2. VARIABILITY AND FORECASTING OF THE SUMMER MONSOON RAINFALL OVER INDIA

D. A. Mooley and J. Shukla

1 INTRODUCTION

The phenomenon of monsoon is global in character, affecting a large portion of Asia, parts of Africa (Sahel), and northern Australia, which cover more than 50% of the world population. The huge land mass of Asia in juxtaposition with the huge water mass of the Indian Ocean makes the Asian monsoon spectacular.

India, with a population of 15% of the world population, is located in the central portion of South Asia and is predominantly within the monsoon regime. The Indian summer monsoon (also known as southwest monsoon) contributes about 80% to the annual rainwater potential. The high concentration of rainfall during the monsoon (June through September) and meager irrigation facilities make the economy and prosperity of India vitally dependent on the performance of the summer monsoon. The summer monsoon rainfall largely provides the country's water requirements for agriculture, industry, generation of hydroelectricity, and drinking water.

Summer monsoon is a regular phenomenon only in the sense that it comes every year. But its onset, its activity during the season, and its withdrawal are subject to variations that sometimes are large. A delay in the onset of the monsoon results in low levels in hydroelectric reservoirs, with a consequent reduction in the generation of hydroelectricity and the imposition of a power cut on the industry. Anomalies in the activity of the monsoon during the season affect the crop production of the country. Early withdrawal has an adverse effect on crops and the country's water potential. Large-scale failures of the monsoon upset the country's economy and result in intense suffering for the masses (Report of the Indian Famine Commission 1880, 1898, 1901; Bhatia 1967; Srivastava 1968). So that the increasing demands for water are met, it is necessary to plan the country's water resources by taking into account the variability of the summer monsoon rainfall over India as a whole as well as over the different meteorological subdivisions of the country.

The variability of the summer monsoon over India, over its meteorological subdivisions, and over some of its rain gauge stations, the influence of global boundary conditions and other factors on the monsoon, and the forecasting of the Indian summer monsoon rainfall are reviewed in this chapter.

2 VARIABILITY IN ONSET AND WITHDRAWAL OF THE MONSOON

During the season March to May, which is usually referred to as the hot weather season, gradual heating of the land mass of South Asia takes place and the intertropical convergence zone (ITCZ) continues its northward advance. Toward the end of this season, a trough of low pressure extending from Northeast Africa to Southeast Asia develops, troughs in the midlatitude westerlies shift to higher latitudes north of India, low-level cross-equatorial airflow strengthens, and finally the moist monsoon current advancing from the southwest arrives over the southernmost parts of India, bringing with it characteristic wind, cloud, and weather patterns. The cloud patterns can be clearly seen in satellite picture. While declaring the onset of monsoon over southern Kerala or any part of India, the weather forecaster takes into account the rainfall, the wind and temperature fields (both at the surface and at the upper levels), the moisture field, the cloud patterns, the type of clouds, and the state of the sea. The emphasis is, however, on rainfall. Although the criteria for the onset of the monsoon lack objectivity, they have been in use over a long period and have stood the test of time (Ananthakrishnan et al. 1967, 1968, 1981). There is a fair amount of consistency in the date of arrival of the monsoon over India each year.

In spite of the difficulties of defining precisely the onset of the monsoon, the Indian Meteorological Department has been fixing as best as possible the date of onset of monsoon over South Kerala and other parts of the country in view of the practical and meteorological importance of the event.

Before we consider the variability in the onset or withdrawal of the summer monsoon, let us examine the normal dates of onset/withdrawal of the monsoon for the different parts of India, as prepared by the Indian Meteorological Department (1943). These dates are given as isolines over a map of India. The maps showing the normal dates of onset and withdrawal of the monsoon are given in Figs. 2.1 and 2.2. These diagrams were prepared on the basis of average

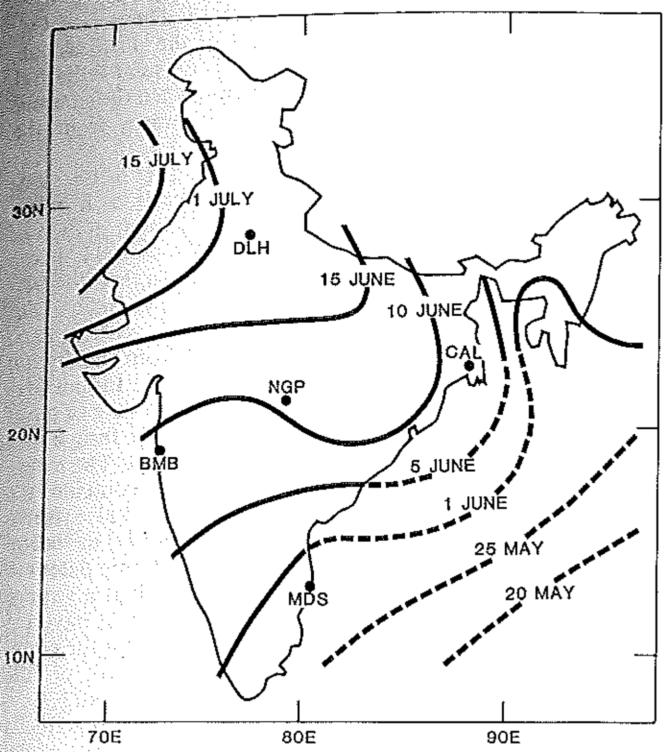

Fig. 2.1. Mean dates of onset of the summer monsoon over India. Broken lines denote isolines based on inadequate data. (From Indian Meteorological Department 1943.)

