres _ _ _ _ _ Trough line.



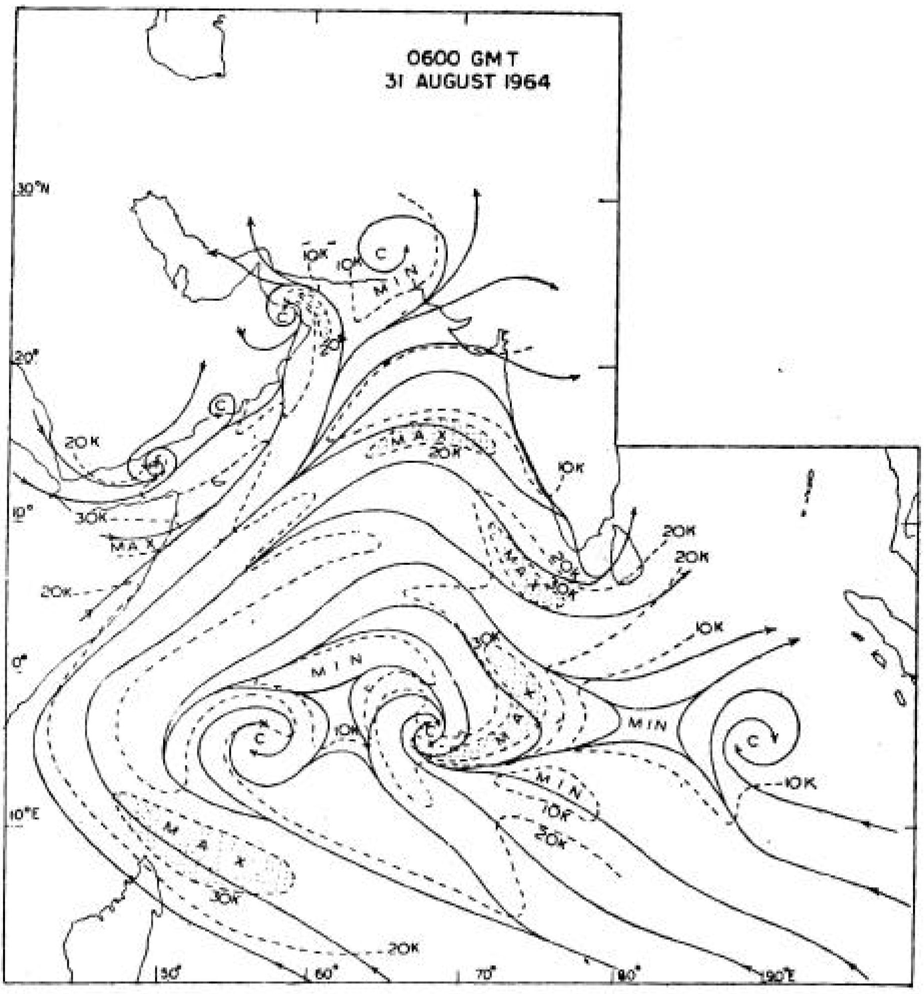


Fig 9.24 Streamlines analysis of surface

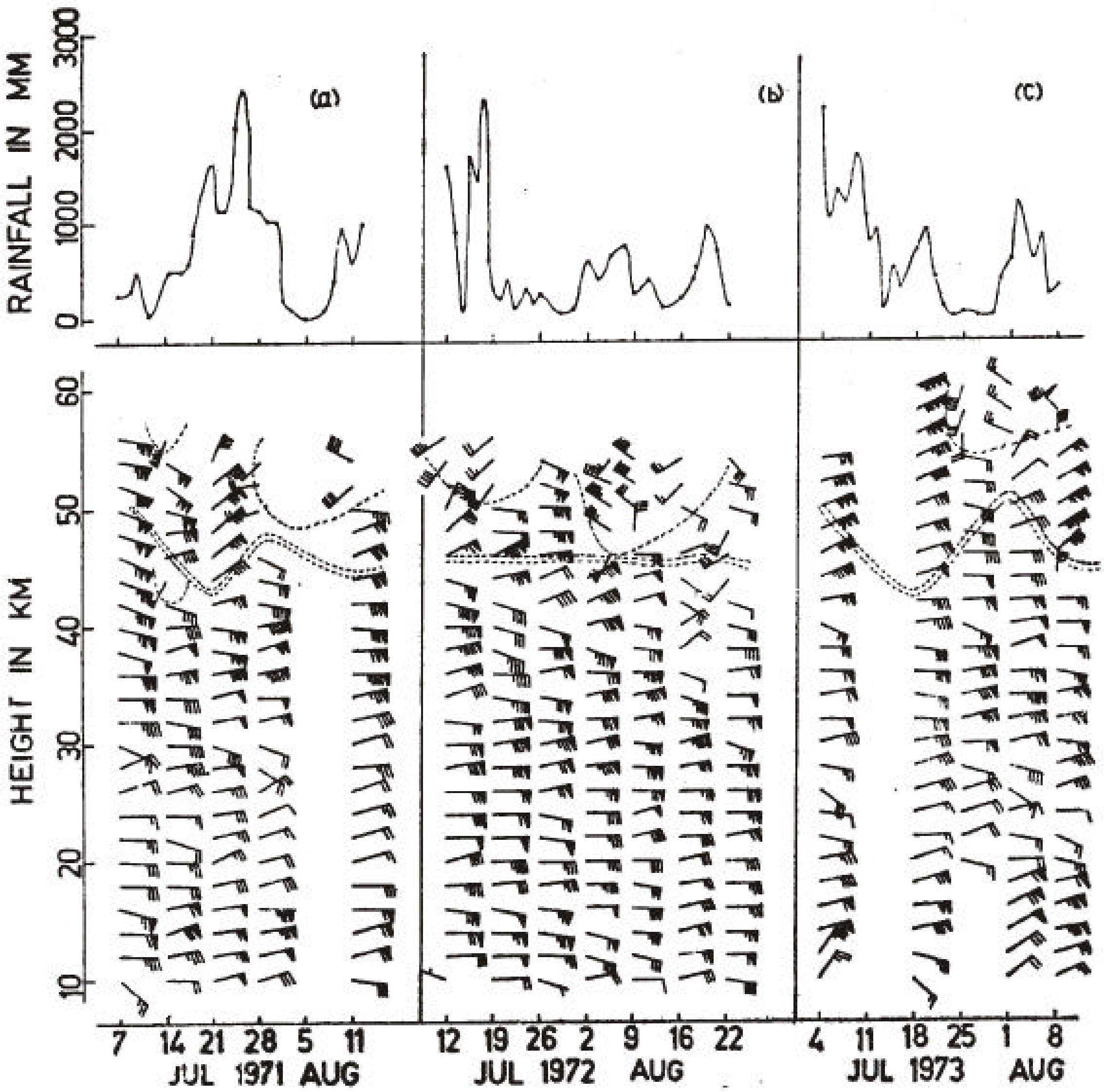


Fig 9.25 Vertical time section of upper wind over thumba during 'break monsoon' and rainfall Kerala and Lakshadweep.

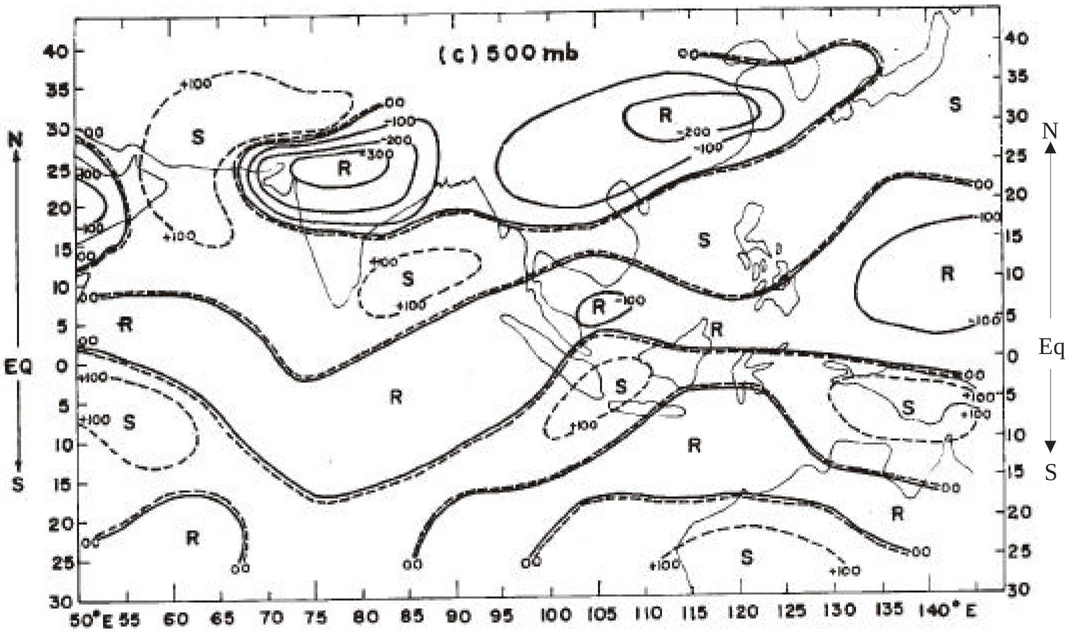


Fig:.9.26 (a) Distribution of vertical velocity in mb per day at 500 mb level. Negative values (heavy continuous 1ines) indicate rising (R) and positive values (heavy dashed lines) sinking (S) air. (Saha,1968).

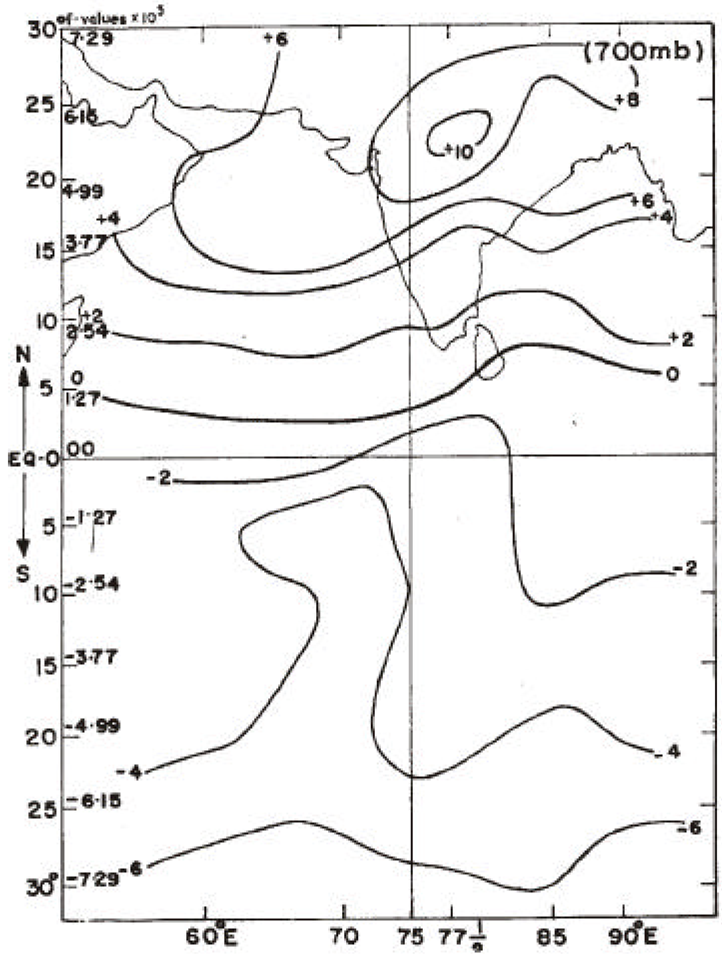


Fig.9.26 (b) Distribution of absolute vorticity at 700 mb level over the Indian Ocean and neighbouring continental areas at 12 GMT 7 July 1963. Unit: $10^{-5} sec^{-1}$.(Saha, 1968).

9.8.7 Computations of vertical velocities for individual days by kinematic methods have brought out the features associated with prevailing synoptic systems and the limitations of such methods due to sparse network of upper wind stations and errors in wind measurements. Krishnamurthy's (1966) evaluation for 2 to 4 August 1961 using a balanced model, showed ascent in the northeastern parts of the monsoon trough upto a maximum of 1 cm sec^{-1} and subsidence to the southwest. There were considerable variations in the distribution of vertical velocities during the three days, associated with synoptic scale perturbations. Saha's (1968) computation by simple kinematic techniques, for 7th July 1963, gave vertical velocities still increasing in magnitude at 100 mb. His charts of vertical velocity at 500 mb and vorticity at 700 mb are given in Figs. 9.26 (a and b).

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CHAPTER 10

OROGRAPHIC EFFECTS

10.1 Orography around India has influenced without doubt the way the monsoon circulation and rains occur. The northern part of the Indian sub–continent is bounded by the Himalayas in the north, the hill ranges from Arakan Yoma to the Patkai hills in the east (near 95 E) and the ranges in the west from the Hindukush mountains to the Kirthar range Further, the Tibetan Plateau lies to the north of the Himalayas at an elevation between 4000 and 5000 metres. Actually, a chain of ranges higher than 2000 m runs between Latitudes 30° and 40° N from about 40° E to 100° E. What would have been the circulation over the Indian sub–continent if the east–west mountain ranges had not been in existence, is difficult to visualise with any certainty. This will have to wait until three dimensional numerical modelling of general circulation has developed to sufficient accuracy At the moment, we can only speculate on general considerations and analogy of other areas.

10.2 Meridional movement of air and exchange of heat etc. in the lower troposphere is practically blocked across the latitudinal belt 30 –40° N by the mountain ranges. This has increased the heating to the south, particularly in areas out of reach of the maritime air mass and perhaps contributed to the– intensity of the heat low The heat low itself would not have disappeared as this occurs over all continents and right up to Siberia, though the position of the low could have been different.

10.3 Along any extensive barrier the most favoured flow would be parallel to the boundary This tendency is noticed in the mean flow pattern as southerlies skirting the eastern hills and as easterlies along the Himalayas. If development of monsoon or equatorial westerlies to south is a character of global circulation, though in this case it extends to a higher latitude, the formation of the monsoon trough between the westerlies to south and easterlies along the Himalayas seems logical. Banerji (1931) regarded the monsoon trough as a dynamical development rather than thermodynamical, which is the case of the heat low. Even during breaks, the westerly flow is roughly along the Himalayan range. While the latitudinal position of the monsoon trough is dictated by other considerations, easterly or westerly flow along the Himalayas is the most frequent configuration.

10.4 Northern part of the sub–continent has been compared to a box, with the hill ranges forming the three sides, open on one side, to south. Air entering this box from the south has to rise above the bounding .walls to get out of the area. No computations have been made of the vertical motion induced by this model, However, the amount of vertical motion, as manifested by rainfall, is much less when the mean pattern alone prevails than when other perturbations of transitory nature are super imposed.

10.5 The Tibetan Plateau which has an average elevation of 5.3 km is a source for heating the upper layers. Numerical modelling has confirmed that this heating increases the strength of the upper tropospheric easterlies. In such two–dimensional models the heat advected away by the easterly flow is perhaps not taken into account. Still, the strengthening of the upper tropospheric easterlies on account of the heat ing of these layers by the mechanism of the Tibetan Plateau is not in doubt. What is the effect of this stronger upper tropospheric easterlies on the development of lower tropospheric westerlies? Superposition of this upper tropospheric pressure gradient, should weaken pressure gradient required for the lower westerlies. Nevertheless, considerations of meridional circulation and angular momentum balance may possibly be inducing stronger westerlies below.

10.6 A relatively not–so–complicated a problem is the effect of hill ranges on the prevailing circulation pattern. No observations exist, of the modification in three dimensions of pressure gradient or winds near mountain ranges Rainfall shows great variations from the windward side to the summit and leeward side, that should have been due to corresponding vertical motion induced by changes in horizontal flow.

Rainfall augmentation on the windward side seems to extend to considerable distance in the plains. Flow across a range involves at least ascent of some layers of air through the height of the hill range. At some higher level, this ascent should become zero. The vertical velocity at the boundary of the hill slope should steadily decrease aloft or first increase and then fall to zero. Occasions of heavier rains are apparently due to the latter type of distribution, The depth of the atmosphere in which induced ascent occurs also varies. These variations would depend upon wind direction, speed, wind shear and stability of air. Lee flow has been studied in great detail than the effects on the windward side. Lee flow can also break down into turbulence under some conditions. From rainfall pattern, it has come to be recognised that a synoptic pattern favourable for ascent of air should exist for enhancement by orographic lifting.

The long range of Western Ghats, 1 to 2 km high, roughly running north to south, barring the fairly steady westerly flow of the monsoon is an ideal situation to study the problem. Horizontal and vertical shears and speeds vary very much but lapse rate is near moist adiabat. In a few cases, weak easterlies or southerlies may occur near the ground.

10.7 Sarker (1967) investigated theoretically mountain waves over the Western Ghats by a two dimensional model. Neglecting effects of earth's rotation and friction, in a steady flow, with undisturbed flow parameters being a function of height (z) only and disturbed quantities being small, the perturbation equation is derived as

$$\frac{\partial^2 W(Z, r_2)}{\partial z^2} + \left[f(x) - \hbar^2\right] W(z,\hbar) = 0$$

where

$$\text{f(Z)} = \frac{\text{g}(g^* - g)}{\text{U}^2 T} - \frac{1}{U}\frac{d^2U}{dZ^2} + \left(\frac{g^* - g}{T} - \frac{g}{cRT}\right)\frac{1}{U}\frac{dU}{dZ} - \frac{2}{cRT}\left(\frac{dU}{dZ}\right)^2 - \left(\frac{g - Rg}{2RT}\right)^2$$

The vertical velocity (perturbation) is given by –

$$w(x,z) = \text{ Real part of } \int_0^\infty W(Z,\hbar) e^{i\hbar z} \exp\left(\frac{g - Rg}{2RT} z\right) d\hbar$$

$$\cong \text{ Real part of } \left(\frac{r_0}{r}\right)^{\frac{1}{2}} \int_0^\infty W(Z,\hbar) e^{i\hbar z} d\hbar$$

U, T and r are undisturbed westerly wind, temperature and density, g^* dry or moist adiabatic lapse rate, g actual lapse rate and $k = g/(g - Rg^*)$ and $\hbar$ is the wave number of the ground profile.

Vertical profile of f(z) determines the vertical velocity distribution. The lower boundary conditions is that the flow is tangential to the surface. f(x)becomes indeterminate above 8 km due to easterly flow setting in. For convenience, f(z) is set at a constant value above 8 km, the choice of this constant having only a small effect on lower levels. Numerical integration was carried out using intervals of 0.25 km for height.

Computations were made of vertical velocities and rainfall assuming moist adiabatic lapse rate, in five July cases. Sinusoidal lee wave of only one frequency developed in every case, wave length being in the range of 19 to 32 km. The pattern of vertical velocity may be illustrated by the case of 5 July 1961. The vertical wind profile is shown in Fig. 10.1. This gave a lee wave of 19 km wavelength. Fig. 10.2 shows the vertical velocity distribution. Even at a distance of 60 km upwind from the crest, orographic ascent commences and vertical velocities increase upto 15 km from the crest. They decrease thereafter, upto 5 km on the lee-side (i.e. downwind). Compared to 1 km height of the range, on the windward side, vertical velocity increased upto 1.5 km, decreased aloft to zero at about 5 km and descent occurred higher up. Between 5 and 15 km on the lee-side from the crest, there is descent generally upto a height of 4 km, above which is again ascent. This is due to the lee wave of length 19 km. Thereafter, there is a zone of descent of air at all levels and subsequently positive and negative vertical velocities occur alternately.

There is fairly good agreement between computed and observed rainfall amounts as shown in Fig;. 10.3. Sarker observes that the rainfall on the windward side originates from levels below about 5 km but on the lee-side from above that level. He has not applied his model to weak monsoon conditions, though by an earlier model he studied a case of 2 mm hr^{-1} rain at Bombay as an example.

Rai (1958) suggested that when wind speed in lowest 2 or 3 km decreased with height, air flowing across Western Ghats would be subject to considerable turbulence, as distinguished from lee waves, and this would be more favourable for rainfall at Poona.

10.8 As mentioned earlier, the effect of orography on rainfall is strikingly brought out in the monsoon season. Hills and mountain ranges get more rainfall than the neighbouring plains. But variations from day to day, with elevation and in different areas bring out the complexity of this problem.

10.9 In Khasi hills of Meghalaya, is located the rainiest place Cherrapunji, with annual rainfall of 1142 cm.,* of which 837 cm falls from June to September. The highest rainfall in a day at this station is 104 cm. In 1876 the annual rainfall was 2299 cm, the highest so far. The Garo-Khasi hills are 300 km in length and 70 km in width, running east to west. The highest peaks are below 2 km while Cherrapunji is 1313 m, high, on the southern slopes. The slope in this area is about 1:8,starting from the plains about 10 km to south. The mean air flow is from south upto 3 km, becoming easterly aloft. The Baral range to the east prevents air flow around. Rainfall in the monsoon season in plains, at Mymensingh 50 km to south of the land-rise (but to the west of Cherrapunji longitude) is 160 cm. Shillong (1500 m) just to the north of the crest of the hill range, 35 km north of Cherrapunji, has a rainfall of only 159 cm, Tura (370 m) on the western slope of the Garo hills (of peak of 1400 m) has 247 cm of rain, more than Shillong.

Mawphlang (25°27' N, 91°46' E, height > 1500 m) near the crest but still on the windward side in relation to southerly flow, gets 249 cm. Jowai (1390 m) just below the crest on the leeward side has 256 cm of rain. On the other hand, Mawsynram (14017m) which has averages for about a decade, shows rainfall of 918 cm in the season, more than Cherrapunji. Mawsynram is located in the same latitude but a little to west, on a range on the other side of a rivulet. The various rainfall reports given above and observations of meteorologists who have visited the area during rains, suggest that the phenomenal rains at Cherrapunji prevail over a short length of the range, up a small portion of the slope. The funnel shaped catchment opening to south on either side of which Cherrapunji and Mawsynram are located seems to increase the convergence in air rushing from south. Apart from this, it appears that it is on the central slope of

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* Rainfall figures for observatory stations are taken from "Climatological Tables of Observatories in India (1931-1960)" and those for state raingauge stations from I. Met. D. Memoirs Vol. XXXI Part III "Monthly and annual normals of rainfall and rainy days".

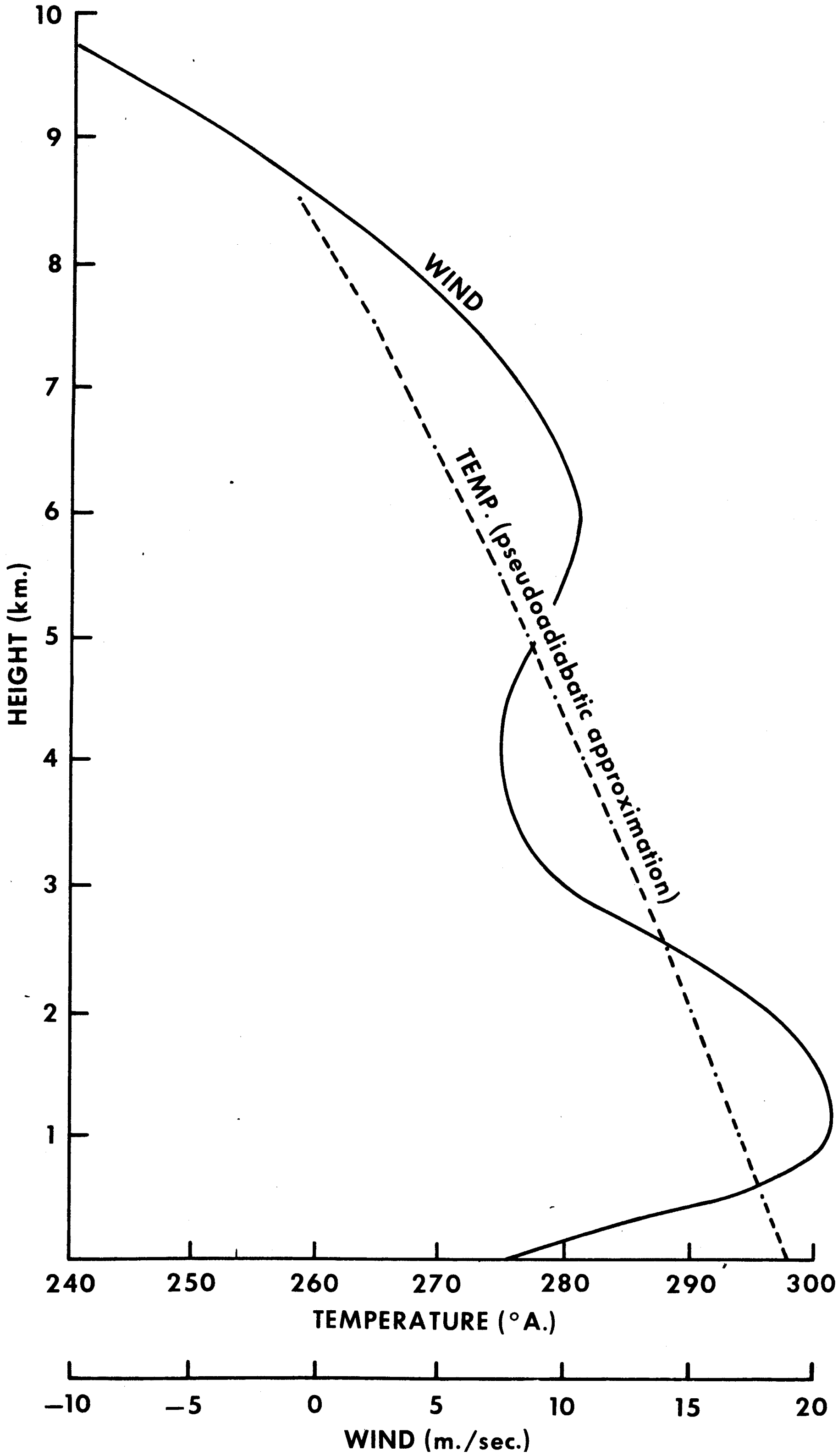

Fig. 10.1 Average wind and temperature profiles for July 5, 1961. Positive wind is westerly, negative is easterly. (Sarker, 1966).

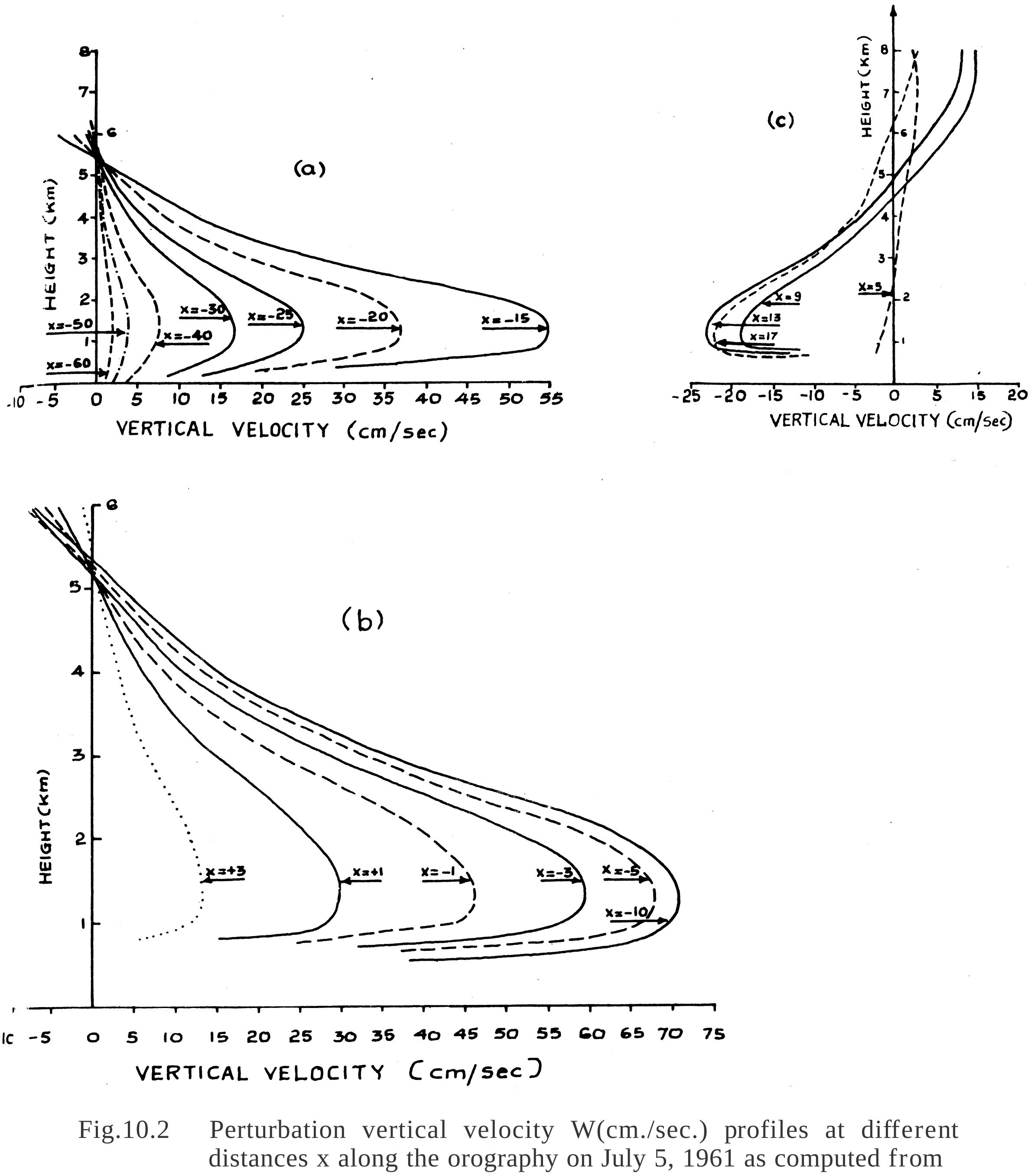

Fig.10.2 Perturbation vertical velocity W(cm./sec.) profiles at different distances x along the orography on July 5, 1961 as computed from the present modified model, x = 5 is the erest of the mountain and x = -60 is the coastal position

(a) Profile for -60 km ≤ x ≤ -15 km.

(b) Profiles for -10 km . ≤ x ≤ : 3 km.

(c) Profiles for 5 km ≤ x ≤ 17 km. (Sarker, 1967).

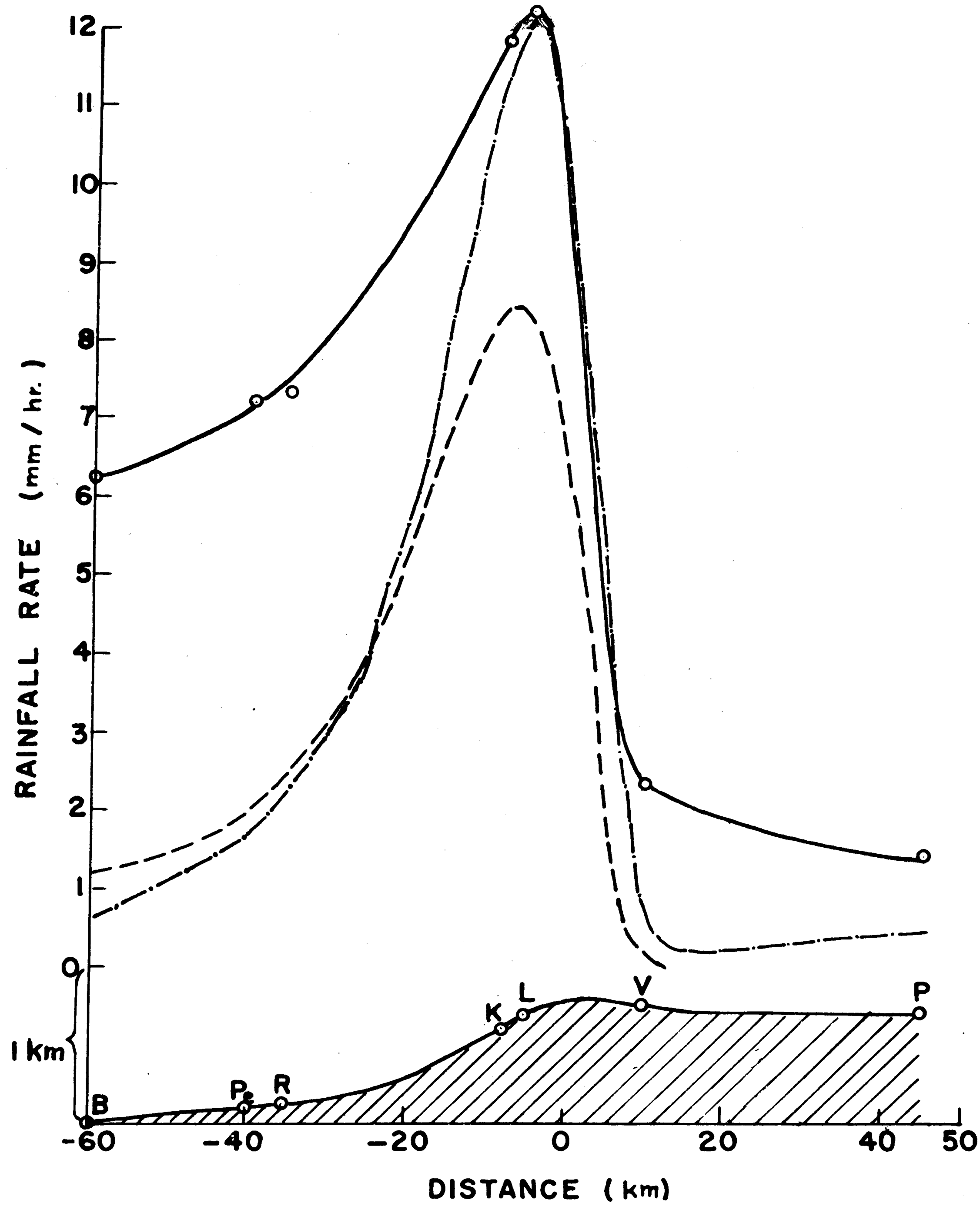

Fig.10.3 Observed (upper solid curve), computed orographic from the approximate model, (dashed curve) and computed orographic from the present modified model (dashed-dotted curve) rainfall distribution for July 5, 1961 along the orographic profile (shaded) from the coast at Bombay (B) inland through Pen (Pe), Roha (R). Khandala (K), Lonavla (L), Vadgaon (V), and Poona (P). (Sarker, 1967).

the hill ranges that the forced uplift and rainfall are maximum, Towards the crest, this is much less effective. Here the decrease of rainfall on the leeward side is not so drastic as in some other cases. However, Lanka (25°55' N, 92°58' E) in a bowl to the northeast, records only 79 cm. This bowl of elevation less than 150 m is surrounded by hill ranges on all sides except for an opening 20 km wide to northwest.

Assuming ascent of surface air upto 5000', ascent decreasing aloft to zero by 20,000', Murray (1948) computes "that the rainfall per hour is 0.6V / D inches. where V is the mean wind component in mph up the slope and D is the horizontal projection of the slope; D taken as 30 miles, V = 10 mph, is expected to give 4.8" in a day". Comparing with actuals, he concludes that some extra rain falls owing to instability.

10.10 The general crest of the Western Ghats is above 1 km, except being somewhat lower between 14^o and 16^o N and almost reaching 2 km south of 12° N. The crest is within 65–100 km of the coast. On the leeward side, to the north of 14° N, after the first sharp drop from the crest, the land, slopes gently eastward.

The seasonal rainfall of some stations in the Western Ghats is given in Table 10.1.

Table 10.1
Normal Rainfall in Southwest Monsoon period in the Western Ghats

Station	**Latitude (^{0} N)**	**Longitude (^{o} E)**	**Height (m)**	**Rainfall (cm)**	**Remarks**
Waghai	20 46	73 30	335	212	
Igatpuri	19 43	73 35	607	316	
Matheran	18 59	73 17	756	493	
Khandala	18 46	73 22	539	449	
Amboli (Ratnagiri)	18 46		720	707	(16 years' average)
Mandangad	17 59	73 15	610	366	
Mahabaleshwar	17 56	73 40	1382	593	
Gaganbavda	16 33	73 50	690	583	
Radhanagari	16 20	73 59	610	357	
Saklespur	12 57	75 47	895	188	
Tirthahalli	13 41	75 14	740	265	
Koppa (Balgadi)	13 33	75 21	806	260	
Sringeri	13 25	*75* 15	743	328	
Bhagamandala	12 23	75 31	876	516	
Peermade	9 34	76 59	(1200–1389)	377	
Neriamangalam	10 3	76 47	600	383	
Anamalai	10 35	76 55	1372	315	
Davala	11 28	76 24	1067	329	
Agumbe	13 32	75 6	659	718	

These have been selected for the highest rainfall in each district. The rainfall at coastal stations along this length are also given in Table 10.2

Table 10.2
Normal Rainfall in the Southwest Monsoon period in the plains of West Coast

Station	**Latitude** (^{O}	**' N)**	**Longitude** (^{O}	**' E)**	**Height (m)**	**Rainfall (cm)**
Surat	21	12	72	50	12	113
Dharampur	20	33	73	11	38	232
Dahanu	19	58	72	43	5	250
Bombay – Colaba	18	54	72	49	11	197
Bombay – Santacruz	19	7	72	51	15	256
Shahapur	19	27	73	20	79	245
Alibag	18	38	72	53	7	198
Harnai	17	49	73	6	220	262
Ratnagiri	16	59	73	20	35	279
Devgarh	16	22	73	22	36	237
Vengurla	15	52	73	38	9	295
Sawantwadi	15	54	73	49	96	345
Marmugoa	15	25	73	47	62	232
Bhatkal	13	59	74	33		349
Honavar	14	17	74	27	26	320
Karkal	13	13	75	0		403
Mangalore	12	52	74	51	22	289
Kozhikode	16	57	82	14	5	235
Alleppey	9	33	76	25	4	189
Trivandrum	8	29	76	57	64	83

We should try to elucidate the effect of orography on the variations of coastal rainfall and that at the available hill stations. Rainfall at both types of stations (coastal and hill stations) would depend on the synoptic patterns affecting the various parts which are known not to be the same. Apart from this, rainfall seems to depend mainly on the ground contour in the vicinity, so that any interpretation in terms of broad orographic pattern has to be attempted with the greatest care, Colaba and Santacruz differ in rainfall amounts by 25 per cent though within 30 km. A hillock of 300 m within 5 km of Santacruz apparently accounts for this increase. The profiles of the normal rainfall at stations along the West Coast as presented by Ramakrishnan and Gopinath Rao (1958) are given in Fig. 10.4. They show a maximum at 14^{o} N and a secondary maximum at 18^{o} N George (1962) finds the average rainfall in one–degree squares in the coastal belt has a maximum in all months between 13^{o} and 14^{o} N (Fig. 10.5). The secondary peak at 18^{o} N reported by Ramakrishnan and Gopinath Rao (1958) is due to the inclusion of Dapoli which has been omitted by George, as it is not very close to coast.

Raghavan (1964) found the relationship shown in Fig. 10.6 between coastal rainfall and distance of 150 m contour from the place, giving a correlation of 0.6. The ground to 150 m contour exhibits a steady slope and this has effect on the coastal wind convergence through increased friction. Between 14° and 15^{o} N, the hill ranges are closest to coast and rainfall is a maximum near about 14^{o} N.

Rainfall enhancement on account of orography may depend on slope of ground, height, configuration of land around and synoptic systems causing rainfall. Height alone is not a decisive factor. Agumbe (659 m) in the Western Ghats records 718 cm of rain in June to September as compared with 837 cm at Cherrapunji (1313 m). In the Ghats themselves, Mahabaleshwar at nearly twice the height of Agumbe records 16 per cent less rain. The slope near Agumbe is l/6, about the same as at Cherrapunji, while it is l/l5 at Mahabaleshwar, The rainfall at Matheran is quite high for its location on the plateau of a minor range almost at the beginning of its lee side.

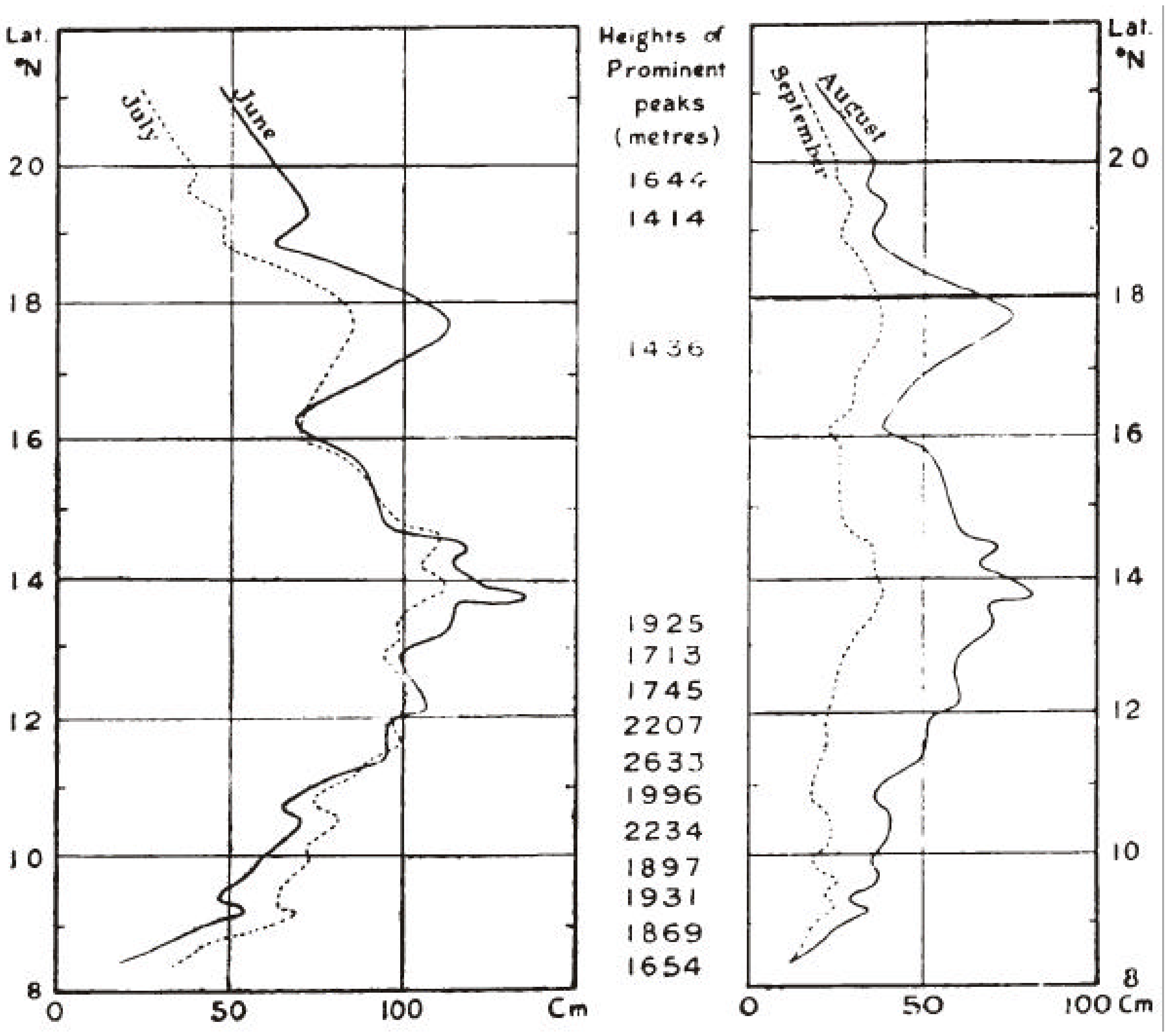

Fig. 10.4 Normal southwest monsoon rainfall at stations along the West Coast.

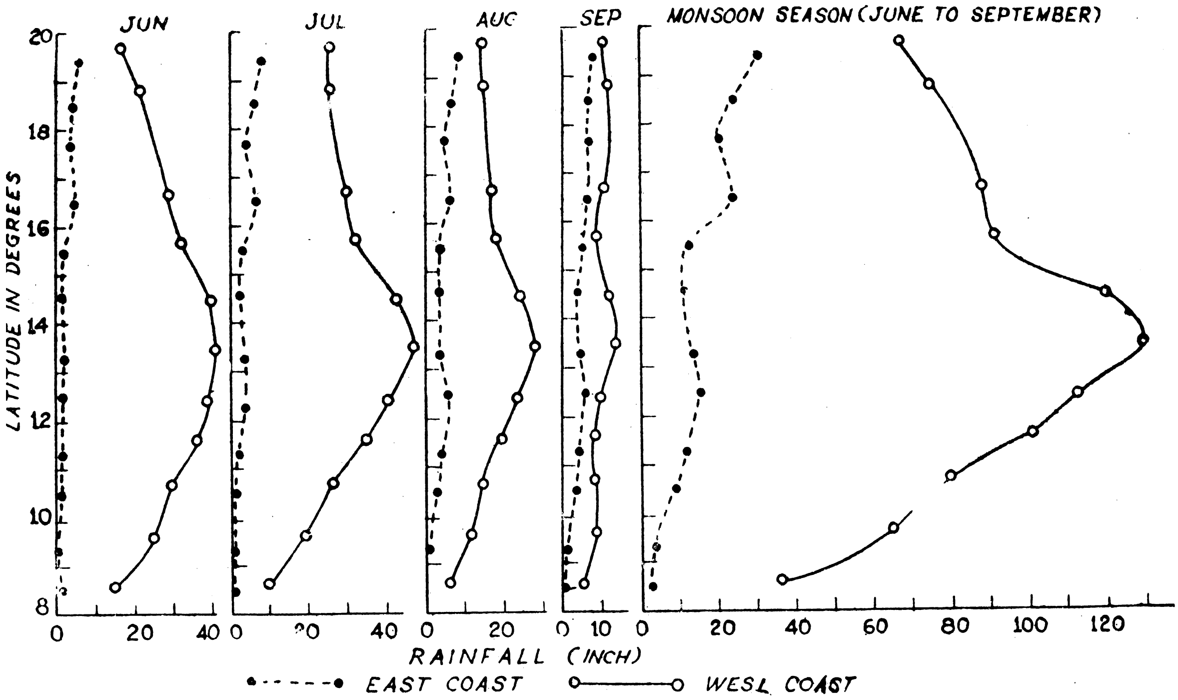

Fig. 10.5 Rainfall profiles of monsoon season.

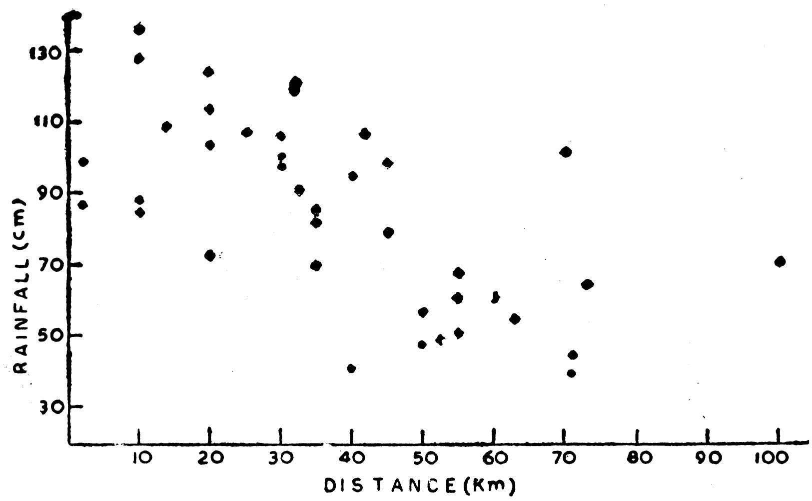

Fig. 10.6 Distance of the Ghats 150 m a.s.l. and normal rainfall in July in the West Coast

Bhatkal and Honavar record the highest rainfall among the stations right on the coast, the main hill slope commencing within 10 km of coast. Sawantwadi gets 50 cm more rain than Vengurla, though on the same latitude and in the plains, either due to the small elevation nearby or its being 15 km nearer the main hill range. Karkal, a plain station, gets about 100 cm more than along coast, at the same latitude. This is mainly due to its being only 12 km from main hill slope while the coast is 25 km away. Besides, there is a hillock of over 300 m at 7 km distance to east from Karkal.

Saklespur (895 m) in the valley of river Hemakoti, records only about two thirds of Mangalore rainfall on the coast. Sringeri (743 m) and Agumbe (695 m) make a striking contrast with rainfall amounts of 328 and 718 cm respectively. The latter is at a lower elevation by 80 m. The distance between them is only 22 km. But the important difference is in their positions with reference to the hill slope. Agumbe is just on the main hill slope (l/6) or just at its end. Sringeri is 10 km behind the crest of over 900 m.

The ratio of Agumbe rainfall to the coastal amount in the same latitude is 2:4. Amboli, Mahabaleshwar and Gaganbavada also give a ratio of 2:5 with respect to coastal rainfall in their respective latitudes. This is much lower than that of Cherrapunji rainfall to Mymensingh which is more than 5. Ascent of air near hill range barriers has necessarily to commence some distance ahead of the slope on the windward side. This kind of ascent is perhaps more prevalent in the coastal plain near the Western Ghats than in the windward plains of Khasi hills. Hence the ratio of enhancement of rainfall on the hills compared to the plains would be more in the latter case.

10.11 Within 80 to 100 km leeward from the crest of the Western Ghats, rainfall decreases to 30 cm. We saw that Gauhati on the lee–side of the Khasi hills has 121 cm of rain and Lanka 79 cm. The lee–side decrease is much more in the case of the Western Ghats. Certain synoptic patterns are quite favourable for causing rain in the Brahmaputra valley without flow–pattern being affected by lee–side effects, as in break monsoon situations. Similarly rainfall on the leeside of the Western Ghats to the north of 19^{o} N is a little more than to the south, as this area can get rains from the travel of monsoon depressions, particularly in September, un influenced by lee–side effects on the flow by the Western Ghats. South of 11° N, the rainfall on the lee–side is still less but this is also a continuation of the extreme dry belt from Tuticorin, on the average drier in this period than even the western Indian desert.

10.12 In the Himalayan area, the mean flow during the season is more or less parallel to the Himalayan ranges both in the lower and upper troposphere. The seasonal rainfall at a few stations which are the highest among the few raingauge stations in the various regions are given in Table 10.3.

Table 10.3
Seasonal Rainfall in Himalayas (June to Sept.)

Station	**Latitude**		**Longitude**		**Height**	**Rainfall**
	(°	**' N)**	**(°**	**E)**	**(m)**	**(cm)**
Denning	27	52	94	50		349
Pasighat	28	6	95	23	157	315
Kurseong	26	53	88	17	1476	340
Mussoorie	30	27	78	5	2042	197
Nainital	29	23	79	27	1953	227
Chaukuri	29	45	80	15		197
Landsdowne	29	50	78	41		177
Simla	31	6	77	10	2202	117
Dharmsala (upper)	31	49	76	5	899	260

Even the few raingauges are located at inhabited stations which are obviously not at the most rainy points. It is generally considered that the rainfall decreases along the Himalayas from east to west. Weak circulations prevailing over or near these mountains give rise to wind flow from south which brings into play effects of orographic uplift, Such systems are most prevalent during break situations.

10.13 In the central parts of the country, Pachmarhi (1075 m) is often in the southwest sector of monsoon depressions and gets a seasonal rainfall of 204 cm. Stations around, at about 600 m or below, record on the average 120 cm. The enhanced rainfall is 1.7 times, comparable to some stations on the Western Ghats but less than 2.5 at places like Amboli. Orographic effect is thus noticed in depression rains as well, though to a smaller extent.

10.14 Day–to–day variations of rainfall on the hill slopes in relation to nearby places, show interesting features. At Mahabaleshwar, rainfall is mostly more than at coastal places. On a few occasions rainfall is more along the coast than on the hills, e.g. 2 July 1939 when Mahabaleshwar, Bombay and Ratnagiri recorded respectively 7, 23 and 13 cm. The heaviest rainfall in a day is 58 cm (5 July 1974) at Bombay but only 38 cm (July 1958) at Mahabaleshwar. Even when coastal rainfall is insignificant, it may rain heavily on the Western Ghats. On 14 July 1939 Mahabaleshwar recorded 16.5 cm while Bombay had 0.3 cm and Ratnagiri 0.8 cm. Lastly, even in July, both the Ghats and the coastal plains may be practically dry on some days. These large variations are interesting as the July flow pattern over the Western Ghats has almost always ascent up the western slopes. A significant association between rainfall at Mahabaleshwar and the strength of the lower tropospheric westerlies along Konkan has been noticed (Srinivasan et al, 1972). Four types of synoptic patterns seem to exist, causing more rain on the Ghats or more along the coast in a general area of rains, good rains on the hills but little in the plains and little rains in both areas. Synoptic pattern is thus important in enhancing rains even through orographic features.

Cherrapunji area has less steady flow pattern. When there is a depression over the north Bay, Assam goes dry with easterly flow and so does Cherrapunji. In 'breaks' also, the flow will be more parallel to the Khasi hills, though any perturbation superimposed may cause southerly winds. When the monsoon trough is in the vicinity, favourable southerly flow may prevail. Owing to the one order of magnitude difference in rainfall between Cherrapunji and neighbouring plains, it is difficult to trace any relationship between the two.

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CHAPTER 11

OCEANIC FEATURES

11.1 Most intensive atmospheric and oceanographic observations were made in the north Indian Ocean during the International Indian Ocean Expedition. Though for only two years, these data give a broad picture of the normal pattern.

11.2.1 In July 1963, sea surface temperatures were least (23–24° C) along and off the coasts of Africa (north of the equator) and Arabia (except in the Gulf of Aden), owing to strong upwelling caused by the wind stress due to strong winds parallel to the coast. They increase to 31° C in the Gulf of Aden and towards east in the Arabian Sea to 28°–29° C between 65^o and 70° E, after which there is slight decrease of 1° C upto the west coast of the Indian peninsula, south of 20^o N.

Thus, in the Arabian, Sea there is more or less an east to west gradient of sea–surface temperature, with maximum gradient to the west of 60^o E. At the equator, west of 50^o E, the sea surface has a temperature of 25–2.6° C and between 50 and 60^o E, 26°¯27° C. In the Bay of Bengal, there is very little gradient, temperature decreasing by 1° C from west to east, north of 9° N. Only south of the equator do the isotherms run east to west, though in the southern tropical areas off the coast of Africa, the meridional alignment persists.

In June, the gradient in the Arabian Sea is a little less. In August, the waters off the coasts of north Africa and Arabia are colder by 1° C and the maximum temperature in the east Arabian Sea, north of 15^o N and along 70^o E, is less marked. South of 10° N and east of 60° E, rather warm waters develop. In the Bay, there is little gradient south of 10^o N and cold waters lie off Andhra coast. Sea temperatures are a little lower off Irrawaday delta and north Tennasserim coast so that the gradient from about 87^o E to east is a little more marked.

In September, the warmest patch north of 10^o N in the Arabian Sea is off the west coast of India. In the Bay, north of 10° N, temperature decreases from west to east. Slightly warmer waters are to the south of 6^o N and east of 60^o E with small patches of rather cold water around 72^o E and 83^o E, near the equator.

The mean sea temperatures are controlled by oceanic circulations including upwelling and radiation balance. In both the cases interaction with atmospheric circulations is important. While considering the averages for one year, effects of synoptic peculiarities in the atmospheric circulation of that year may also be reflected. Some of the smaller features noticed in the sea–temperature patterns of 1963 could be of that type. The broad features of the sea surface temperature appear to be the increase in temperature towards east Arabian Sea both from the west and the equator side, to the west of 60 longitude. Air mass coming from these sides into the eastern parts of the Arabian Sea could be convectively modified to carry moisture and heat upwards from the sea surface. In the Bay, the point of interest is that western Bay is warmer, favourable for intensification of lows coming from the east.

11.2.2 Riehl (1954) points out that the upwelling of cold water along the west coast is responsible for the tremendous depressions of temperature just west of the African and American continents, along 15^o S, which are accentuated in winter. In the Arabian Sea it is along the east coast of Africa and Arabia that most marked upwelling takes place, apparently due to the differences in wind flow.

11.2.3 Table 11.1 presents some statistics regarding atmospheric and oceanic temperatures etc. at selected points in the Indian Ocean gathered from the analysis of ships' reports of July 1964.

Table 11.1
Sea surface temperatures in July 1964 in the Indian Ocean

Position				No. of ships' reports	Mean air Temp (°C)	Mean dew point (°C)	Mean sea surface temp (°C)	Percentage of occasions when sea surface–air temperature was > 1° C	Most probable wind direction (in degrees)
Lat.		Long.							
21.8	N	62.5	E	62	27.4	24.7	28	28	220
19.5	N	58.4	E	58	25.3	23.6	23.7	0	200
17.9	N	66.7	E	72	27.9	24.5	27.9	10	250
17.5	N	40.4	E	130	32	27.1	31.2	5	330
16.5	N	85.8	E	32	28.9	25.2	29	30	210
14.1	N	71.8	E	43	27.8	24.9	27.8	15	270
11.2	N	60.5	E	91	26.3	22.9	26	4	230
9.7	N	51.5	E	74	24	21.3	23.1	15	210
8.3	N	70.6	E	103	27.6	23.8	27.9	20	280
5.8	N	86.5	E	176	27.6	24.5	28.1	23	240
1.0	N	50.0	E	10	25.5	21.5	25.2	Data inadequate	210
1.5	N	66.6	E	41	27.9	23.9	28.3	35	Variable
1.6	S	84.5	E	41	27.2	24.5	28.5	46	230
6.5	S	78.3	E	42	26.5	23.9	27.8	41	130
7.6	S	44.7	E	29	24.2	19.9	24 .8	26	150

(from "Marine Climatological Summaries – Vol. 4, 1964", I. Met. D.)

Close to the equator, in the Bay and east of Long, 70^{o} E in the Arabian Sea, there is a substantial percentage of occasions when sea surface is warmer than the air layers in contact by more than 1° C. In such cases, the warm sea may cause convection in the overlying air and lead to the moist air mass building up in depth. The dew points at the western end of the Arabian Sea or near the equator are rather low compared to the conditions of the monsoon air mass as it reaches the west coast of India. Travel over the warm sea contributes to increase of dew point. Whether spell of monsoon activity is related to the sea surface being warmer than the air mass requires to be studied. If there is any such relationship, bursts of air at high speed from across the equator with rather low air temperatures, perhaps from high latitudes in the southern hemisphere, may be most suited for convective modification in the east Arabian Sea. Thus there can be substantial contribution of air flow from across the equator but addition of moisture is taking place only in the eastern part of the Arabian Sea.

11.3.1 Two Atlases of the Indian Ocean have recently been published using data of the International Indian Ocean Expedition. Volume 1 (Ramage, Miller and Jefferies 1972) describes the surface climate of 1963 and 1964, Volume 2 (Ramage and Raman 1972) depicts upper air features based on long term radiosonde and wind summaries as well as aircraft averages. In broad terms they could be taken as representative of the mean patterns over the Indian Ocean, Further description relates to the area between 15° N and 20° S, unless otherwise stated.

At the surface, there was no westerly flow in April 1963 but in 1964 westerly component was developing between equator and 10° S in the belt 60 –95^{o} E. Presumably the Indian Peninsular trough had developed this year and also westerly flow north of equator around this. In May 1963, westerly component was developing between 0–5^{o} S west of 70^{o} E, and in all longitudes to north of equator. In 1964 westerly component developed east of 50° E between 0 and 5° S and as in previous year in all longitudes further to north. In both years in June there was no westerly component south of equator except in 1964 between 80° and 90° E. In July also, there was no westerly component south of the equator, except between 75^{o} and 95^{o} E (weak component) in 1964.

In August, there was no westerly component to south of equator. In September, but for 70^o to 85^o E between 0° and 5° S in 1964, there was no westerly component. By October there is a distinct tendency for westerlies to develop in both years in 0 to 5^o S, Thus , on the whole, westerlies do not develop at surface south of equator, when analysed in latitudinal intervals of 5°.

11.3.2 Rao and Raghavendra (1967) analysed 50,000 ships' observations of 1964 between 15^o N and 15^o S in the Indian Ocean between 40^o E and 100^o E, Averages of pressure, wind, cloudiness and occasions of rainfall have been computed for 2 ½ latitude belts for each month. Their results are shown in Figs. 11.1 to 11.3. Data for 1965 gave results similar to 1964 In May, there is a pressure minimum between 2.5^o S and 5^o S, while wind shifts from an easterly component between 5^o S and 7.5^o S to a westerly component in the next 2 1/2 degree belt to north. This wind shift without pressure minimum may be called a wind trough. In June and July, a wind trough is noticed at 2 $½^o$ S. August shows a pressure trough between 2 $½^o$ S and 5^o S and September a wind trough at 2 $½^o$ S. The conclusion is that a very weak pressure or wind trough occurs near about 2 $½^o$ S and though a weak westerly wind component starts developing just south of the equator, the main development is to the north. 1964 was a year in which surface westerlies were better developed between 0^o and 5° S than in 1963, as seen from the analysis of Ramage et al. In the analysis of Rao and Raghavendra, it is seen that the westerly component at surface which builds up just south of the equator in April and May, decreases thereafter. Generally in this period there is practically continuous decrease of pressure from 15° S to 15° N, The higher frequency of rainfall in July from 7 ½° S to 5^o N and progressive decrease to north till 15 is interesting.

11.3.3 Raman (1965) reported occurrence, over the Indian Ocean area, of two distinct trough systems with embedded cyclonic vorticities, in the lower and middle troposphere, one in the northern and another in the southern hemisphere in all months of the year. The zone between the two troughs is characterised by persistent westerlies. In the southwest monsoon period, the trough north of the equator is the monsoon trough across northern India extending into north Bay. Over the Indian Ocean these troughs seem to be associated with warmer sea surface temperatures.

11.3.4 Flohn (l958) pointed out that in the Pacific and Atlantic in the northern summer, southern trades after crossing the equator, do not develop westerly component, at least not between 0^o and 5^o N In the Indian Ocean, the southeast trades turn to southwest at about 2^o – 3^o S. Similarly in northern winter, the northeast trades become northwest or west at about 2°–3^o N. At 200 mb air crossing the equator continues as southeast in winter (northern hemisphere) and northeast in summer (northern hemisphere) in the Indian Ocean. area Westerly component is acquired after flowing around the sub–tropical highs of the winter hemisphere.

11.4 Bunker (1965) traced a low level jet off Somalia and thence across the central parts of the Arabian Sea to the coast of India, decreasing in speed progressively to east. A maximum speed of about 50 kt is attained at the top of a 1000 metre thick layer of air, cooled by contact with cold upwelling water which may have as low a temperature as 13° C. The jet is a result of the strong pressure gradient at right angle to the Somalia coast between the air over the heated land and cold off–shore waters. The upwelling and very low sea surface temperatures are due to the strong wind transporting large quantities of surface water eastwards. The wind profiles in this low–level thermal wind jet are shown in Fig. 11.4. The jet crossing equator pointed out by Findlater (1969), Bunker's jet off Somalia and the low level jet over the peninsula brought out by Joseph and Raman (1966) may all be different segments of the same feature. As the colder air is warmed during its further passage over the Arabian Sea the horizontal temperature gradient decreases and the jet's maximum value decreases progressively to east, dropping to less than 30 kt near India.

The air off the coast of Somalia has a very stable lapse rate but conditionally unstable. Over the western section of the Arabian Sea, the skies are generally clear with only some scattered

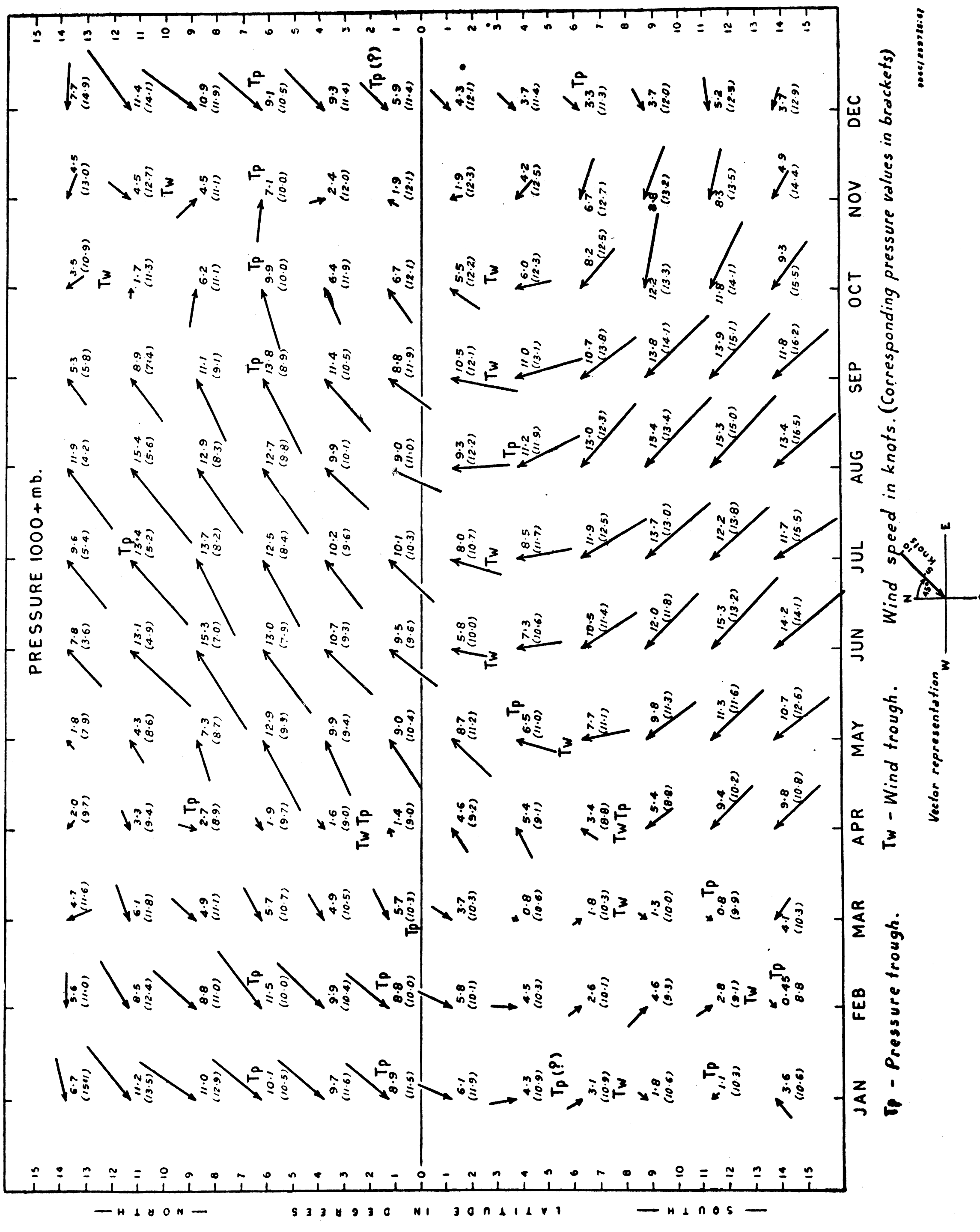


Fig. 11.1 Mean pressure and vector mean wind in 2½ degrees Latitude Belts - 1964 (Rao and Raghavendra, 1967)

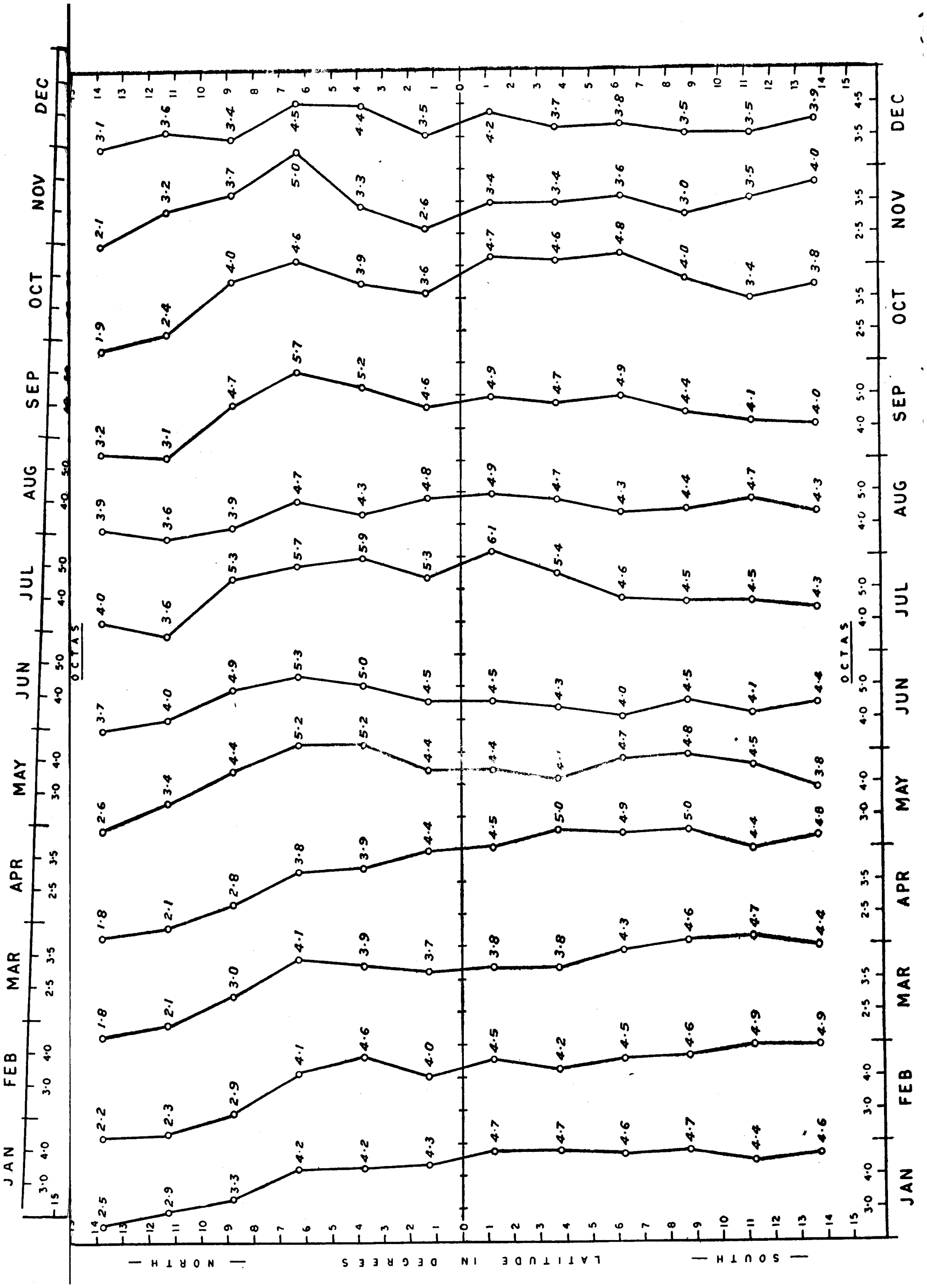


Fig.11.2 Average cloud amount in 2 degrees Latitude Belts - 1964. (Rao and Raghavendra, 1967)

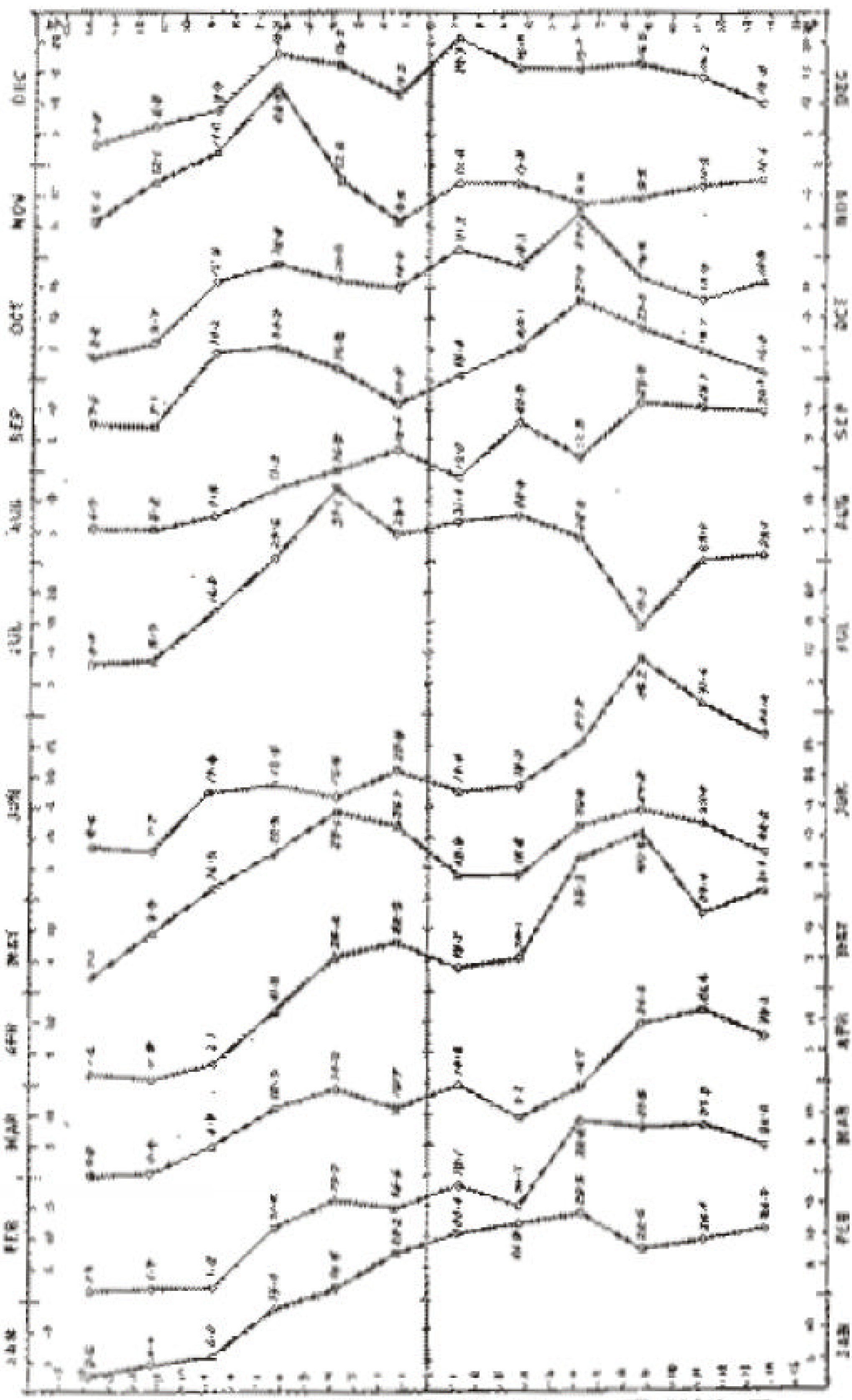


Fig.11.3 Percentage occasions of rainfall in 2 ½ degrees Latitude Belts - 1964. (Rao and Raghavendra, 1967).

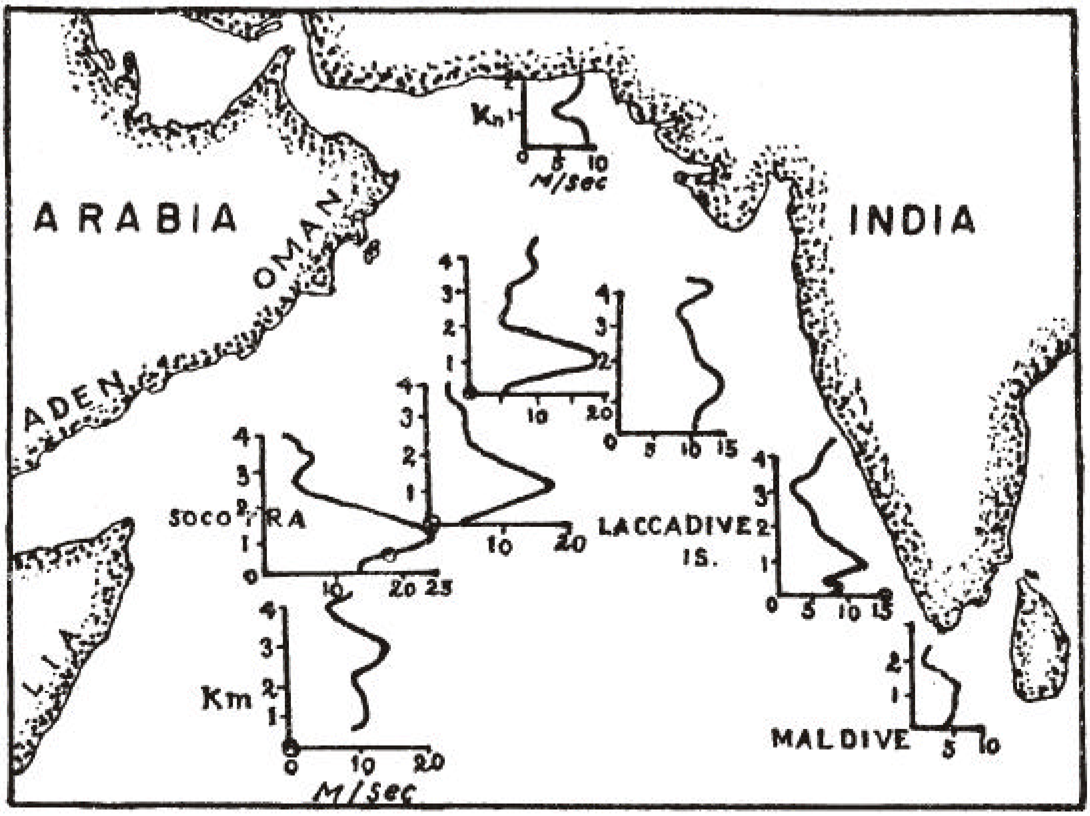
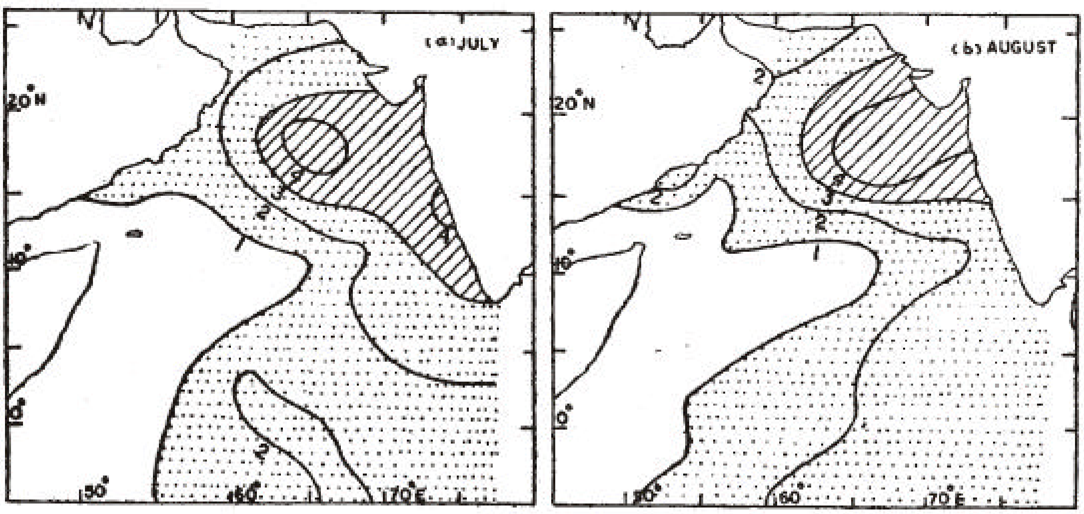


Fig. 11.4 Wind profiles in low level thermal wind jet. (Bunker 1965).



Mean cloudiness (a) July and (b) August during 1967-69.

Fig. 11.5

cirrus and occasional patches of lower clouds. East of 60° E, variable amounts of cumulus occur, as well as increasing altostratus and cirrostratus clouds. As the Indian coast is approached from the west, the amounts of cumulus increase and groups of Cb are formed. Over the Indian coast, the showers are widespread. According to Bunker, in the west half of the Arabian Sea, the jet winds are accelerating and there appears to be widespread subsidence. In the eastern half, the jet is decreasing in magnitude and there is small mean convergence which accounts for the difference in weather between the two parts, caused by the lifting of the conditionally unstable air.

11.5.1 Studies of satellite cloud pictures have provided continuous sampling of the cloud distribution over the Indian Ocean areas. Sikka (l97l) has studied the variations of cloudiness in different parts of the tropical belt from 0 to 140^o E, bringing out the differences between the Indian and adjoining areas. Between 60 and 90° E, axis of maximum cloudiness lies near 10° S in Jan. – Feb. and cloud amount decreases northward across equator. By April, this maximum moves nearer the equator and weakens, while simultaneous increase is noticed at the equator and upto 5° N. These conditions prevail in May too, though the northern hemisphere cloudiness increases upto 10° N. With the onset of the monsoon, there is a sudden increase of cloudiness between 10^o and 20^o N between May and June and the axis of maximum is near 12°–15° N. The maximum intensifies and approaches its northern–most position of 18° N during July–August. The maximum begins to weaken in September and shifts southward to about 12^o N. In October, the field of cloudiness becomes flat between 10° N and 10° S and remains so in November and December. The cloudiness maximum does not seem to migrate across the equator twice a year. What is observed is that the field of cloudiness becomes flat in the transitional months in the equatorial belt due to weakening of the maximum of the coming winter side and growth on the coming summer side. Between 90^o and 100^o E, July–August maximum is at 25° N. In both northern and southern summers, significant cloud amounts extend upto 5 latitude on the winter side. Between 100^o and 120^o E, maximum cloudiness shows a regular seasonal progression across equator. Summer maxima lie closer to equator, near 5° S in January–February and 10 o N in July–August; but there is significant clouding on the winter side of the equator as well. Further east, between 120° and 140° E, seasonal migration is noticed between 5^o S and 10° N as the two summer positions. Between 40° and 60° E, the striking abnormality is the lack of cloudiness in the equatorial region, even in the southwest monsoon period. In the southern summer, cloud maximum is near 8°–10° S which weakens by April. Between 0° and 40° E, the axis of maximum cloudiness oscillates between l0°–12° S (in February) and 8° N (in August).

11.5.2 Srinivasan (1968) finds that during the monsoon, three well defined cloud belts in the Indian Ocean – northern belt between 15° and 25° N, near equatorial belt between 10 o N and equator and southern hemisphere belt near 5° or 10° S. This characteristic cloud distribution is apparently present irrespective of the activity of the monsoon on the land area and is noticeable with some fluctuations on most of the days. The cloud field near the equatorial region of Indian Ocean is somewhat quasi–stationary with areas of maximum cloudiness at intervals at 15° to 20° longitude, with peaks at 45 –50^o E, 60 –65° E, 70°–80° E and 9O°–95° E. This is unlike the Atlantic and over Africa, in the form of single east–west oriented zone. He attributes this to sea–surface temperature and meridional distribution of ocean–land masses.

11.5.3 Thiruvengadathan and Jambunathan (l97l) have presented a picture of mean cloudiness over the Arabian Sea as worked out from satellite cloud pictures for July and August for the three years 1967 to 1969. The cloud amounts were estimated in 2½ degree grids, taking as clear when thin Ci clouds alone were present. Fig. 11.5(a) shows that, in July, a cloud maximum lies around 65° E between 15 –20° N, i.e. about 500–700 km away from the west coast. This cloudiness decreases rapidly towards southwest and south. The area of greater cloudiness extends to west coast with a secondary maximum close to Karnataka coast. Minimum cloudiness extends from Somalia coast to about 67^o E. In August (Fig.11.5b) the secondary maximum near Karnataka is absent and cloudiness is less to the south of 15° N along and off west coast. Fig.11.6 presents

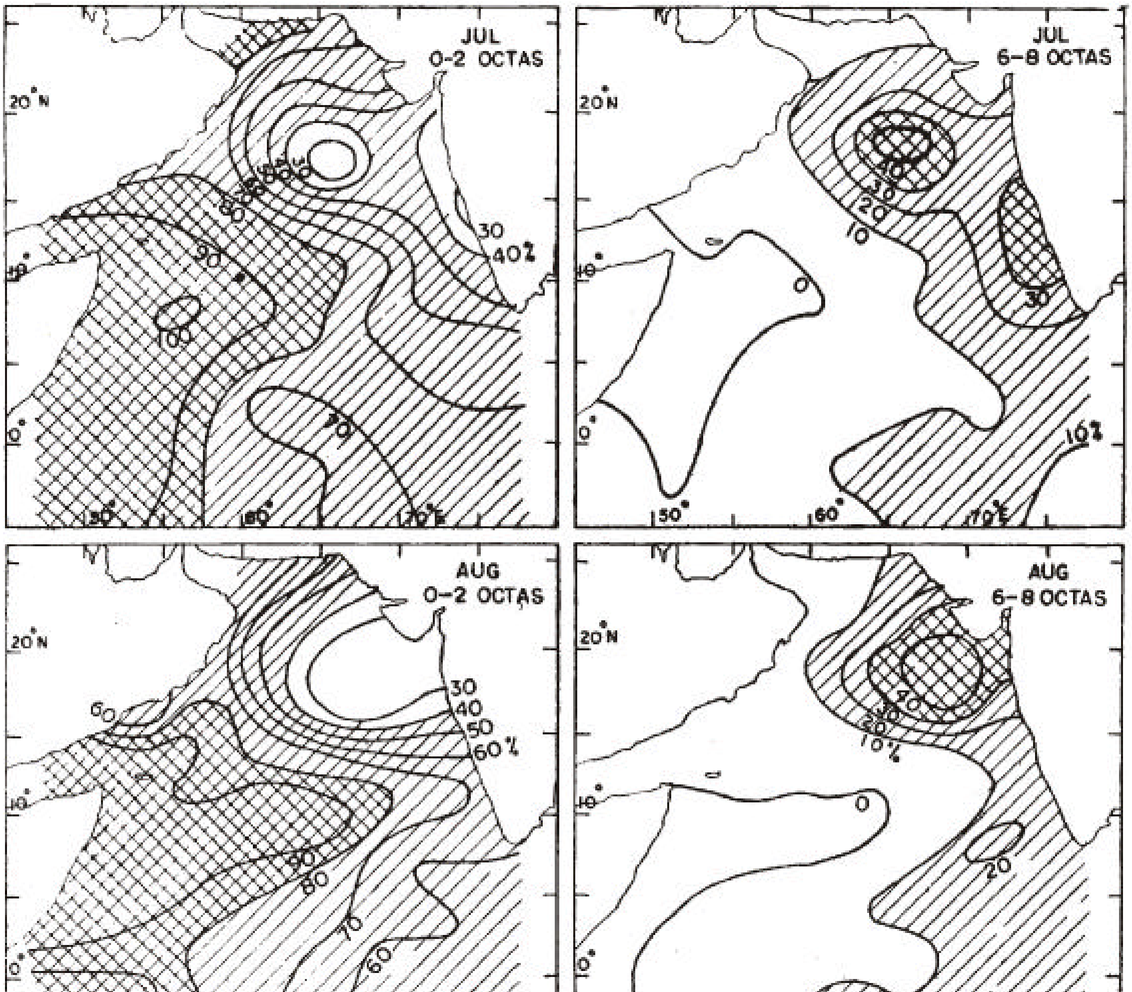


Fig. 11.6 Percentage frequency of cloud amounts 1967-69.

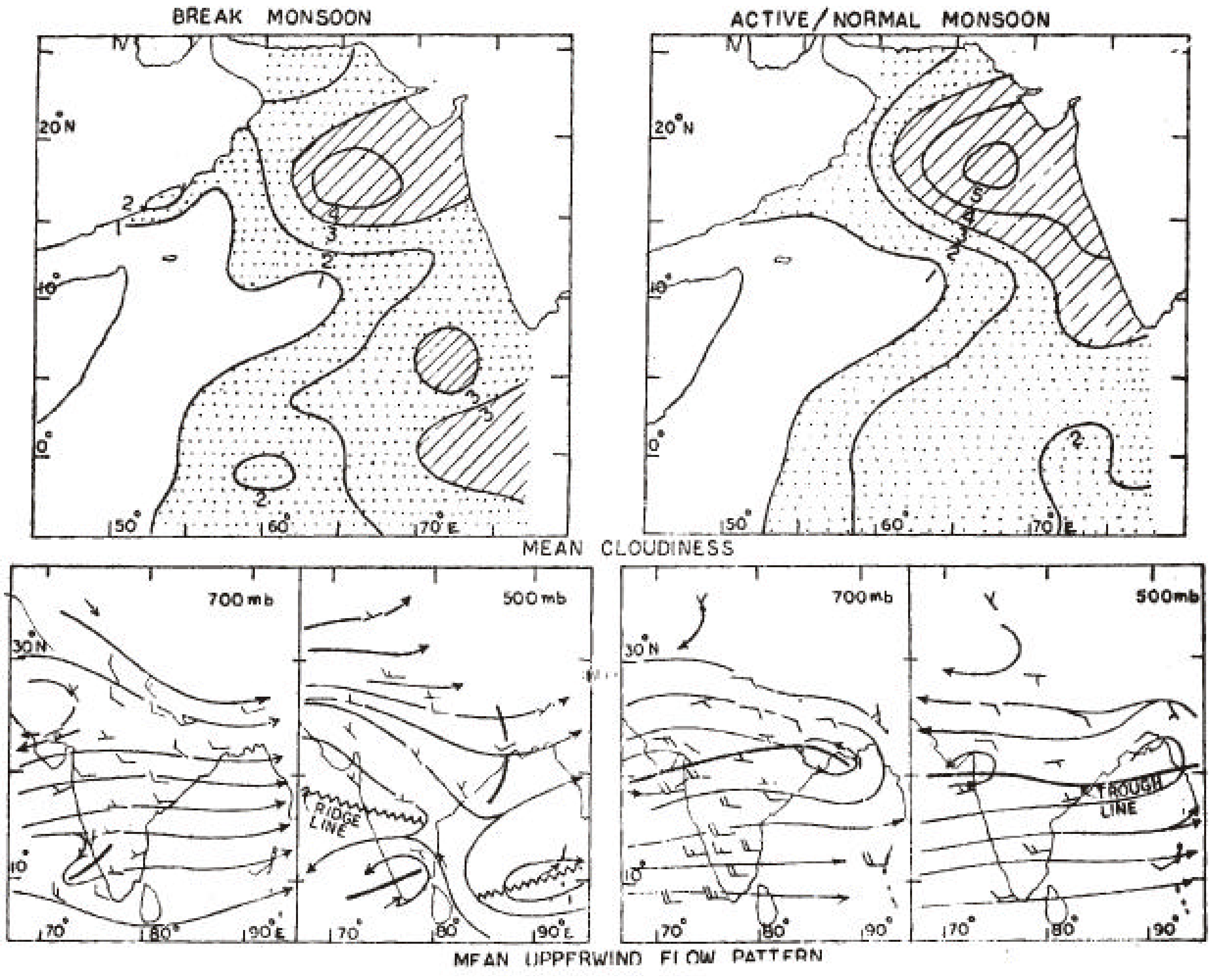
frequencies of 0–2 and 6–8 oktas of clouds. Generally it would appear that maximum cloudiness is after the air reaches the area of highest sea surface temperature. Fig. 11.7 showing the Arabian Sea cloudiness during 'break', still shows the maximum cloudiness from near 65° N to northern parts of west coast, though the amount along southern part of the coast has decreased like August. During 'breaks', cloudiness seems to increase south of 8 o N and east of 70 E,associated with low pressure systems that move in that area. Besides the amount of cloud, types and their vertical extent would vary between different areas and in different synoptic situations. The cloud maximum area over the Arabian Sea seems to result from Cu, Cb and dense layers of medium and high clouds. In the same area, during 'breaks' only Cu clouds are present without middle layers.

11.5.4 Nene (l97l) found from two years' data, that in the Bay, minimum clouding in July and August is to south of 10 o N and west of 90^o E and the maximum in the north, off Orissa–Bengal and Burma coasts. In a case of weak monsoon, maximum clouding was in east central Bay (13–16 July, 1968).

11.5.5 Hamilton's (1974) presentation of the average cloudiness from satellite data over the Indian Ocean from June to September 1967 confirms the extension of the cloudiness along west coast of India upto 62^0 E Cloudiness is high off the coast of Burma, Andaman Sea and the Bav north of 13° N. In the Indian Ocean, cloudiness increases at about 5^o S between 60^o and 70^o E and also to the south of Sri Lanka. The former is probably associated with the trough just south of equator. There are variations in the cloudiness at equator, generally in phase with further north over the Indian region. From April to May 1967, cloudiness increased most near equator and to north upto 5° N between 60° and 80^o E, over Sri Lanka, off Kerala coast, Andaman Sea and adjoining Bay. From May to June there was striking decrease in cloudiness over equator between 60^o and 80^o E but increase to west between 55^o and 65^o E which persisted in July. Cloudiness in this region near the equator may be related to monsoon activity over India. Other changes from May to June are general increase of cloudiness over the Arabian Sea, over India and Bay to north of 10° N, with decrease over Andaman Sea and south Bay.

11.5.6 Volume I of the Meteorological Atlas of IIOE (Ramage, Miller, Jefferies 1972) gives the cloudiness and percentage of rainfall reports from ships' observations of 1963 and 1964. North of the equator, in July, average cloudiness is 4 oktas or more east of 60^o E in the Arabian Sea, increasing to 6–7 oktas to the east of 65^o E. In the Bay, it is more than 5 oktas increasing to 7 oktas in east Central Bay. Differences in types of clouds must be more than the differences indicated by the amounts. This is brought out by the frequency of rainfall. North of 10° NP rainfall frequency is 3 per cent of the reports to the west of 65^o E but increases towards the Indian coast (still over sea) to 12 percent to 16 per cent, On the Bay side, to the east of 80^o /85^o E in the central and northern parts, the effect of low pressure systems approaching from the east, the seasonal position of the monsoon trough and development of depressions are seen as 15–20 per cent of rainfall reports. West of 85^o E, it is less than 5 per cent. Between 5^o N and 10^o N, rainfall frequency is more than in the 5 belt just to the north, except between 70^o E and 80^o E where it is less. In the 0–5° N strip, it increases still more between 55^o E and 90^o E but slightly decreases between 0 and 5° S. On the whole, the 5 belt on either side of the equator, to the east of 50^o /55^o E is as rainy as the area off the west coast of India or the eastern parts of the Bay. The equatorial areas cannot be regarded as dry.

11.6 The location of ITCZ and its utility in synoptic practice is an important aspect. A simple concept of ITCZ is a zone of convergence of the air from the two hemispheres, located in a trough in the summer hemisphere. At the height of the monsoon, a trough runs regularly along the Gangetic valley, called the monsoon trough which is normally identified with the ITCZ over is–he Indian area at surface level. While the air to the south of the trough line can be traced back to the southern hemisphere, that to the north is a continuation of the same air mass at least upto 2 km after some modification over the Bay and northern plains. To that extent it cannot be regarded as a line of convergence of the air from the two hemispheres. The hill ranges to the east prevent air



Mean cloudiness and upper wind flow pattern (700 and 500 mb) during 'break' and active/normal monsoon conditions.

Fig.11.7

from further east flowing to the north of the monsoon trough. Sawyer (1947) placed the inter–tropical front between continental air to northwest and the monsoon air on both sides of the monsoon trough. As discussed earlier, the monsoon trough is not the cause of much weather, though its position affects the development of other systems which have a profound influence on rainfall.

During the period of advance of monsoon, it is usual to identify the northern limit of monsoon air mass as the ITCZ. While in some areas there may be a trough along this line, it may only be a shear zone in other places with stronger winds to the south. This line usually takes a more northeast–southwest orientation in the Bay of Bengal due to the earlier advance in east Bay than in the Arabian Sea and the Peninsula.

The existence of another trough south of the equator does not require any modification of a single ITCZ north of equator. In the southern trough the same air is flowing across the trough while it is only in the north that we can identify a meeting of the air of the two hemispheres. The inter–tropical character of this boundary does not provide any extra–kinematic properties to this line, that, if it is treated as a shear zone or trough in one hemisphere, When the ITCZ is not too far north, any surges from across equator may provide some clue to its activity or displacement. Long bands of clouds are now and then seen in satellite cloud pictures over the seas during the phase of advance of monsoon which could be associated with position or activity of the ITCZ.

11.7 The eastern part of the Arabian Sea has an important role in developing an air mass moist in great depth with a lapse rate slightly in excess of saturated adiabat. Hence the synoptic systems responsible have to be clearly identified. If the low level jet is the main factor, the initial cooling off Somalia coast would become decisive. It would be an interesting accident of nature, perhaps made possible by the shape of the Arabian Sea and western land border of the Indian Ocean, that air is first cooled off this coast and then warmed further east, in the process pumping moisture into a deeper layer.

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CHAPTER 12

CLOUD AND RAINFALL CHARACTERISTICS

12.1 The monsoon is one of the most rainy systems over tropics and there have been studies on some aspects of the cloud systems and type of rainfall. Advent of radars enabled mesoscale features of the rain systems to be studied in relation to local physical features.

12.2 Precipitation during the monsoon could be in the form of thundershowers, showers or rain. Each type may occur in certain situations or regions or two or more types may occur together. The most striking feature of monsoon rain is the almost complete absence of thundershowers along the west coast to the south of 20^o N, once the monsoon is established. The first advance everywhere is with a thunder storm. But, thereafter, along the west coast it is as cumulus showers or some times as continuous rain from a deep layer cloud (even 9 km deep) interspersed with showers. Squalls occur when the pressure gradient is steep. Bombay has only one thunderstorm in July as against 24 rainy days. Mild thunderstorms occur during weak monsoon or when there is incursion of dry air in mid–troposphere (The description hereafter is with reference to July conditions). The west coast and even the Lakshadweep Islands record not more than one to two thunderstorms, while Port Blair in the same latitude has seven thunderstorms out of 21 rainy days.

This characteristic seems to extend leeward of the Western Ghats upto 77^o E between 13^o and 20^o N, Poona and Belgaum practically record no thunderstorms. In the rest of the Peninsula to the south of 14^o, the few days of rain are as thundershowers, particularly during 'break' monsoon conditions. Madras records five thunderstorms and seven days of rain.

The situation along the west coast perhaps extends out into the sea for 300 to 400 km. This is the zone where the deep moist air mass with saturated adiabatic lapse rate is built from the rather shallow moist layer further west. Towering cumulonimbus clouds well above the freezing level (6 km) would have been a very favourable mechanism but apparently it is not the case. Bhaskara Rao and Dekate (l97l) consider that the growth of cumulus clouds above 6 km is inhibited by the wind shear due to the easterlies aloft.

In the formative stage of monsoon depressions, thundershowers perhaps predominate. Later, the precipitation in their southwest quadrant is more frequently as rain. Near the mean position of the monsoon trough, thundershowers are more; Allahabad has 13 thunderstorms as against 14 rainy days and Delhi 6 out of 10. Further to the north, thundershowers are quite frequent. Over Rajasthan and Assam mostly thundershowers account for the rainfall. But their proportion is less over Gujarat. Southward from mean depression track, thunderstorms decrease so that Hyderabad records only one thunderstorm against 11 rainy days. The frequency of days of thunder as given by Rao et al (l97l) is shown in Fig. 12.1 while the diurnal variation as given by Raman and Raghavan (l96l) in Fig. 12.2.

12.3 Roy (1969) studied the occurrence of rainfall over Calcutta in relation to height of cloud tops as seen on radar, Clouds with tops at or below 6 km were regarded as 'warm', as the mean freezing level is about this height. Cold clouds, viz. with tops reaching 7 km or more, were sub–divided into Cumuliform or stratiform, according as the showers lasted for less than half an hour or precipitation was for a longer period, often for one hour or more covering quite a wide area. Table 12.1 shows the features of rainfall with the three types of situations.

12.4 Prasad (1970) has studied the diurnal variations of rainfall. Table 12.2 shows the hours of Maximum and minimum rainfall at a few stations. These are very variable between different stations. Secondary maxima and minima can also be identified. These variations can be

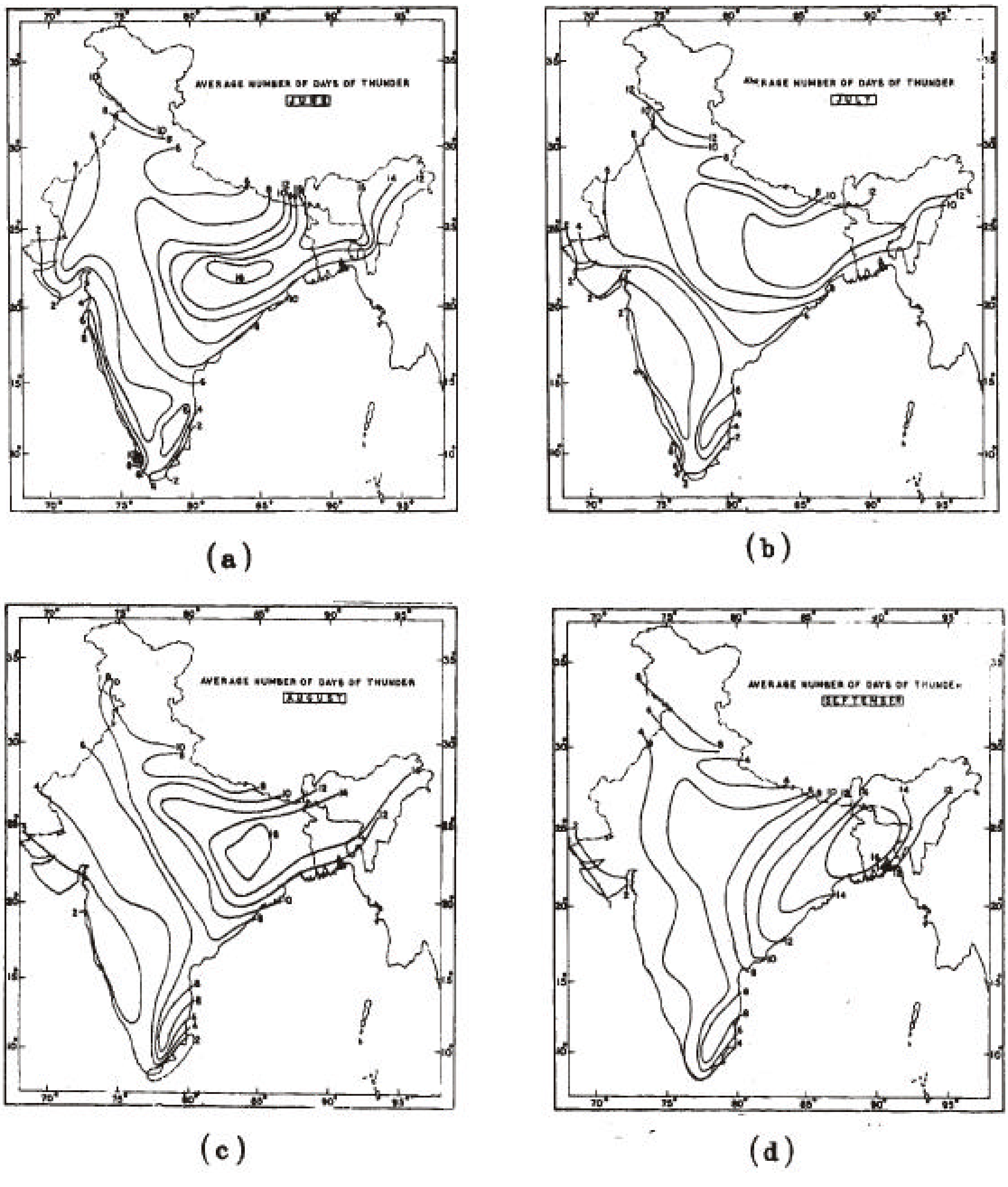

AVERAGE NUMBER OF DAYS OF THUNDER
JUNE
(a)
AVERAGE NUMBER OF DAYS OF THUNDER
JULY
(b)
AVERAGE NUMBER OF DAYS OF THUNDER
AUGUST
(c)
AVERAGE NUMBER OF DAYS OF THUNDER
SEPTEMBER
(d)

Fig. 12.1

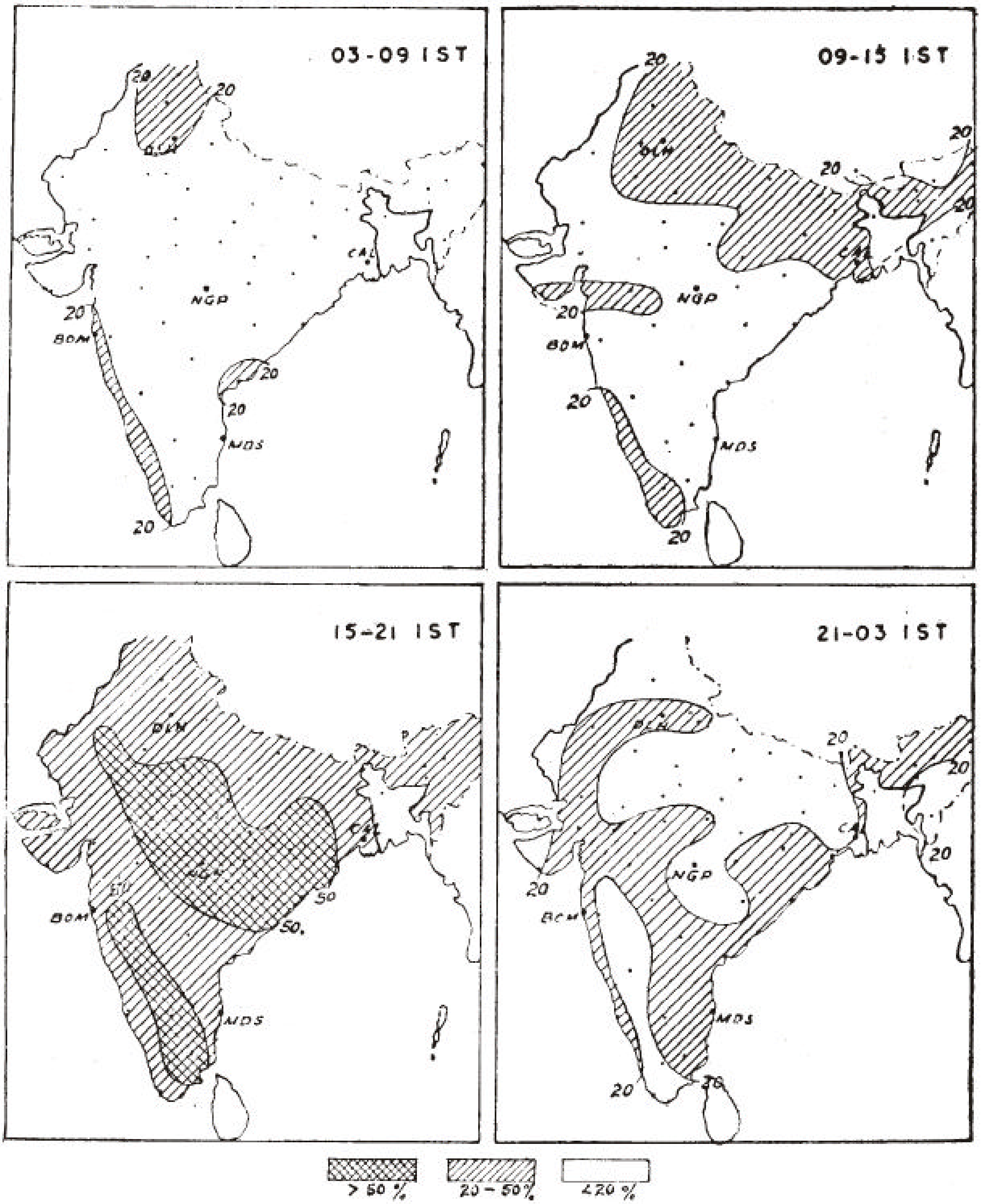


Fig.12.2 - Percentage of thunderstorm occurrences in different periods of the day during Monsoon (June-September).

Table 12.1
Association of rainfall at Calcutta with cloud types
(July to September, 1962 to 1964)

		Type of cloud		
		Cold stratiform	**Cold Cumuliform**	**Warm convective**
Rainfall and percentage of total		236 cm (50%)	212 cm (45 %)	22 cm (5 %)
Duration of rainfall (minute) and percentage of total		19582 (55 %)	11881 (33 %)	4183 (12%)
Rate of rainfall		7 mm/hour	11 mm/hour	3 mm/hour
Number of rainspells		94	498	198
Average duration of rainspell (minutes)		208	24	22
Distribution of rainspells in the periods				
(i) 00 to 08 hrs. IST				(i) 91 (46%)
(ii) 08 to 16 hrs. IST				(ii) 60 (30%)
(iii) 16 to 24 hrs. IST				(iii) 47 (24%)
Delhi	(Percentage of total rainfall	52	46	2
	(Rate of rainfall	2 mm/hour	–	1.3 mm/hour

Table 12.2
Times of maximum and minimum rainfall during monsoon

Station	Times of (I ST) Maximum	Minimum	Ratio of average rain at maximum time to minimum time in July
	Rainfall (mm)		
Cherrapunji	4–5 J N, JL	16–17 J N, JL	6
	0–1 A , S	15–16 A	
		17–18 S	
Mahabaleshwar	14–15 J N, JL	8–10 J N, JL	1.5
	15–17 A, S		
Saugor Island	3–4 J N	17–18 J N	5
	6–7 JL, A	20–21 JL	
	4–5 S		
Bombay	6–7 J N	18–19 J N	2
	7–8 JL	17–18 JL,A	
		20–21 S	
New Delhi	13–14 JL	22–23 JL	10
	4–5 A	20–21 A	
	8–9 S	0–1 S	
Jamshedpur	16–17 JL	0–1 JL, S	5
	15–16 A	8–9 A	
	13–14 S		

(J N = June JL = July A = August S = September) plausibly interpreted as influences of solar heating, radiational cooling from cloud tops, effect of local wind, flow patterns etc. as convenient. But there is little scope of establishing the relationship without further studies.

12.5 Rao's (1958) compilation of the base and tops of clouds over south Asia from the Comet aircraft reports is given in Table 12.3.

Table 12.3
Heights of monsoon clouds in South Asia (Kilofeet))

Height	High CLOUD S				Medium Clouds						Cumuliform clouds	
	Ci		Cs & Ci Cu		Lowest Layer		Middle Layer		Top Layer		Cb	
	Top	Base	Top	Base	Top	Base	Top	Base	Top	Base	Top	Base
Maximum	55	50	48	46	19	15	30	23	30	29	55	12
Minimum	32	25	30	20	9	7	18	15	22	22	30	1
Mean	41	36	39	34	13	11	22	19	26	25	40	4
No. of Observations	22	52	77	130	28	66	33	73	23	19	112	100

Deshpande (1964) analysed the tops of Cb clouds from commercial high–level aircraft reports and IAF reconnaissance flights for six monsoon seasons. Table 12.4 gives his results along with Rao's (1955) result 1 for south Asia and radar study for northern India by Kulshrestha (1962).

Table 12.4
Percentage frequency of heights of Cb tops during June to September

Height (Feet)	Aircraft reports 1957–62 (India)	Aircraft reports 1952–53 (South Asia)	Radar study 1958–59 (North India)
Below 30,000	3	0	3
30,000 – 34,900	6	12	28
35,000 – 39,900	21	22	
40,000 – 44,900	43	34	50
45,000 – 49,900	20	17	
Above 50,000	7	5	15
Total No. of reports	322	112	238

12.6 Bunker and Chaffe (1969) have analysed the innumerable cloud photographs, both time–lapse and still, taken from aircraft during; the IIO E over various parts of the Indian Ocean, Extracts from their monumental work, "Tropical Indian Ocean Clouds", are presented hereafter.

12.6.1 Their averages of cloud heights, amounts and extremes are given in Tables 12.5 and 12.6.

Table 12.5
Average Cloud Heights and Amounts (IIOE data)

Area	Height of Cumulus and Cumulus Congestus, (metres)	Amount of all Cumulus (oktas)	Height of Cumulonimbus (metres)	Middle cloud		High Clouds	
				Height (metres)	Amount (oktas)	Height (metres)	Amount (oktas)
70°E to India, 12°N to 20°N	2400	4	8200	4000	5	9300	4
65°E to 70°E, 11°N to 19N	1700	3	–	3500	4	9100	3
60°E to 65°E, 10°N to 18N	1600	3	–	4100	3	8700	3
55°E to 60E, 07°N to 15N	1400	3	–	4200	1	8800	3
50°E to 55°E, 05°N to 13°N	1500	1	–	4000	1	9300	1
45°E to 50°E, 02°S to 06°N	3300	5	–	3700	4	9000	2
70°E to 85°E, 04°S to 04N	3200	3	7900	5400	5	8000	3
85°E to 95°E, 02°S to 06N	3900	4	6000	5400	6	7400	4

Table 12.6
Extremes of clouds heights and Amounts observed during the Southwest Monsoon IIO E data)

Area	Height of Cumulus Tops (metres)	Amount of all Cumulus (oktas)	Middle cloud		High Clouds	
			Amount (oktas)	Height (metres)	Height (metres)	Amount (oktas)
70E to India,	Max. 10,200	8	8,800	8	13,400	8
12N to 20N	Min. 600	0	1,900	0	6,000	0
65E to 70E,	Max. 10,000	7 .	8,000	8	13,400	7
11N to 19N	Min. 700	0	2,200	0	8,500	0
60E to 65E,	Max. 6,300	6	6,200	7	11,800	8
10N to 18N	Min. 700	0	2,200	0	7,500	0
55E to 60E,	Max. 3,800	7	5,800	5	10,800	5
07N to 15N	Min. 700	0	2,500	0	7,000	0
50E to 55E,	Max. 3.000	6	5,800	5	11,800	7
05N to 13N	Min. 800	0	2,200	0	8,400	0
45E to 55E,	Max. 6,600	8	6,600	8	10,200	7
02S to 06N	Min. 900	0	2,400	0	8,500	0
70E to 85E,	Max. 10,400	6	6,400	6	10,400	8
04S to 04N	Min. 700	0	2,800	0	5,800	0
85E to 95E,	Max. 7,800	8	6,000	8	8,000	8
02S to 06N	Min. 1,000	1	2,800	0	6,400	2

Clouds amounts increase from off Somalia to India. The maximum heights over this part of the Indian Ocean are less than over India and the Bay of Bengal analysed by Rao (1958).

12.6.2 Fig. 12.3 shows areas over the north Indian Ocean where aircraft observations showed at least once presence of growing Cu and Cb clouds. Sampling was not uniform over different areas and there are regions from which no observations were available. This is borne out by the depiction of conditions over the north Bay, This figure brings out that a three degree belt off the west coast of India is the most convective region with an extension in the central Arabian Sea upto 60° E; the rain area shown in Fig. 12.4 confirms these features, though the width of the area to west is less. The westward extension of clouds (Fig. 12.5) in the northern parts of the east central Arabian Sea has been corroborated by satellite pictures. This feature requires to be satisfactorily explained.

12.6.3 From Fig. 12.6, Bunker and Chaffe (1969) observe that haze is the most widespread monsoon weather phenomenon over the Arabian Sea. Though infrequent and less dense along the equator, its presence over mid–ocean far away from desert areas is interesting.

Srivastava and Roche (1966) found that near the sea surface, where the relative humidity was about 70 %, visibility was good. The air became hazy near the cloud base with relative humidity of 90 %. During the vertical ascents of aircraft a sharp transition to hazy conditions was noticed as the relative humidity increased to over 80 %. This suggests that the haze was due to condensation of water vapour on sea–salt nuclei.

'Dry' haze was encountered over the western Arabian Sea, Somalia, Arabia and Iran with relative humidity less than 50 per cent. Desert areas are the source of this haze which extends upto 4 or 5 km.

12.6.4 Bunker and Chaffe (1969) find that the existence of many cloud layers above the tops of cumulus is an outstanding feature of the southwest monsoon. Discontinuous patches and layers of middle and high clouds occur hundreds of kilometres west of the Indian coast. Except in intense monsoon rain or close to the coast, these layers are 1 to 10 km above the tops of cumulus. This is illustrated by the cross–section of clouds in Fig. 12.7 for a flight of 26th June, 1963, which has been analysed for the different layer clouds. In view of the large gap between cumulus tops and lowest middle clouds, they do not consider that the middle and high clouds, were produced locally from water vapour transported by cumulus. The moisture was from regions east of the Arabian Sea and it is likely that the cirrus had originated from cumulonimbus over the east coast of India where a monsoon depression had developed.

The cloud photographs give the impression of a chaotic sky with the clouds not conforming to the categories of the international cloud codes. The altostratus and altocumulus just above the flight level fit M9 category best, but resemble fractostratus of the low cloud group L7. This layer of clouds may have spread out or sheared from the tops of cumulus that reached about 5 km near the Western Ghats, The bases of altocumulus at 6 km shown at 300 km in Fig. 12.9 merge into a featureless sheet of altostratus and can be classified as M5 category. Cirrus between 9and 11 km may have been from wave motion or anvil cirrus distorted by wind shear during passage over the Arabian Sea.

12.6.5 Over the Indian Ocean, cumulus clouds were usually organised into rows. Such organisation occurs over limited areas but the regime may break down within a few tens of kilometres after having prevailed over a few hundred kilo metres.

12.6.6 Over the Arabian Sea, on several occasions, rain was organised into long bands nearly parallel to the wind direction. The case of 28th June 1963 is illustrated in Figs. 12.8 and 12.9. The time lapse films reveal that rain between 19° and 11° N was concentrated into long

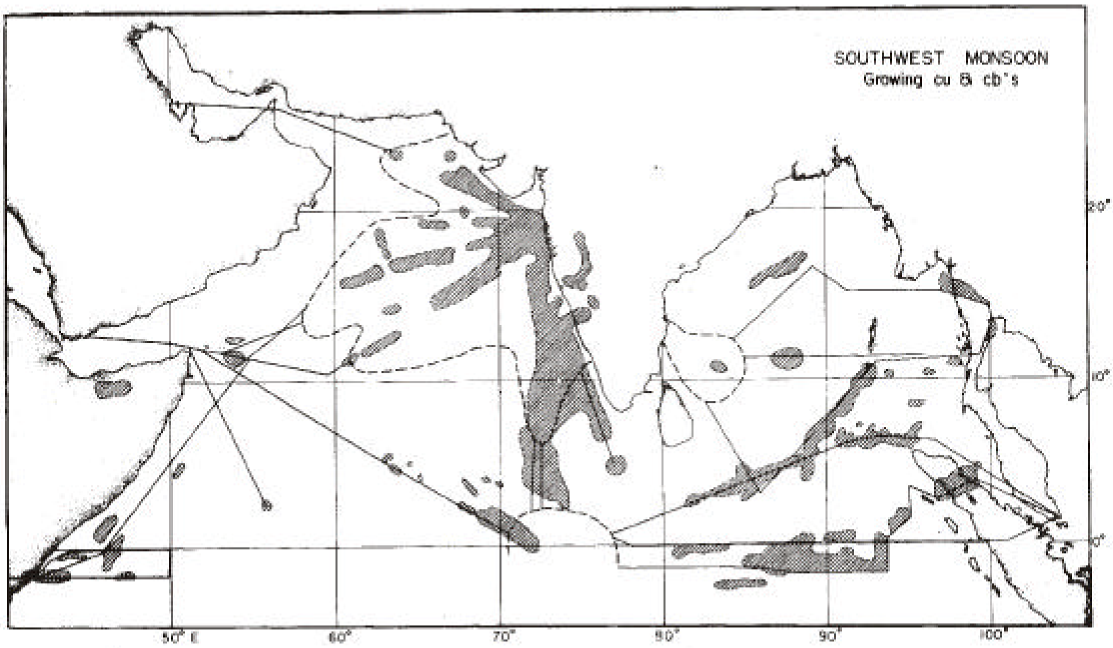


Fig. 12.3 Cumulus congestus or cumulonimbus occurrence during the southwest monsoon. The shaded areas indicate that at least one observation was made of these clouds.(Bunker & Chaffe ,1969).

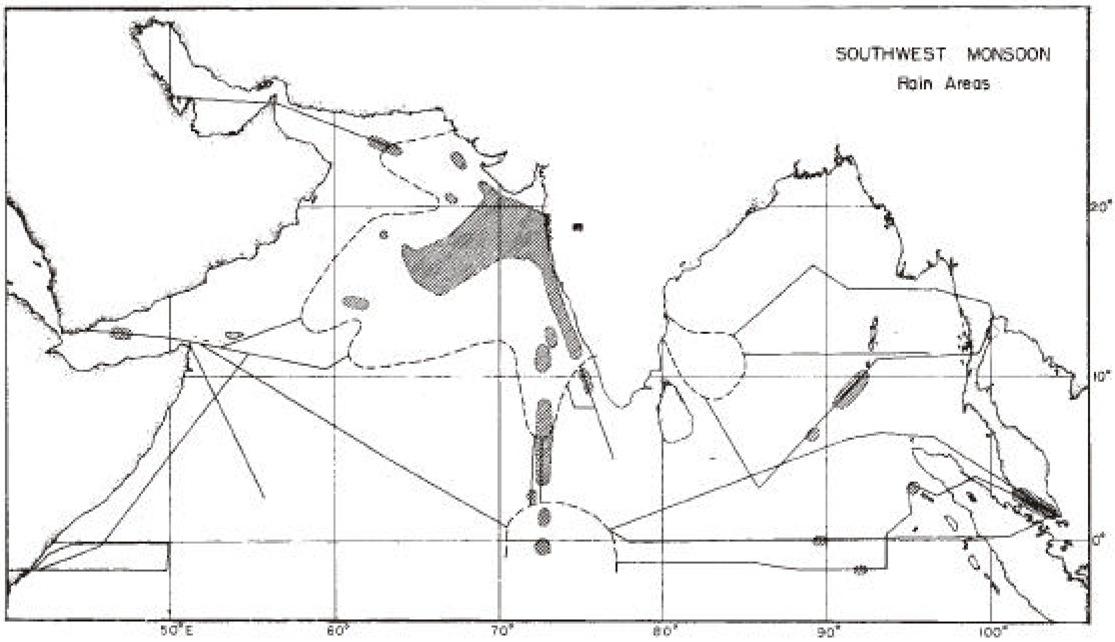


Fig.12.4 Rain occurrence during the southwest monsoon. The shaded areas indicate that at least one observation was made of rain. (Bunker & Chaffe , 1969).

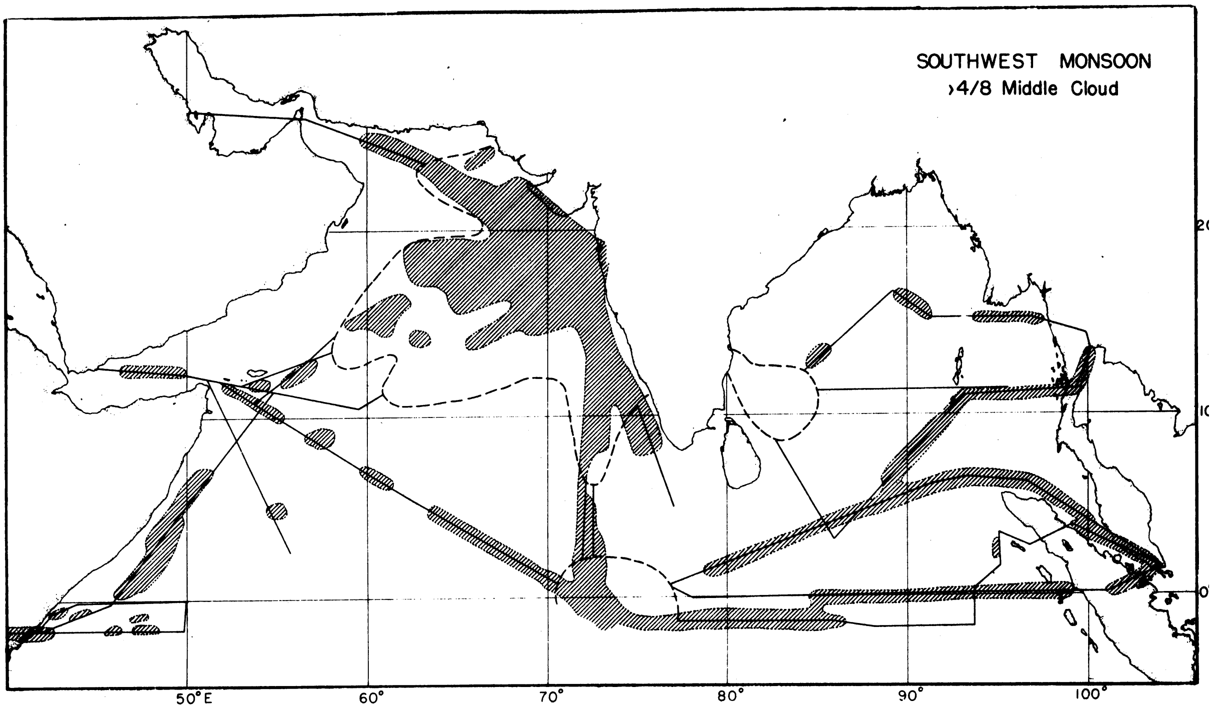


Fig. 12.5 (a) Middle-cloud occurrence during the southwest monsoon. The shaded areas indicate that at least one observation was made of more than 4/8 middle clouds. (Bunker and Chaffe, 1969)

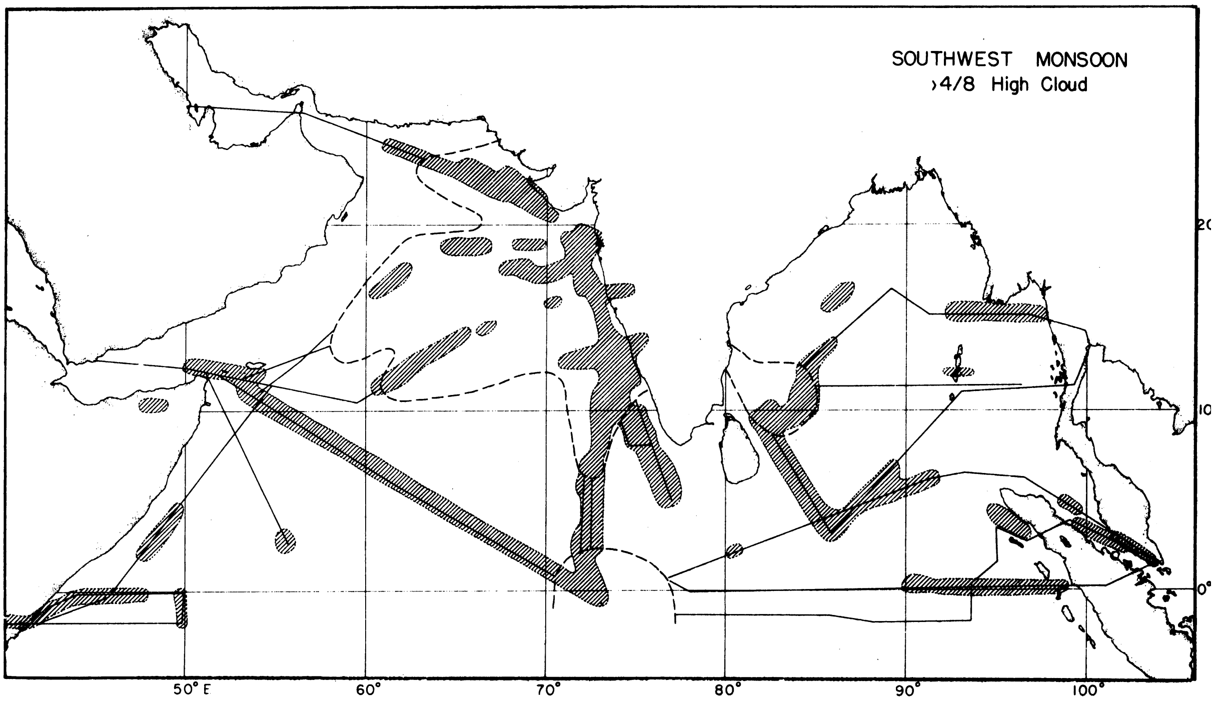


Fig. 12.5 (b) High-cloud occurrence during the southwest monsoon. The shaded areas indicate that at least one observation was made of more than 4/8 high clouds. (Bunker and Chaffe, 1969)

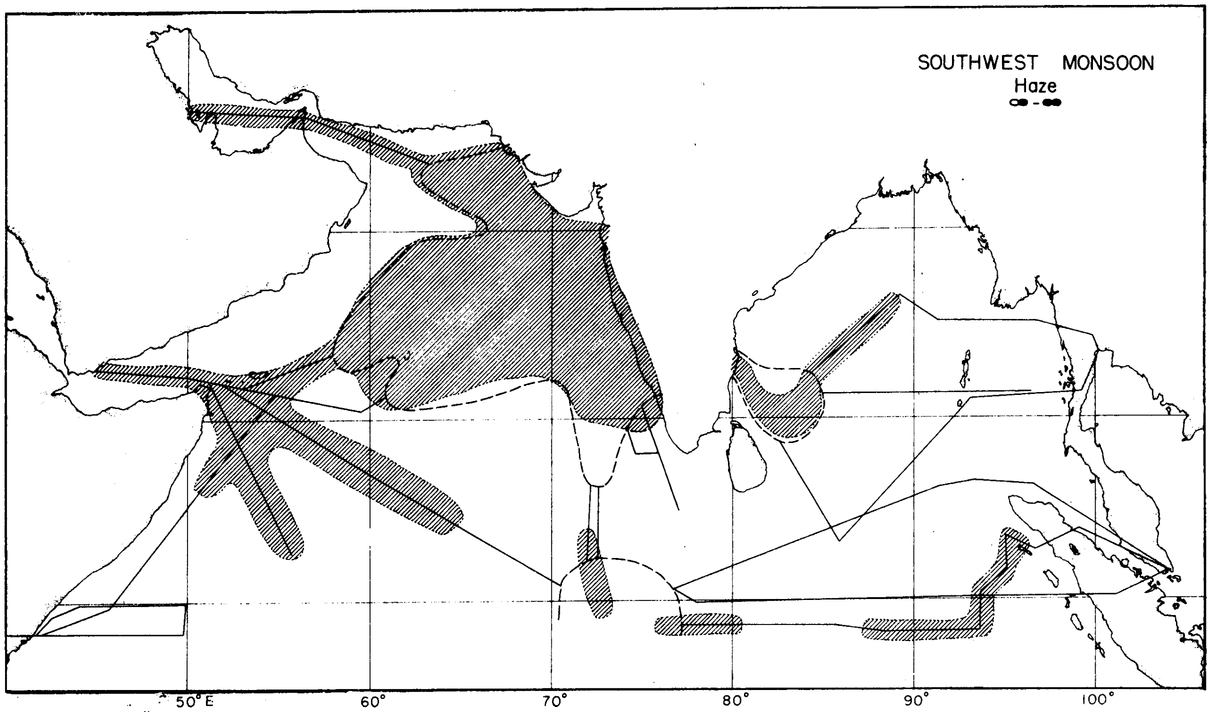


Fig.12.6 Haze occurrence during the southwest monsoon. The shaded areas indicate that at least one observation was made of medium to heavy haze.
(Bunker & Chaffe , 1969)

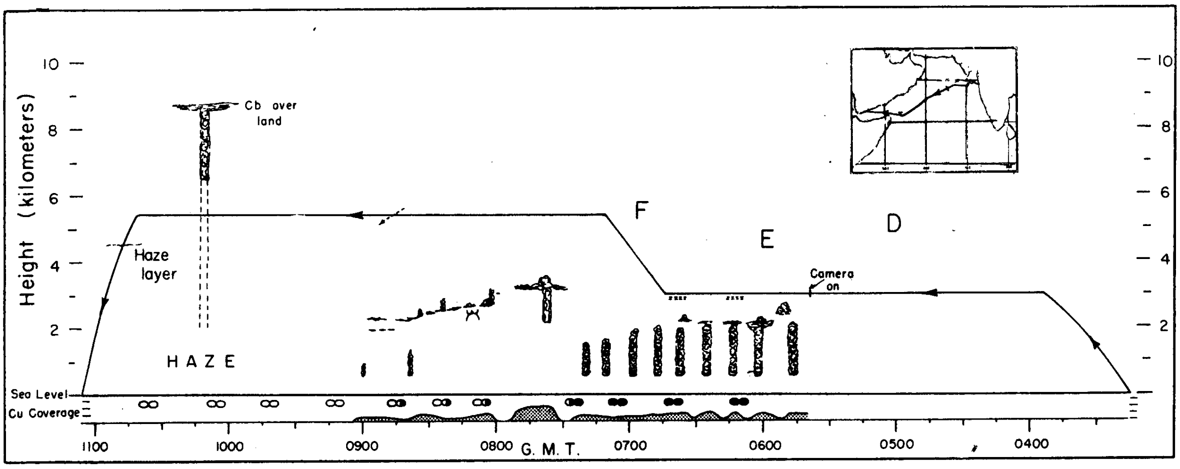
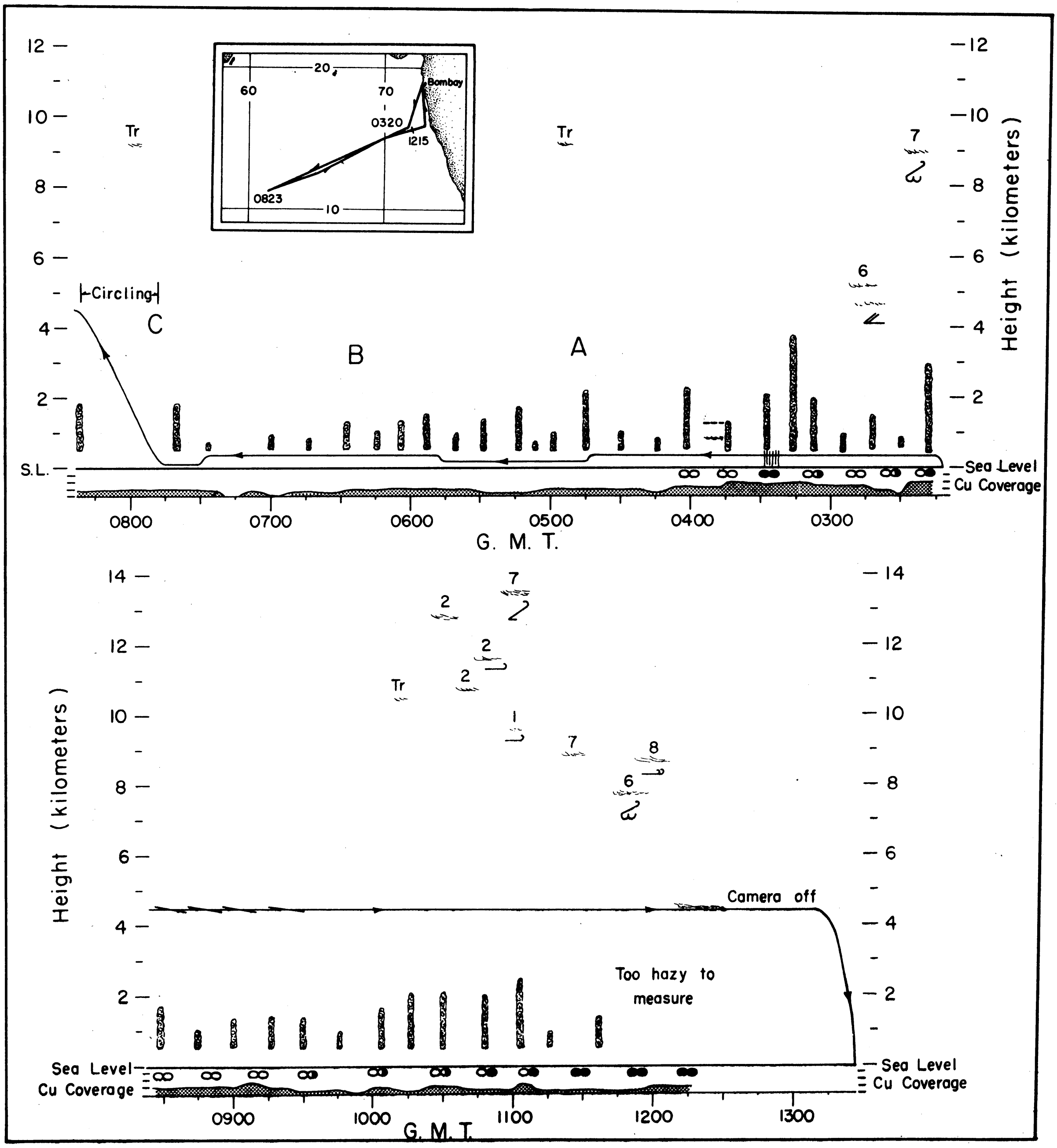


Fig.12.7 (a) Cloud cross section of 26 June 1963 RFF flight from Bombay to Aden. (Bunker & Chaffe, 1969)



(b) Cloud cross section of 26 June 1963 WHOI flight (Bunker & Chaffe, 1969)
Fig. 12.7

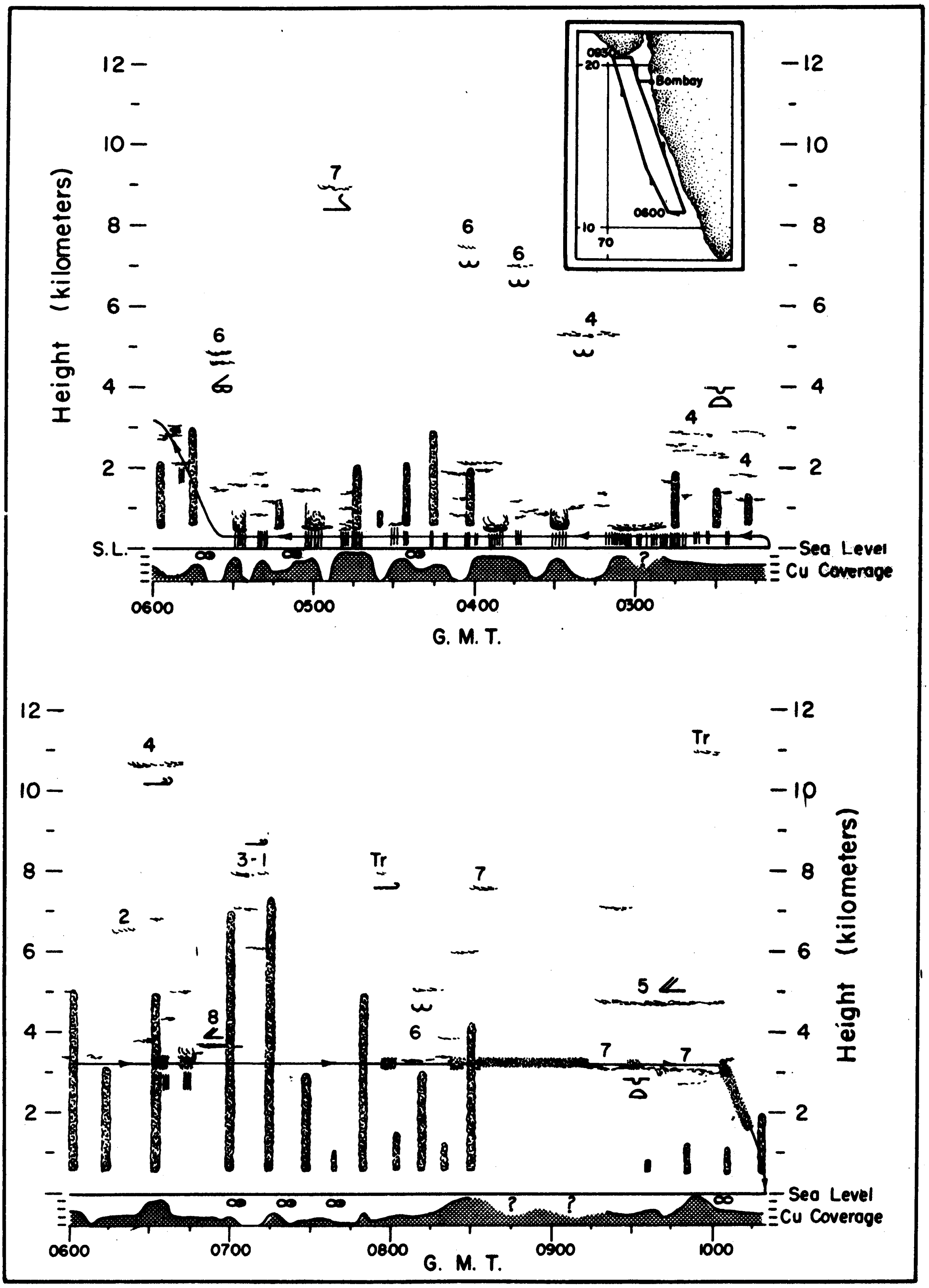
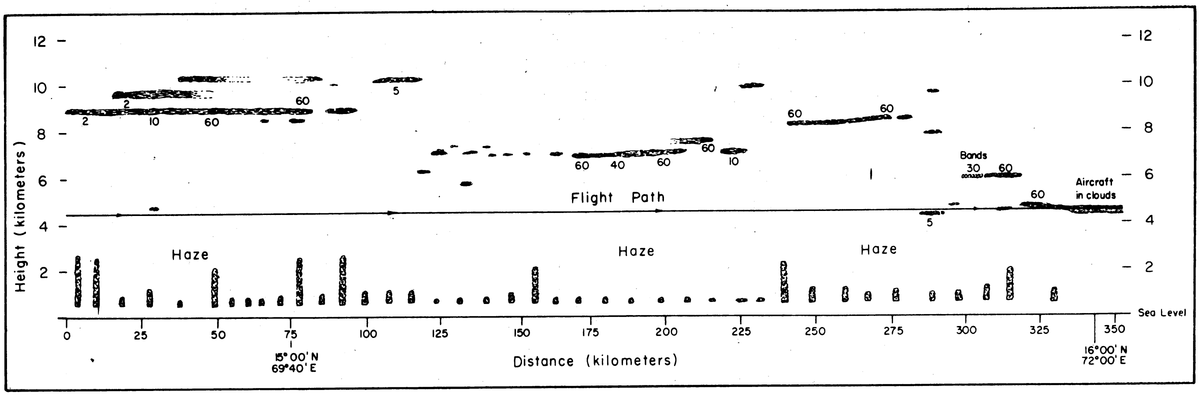
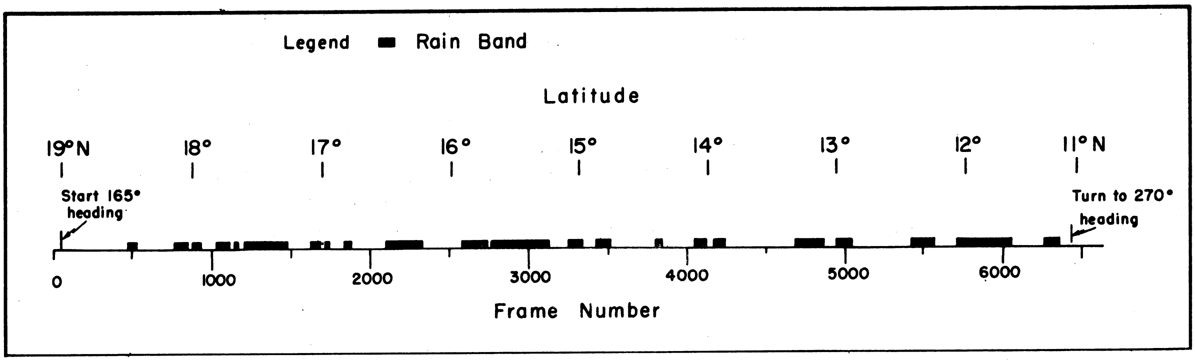


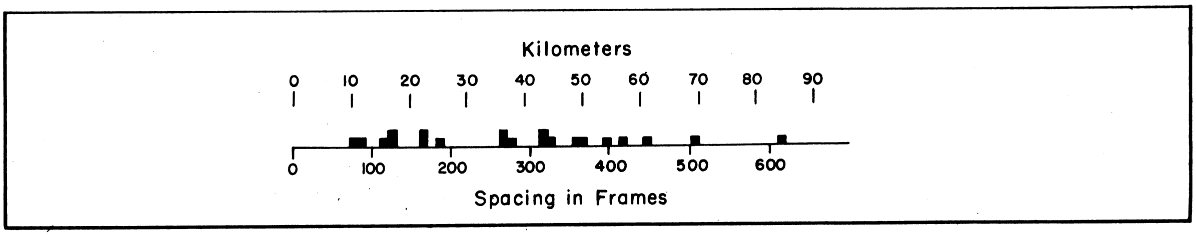
Fig. 12.8 Cloud cross section of 28 June 1963 WHOI flight along the west coast of India, (Bunker & Chaffee, 1969)



(A) Detailed cloud cross section of 26 June 1963 WHOI flight over the Arabian Sea. The distances that cloud layers extend from the aircraft toward 335 degrees is indicated in kilometres near the clouds. (Bunker and Chaffee , 1969).



(b) Occurrence of rain during the 28 June 1963 WHOI flight southward, 50 kilometers off the Indian west coast. Frame numbers of the 16 mm film are labeled along the abscissa. Solid bars indicate where rain was observed (Bunker and Chaffee. 1969).



(c) Frequency of spacing between centres of rain bands observed on 28 June 1963. The abscissa gives the distance between centres in frame counts and kilometers. The ordinate gives the number of occurrences of spacings with the indicated distances. (Bunker and Chaffee, 1969).

Fig.12.9

bands parallel to the wind direction. The bands extended downwind, upto the 20 to 30 km distance that could be seen in rain–free areas. The bands varied greatly in width, rain intensity and absorption of solar radiation. Short wave radiation under one band was 0.18 langley min^{-1} This large variation occurred, because in rain bands, clouds extended from a few hundred metres upto the cirrus level, while between the bands the skies were clear except for a few traces of cirrus. Zones bordering small rain bands (2 km to 10 km wide) were not clear but had varying amounts of cumulus and middle and high clouds. Twentytwo rain bands were found between 19^{o} N and 11^{o} N. Some of the wider rain bands may have enclosed non–precipitating zones. In the intense bands, it was so dark that it could not be determined whether rain was continuous. Rain bands amounted to 37% of the total area. The average width of the bands was 16 km with a standard deviation of 14 km, extremes being 46 and 2 km. The intermediate zone varied between 65 and 6 km, average being 23 km with a standard deviation of 17 km. No pattern could be detected in the spacing of the bands. The frequency of the spacing between rain bands showed bimodal distribution. Convergence fields existed on several scales that interacted with organised convection to amplify the intensity and width of activity in certain regions. From Doppler radar measurements made every five minutes during the flight, no systematic differences were apparent in wind direction or speed between rain and non–rain areas. The spacing of the rain band seems to correspond to the dimensions of off–shore vortices estimated by George (1956). He had no way of identifying the extent of the rain vortices at sea. All this is definitely much more information on the features of rain bands off the west coast than has been gathered regarding the mesoscale characteristics of rain cells over Indian land area. Still more information about the growth, decay and movement of these rain bands is worth collecting.

12.6.7 Root mean square values of the fluctuations of the vertical and horizontal components of the turbulent wind and of temperature and humidity are given in the Table 12.7. The observing period was one minute and values were available each fifth of a second. In the cumulus congestus observed on 22 August, 1964, the maximum updraft and downdraft were respectively 553 and 510 cm sec^{-1}.

Table 12.7
Turbulence Values

Date	**Time GMT.**	**Latitude**	**Longitude**	**Height (m)**	s **W cm/sec**	s **u cm/sec**	s **T ° C**	s **q g/Kg**	**Remarks**
9.8. 1964	1338	20.3 N	69.2 E	2990	80	33	.13	.19	As and Ac
18.8. 1964	0740	17. 4 N	70.0 E	4835	20	24	.04	.19	Between Stratus
22.8. 1964	1139	11.9 N	74.2 E	2085	174	127	.28	.45	Cu Congestus

12.7 Ramage (1963) put forward that the monsoon rain would appear to be associated with low level divergence, whereas the shower type of precipitation typical of fair weather regions within the monsoon may be associated with low level convergence. In his model of monsoon rains, convergence is mainly near 500mb, air rising from there upto about 300 mb. Below 600 mb there is descent of air along with falling rain. The very heavy, rain in monsoon depression over extensive areas is unlikely if the ascent of air commences from about 600 mb, without excessive values of vertical velocities, Bhattacharjee and De (1965) report that the tops of echoes in the cyclonic storm of 21st September, 1962 as observed on the radar was only about 5 km.

12.8 Narayanan (1967) has reported on three radar echoes of rain squalls of the monsoon season over Bombay of 6–8 km height. Fig. 12.10 shows the sequence of two squall lines. The

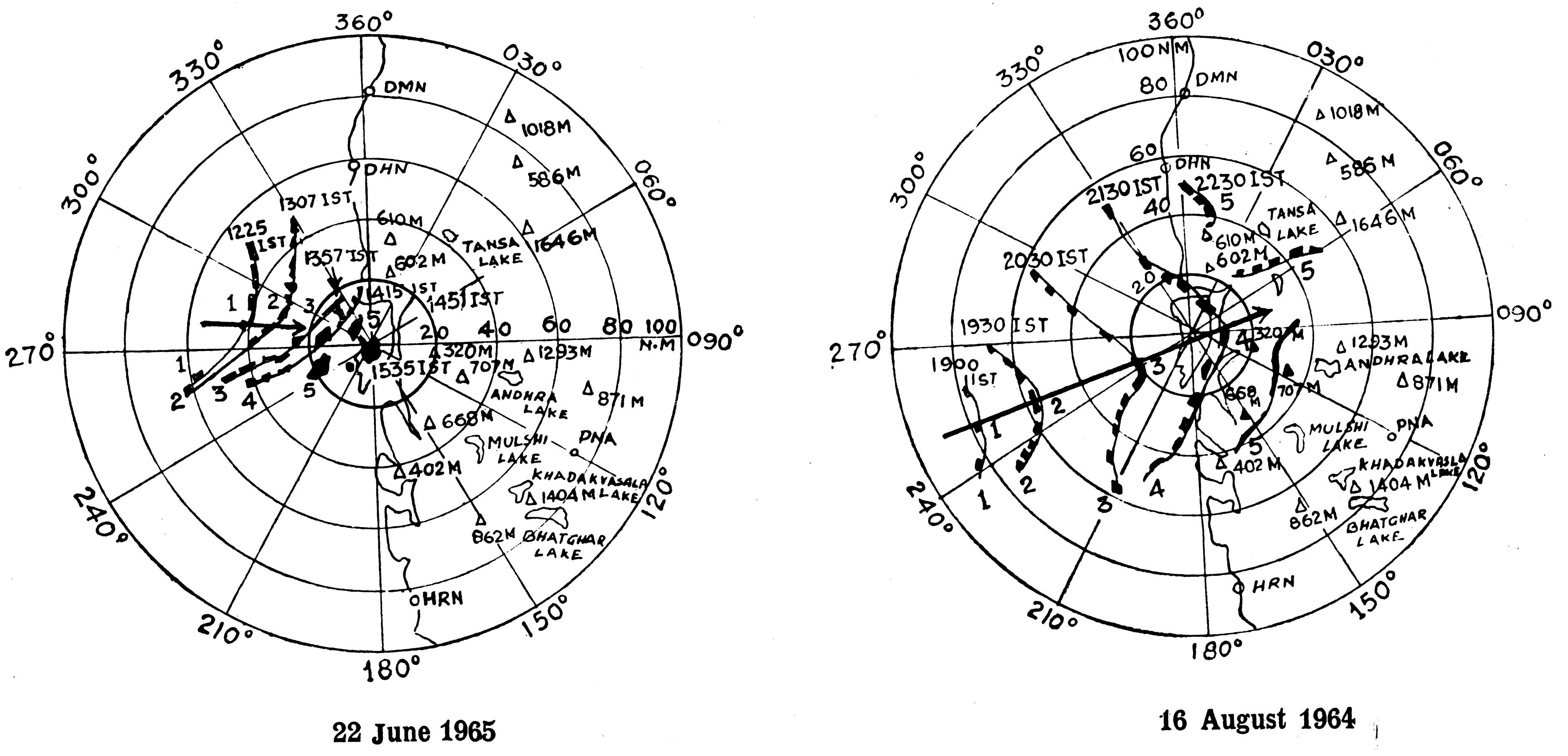


Fig.12.10 - 100 n.m. polar diagram showing the movement, pivotal motion and splitting of the rainfall echo on reaching closer to the Bombay coast

movement (with speeds ranging from 13 to 30 kt.) agreed with the vector mean wind in the first 6 km. The echoes moved faster over the sea during the developing stage but slowed down to nearly half the original speed in the dissipating stage on approaching the coast. One end of the line moved faster than the other, showing pivotal motion. They seemed to split into two or three parts which moved in different directions close to coast or over land.

12.9 Narayanan (1970) finds radar line echoes approaching Trivandrum, 60–120 km in length and 10–20 km in width. They are seen from 100 km off the coast and after moving inland break up into individual cells. Two to three bands about 100 km long are generally noticed at the same time with a spacing of 15 km. Echoes move at 30–40 kmph.

12.10 Monsoon echoes observed over Poona with 9.1 cm radar by Mani and Venkiteshwaran (l96la, 196lb) and Gupta, Mani and Venkiteshwaran (1955) have been of two types. One, associated with rain cells of comparatively short duration of less them an hour and the other associated with steady rain lasting four or five hours. The first type generally appears well below the freezing level and comes from west over the station and rapidly disappears to the east. The second type observed during steady continuous rain shows 'bright band' throughout the period of precipitation.

12.11 Billa and Hem Raj (1967) studied the movement of five rainfall patches over a triangular network at Poona and estimated that the horizontal extent varied between 78 and 276 km and the speed of movement between 9 and 23 kt. Two patches moved to east and the rest to west, in agreement with winds at 850 mb.

12.12 Kundu and De (1967) report that formation of line type echoes at Agartala is maximum in monsoon, 124 in one season. On average, the length was 100 km (2 % of over 200 km) and width 10 to 20 km. Most frequent movement was towards east and least frequent towards a direction between southeast to west. Average speed was 10 knots. Movement was best correlated with 3 km winds. Line echoes had a life of 3 to 4 hours, sometimes even 7 to 12 hours, 1730–2030 I ST and 0530–0830 IST were the times of least frequent occurrence while 1230–1730 I ST and 2030–0530 IST were most opportune times. Height of tops ranged from 3 to 9 km and only 10 % aloft. Secondary echo formation was either in front or behind rather than late rally, more in front.

12.13 Mukherjee, Arunachalam and Rakshit (1964) report that at Gauhati Cb cells were seen on all days on radar. Except in vigorous monsoon Cb cells start growing over the southern hills, sometimes around 1000 IST, reach a maximum by 1400 hours and decay before evening. This appears to be due to anabatic winds. Movement was from south or west.

12.14 Ghosh (1967) studied the radar echoes at Delhi from June to September 1959, The mean height of tops in thunderstorms was 13 km (highest 17 km) and other convective clouds 7 km (highest 12 km). Many clouds of the latter type grew above 0° C isotherm level. He concludes that when echoes have tops above the level of –25° C, thunderstorm occurs with 85% probability.

Seshadri (1963) has tabulated the distribution of heights of tops of convective clouds as observed on radar for a two year period (Table 12.8 and 12.9).

Table 12.8
Distribution of tops during different times over Delhi

Ht. Kft.	00–06	IST	06–12	IST	12–18	IST	18–24	IST
	No.	%	No.	%	No.		No.	
Upto 20	44	44.0	21	20. .9	13	10.6	42	36.2
21–30	45	40.5	55	53. .8	43	35.0	45	39.2
31–40	15	13.7	26	25. .3	53	43.0	24	20.4
= 41	3	3.6	Nil		14	11.4	5	4.3
Total	**107**		**103**		**123**		**116**	

Table 12.9
Frequency of maximum heights of Cb over Delhi

Height (Kft)	No. of occasions	%
11–20	17	11.1
21–30	40	26.8
31–40	70	44.5
41–50	27	17.6

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CHAPTER 13

BALANCE OF MASS, RADIATION, ANGULAR MOMENTUM ETC

13.1 General circulation studies have shown the tropics as source region of heat and angular momentum in maintaining global balance. This mean pattern of tropics is completely disturbed by the monsoon perturbation. In its areal extent and intensity the southwest monsoon is on a much bigger scale than anything else on the globe. The monsoon area around India is at least 40^o in longitude and 30^o in latitude, besides the areas further east upto Japan and coast of Siberia. The extensive thick clouds reduce the solar radiation reaching the surface of the land and sea, so that the role of this tropical belt as a source of heat is reduced. Considerable mass exchange across the equator is an additional factor in the balance of all parameters. Usually trade wind areas have only a shallow moist layer overlain with dry air aloft. The intense convection in the southwest monsoon region carries moisture into the mid– and upper troposphere and rate of transport may also be high. The usual northeasterlies at the surface, of the tropics, are replaced by somewhat stronger westerlies. In this area, a source region of westerly angular momentum becomes a sink. Hence the balance of various para meters in the monsoon region is worth study but only some aspects have received attention.

13.2 Mass Balance

13.2.1 Mass balance of overlying atmosphere is always necessary within two or three percent as known from pressure variations. The southwest monsoon area is not a closed system with more or less the same air circulating over and over. No well–defined boundaries can be fixed on any physical basis for this monsoon area. All that is relevant is to identify the inflow and outflow at different levels into the Indian area, their connections particularly through vertical motions and role on maintaining balance of other parameters.

13.2.2 Rao (l96l) first discussed from the upper wind data of Nairobi and Singapore, the mass flow across equator. He brought out that to compensate for the northward flow in the lower troposphere, there was marked flow into the southern hemisphere in the upper troposphere. Findlater (1969) made a very important contribution to the monsoon dynamics by identifying the massive flow across equator in a narrow belt between 35^o E and 60^o E longitudes. Upper tropospheric flow into the southern hemisphere is more uniform than the flow into the northern hemisphere in the lower troposphere, which is interrupted in the central longitudes on account of the clockwise circulation just south of the equator near 70^o E which may also be responsible for increased flow further west. This kind of flow with southerlies in lower troposphere and northerlies in upper troposphere extends generally over the Indian area upto about 25^o N. This may be called monsoon meridional circulation. It exists in longitudes outside this monsoon area also but only within 10° from the equator. This circulation would be favourable for ascent of air in the northern half (north of about 10 N) and descent in the lower latitudes, even on equator. Only when averaged over the whole globe, meridional flow at one level has to be compensated by reverse flow at some other level. But apparently as the monsoon region is fairly large, at least partial compensation seems to be maintained within the area by opposite meridional flow between lower and upper tropospheres. For mass balance over a limited area, it is divergence of zonal and meridional winds that is important and not the actual flow.

13.2.3 Koteswaram (1958) proposed a meridional circulation of a monsoon cell between equator and monsoon trough and the usual Hadley cell to the north. In that he represents the cross–equatorial flow only as skin–deep. Except in a small segment near 70^o E along the equator, there is flow from the southern hemisphere at least upto 700 mb. Wind data show that the monsoon meridional circulation extends across the equator both in lower and upper troposphere much more than what has been indicated in Koteswaram's model. The air circulating into the monsoon cell is not subsided air from between 0^o and 10^o N but mainly from the winter hemisphere.

13.2.4 Asnani (1973) pictures a separate equatorial cell between the two Hadley cells of the hemispheres. This equatorial cell is further divided into sub– cells, with subsidence near 10° N and again south of equator. The meridional circulations are not closed systems when considered over any longitudinal belt and vertical motions would not only be controlled by the gradients of meridional motion but also by vergence in zonal motion. In order to explain differences in weather in any latitudinal belt from one meridion to another, it is not necessary to picture variations in meridional circulations.

13.2.5 Raman, et al (1965) have presented cross–sections along 55° E, 75° E, 90° E of the mean meridional wind components in July (Fig. 13.l). The monsoon cell with southerlies in the lower levels and northerlies above 400 mb is along 90° E. On the other hand, at 55° E the Hadley cell is noticed from 35° to 5° N. South of 25° N or so, this Hadley cell starts from mid–troposphere, as the monsoon cell occupies the lower troposphere. South of 5° N, southerlies are present in the entire troposphere. 75° N is regarded as the transition between the two types, as the meridional components are on the whole weak with a shallow monsoon cell and a marked flow from north in upper troposphere to the south of 10° N. These confirm the earlier findings of Rao (1961, 1962). On a strong monsoon day of 7th July 1963 (Fig. 13.2), along 75° E Raman et al (1965) find the monsoon cell fills the whole of the troposphere upto 15° N and is more marked. Further north, the lower tropospheric monsoon cell is a little more deeper than in the mean and the Hadley cell is also more intense. With weak monsoon (Fig. 13.3) on 19th July 1963, the monsoon cell is only to the south of 5° N, while to the north of 20° N, the northerlies of the Hadley cell fill from 850 to 100 mb. They conclude that in strong monsoon, the monsoon cell present in the mean over the Bay, extends westwards to 75° E.

13.2.6 An east–west circulation (i.e. in zonal planes) has also been visualised in the monsoon. There is considerable convection in longitudes to the east of 70° E. Subsidence occurs to west of that longitude at least in upper and mid–troposphere in the regime of the western sub–tropical anti–cyclone. Some workers seem to regard the lower tropospheric westerlies as ascending and forming the upper easterlies which subside into the lower tropospheric westerlies further west as a semi–closed cell. But the origin of most of the upper tropospheric easterlies is not likely to be from ascent of lower tropospheric westerlies as these easterlies are present during weak monsoon or 'break' also.

13.2.7 Krishnamurthi (l97l) shows the existence of three large–scale circulations in northern summer as in Fig. 13.4. Over the Pacific and Atlantic Oceans, northeast trades converge near the ITCZ and the ascending air moves north poleward ahead of mid–oceanic upper level troughs and descend off the west coast of continents (Africa and southwestern United States). The vertical circulation over south east Asia is opposite, in the sense that the lower level southwest monsoon circulation is shown to ascend near the Himalayas and return equatorward in flows around Tibetan High and the easterly jet. A representation of these circulations in x, p – plane would constitute 'Walker Circulations'. These circulations are thermally direct.

13.2.8 As pointed out by Rao and Desai (1973), the monsoon is not a closed cell, in the sense of the Hadley cell. The limits of the monsoon circulation cannot be defined longitudinally and attitudinally to demonstrate a closed circulation. Hence a monsoon circulation cannot be postulated as if air rising in one part sink ing in another part to an equal extent. Reversal of flow in zonal and meridional directions from lower to upper troposphere has led some to suggest closed circulations. The inflow into the monsoon area and outflow may be linked through the general circulation in other areas. The role of the monsoon would be to distribute the excess heat gained, particularly in the continental areas, to other regions of the northern hemisphere and also the southern hemisphere.

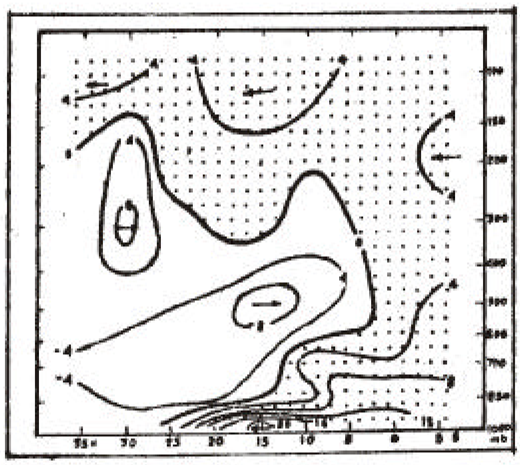


Fig.l3.l (a) Meridional component (kt) 55°E Mean July.

Fig:.13.l (b) Meridional component (kt) 75°E Mean July.

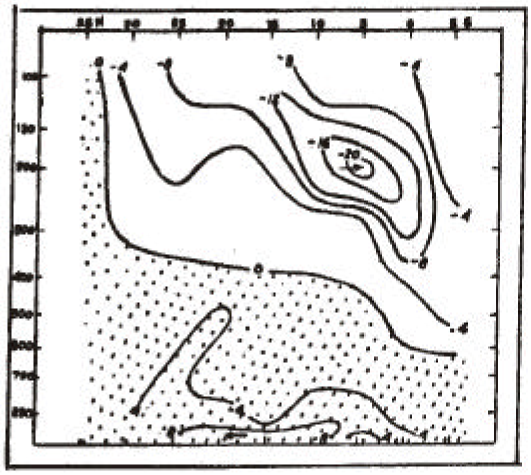


Fig.l3.l(c) Meridional component (kt) 90°E Mean July.

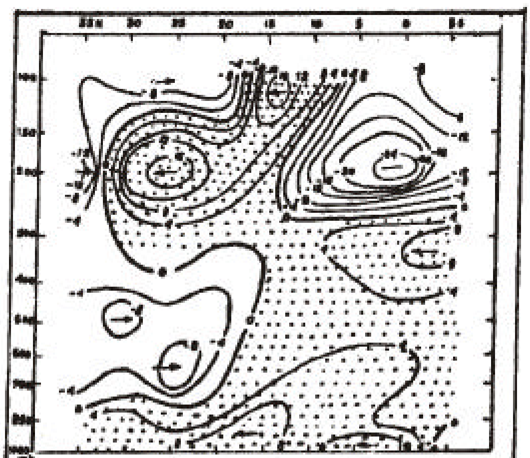


Fig.13.2 Meridional component (kt) 75°E Strong monsoon - 7 July, 1963.

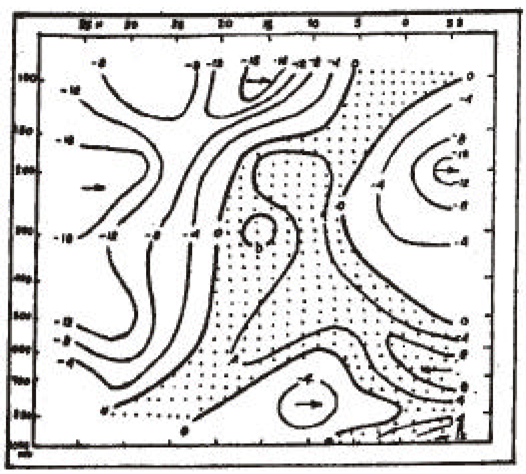


Fig.13.3 Meridional component (kt) 75°E Weak monsoon - 19 July 1963.

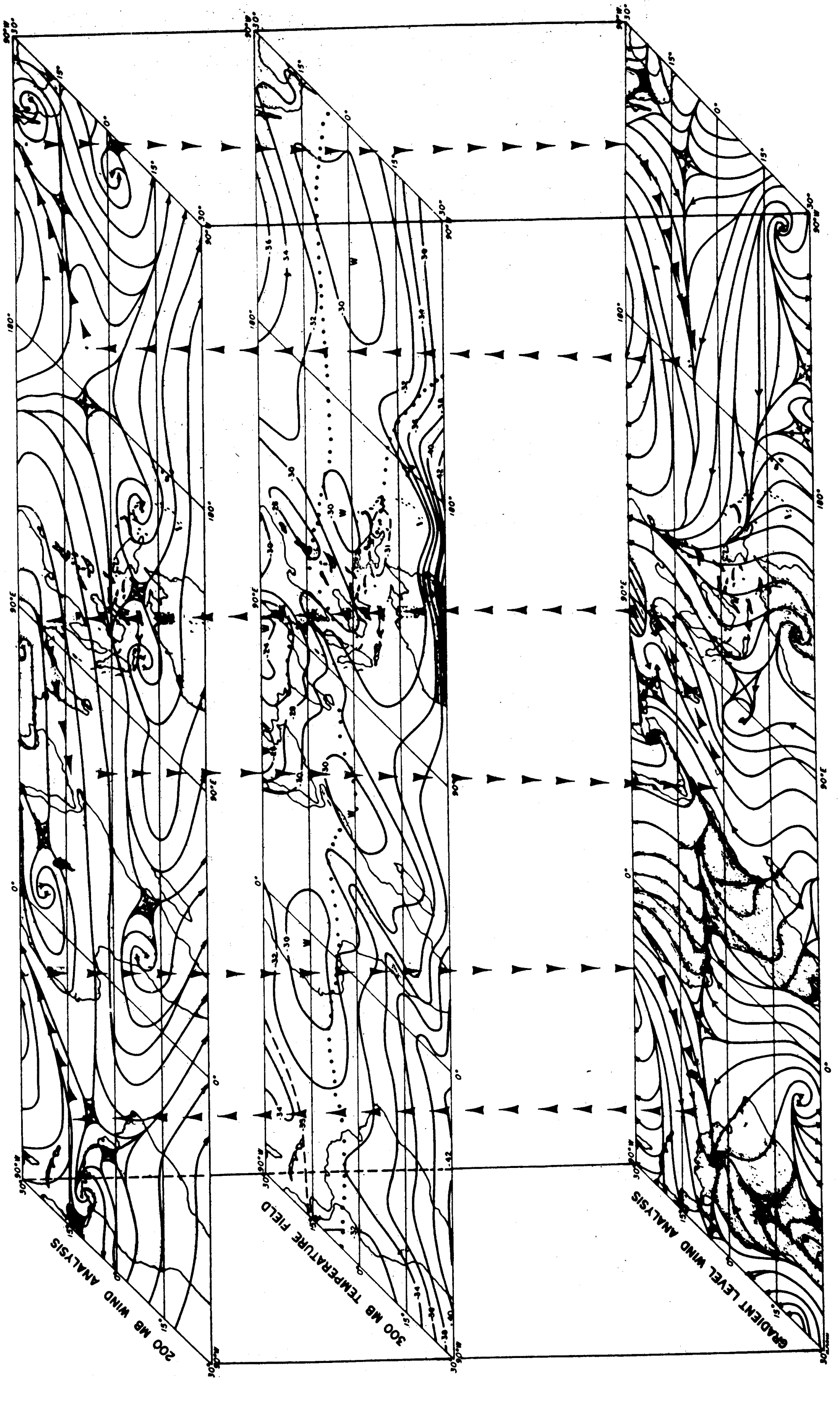


Fig.13.4 A schematic model of typical circulation features at the 200 mb and the gradient wind level (2000 feet). The 300 mb temperatures (middle panel) indicate a zonally asymmetric pattern. Zonally asymmetric vertical circulations are indicated by arrows. (Krishnamurti, 1971).

13.3 Radiation

13.3.1 Ganesan (1970) combined measurement of global solar radiation at ten pyranometer stations over India with the greater amount of sunshine data and mapped the global radiation received at the surface in July. He computed the outgoing long wave radiation from air temperature and vapour pressure at surface and sunshine data. Assuming suitable values of albedo, the net radiation map was constructed. The maps for July are given in Fig. 13.5. Both global incoming and outgoing radiations are reduced, due to cloudiness and water vapour content of the atmosphere so that net radiation gain at surface is still round about 300 langley day^{-1}. This is about 20 % lower than in the hot month of May when clouds are much less. Much of this net radiation would be used in evaporation from wet ground.

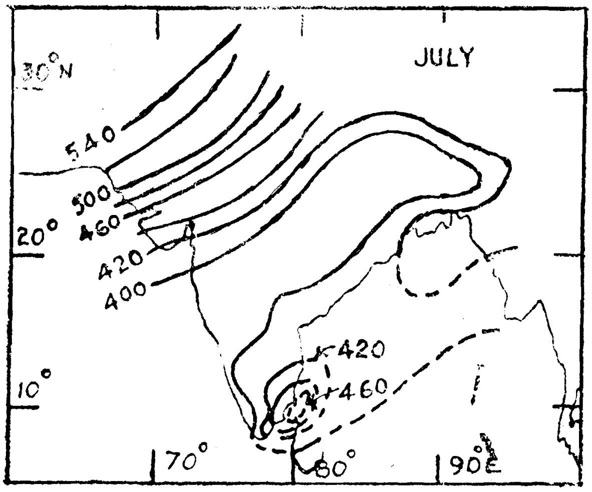
13.3.2 Godbole and Kelkar (1969) have computed the net infra–red radiative heat flux for July 1962, using the mean upper air temperatures and water vapour at levels upto 200 mb, using radiation tables of Elsasser and Culbertson. Fig. 13.6 shows the distribution of net radiative heat flux at surface taking into account clouds. There is broad agreement between Figs. 13.5 and 13.6, calculated by different methods. Northwest India with high surface temperature, small cloud amount and low moisture content shows relatively more radiation loss. The minimum is along west coast with low temperature, high moisture content and large cloud cover. Their diagram (Fig. 13.7) of radiative cooling between surface and 300 mb is interesting in bringing out that the monsoon effect of high moisture content and large amount of deep clouds reduces the heat loss by radiation, making up for the high albedo on account of clouds which reduces the absorption of solar radiation in the earth–troposphere system.

13.3.3 In the Meteorological Atlas of the IIO E, Vol. I, Ramage, Miller and Charmian Jefferies have presented maps of the net radiation gain at sea surface and the heat loss as latent and sensible heat, calculated from various ocean and atmosphere parameters. In the Bay of Bengal and the Arabian Sea, the net radiation gain at sea surface is 250 to 300 langley day^{-1}, increasing to 400 langley day^{-1} near Arabia Coast, generally agreeing with the computation over India. In the central parts of the Arabian Sea, the heat loss as latent heat from the sea surface is 30 % more than the radiation gained. Even in most parts of the Bay, the sea loses more heat by evaporation than the net gain of radiation. The large evaporation over the Arabian Sea and the Bay in the monsoon season is due to the strong winds. In the Indian Ocean at 15^{o} S, 65^{o} E, the radiation gain is only 150 langley day^{-1}, while the heat loss is as much as 500 langley day^{-1}. As noted in the Atlas, the amount of solar radiation reaching the sea surface depends upon the Sun's altitude and cloudiness, whereas the amount of heat abstracted from the sea by the air is directly related to the wind speed and air–sea temperature differences in the case of sensible heat, and to the wind speed and the vertical vapour pressure gradient in the case of latent heat.

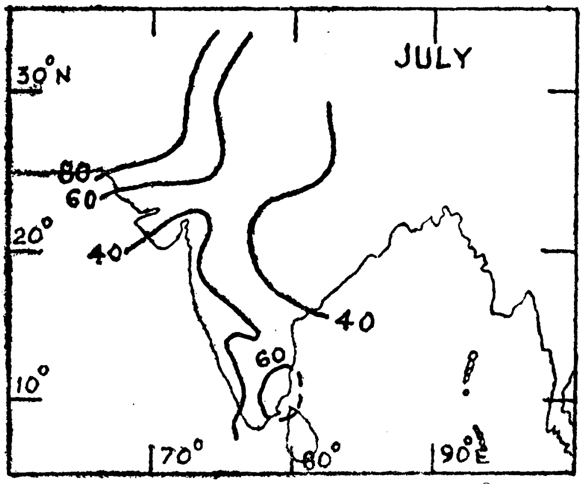
13.3.4 Bryson (1967) ascribes an important role to the dust in the radiation balance over northwest India. Das (1962) found that to make up for the warming of air by subsidence over northwest India, cooling at 2.4° C day^{-1} would be required to maintain a steady circulation. Considering water vapour and CO_2 as radiators, 1.8° C day^{-1} could be accounted for. Bryson et al (1969) found from radiation measurement over northwest India that the observed cooling was 2.4° C day^{-1} and that due to radiative divergence was 1.6° C day^{-1}. They considered that the additional cool ing could be due to dust over the arid regions of northwest India.

13.4 Heat Transport

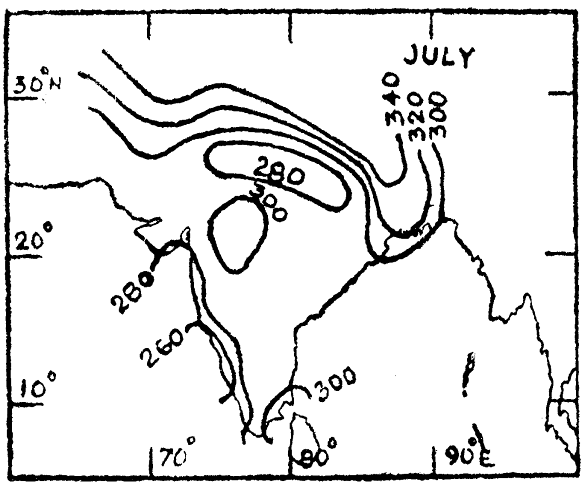
13.4.1 Anjaneyulu (1969) has discussed the flux of (CpT + Agz + Lq) in an ellipse of area 19.6 x 10^{15} cm^{2} and lateral boundary 6.6 x 10^{8} cm, around the monsoon trough (Fig. 13.8). The area had inflow upto 2.1 kt below 600 mb and outflow upto 2.4 kt between 500 mb and 100 mb. Table 13.1 below summarises the fluxes, eddy transport being insignificant. In the mean, the column of atmosphere above a square centimetre in this area is gaining about 8000 cal day^{-1} in the lower troposphere and losing the same amount aloft by advective processes. This may be comparec with the net radiation gain of about 500 langley day^{-1} at surface in the area. This corroborates that radiational processes are one order of magnitude less important.



(a) Global radiation in Cal/cm^2 /day



(b) Out Going radiation in Cal/cm^2 /day



(c) Net radiation in Cal/cm^2 /day

Fig. 13.5

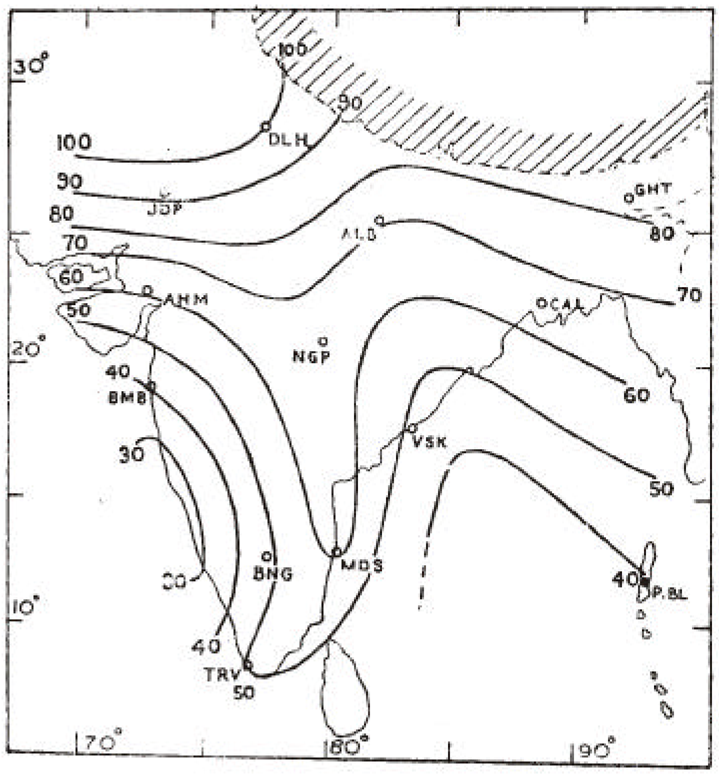


Fig.13.6 - Distribution of net radiative heat flux (with clouds) at the surface. Mean for July 1962. Units : 10^{-3} ly per min.

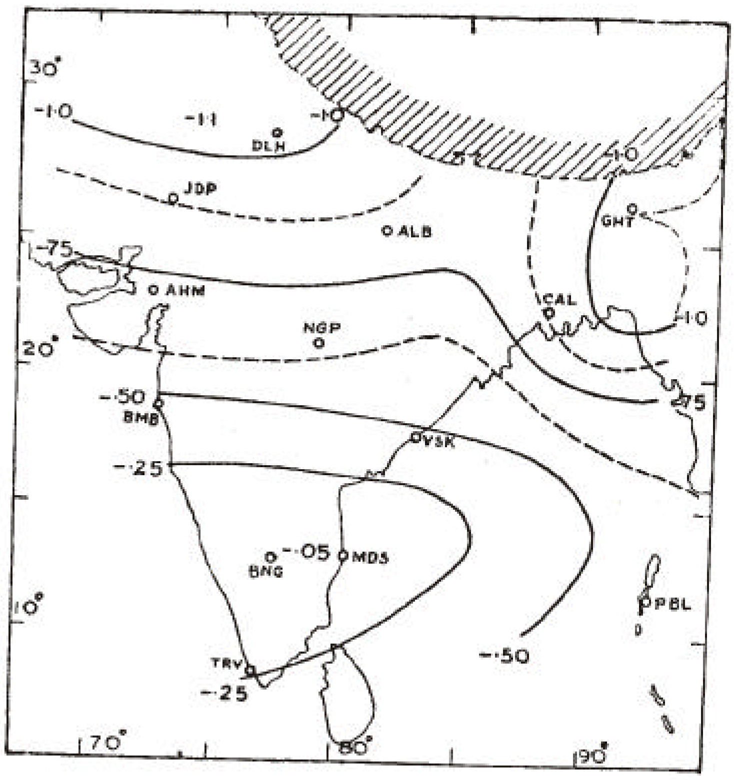


Fig. 13.7 - Distribution of radiative cooling in a layer from the surface to 300 mb. Units : °C per day.

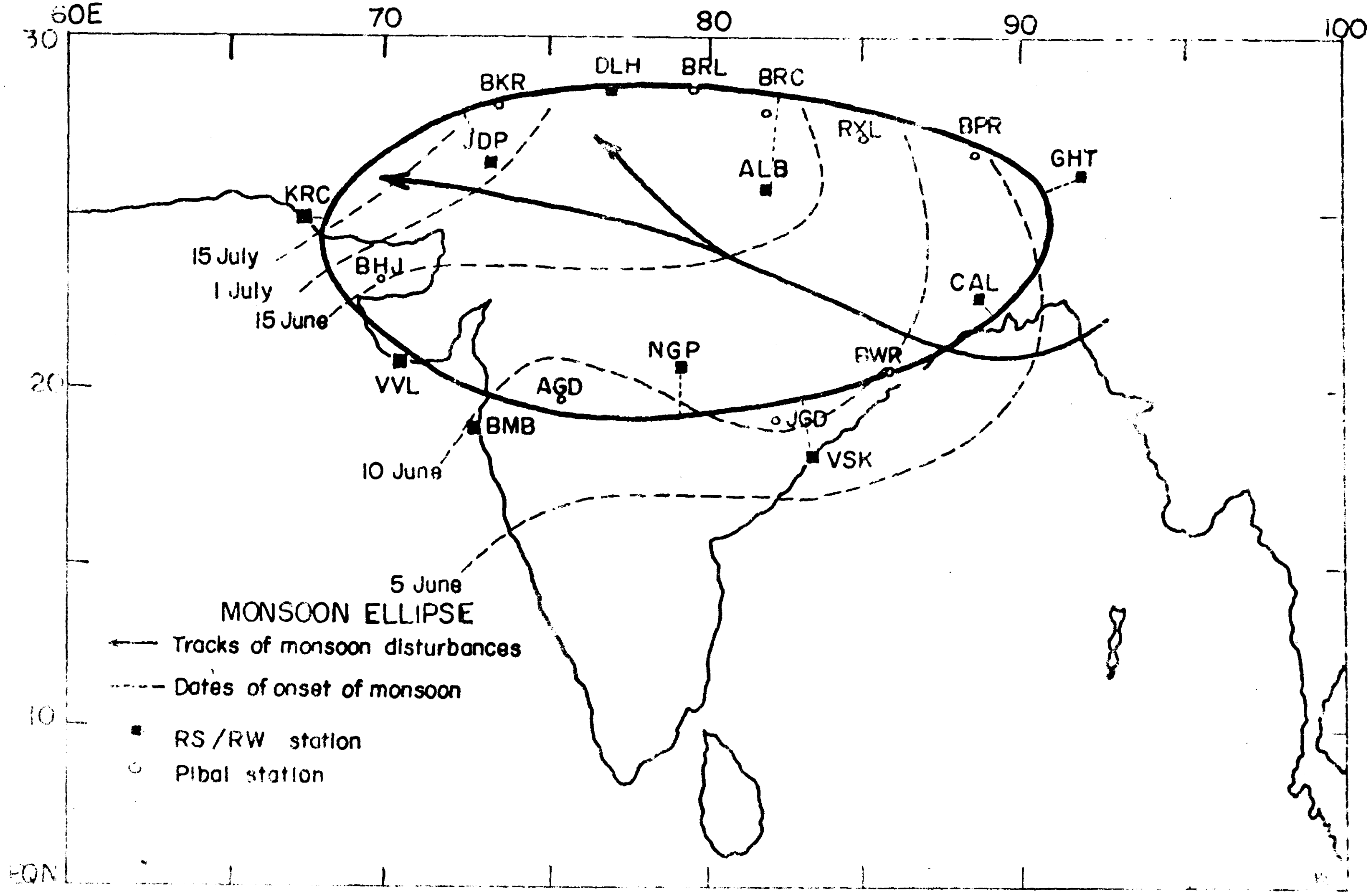


Fig 13.8 - Monsoon ellipse with aerological stations shown around it. Mean tracks of monsoon disturbances (thick lines) and normal dates of onset of monsoon (broken, lines) shown. (Anjaneyulu, 1969).

Table 13.1

Lateral flux into monsoon ellipse (positive outward)

Pressure (l00's mb)	Normal component of motion averaged around the boundary (knots)	Lateral mass flux 10^{16} gm day^{-1}	Flux of Lq (Unit: 10^{18} cal day)	Flux of CpT+Agz+Lq (Unit: 10^{18} cal day^{-1})
Surface – 9	– 2.1	– 0.63	– 7.0	– 53.9
9–8	– 1.9	– 0.57	– 5.1	– 49.9
8–7	– 1.6	– 0.48	– 3.1	– 40.0
7–6	0.4	– 0.12	– 0.8	– 10.1
6 – 5	0	0	– 0.2	– 0.4
5–4	0.3	0.09	+ 0.1	+ 6.9
4–3	1.2	0.35		+ 24.5
3–2	2.2	0.66		+ 54.6
2 – 1.5	2.2	0.34		+ 28.9
1.5 – 1.0	2.4	0.36		+ 32.1
Surface – 5	– 6.0	–1.8	–16.2	–154.3
5–1	+ 6.0	1.8	+ 0.1	+147.0
Surface – 1	0	0	–16.1	– 7.3

than dynamical processes in thermal balance, for short periods. The net heat inflow in the lower troposphere is transported into the upper troposphere by mean vertical motion. Vertical velocity of 1.5 cm sec^{-1} at 500 mb caused by inflow below is sufficient to carry about 150 x 10^{18} cal day^{-1} into upper troposphere for export. The role of moisture flux is one order of magnitude lower than enthalpy and potential energy, The computations of Anjaneyulu clearly demonstrate the role of the monsoon trough in the mean picture as a source of heat energy, feeding the upper troposphere through convection for export.

13.4.2 Bunker (1965) has studied the heat balance and flux of sensible and latent heat along the trajectory of air in the low level jet over the Arabian Sea from 1000 mb upto 600 mb. He used the equations

$$\Delta\sum.q_E = \Delta t\left[SH_B - SH_T + LH_B - LH_T - IRE + SA\right]$$

$$\Delta\sum.q \ = \Delta t\left[SH_B - SH_T + SA - IRE\right] + L(PPT)$$

$$L(PPT) = a \quad LH_B \Delta t$$

$\Delta\sum$, q_E and $\Delta\sum$ are the differences in potential plus internal plus latent heat content and potential plus internal heat content of the layer at the beginning and end of the passage across the region in a transit time Δt . SH and LH are the vertical transports of sensible and latent heat by all processes. Subscripts 'T' and 'B' refer respectively to the top and bottom surfaces of the layer. IR E is the rate of heat exchange from the layer by infrared radiation. L (PPT) is the heat released by condensation of water vapour, SA is the rate of heat absorption from solar radiation. a is the fraction of latent heat flux converted to sensible heat by condensation and is estimated from the amount of water vapour condensed in ascent of saturated air. The equations were solved for SH and LH assuming that no heat was transported through 600 mb level except by radiation and there was no water vapour transport at this level.

The fluxes of sensible and latent heat in the western (west of 60° E) and eastern Arabian Sea computed by Bunker are given in Table 13.2.

Table 13.2
Sensible and latent heat fluxes (vertical)
(Unit: Cal cm^{-1} sec^{-1})

Level (mb)	Western Region		Eastern Region	
	Sensible	Latent	Sensible	Latent
600	Assumed aero			
700	–0.15	+0.24	+0.20	+0.34
850	+0.23	+1.9	+0.32	+0.87
1000	+1.5	+4.8	+0.44	+2.8

Latent heat fluxes are upward at all levels in both regions, The sensible heat fluxes are all much smaller and at 700 mb in the west there is a small downward flux. The latent heat fluxes at 1000 mb level are equivalent to evaporation rates of 7.0 and 4.1 mm day^{-1} respectively in the western and eastern parts. These figures would have been higher if moisture transport through 600 mb had been taken into account which would have been the case in the eastern region. These computations bring out that monsoon air mass while passing over the Arabian Sea is acquiring considerable moisture and some sensible heat.

The heat budget of the air in the two parts of the Arabian Sea according to Bunker (1965) are given in Tables 13–3 and 13.4

Table 13.3
Heat Budget West of 60° E (Unit: $Calories^{-1}$)

Layer (mb)	**Sensible heat flux**	**Latent heat flux**	**Infrared radiation absorption**	**Solar radiation absorption**	**Net accumulation**
600 – 700	–13	21	– 5	1	4
700 – 850	33	144	– 53	20	144
850 –1000	109	248	– 63	36	334
Total	129	413	–121	57	478

Table 13.4
Heat Budget East of 60° E
(Unit: Calories day^{-1})

Layer (mb)	**Sensible heat flux**	**Latent heat flux**	**Infrared radiation absorption**	**Solar radiation absorption**	**Net accumulation**	**L (PPT)**
600 – 700	17	29	– 33	12	25	3
700 – 850	10	46	– 45	24	35	23
850 –1000	10	167	– 47	34	164	24
Total	**37**	**242**	**–125**	**70**	**224**	**50**

Air accumulates heat nearly twice as fast west of 60° E as to the east. This is from the sea surface as the air is there much cooler and drier. The smaller radiation loss from the top layer of the western air is due to its dryness. The small loss from the eastern bottom is due to greater back radiation of the middle and high cloud cover. More short wave radiation is absorbed in the east because of its greater cloudiness and haziness, The computed heat release by condensation, L(PPT), corresponds to rainfall of 0.8 mm per day. The rainfall over the sea is from widely scattered areas of Cb clouds and may be expected to be about two orders of magnitude smaller than the monsoon rains along the west coast in the vicinity of hill ranges. The most important feature is the dominant role of the exchange from the sea surface.

13.5 Moisture Ba1ance

13.5.1 Pisharoty (1965) studied the moisture balance of the atmosphere over the Arabian Sea upto 450 mb, He computed the moisture inflow into an area bounded by equator and 26° N, and 75° E and 42° E. Using data of six upper air stations and some reports along equator.

$$\frac{1}{g}\oint\int_{1000mb}^{450mb} q.v.dl.dp$$

was evaluated, where 1 is the length along the boundary, q specific humidity and v the wind speed normal to the boundary. In July 1963, there was inflow across northern and southern boundaries and in next July inflow was also at western boundary, But the outflow– at the other boundaries caused a net divergence equal to 20% of the total transports (neglecting sign) through the four walls. Evaporation from the Arabian Sea (0.28 gm cm^{-2} day^{-1}) was estimated as only half the net flux divergence of water vapour. Sparse data, lack of knowledge of the strong flow across equator, later identified by Findlater, and whether required mass balance was ensured or not, were the drawbacks of the result.

13.5.2 Saha and Bavadekar (1973) recomputed the water vapour budget of the Arabian Sea, including evaporation, using more data and found evaporation was substantially more than the net flux divergence of water vapour. The mean rainfall over the area was computed from the excess of evaporation over the net flux divergence, which seems to agree with precipitation estimated from available rainfall charts. Their results are summarised in the Table 13.3

Table 13.5
Water vapour flux over Arabian Sea (10^{10} tons day^{-1})

		Across				**Net Flux divergence**	**Estimated evaporation**	**Precipitation computed from water vapour budget**	**Estimated precipitation from rainfall charts**
		42° E	**20o N**	**75° E**	**Equator**				
Sep.	1963	0.85	–1.09	–2.35	3.60	1.01	2.1	1.1	1.0
Jun,	1964	0.11	–1.21	–3.38	4. 99	0.51	3.5	3.0	2.5
Jul.	1964	0.50	–1.53	–4.36	5.87	0. 48	3.2	2.7	2.7
Aug.	1964	–0.67	–1.46	–3.89	5.75	–0.27	2.5	2.7	2.1
Sep,	1964	0.15	–1.27	–3.01	4.95	0.82	2.4	1.4	1.7

Net flux divergence of 1.2 x 10^{10} tons obtained by Pisharoty, for July 1964 may be compared with 0.48 x 10^{10} tons reported by Saha and Bavadekar Evaporation for the same month obtained by the two authors are respectively 1.2 x 10^{10} and 3.2 x 10^{10} ton day^{-1} Water vapour flux etc of 1 x 10^{10} tons day^{-1} over this area of 7.7 x 10^{6} km2 corresponds to an amount of . 13 cm cm^{-2} day^{-1}. The mean evaporation over a square centimetre per day of the Arabian Sea in July according to Pisharoty is 0.l6 cm, Saha 0.42 cm and Bunker in West Arabian Sea 0.7 cm and East Arabian Sea 0.41 cm. Precipitation rate for July according to Saha is .35 cm per sq, cm per day while Bunker's figure for the East Arabian Sea is .08 cm per sq cm per day.

Though over most oceans, evaporation is least in summer, over North Indian Ocean , one of the maxima is in the Southwest Monsoon (Venkiteshwaran, 1956 and Privett, 1959). It is a maximum in June and July and just begins to decrease in August in most areas. The strong winds of the monsoon are responsible for this feature.

13.5.3 Anjaneyulu (l969) computed that in July and August, 28 x 10^{15} gm day^{-1} water vapour accumulates in the monsoon ellipse mentioned in earlier para. This is equivalent to mean rainfall of 1.4 cm day^{-1}

13.6 Kinetic Energy Balance

13.6.1 Anjaneyulu (l97l) showed that the same .monsoon trough ellipse is an exporter of kinetic energy westward in the upper troposphere. Monthly means (July and August) of the aerological stations around the ellipse were used, so that the time variations were neglected Table 13.6 gives the lateral flux and production of kinetic energy as per his computations The ellipse had inflow in lower troposphere and slightly excess outflow aloft.

Table 13.6

Production and lateral flux of kinetic energy in monsoon ellipse

Pressure (100's mb)	Lateral Flux of KE (Export – positive : Unit: 1015 cal day –1)			Production of KE (Production – negative : Unit 1011 cal day–1)		
	Mean motion	Standing eddy motion	Net	Mean motion	Standing eddy motion	Net production
Surface – 9	– 0.21	– 3.03	– 3.24	– 21.79	+ 63.83	+ 42.04
9–8	–2.30	+ 0.23	– 2.07	+ 16.65	– 6.52	+ 10.13
8–7	–1.33	+ 2.70	+ 1.37	+ 23.88	+ 51.13	+ 75.01
7–6	– 0.19	– 0.56	– 0.75	– 5.60	– 14.27	– 19.87
6–5	0	– 0.23	– 0.23	0	+8.13	+ 8.13
5–4	+ 0.13	– 0.03	+ 0.10	– 6.63	+ 21.94	+ 21.31
4–3	+2.41	+ 1.48	+ 2.89	+ 8.54	– 23.67	– 15–13
3–2	+ 8.98	+ 8.30	+ 17.28	– 38.90	– 56.06	– 94.96
2–1.5	+11.24	+ 28.37	+ 40.21	– 52.13	–109.64	–161.77
1.5–1	+20.19	+ 59.12	+ 79.31	–105.61	–137.59	–243.20
Surface–5	– 4.23	– 0.84	– 5.07	+ 13.14	+102.30	+ 115.44
5–1	+42.95	+97.84	+140.79	–188.73	–305.02	–493.75
Surface–1	+38.72	+97.00	+ 135.72	–175.59	–202.72	–378.31

The equation for kinetic energy is

$$\frac{\partial}{\partial t}(r_k) + \nabla.rkV = -V.\nabla p + r.V.\overline{F} = -\nabla.Vp + p\nabla.V + rV.\overline{F}$$

where $K=½\ \vec{V}.\vec{V}$ and $\overline{F}$ is frictional force. Assuming a quasi–steady state in a closed volume, this becomes

$$\iint \frac{\overline{K}\overline{C_n}}{g} dsdp + \iint \frac{\overline{K'C'_n}}{g} dsdp = S\int \overline{C}_n (Z_a - \overline{Z})dp - S\int \overline{C'_n Z'} dp$$

$$+A\int \left[Z * (\nabla.\vec{V}) *\right]_a dp + D_s + D_T$$

Bar indicates mean at the boundary of the volume and prime deviations from that mean. C_n is the component of wind normal to the ellipse. Subscript 'a' indicates average value over the area and the asterisk the deviations from the areal average. S is perimeter of the lateral volume and Z the contour value. D_s represents the dissipation of kinetic energy due to surface friction and D_T the internal dissipation within the troposphere.

The first three terms on the right hand side represent production of kinetic energy which basically takes place when air parcels move across isobars. The first two are export of potential energy. The third term on the right hand side could not be evaluated. This represents vertical flux of potential energy and conversion of potential energy into kinetic energy by the physical process of rising warm air and sinking cold air. In the upper troposphere there is a weak flow from the higher pressure to north (sub–tropical ridge) towards lower pressure, which may be responsible at least partly for producing kinetic energy. Surprisingly there is no production of kinetic energy in the lower troposphere even in the frictional layer in the trough but only dissipation.

Substantial export of kinetic energy (the two terms on the left) takes place in the upper troposphere, especially above 200 mb, both on account of net mass export from the ellipse and standing eddies. The latter is perhaps the effect of easterlies (with a slight northerly component)being stronger to the south and southwest, The dissipation of kinetic energy due to surface friction $D_S = C_d\ r_0 V_0^3$ A is calculated as 69 x 10^{15} cal day^{-1}, taking $C_d = 5 \times 10^{-3}$. D_T the internal dissipation in the troposphere, is assumed as equal to D_S.

The kinetic energy balance is summarised in Table 13.7.

Table 13.7
Kinetic energy balance around the monsoon trough (Unit: 10^{15} cal day^{-1})

	Source	**Sinks**
Production	378.31	–
Export	–	135.72
Dissipation by friction	–	138.00
Imbalance	–	104.59

The study shows excess of production by about 28 %. Time eddies have not been taken into account. Monsoon depressions which increase kinetic energy in lower troposphere have been left out of picture. Export in upper troposphere may also be affected at such times. More than half the dissipation by friction is at surface and lower troposphere. There is also other dissipation in lower troposphere as seen in Table 13.6. As production is only in upper troposphere, some mechanism should exist to replenish the kinetic energy in the lower troposphere from the upper, which is unlikely to be through subsidence or this anomalous conclusion may have arisen on account of neglect of time eddies (monsoon depressions and lows) and vertical circulations.

13.6.2 Keshavamurthy (1968) has also discussed the energy conversion in the Indian monsoon. In extratropical atmosphere, the conversion of energy is as given below :

Radiation and other diabatic sources → Zonal available potential energy → Eddy available potential energy $\xrightarrow{\text{By baroclinic instability}}$ Eddy Kinetic energy $\xrightarrow{\text{By inverse of}}$ Zonal kinetic energy

Zonal available potential energy feeds zonal kinetic energy through the intermediaries of synoptic eddies. But in the case of the monsoon, the mean meridional circulation is strong enough to convert available potential energy into zonal kinetic energy, eddy mechanisms being at least one order of magnitude lower.. Palmen et al (1958) express kinetic energy released by the mean meridional circulation over a longitudinal belt l per unit length of the meridian as

$$\frac{lf\cos f}{g}\int \bar{v}\bar{v}_g\,dp$$

v_g geostrophic wind speed along x–direction Keshavamurthy's (1968) calculation of $\overline{u}\overline{v}$ is shown in fig 13.9. He assumed $\bar{u} \approx u_g$. The monsoon meridional cell extending up to about 25^o N, produce kinetic energy in upper troposphere and to a much very smaller extent mid–troposphere, but very little in lower troposphere where meridional flow is reversing in direction. Further to the north the meridional flow produces kinetic energy only above 100 mb and causes some dissipation between 400 and 100 mb. Production of kinetic energy in upper troposphere between 21^o and 30^o N the area considered by Anjaneyulu, is confirmed. But much greater production by the meridional motion is noticed further south, though geostrophic assumption may be introducing some errors. Fig. 13.10 shows the rate release of kinetic energy different latitudes by mean meridional motion. The total is 3 x 10^{10} Kw (or 73 x 10^{13} calories sec^{-1} or 6.2 x 10^{17} calories day^{-1}) in the area of 13 x 10^{16} Sq. cm between equator and 28^o N and between 50^o and 100^o E. This is about the same order as the production (3.3 x 10^{10} Kw) over an equal area of the winter Hadley cell (1958). The production of KE by this mean meridional from Palmen et al (1958), The production of K E by this mean meridional circulation works to 5 cal cm^{-2} day^{-1} and Anjaneyulu's by mean motion to 9 cal cm^{-2} day^{-1}. The decrease in the strength of the monsoon westerlies during break monsoon coincide with weakening of the meridional circulation. During strong monsoon the meridional ceil is stronger which is quite consistent with the production of kinetic energy by the mean meridional motion.

13.6.3 Keshavamurthy (1970) also studied the maintenance of kinetic energy in the area of the monsoon trough. Assuming quasi–steady state, transport of kinetic energy through boundaries should be balanced by production and dissipation. Production of kinetic energy comes out as flux of potential energy through the boundaries (including vertical) and conversion of available potential energy by the process of rising warm air and sinking of cold air. Keshavamurthy evaluated the last process and time eddies unlike Anjanevulu, but not fractional loss. Computations were made for the area from 20^o to 25° N and 50° to 95° E and upto 500 mb only, as this is the upper limit of the monsoon trough. His results are given in Table 13.8.

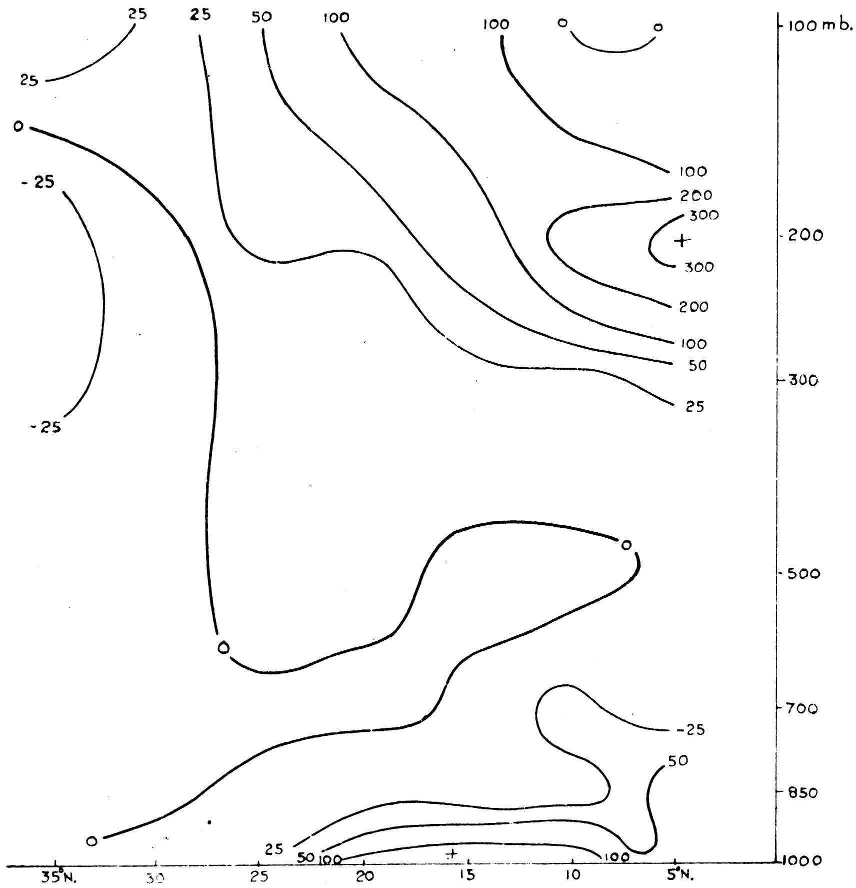


Fig.13.9 $[u]$ $[v]$ (kt^2) as a function of latitude and pressure. Square bracket indicates averaging between 50°E and 100°E. (Keshavamurthy, 1968).

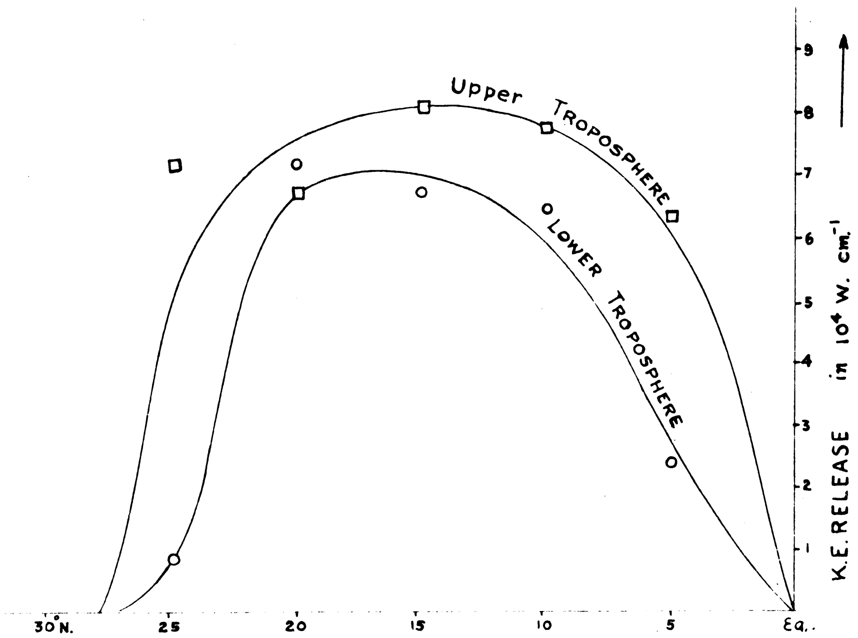


Fig.13.10 Kinetic energy released by the mean meridional circulation per unit length of latitude. (Keshavamurthy, 1968)

Table 13.8
Rate of production of Kinetic Energy in monsoon trough area (July–August) in lower troposphere (Unit : 10^{-3} cal day^{-1} cm^{-2})

Mean advec–tion	Transient advection	Production by horizontal pressure force	Horizontal flux of potential energy across sidewalls	Vertical flux of potential energy across top–bottom boundaries	Conversion of potential energy by rising of warm air and sinking of cold air
52	11	672	1680	5460	–4620

The mean advection and last four terms of the table have been computed from mean winds but the transient advection from data of 1963 and 1964. Production by horizontal pressure forces is the most important factor making up for dissipation of kinetic energy by friction. Least in magnitude are time eddies. Though small, the distribution of this term at 850 mb is interesting (Fig. 13.ll). It is negative over North Bay and Central India where depressions and lows draw kinetic energy from the mean motion but positive (feeding the standing eddy) in areas where they dissipate.

Vertical motion is consuming energy in the lower troposphere, meaning ascent of cold air and descent of warm air. This effect was not evaluated by Anjaneyulu who found excess of production of other types over dissipation. Computation seems to be from the mean motion which cannot be regarded as complete evaluation. The large vertical flux of potential energy into the region is mentioned as occurring at the bottom, which is explained as due to large orographically induced vertical velocities. This has also affected evaluation of the last term in the table. The upper troposphere in which Anjaneyulu found large production of kinetic energy is not included in this study. He found in lower troposphere, dissipation o£ kinetic energy by movement against pressure force neglecting ascent and descent of air, whereas Keshavamurthy finds production of kinetic energy by pressure force. The dissipation rate according to Anjaneyulu is 5 cal cm^{-2} day^{-1}, while production in Keshavamurthy;s study is 0.7 cal cm^{-2} day^{-1}

13.6.4 Krishnamurthy (l97l) finds at 200 mb, on global average, kinetic energy flux from June to August is directed towards the winter hemisphere across the equator.

13.7 Angular Momentum Balance

13.7.1 Rao (l96l) first pointed out that the Indian monsoon area becomes a sink for westerly angular momentum on account of surface westerlies, instead of being the usual source region with easterlies in the tropics. There was transport of angular momentum from across equator into this region in the upper troposphere.

13.7.2 Keshavamurthy (1968) made an analysis of angular momentum balance in the region bound by equator, 20° N and the longitudes of 50° E and 100° E. This is mostly an oceanic region with part of Indian Peninsula, Burma and Sumatra. The angular momentum lost by friction at the surface to the earth, is to be made up by transports through the walls, atmospheric pressure differences between the boundaries to the east and west and torque on mountains in the region due to difference in atmospheric pressure on their eastern and western faces. Through the southern and northern faces, transport of the earth's angular momentum by the mean meridional currents is separately evaluated and called Ω transport term. This is to be included as the air parcels show east–west motion in addition to part–taking in earth's notation and angular momentum on account of both the motions is to be conserved.

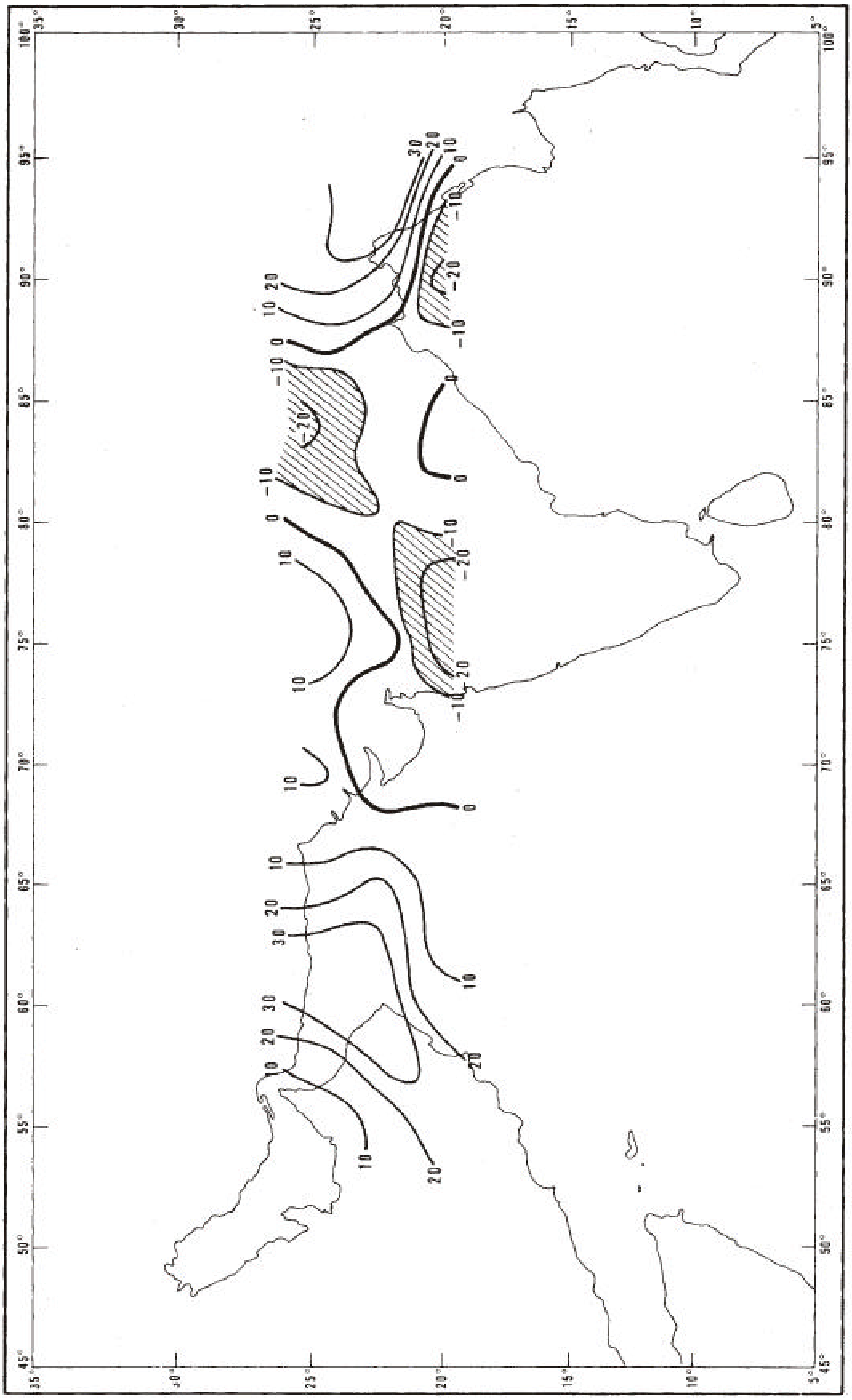


Fig.13.11 Contribution from the transient motion to standing kinetic energy (kt ^{2} deg $^{-1}$) at 850 mb. (Keshavamurthy and Awade, 1970).

Keshavamurthy's evaluation of the budget of angular momentum in the area mentioned earlier is given in Table 13.9.

Table 13.9
Budget of angular momentum (Westerly momentum July–August 1963–64)
Equator and 20° N, 50° E and 100° E
(Unit : 10^{25} gm cm2 sec–2)
(Gain within boundary or import–positive)

			Lower troposphere	Upper troposphere
Ω transport			6.8	–7.0
Frictional torque			–7.0	–
Zonal pressure gradient term			–0.4	–2.6
Mountain torque			2.1	–
Flux diver gence	East– West	Mean	–0.3	+5.0
		Eddy	–0.4	–5.0
	North– South	Mean	0.3	+4.8
		Eddy	–0.1	+0.5
Balance			1.0	+ 0.2

Interestingly, balance is more or less reached separately in lower and upper tropospheres. This is necessary as no significant transport is possible from one to the other through convection as winds are very weak in mid–troposphere. This is different from heat which has to be transported from the lower part of the troposphere to the upper. The northward meridional motion in the lower part of the monsoon cell circulation generates enough angular momentum through Ω transport to make up for the frictional torque, The mountain torque is rather large for ranges below 2 km. In the upper troposphere due to the higher speed of easterlies at the western and southern boundaries and also greater meridional component at equator, there is net flux convergence of westerly angular momentum (actually net divergence of easterly momentum). This is dissipated as Ω transport in the southward meridional flow of the upper arm of the monsoon cell. Lower atmospheric pressures on the west than on the east, along any latitude in upper troposphere, also cause dissipation of angular momentum.

Westerlies are known to develop just south of the equator in the season both at surface and lower troposphere. This cannot be accounted as due to conservation of angular momentum as the air is still moving equatorward. The study by Keshavamurthy brings out that while the mean westerly flow is determined by the pressure gradient, the frictional loss is made up by Ω transport or in other words by conservation of angular momentum.

13.7.3 Newton (l97l) found the transport of angular momentum across equator towards the summer hemisphere a maximum in June, July and August, and equal to 13 x 10^{25} gm cm^2 sec^{-2} . This is an average for the whole globe but the transequatorial flux of linear momentum in Northern Hemisphere summer is a maximum between 60^o and 120^o E (Fig. 13.12), Momentum flows toward the hemisphere in which the rising branch of the Hadley circulation (of winter hemisphere) is located, most prominently in the longitudes of the strong monsoons.

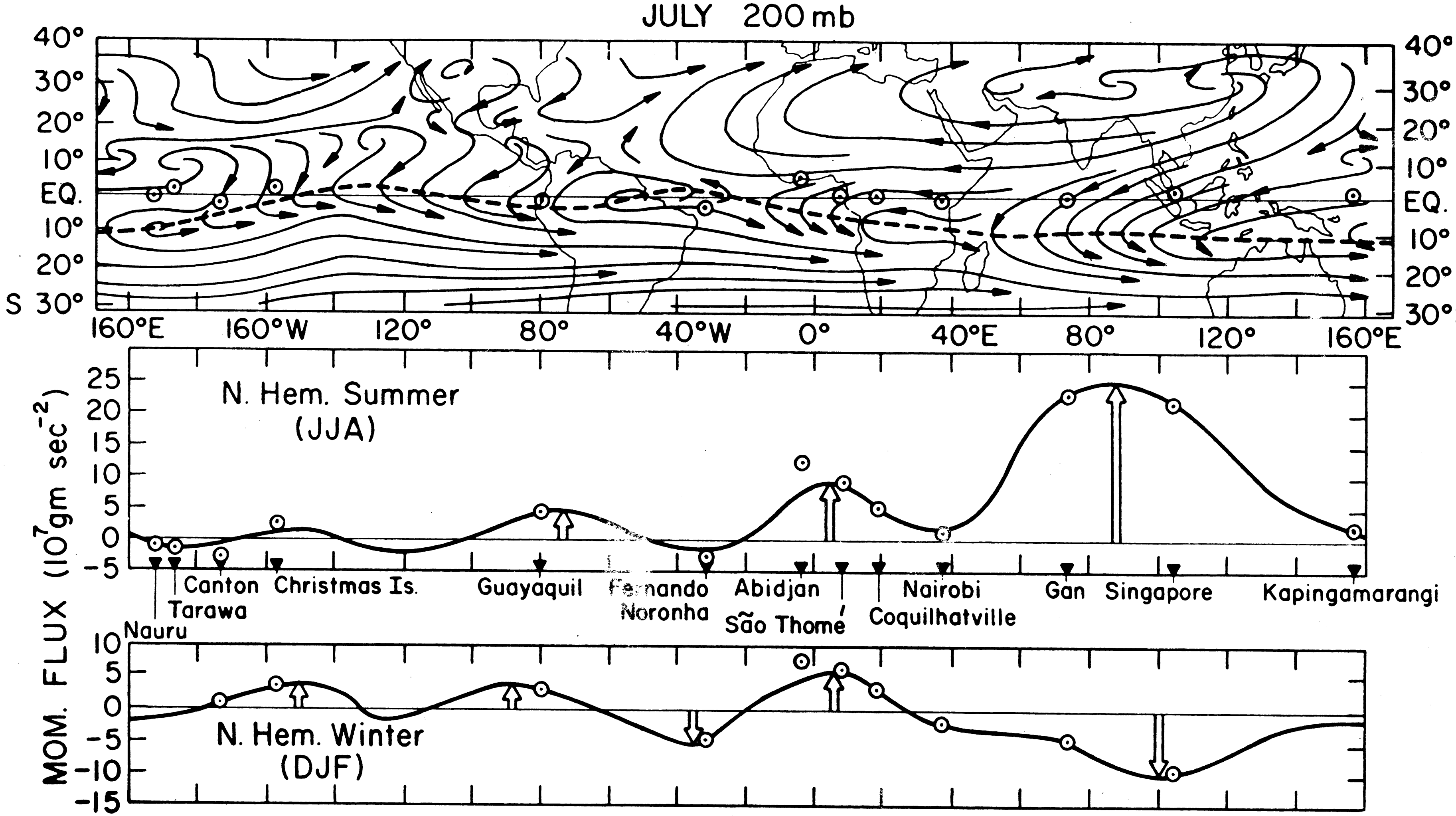


Fig.13.12 Mean streamlines at 200 mb in July, and trans-equatorial flux of (linear) momentum in Northern Hemisphere summer and winter. Encircled dots show station locations (top) and mean momentum flux at stations based mostly on calculations by Henning (1968). Streamlines were taken mainly from Frost and Stephenson (1963) and Sadler (1970). (Newton, 1971).

13.7.4 Krishnamurthi (l97l) from spectral analysis of global data, shows that the ultralong waves of wave numbers 1 and 2 at 200 mb between 20^{o} S and 45^{o} N, have a strong southwest to northeast tilt. The tilt is small in the earlier part of June, becomes very pronounced by about the middle of June and persists upto end of August. Quasi–periodic oscillations occur in intensity and tilt, with a period of 3–5 days. This persistent tilt contributes to the northward flux of angular momentum. This is from the standing part of the waves, travelling parts having a much smaller amplitude.

The strong southwest–northeast tilt of the ultralong waves appears to vanish near the equator and in the winter hemisphere the tilt is almost nonexistent for the ultralong waves. However, in this area, northwest–southeast tilt is found for shorter waves which transport a significant part of the total westerly angular momentum towards the winter pole. While the tilt of the ultralong waves is almost always in the sense of giving a north poleward flux of momentum, the net convergence of flux changes sign periodically, giving rise to either a stable or unstable configuration of the easterly jet.

13.8 Studies on the maintenance of the mean circulation of the southwest monsoon have been very inadequate. The present position of the studies is summed up as follows. Air from south of the equator is the main feed into this circulation rising in convection to the east of 70° meridian. Part of it is carried away westward in the upper tropospheric easterlies, slowly subsiding to west of 70^{o} E but not returning into the lower tropospheric monsoon flow. While the incoming short wave radiation reaching the ground or sea surface is reduced by cloudiness, the outgoing infra–red radiation is also reduced and there is still a net gain of radiation at the bottom surface, which is substantially used in evaporation over the seas and perhaps even over the wet land in this season. Into the monsoon trough area over northern India, there is inflow of total heat content (CpT + AgZ + Lq) in the lower troposphere which is carried aloft by convection and exported in upper troposphere. Moisture (Lq) contributes only ten per cent of the advection below 500 mb. Quite an amount of the air flowing across India is cooled by upwelling off the African and Arabian coasts to be heated again in its passage across warmer waters of the rest of the Arabian Sea where substantial amount of moisture is picked up. The mode of building up a deep layer of moist air, two to three hundred kilometres off the west coast, is not yet clear. In the monsoon trough area over northern India, the transient disturbances must be producing kinetic energy by pressure forces in the lower troposphere as the mean motion and standing eddies only dissipate kinetic energy. There is, however, production by the mean motion and standing eddies in upper troposphere which is exported out at those levels. There is some evidence that colder air seems to be ascending and warmer air descending in the vertical circulations in the lower troposphere there by consuming kinetic energy. Vertical motions have to be assessed more accurately. The mean meridional monsoon cell circulation produces substantial kinetic energy in the lower and upper troposphere between 5^{o} and 20^{o} latitudes. In maintaining the westerlies of the lower troposphere between equator and 20^{o} N and 50° and 100° meridians, ‘Ω transport (conservation of absolute angular momentum) due to air motion from south to north makes up for the frictional torque at the surface. In the upper troposphere there is a net inflow of angular momentum which is used up in ‘Ω transport’ as the air moves southward in the upper limb of the monsoon cell. Even these few studies are only on the mean conditions of July and August but not regarding the changes in June and September on how mean monsoon patterns develop and dissipate.

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CHAPTER 14

NUMERICAL MODELLING OF THE MONSOON

14.1 Murakami, Godbole and Kelkar (1970) conducted a numerical experiment in simultating the monsoon in a two–dimensional meridional plane along 80° E, extend ing from the Equator to the Pole.The atmosphere was divided into eight layers in sigma–system of coordinates and the top of the turbulence layer coincided with the level '8½'. In the horizontal, 18 grid–points at 5–degree Latitude intervals were used.

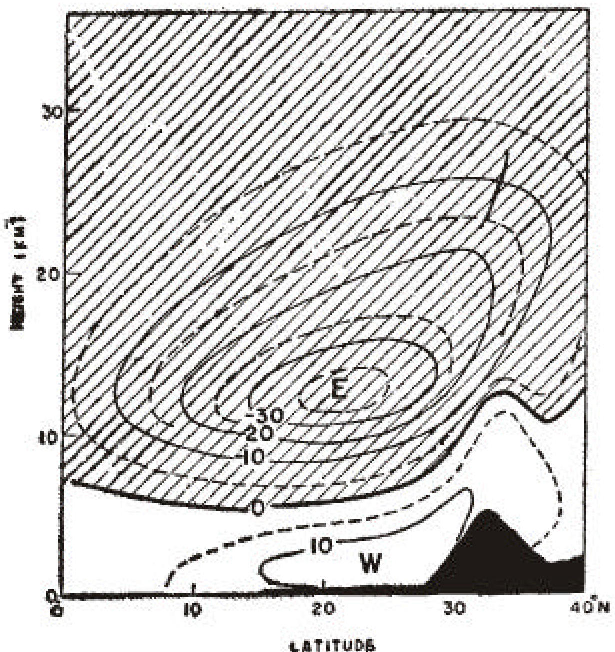
Observationally determined distribution of water vapour, carbon dioxide, ozone and cloud amount has been used for computing the radiational heating due to short and long wave. Albedo at the earth's surface is determined by considering the cloud amount, precipitation and surface conditions. In estimating radiational balance, the distribution of water vapour predicted during the progress of the numerical experiment has not been used but only the observed distribution. Seasonal variation of monsoon circulation, interaction between the southern and northern hemispheres, effect of the Asian continent and oceanic currents have been disregarded. Roughness parameters appropriate to sea, land and mountains have been applied to estimate the forced convection. Vertical distribution of temperature is adjusted to dry (or wet) adiabatic lapse rate whenever it exceeds these values depending on the humidity being less (or more) than 100 %. Temperature and mixing ratio at the land surface are determined by requirements of heat balance between the net downward solar insolation, downward and upward long wave radiation and the sensible and latent heat flux. Sea temperature is kept constant at 300°K (27° C), the observed mean value.

The initial state was calm and the vertical temperature distribution that of the standard atmosphere. Integration over a period of 80 days using 10 minute time steps, determined the zonal profile. Lower tropospheric westerlies and upper easterlies develop in the model, showing speeds of the same order of magnitude as in climatic normals. When mountain effect is not included, westerlies were less than 10 kt and easterlies 20 kt, like the conditions to the east of the Himalayas.

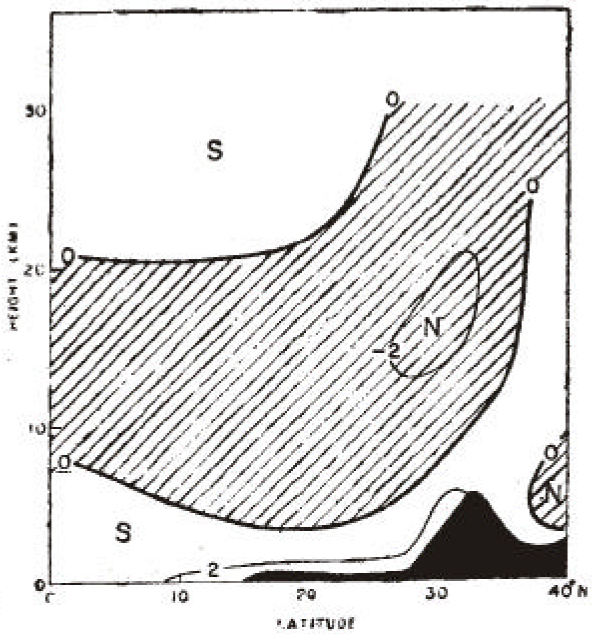
14.2 Godbole (1973) has made further computations with and without Himalayas and moisture in the same manner as in earlier work. The sea surface temperature was varied from 300 to 290°K in the case without moisture and Himalayas. Radiation is the primary force in developing the circulation. However, radiative heating depends upon the temperatures at various levels obtained in the computations. The product of the duration of sunshine and the cosine of the zenith angle of the Sun makes the direct solar radiation a maximum at 35° N. The observationally determined water vapour maximum near 20° N causes relative radiational warming maximum in lower troposphere. The computations derive a thermal and wind distribution in balance with the radiation input. $(E)^{-1}$

Cross–sections of the distribution of zonal and meridional winds and temperature obtained by these computations for the wet model with Himalayas are given in Fig. 14.1. Westerlies in lower troposphere and easterlies aloft of the right order of magnitude are brought out, although their maxima are at a higher latitude. But, the monsoon trough is missed.

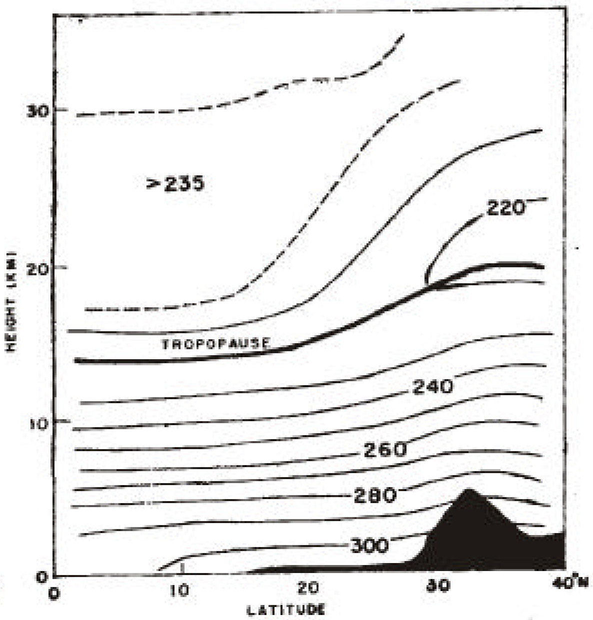
In the dry model with the Himalayas, the easterly and westerly winds are slightly weaker. 'Without the Himalayas', more shallow easterlies capped by easterlies, both weaker (above one fourth in strength compared to 'with Himalayas'), are simulated. Variation of sea surface temperature by 10° C does not make any material difference. 'Without the Himalayas', the temperature difference at 208 mb between 2.5° N and 27.5° N is only 0.3° C; while 'with Himalayas' it is about 12° C, explaining the development of strong easterly jet in the latter case. An interesting feature of the result of omitting Himalayas, is that the upper tropospheric easterly



(a) Vertical cross-section of simulated zonal wind u(m sec-1 along 80° E for wet model with the Himalayas . Easterly regime (E) is hatched
E - Easterly. W Westerly



(b) Vertical cross section of simulated meridional winds u (m sec-1) along 80°E for wet model with the Himalayas. Northerly regime (N) is hatched.



(c) Vertical cross-section of simulated temperature (°K) along 80°E for wet model with the Himalayas.

Fig.14.1

maximum is at a lower latitude than 10^{o} N, nearer the position of easterly jet maximum. While the computations demonstrate the predominant effect of the Himalayas, this two dimensional model must be having considerable limitation in not taking into account the eddies, monsoon depressions and lows and middle latitude troughs in westerlies. Godbole's conclusion that moisture is not an important factor in controlling the monsoon circulation is, perhaps, not justified, as he has assumed the same radiational input for his dry case based on the mean climatological distribution of water vapour and cloudiness.

14.3 The most comprehensive numerical simulation of the monsoon has been by Hahn and Manabe (1975), using a global general circulation model developed at the Geophysical Fluid Dynamics Laboratory (U SA). Results have been obtained with and without the mountain topography of south Asia. Sigma co–ordinate system, eleven unevenly spaced finite–difference levels from the planetary boundary layer upto 31 km and horizontal resolution of approximately 270 km are used. Topography is smoothed, so that the scale of variation in land heights is not smaller than the horizontal grid scale, which systematically lowers and broadens tall, narrow mountain ranges. Climatological sea surface temperatures are imposed at the lower boundary of the seasonal march experiments. The mountain–model (M–model) is time–integrated for 3.5 years. The experiment with no mountains (NM–model) was started from 25 March of M–model setting unitial pressures everywhere to 988 mb.

The M–model is successful in simulating important features of the south Asian Monsoon circulation. In July the computed south Asian low pressure belt extends from Arabia across Asia and into west Pacific, the last somewhat unrealistic. Lowest pressure is near the highest point of the Tibetan Plateau, a discrepancy which may be due to extrapolation of pressure to sea level over Tibet. This doubt is strengthened by the existence of a low even in April at about the same spot. The low over Pakistan is weaker by 5 mb. The monsoon trough from the northwest of the Indian sub—continent to northwest Bay and monsoon depressions are underestimated. Monsoon depressions and lows over the northern part of the Bay move northeastward in the model, while in nature they move westnorthwest. In the surface flow, southeasterlies emanating from the sub–tropical anticyclone of the Southern Hemisphere become south/easterlies on crossing the equator. The Somali jet develops, while convergence into the monsoon trough and easterlies north of the trough are weak. Large–scale anticyclone over the low pressure belt is brought out but the 190 mb anticyclone is 5^{o} too far south. Most discrepancies are noticed in the computed precipitation patterns. The model gives in northwest angle Bay and adjoining land areas, most affected by monsoon depressions, as little rain as in the western desert. Southeastern part of the Peninsula gets very high rainfall, which should have been practically dry. The rainfall computed over Tibet is rather high. The lack of rainfall associated with the western Ghats and northeastern India is explained as due to smoothing of mountain profiles.

The onset of the monsoon has been studied with day–to–day computations by the model. Mean meridional wind component averaged from 80^{o} to 95^{o} E, shifts abruptly, from 26–30 May when most of the surface flow from 5^{o} S to 30^{o} N has a northerly component, to 31 May – 4 June when it has a southerly component. The surface flow pattern over the north Indian Ocean from 21–30 May is very similar to the two months preceding the onset of southwesterly flow. A southwesterly flow suddenly appears over the Bay between 31 May – 4 June as a weak depression over northeast India intensifies, similar to synoptic sequence of monsoon being ushered or pulled northward by depressions. In the model, by 15–19 June, Somali jet becomes established as the heat low in northwestern India becomes well established. By 25–29 June, the temperatures continue to rise over Tibet and southwesterly flow becomes established over the Arabian Sea. Still convergence persists along the equator, but in the next five days this breaks down and much of the surface flow over south Asia originates from the southern Hemisphere. About 20–25 May, the sub–tropical jet at about 25° N rapidly weakens and an intensified zonal flow appears at about 40–45° N, indicating abrupt shift of the jet stream. Soon after this event, abrupt onset of southerly flow takes place over eastern India and the Bay in the lower troposphere. In the climatological dates of onset of monsoon determined from rainfall, in the last week of May, monsoon reaches Sri Lanka and lower Burma and the isochrones advance northwestward upto Pakistan by 15 July. The model shows a similar feature, but lagging 10–15 days behind the normal dates.

The differences found by Hahn and Manabe without mountains are now described. The continental low is centred at 50^o N and 125^o E far to the northeast of M–model and actual position, pressure gradient between 30^o S and 30^o N is less, and geostrophic flow near the surface is northwesterly, a circulation which inhibits sea–to–land flow penetrating far into Asia at the surface. The upper tropospheric anticyclone lies to the south of Tibet in the NM–model while it is over the highest mountains in the M–model. Both the westerly and easterly jets are stronger without the mountains, the centres displaced 10^o to the south. In the NM–model the westerly jet is over like mountains of the M–model. In the lower troposphere from 15^o N southward where few mountains exist, zonal wind profiles of both models are similar. In the M–model, maximum 500 mb temperatures are found over the Tibetan Plateau (none of the smoothed heights reaches 500 mb) with a secondary maximum at 15^o N over the Bay of Bengal and eastward. In the NM–model temperatures over Tibet are 10^o–12° C lower with maximum temperatures at 5^o–10^o N. In both models, maximum latent heat release is responsible for the highest mid–tropospheric temperatures. Role of sensible heating by the earth's surface is less important. But this conclusion may have been influenced by the procedure adopted by the authors for incorporating latent heat released by convection. The saturated adiabat through the surface wet bulb temperature is an upper limit for the temperatures that can be attained by adiabatic convection.

In the NM–model, a narrow band of intense upward motion is centred at 10^o N with descending motion on both sides. At the 500 mb level of the M–model, vertical velocity patterns are more complex. Adiabatic warming of descending air along a band surrounding the Tibetan Plateau is supposed to cause warm temperatures in the region of the monsoon trough, which does not conform with synoptic experience, Anjaneyulu's (1969) computations showed that in the monsoon ellipse, heat inflow in the lower troposphere is transported into the upper troposphere by mean vertical ascent. The ITCZ between the two Hadley Cells is located at 15° N in the M–model and at 12^o N in NM model. The later position is interesting as the surface low is at 50° N.

The Somali jet is more intense in the M–model, though still weaker than observed by Findlater (1969). Mountain effect contributes to the making of Somali jet.

The most significant difference in precipitation between the two models is north of the ITCZ. In the NM–model, a desert like climate is simulated in South Asia, resulting from dry northwesterly continental air flowing southward toward the rainbelt. However, in the M–model, moist southwesterly flow at the surface moves northward toward the south Asian low pressure belt, resulting in substantially more rainfall over south Asian continental regions.

In both models, between 80° and 95° E, the two rainbelts along the winter position of ITCZ at 0–5° S and along its summer position around 10^o–15^o N are not due to any mountain effect. These rain belts seem to interact even after the onset of the monsoon. When rainfall decreased in the southern belt during early July, it increased along the northern rainbelt. This indicates that the dynamics which produce rainfall just south of the equator may be important for temporal variations in the monsoon to the north over south Asia. Also, the rainfall in the monsoon trough and the ITCZ to the south seem to interact.

This global numerical experiment with mountains clearly demonstrates the possibility of simulating the South Asian monsoon, A definite result is that the south Asian low pressure belt and the monsoon trough do not exist if the mountains are not considered. The onset of monsoon took place at different dates in the three years. It is not clear whether these variations are due to a transient behaviour of the model resulting from a thermal imbalance of initial conditions or whether they are due to the natural variability of the model itself.

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CHAPTER 15

LONG RANGE FORECASTING OF MONSOON RAINFALL

15.1 Monsoon rains have a profound effect on agricultural production of India and on the whole economy. There has been a pressing need to indicate, in advance, the performance of the monsoon, as the study of meteorology of India progressed. To meet this requirement, long range forecasting of monsoon rains was taken up by the India Meteorological Department, early in its years of development.

15.2 Attempts were made by Blanford to estimate the prospects of monsoon rainfall during 1882–84. The first official forecast was issued in June 1886. The forecasts were based almost entirely on the relationship that heavy snowfall to the north and west of India caused abnormal pressure conditions and was unfavourable to the advance of monsoon over the areas affected. Eliot used extra–Indian data. Walker changed to an objective method.

15.3 By having homogeneous areas, better forecasting formulae result than by treating the country as a whole. The country has been divided into three zones, northwest India, Peninsula and northeast India. Tables 15.1 and 15.2 show the inter–correlation coefficients among the various sub–divisions of the Peninsula and northwest India (Rao and Ramamoorthy, 1958). Only the correlation coefficients in the Peninsula are below 5 % level of significance.

Table 15.1
Inter–correlation coefficients for sub–divisions in the Peninsula
(June to September rainfall)

Sub–Division	2	3	4	5	6	7	8	9	Coefficient of variation %
1. Gujarat	.51	.41	.30	.59	.55	.40	.38	.39	30
2. Bihar		.72	.46	.64	.59	.74	.55	.44	24
3. Central Province, West			.74	.49	.43	.51	.25	.19	17
4. Central Province, East				.36	.26	.40	.24	.21	13
5. Konkan					.75	.63	.59	.44	17
6. Bombay Deccan						.63	.61	.44	18
7. Hyderabad, North							.73	.58	21
8. Hyderabad, South								.73	23
9. Madras Coast, North								1.00	18

Table 15.2
Inter–correlation coefficients between sub–divisions in Northwest India(June to September rainfall)

Sub–Division	**2**	**3**	**4**	**5**	**Coefficient of Variation %**
1. Uttar Pradesh, West	0.62	.28	.33	.64	24
2. Punjab (I)		.59	.69	.79	32
3. Kashmir			.42	.38	23
4. Rajasthan, West				.73	49
5. Rajasthan, East				1.00	28

15.4 Walker introduced testing of relationship between Indian rainfall and preceding conditions by correlation coefficients. He made a worldwide survey and selected 28 factors, some of which are still in use. South American pressure (April to May) and south Rhodesian Rainfall (October to April) have been important factors in the seasonal forecast formulae. The relationship of Indian monsoon rainfall with pressure at about 30° S, half way round the globe, is itself interesting. Some of the correlation coefficients have been changing markedly in the course of years as shown in Table 15.3.

Table 15.3
Correlation coefficients with southwest monsoon rains over Peninsula

Decade	**South Rhodesian Rain**	**South American Pressure**	**Java Rain**
1881–1890		0.48	–0.11
1891–1900		0.78	–0.68
1901–1910		0.42	–0.48
1911–1920	–0.72	0.27	–0.19
1921–1930	–0.24	0.02	–0.24
1931–1940	–0.78	0.50	–0.07
1941–1950	–0.19	0.58	–0.04
1951–1960	0.31	–0.73	–0.11

The periodic fall in the relationship is perhaps due to a shift in the position of the seasonal highs as pointed out by Rao and Ramamoorthy (1958). They have also shown that the correlation between south American pressure and Indian rainfall is well marked during sun–spot minimum but poor at other times.

Table 15.4
Correlation coefficients between South American Pressure and Monsoon Rains

Region	**Sunspot**	
	Above average	**Below average**
Peninsula Northwest India	+ 0.19	+ .47
	+0.09	+ .63
	n = 31	n = 50

15.5 Later, upper wind flow in preceding months over India itself was correlated with the monsoon rains. Northerly winds in April at Bangalore (l3° N) is positively correlated with peninsular monsoon rain, indicating that the strength of the spring Hadley Cell affects development of subsequent rains. Similarly strong easterly winds at Calcutta in May are seen to inhibit monsoon rains, perhaps slowing the building of monsoon westerlies.

Jagannathan and Khandekar (1962) presented the variations in the correlations coefficients of monsoon rainfall with upper winds over India.

15.6 They also derived a number of correlations between monsoon rain in the Peninsula and preceding heights of pressure surfaces and thickness of isobaric layers, as given in Table 15.6.

Table 15.5
Variation in Correlation Coefficients of monsoon rainfall with upper winds

Factor	**Correlation with**	**Correlation coefficient for period**					
		1911-1920	**1921-1930**	**1931-1940**	**1941-1950**	**1951-1957**	**Entire period**
Agra Easterly winds 2 Km (March)	Monsoon rainfall Northwest India	-	+.02	+.37	+.66	+.18	+.34
Calcutta Easterly winds 2 Km (May)	Monsoon rainfall Northwest India	-	-.62	-.49	-.71	+.38	-.35
Calcutta Easterly winds 4 Km (May)	Monsoon rainfall in Peninsula	-	-.39	-.19	-71	+.41	-.03
Bangalore Northerly winds 6 Km (April)	Monsoon rainfall in Peninsula	-	+.40	+.47	+.56	+.91	+.38

Table 15.6
Correlation Coefficients between Peninsular rain and thermal patterns in upper air

Factor	**Correlation coefficient**	**Probable error of correlation coefficients**
Height of 400 mb in March over New Delhi, Jodhpur and Allahabad	+ .76	.08
Height of 500 mb in March and April over Calcutta	+ .68	.10
Height of 500 mb in May over Bombay, Madras, Vishakhapatnam, Trivandrum	–.67	.11
Thickness 850/400 mb in March over New Delhi, Jodhpur, Allahabad	+ .76	.08
Thickness 700/500 mb in March over Calcutta	+ .79	.07
Thickness between 600/400 mb in April over Calcutta	+.69	.10
Thickness between 850/400 mb in May over Calcutta	+ .55	.14

15.7 Shamshad (1967) found that April mean 700 mb height difference between 45° N and 17 o N along 85^{o} E meridian is negatively correlated with the subsequent June to September rainfall over Pakistan.

15.8 Forecasting monsoon rains based on such correlation coefficients has not been satisfactory. Perhaps, as summed up by Normand (1953) "Indian monsoon rainfall has its connections with later rather than with earlier events. On the whole Walker's worldwide survey ended by offering more promise for the prediction of events in other regions than in India".

15.9 Jagannathan and Bhalme (1973) have brought out evidence that the seasonal rainfall depends upon sunspot activity. The rainfall is larger during sunspot maximum than during sunspot minimum over North Assam, North Bengal, Bihar and sub–Himalayan East Uttar Pradesh and the central parts of the Peninsula. Over the rest of the country, rainfall is larger during sunspot minimum. The variations with sunspot cycle are dominant in the areas where rainfall is influenced by orography. The differences are upto 20 % of the normal rainfall. They have been able to demonstrate that pressure patterns significantly differ between the maximum and minimum epochs. During sunspot maximum, negative pressure departures near the foot of the Himalayas (Fig.15.l), like 'monsoon breaks', cause greater rainfall in adjoining areas. During sunspot minimum epoch, monsoon trough is more marked, and depressions are more frequent (Fig. 15.2) giving better rains in the central parts of the country. Though rainfall in many parts of the country exhibit oscillation corresponding to the QBO, this accounts for only a very small part of the variance. (Jagannathan and Bhalme 1973, Koteswaram and Alvi 1969, Bhargava and Bansal 1969).

15.10 The relationship between the Indian monsoon rainfall and the South American pressure is apparently due to 'Southern Oscillation', a name given by Walker. This oscillation is an air pressure 'sea–saw' between the equatorial low pressure area of the Malay Archipelago and sub–tropical high pressure area centred near Easter Island (27° S, 109°w) (Berlage, 1957). Fig.15.3 shows the correlation coefficients between anomalies of April to August pressures at Djakarta with the rest of the globe. Correlation is positive with south India and negative with the area of the south Pacific High. On account of the steep gradient of correlation coefficients over South America, Walker's tacit assumption that the sum of the air pressure deviations at Santiago, Cordoba and Buenos Aires represents South American pressure seems questionable. Djakarta pressure is not representative of the equatorial low pressure area of the Malay Archipelago. The correlation between the half–yearly pressures at Djakarta and Fernandez (also representative of the South Pacific High) varies with time lag. Largest negative correlation coefficient is of Djakarta half a year before Jwan Fernandez; and maximum positive two and a quarter year later. Fig. 15.4 shows the parallel run of Djakarta and Bombay air pressures, though there are small lags of phase between the two waves. The amplitude of the southern oscillation decreases from the tropics to the equator. The southern oscillation is not confined to the southern hemisphere alone, as already seen from Bombay data.

Walker gave a period between 2 and 2.5 years for the Southern Oscillation, so close to the period of QBO discovered about a decade back. The great variability of the period of the Southern Oscillation reduces the amplitudes found by periodigram analysis.

15.11 In view of the current evidence of the cooling of the atmosphere since the forties of the present century, its effect on the monsoon circulation and rains has become a matter of great interest. Roughly from the middle of the sixteenth century to the middle of the nineteenth century was the 'Little Ice Age' and thereafter upto the forties of the present century a warmer phase prevailed. From Lamb's (1966) presentation, the strength of the July pressure gradient between 10^{o} N and 20^{o} N along 75° E had increased by about 20 % between 1860 and 1900 (Fig. 15.3). Weaker pressure gradient over the Peninsula is associated with less rainfall. In the general circulation patterns prevalent in the past century Djakarta air pressure is not positive at intervals of 2.33 years, but so at double this period. Apparently, the variability of the wavelength is such that the chance of touching the right phase after an interval of twice the mean wavelength is greater than after only one mean wavelength. Berlage (1957) observes that air pressure in the Malay Low and the Easter Island High, perhaps even through the whole tropical and sub–tropical zone of the Pacific Ocean, is low during great solar activity and high during weak solar activity.

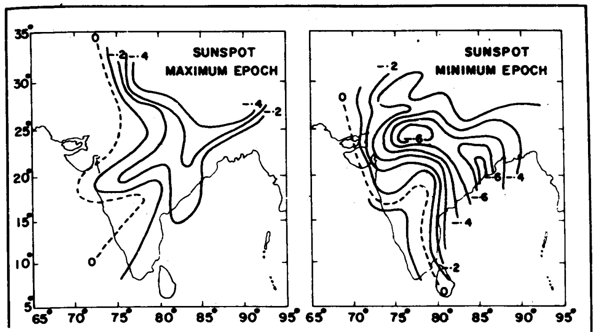

Fig.15.1 Surface pressure departure (mb) during monsoon season (Jagannathan and Bhalme, 1973)

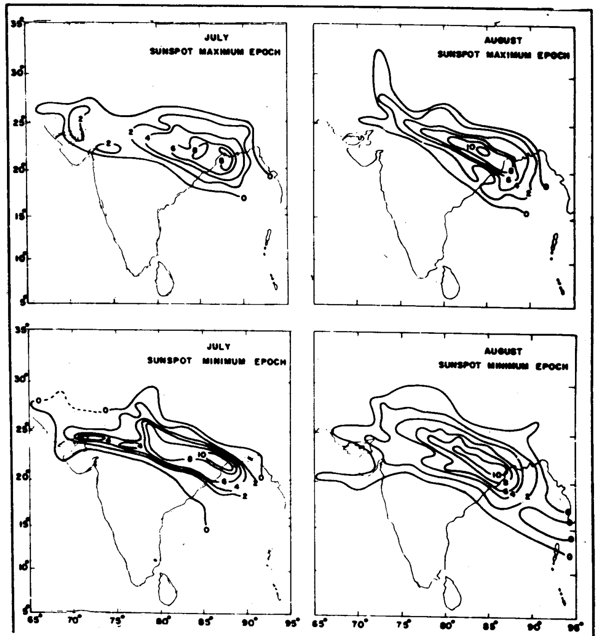

Fig.15.2 Frequency of storms and depressions (1901-1950) (Jagannathan and Bhalme, 1973)

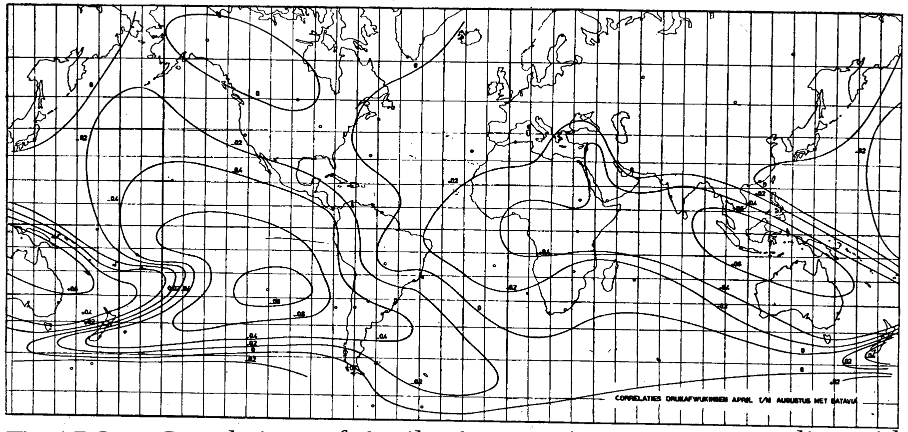

Fig.15.3 Correlations of April - August air pressure anomalies with Djakarta (Batavia) April - August air pressure anomalies. (Berlage, 1957).

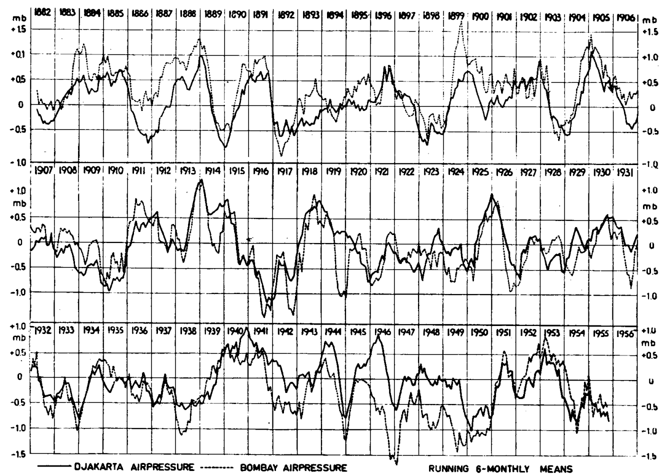

Fig. 15.4 Parallel run of Djakarta and Bombay air pressure, (Berlage, 1957).

The development of periodicity of the Southern Oscillation is explained by Berlage as due to the Peru and South Equatorial Ocean Currents. When air pressure is abnormally low in the Malay Low, it is abnormally high in the Easter Island High. The general air and water circulated through the South Pacific is accelerated. The Peru Current and the South Equatorial Current are accelerated and consequently colder than normal. The negative temperature anomaly created in the east South Pacific Ocean arrives 7½ months later at the Malay Low. Consequently air pressure in the Malay Low increases, while air pressure in the Eastern Island High decreases. When air pressure in the Malay Low is abnormally high, it is abnormally low in the Easter Island High. The air and water circulation through the South Pacific Ocean is slowed down. The Peru and the South Equatorial Currents are slowed down and consequently they become warmer than normal. The positive temperature anomaly created in the east of the South Pacific Ocean arrives 7½ months later in the Malay Low. Now, air pressure in the Malay Low decreases, while Easter Island High strengthens. The extreme equatorial width of the Indo–Pacific region provides a sufficient length of phase lag between a sub–tropical high and an equatorial low.

15.12 Raghavendra (1974) studied the monsoon rainfall of Maharashtra for over 100 years for trends and periodicities. Konkan has a weak indication of 100 year cycle. By fitting orthogonal polynomials upto sixth degree, Konkan and Madhya Maharashtra show a trend of a quadratic curve but none in Marathwada and Vidarbha. Power spectral estimates indicate the existence of a long term slow, increasing trend for Konkan and cycles of 30 years in Vidarbha, 20 and 7.5 years in Marathwada and 15 years in Madhya Maharashtra. The cycles account only for a very small portion of the variance.

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CHAPTER 16

GLOBAL RELATIONSHIPS

16.1 A perturbation of the magnitude of the southwest monsoon should be having linkages with the general circulation over various other parts of the globe. Global circulation patterns preceding the monsoon period have been found to have relationship on the subsequent monsoon activity and these correlations have been used to predict the performance of monsoon. Some other interesting relationship are treated in this chapter.

16.2 In mid–June, the intensified polar jet sweeping south around northeastern Tibet, activates the polar front which has shifted to Yangtze Valley. A considerable air mass discontinuity across the front and very slow moving disturbances ensure copious 'Mei–Yui' or plum rains. At this time, when the sub–tropical jet south of the Himalayas weakens and the polar jet north of the Himalayas strengthens, a blocking high develops over the sea of Okhotsk. The early summer rains over India tend to fluctuate in unison with the 'Mei–Yui. Asakura (1968) deduced that the Okhotsk blocking high results from the union of a large heat source near India and a cold source near the 'Okhotsk Sea. Between 21 May and 19 June, 5–day mean surface pressure anomalies over Northwest India and over the Sea of Okhotsk are negatively correlated (r = –0.56, n = 30).

16.3 In the first half of July, southern China is well out of the range of the polar front and has dry weather. Median dates for the onset of dry spell at Hong Kong and onset of monsoon rain at Delhi (Bhullar 1952) are respectively 5 and 3 July, with a correlation of 0.71 (n = 37), (Ramage, 1952). This shows that major circulation changes are causing the seasonal weather developments at far off places.

16.4 In the first half of June, the frequency of blocking anticyclones between Iceland, Scandinavia and the British Isles reaches its annual maximum (June 11). After that date, the frequency of blocking anticyclones in the European sector drops substantially during the second half of June, July and August (Hess and Bresewsky 1952). Flohn (1958) observes that the rapid advance of the Indian monsoon near mid–June is correlated with frequent occurrences of blocking highs on the western and the eastern coasts of the Eurasian Continent and homologous features in other parts of the atmospheric circulation.

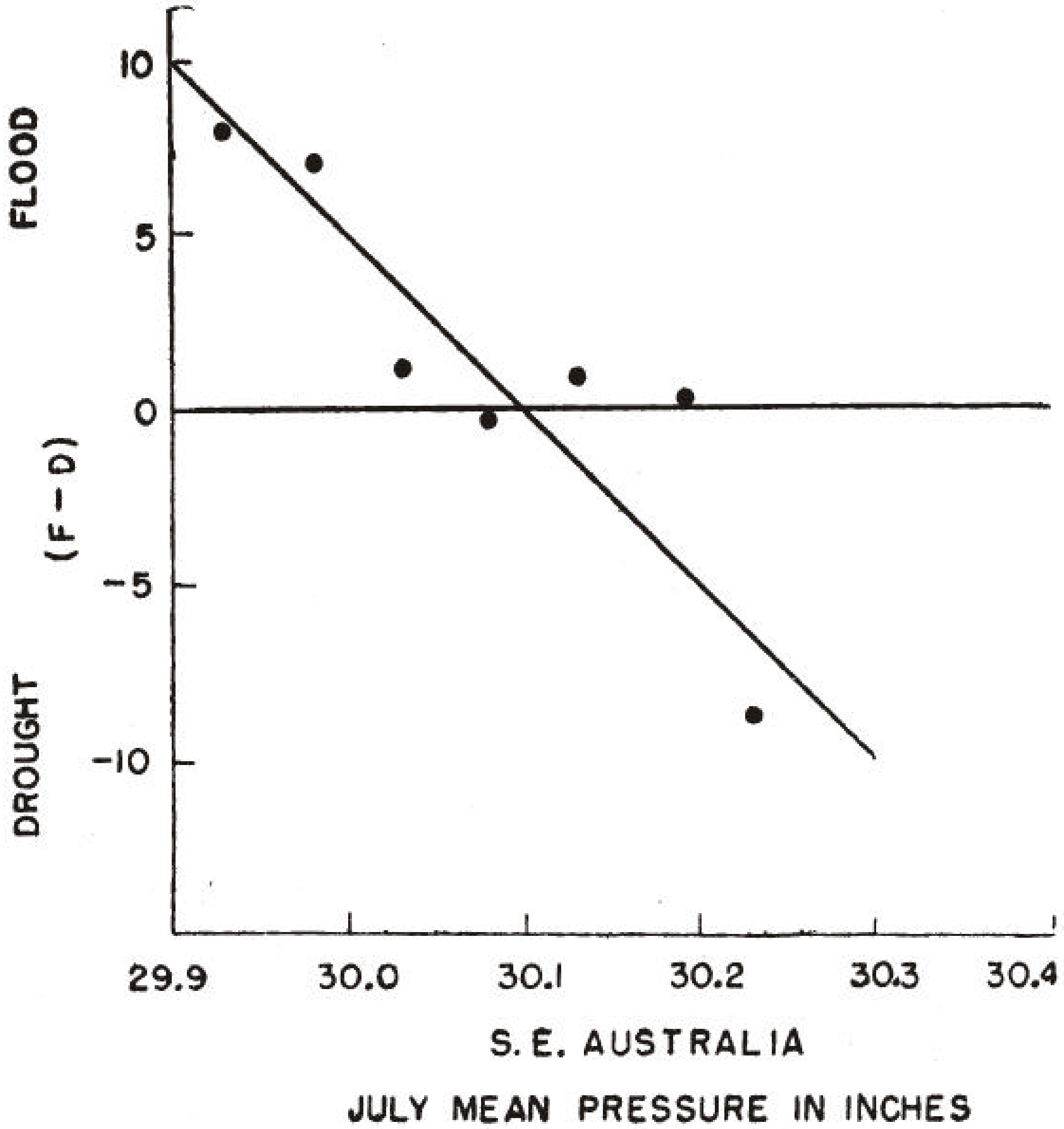
16.5 Wright (1967) regards the date of onset of southwest monsoon over Delhi etc, which is associated with the shift of the jet stream to the north of the Himalayas, symptomatic of the development of the season's circulation over a very wide area of the globe. Perhaps, to make up for this large northward jump of the westerly jet stream from the Mediterranean to China the jet shifts southwards about mid–or late June in the North Atlantic in contrast to the global trend for poleward movement (Lamb, 1972). A short return to a northern position of upper westerlies over the eastern North Atlantic and Europe usually takes place between late August and mid–September. In a further study of the monsoons of the five years (1956–60) Wright found (Lamb, 1972) that the northern summers in these years were arranged in almost the same order as regards :

(i) the date (earliness) of the northward jump of the jet stream, (200 mb winds at 95° E being most relevant in this connection.)

(ii) the longitudinal extent attained by the westerly current over southern Asia.

(iii) the extent and strength of the Madagascar Trade Wind stream in May and June.

(iv) the zonality (westerliness) of the flow over southern Australia about the same time.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FLOOD
DROUGHT
(F – D)
29.9
30.0
30.1
30.2
30.3
30.4
S. E. AUSTRALIA
JULY MEAN PRESSURE IN INCHES

Fig. 16.1

(v) the latitude of the sub–tropical anticyclone in the North Pacific in May–June.

(vi) the meridionality (northerliness) of the flow over the Mediterranean in May and June.

(vii) the wetness and coolness of the summer in London.

700 mb was used in the definition of items (ii), (iv), (v) and (vi). Wright believes that not only (vii) but also (iv) to (vi) define anomalies that tend to persist through the following July and August.

16.6 Ramdas (1958) showed a relationship between the July mean pressure over southeast Australia and excess or deficit of rainfall over India, called 'flood' or 'drought', as in Fig.l6.1.

16.7 Maung Maung Kha (1945) found the following correlations between Burma monsoon rains and some global conditions.

Table 16.1
Correlation coefficient (x l00) between monsoon rainfall over parts of Burma and global conditions

	1	**2**	**3**	**4**	**5**	**6**
1, Arakan Coast rain July-August	+ 100	+ 45	+ 54	- 52	+ 54	- 45
2. Calcutta+Colombo March Pressure	+ 45	+ 100	- 1	- 14	+ 46	- 25
3. Tokyo February to March Pressure	+ 54	- 1	+ 100	- 57	+ 4	- 14
4. Cheyene+San Francisco+Honolulu Apr. pressure	- 52	-14	- 57	+ 100	- 22	+ 37
5. Khartoum January Temperature	+ 54	+ 46	+ 4	- 22	+ 100	- 40
6, Moscow March to April pressure	- 45	- 25	- 14	+ 37	– 40	+ 100
1. Irrawaddy Delta July-August rain	+ 100	+ 57	- 42	+ 49	- 50	+ 48
2. Tokyo February to March pressure	+ 57	+ 100	- 57	+ 28	- 3	+ 12
3. Cheyenne+San Francisco+Honolulu Apr. pressure	- 42	- 57	+ 100	- 7	+ 17	- 12
4. Cape Pembroke May Pressure	+ 49	+ 28	- 7	+ 100	- 38	+ 41
5. Berlin and Moscow and Ustzylma Apr. Temp.	- 50	- 3	+ 17	- 38	+ 100	- 36
6. Durban April to May temperature	+ 48	+ 12	- 12	+ 41	- 36	+ 100
1. Tenasserim July-August Coast Rain	+ 100	+ 59	+ 52	- 52	+ 58	+ 40
2. Copenhagen-i-Berlin January pressure	+ 59	+100	+ 56	- 47	+ 13	+ 18
3. Bodo+Gjesvar January Pressure	+ 52	+ 56	+ 100	- 22	+ 30	+ 14
4, Father Pt,+Charlotte Town Jan. Temperature	- 52	- 47	- 22	+ 100	- 23	- 14
5. Leon January temperature	+ 58	+ 13	+ 30	- 23	+ 100	- 9
6. Rangoon May pressure	+ 40	+ 18	+ 14	- 14	- 9	+ 100

Such relationship even if they are not useful to predict the rainfall by regression equations, bring to light features relating to evolution of monsoon rains, which have to be studied further.

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