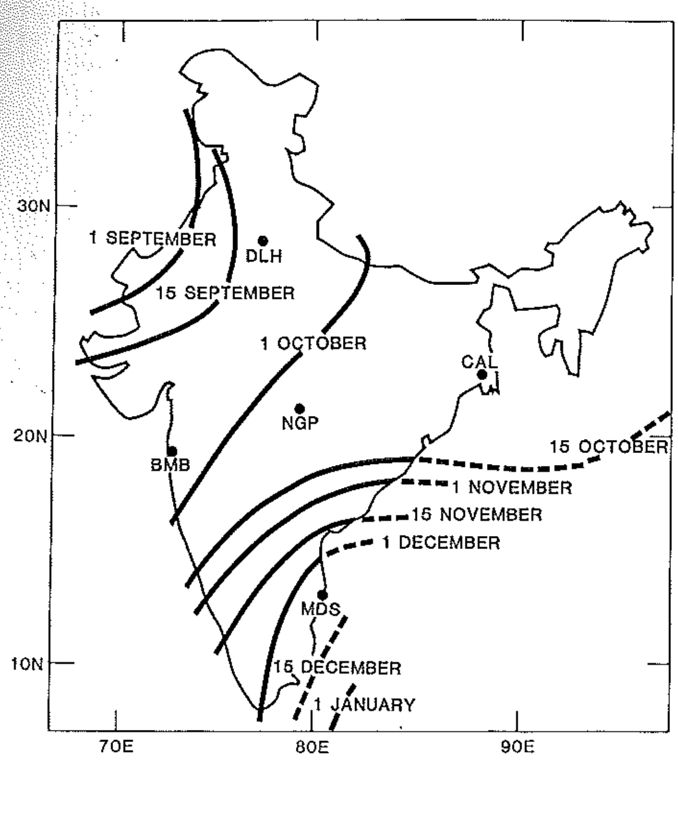

Fig. 2.2. Mean dates of withdrawal of the summer monsoon from India. Broken lines denote isolines based on inadequate data. (From Indian Meteorological Department 1943.)

pentad rainfall data of all observatories. Average pentad rainfall curves were prepared for each station. The middle date of the pentad with an abrupt rise in rainfall was taken as the onset date. The onset dates were plotted on a map and isolines of onset dates were drawn. The middle date of the pentad with an abrupt fall in pentad rainfall was taken as date of withdrawal of monsoon at the station. These dates of withdrawal were plotted and isolines of dates of withdrawal were drawn. Over the Bay of Bengal, broken isolines of onset/withdrawal have been drawn on account of lack of data and consequent uncertainty. The main features to be noted from Fig. 2.1 are normal arrival of the monsoon in the extreme southeast of the bay by mid-May, the onset of monsoon over South Kerala by June 1 and over northwestern Rajasthan in the first week of July, and the advance of the monsoon from southeast to northwest except over the portion of the country west of 80°E and south of 25°N, where the advance is from south to north. Thus, within five weeks of the onset over the southernmost tip of India, the monsoon normally spreads to the whole of the country. Ananthakrishnan et al. (1981), who examined the pentad rainfall of six stations in the Andaman-Nicobar group of islands in the southeast Bay of Bengal and followed the movement of the ITCZ in April to May, showed that the mean date of onset of the monsoon is May 3 for the southern stations (south of 10°N) in the group of islands and May 8 for the nothern group (between 12°N and 14°N). Thus this study supports the much earlier advance of the monsoon over the extreme southeast Bay of Bengal than over southern Kerala, as suggested by Fig. 2.1.

Figure 2.2 shows that the summer monsoon commences withdrawal from the extreme northwestern parts of India by September 1 and withdraws from northern and central India and north peninsular India by October 15. The period October 16 to 31 is the onset of the northeast (or winter monsoon) over south peninsular India and can be considered as the period when the summer monsoon has completely withdrawn from the country.

Figure 2.3 shows the date of onset of the summer monsoon over southern Kerala as fixed by the Indian Meteorological Department in each of the years during the period 1901 to 1984. These dates of onset have been taken from the *Monsoon Summary of India* published by the Indian Meteorological Department. For the recent years, these dates have been obtained from the *Indian Daily Weather Report* issued by the Indian Meteorological Department. The mean date of onset of monsoon over southern Kerala is June 2 and the standard deviation (S.D.) is 8 days. The mean, median, and modal dates of onset coincide, the common date being June 2. Figure 2.3 also shown the mean date and limits of 1 standard deviation on either side of the mean. The extreme dates are May 11, 1918, and June 18, 1972. It is a strange coincidence that both of these extremes have occurred in well-marked drought years for the country; this, however, indicates that early or late monsoon onset is not related to the overall behavior of the monsoon. It can be seen that practically all the dates of onset lie within 2 standard deviations from the mean.

The northward advance of the monsoon along the west coast may be slow when not associated with any synoptic system. But a northward-moving low pressure, depression, or cyclonic storm off the Indian west coast about the time when onset is taking place or about to take place over South Kerala expedites the advance of the monsoon over the west coast. Sometimes, after the low-pressure system has moved away, the advance of the monsoon is not maintained and the monsoon

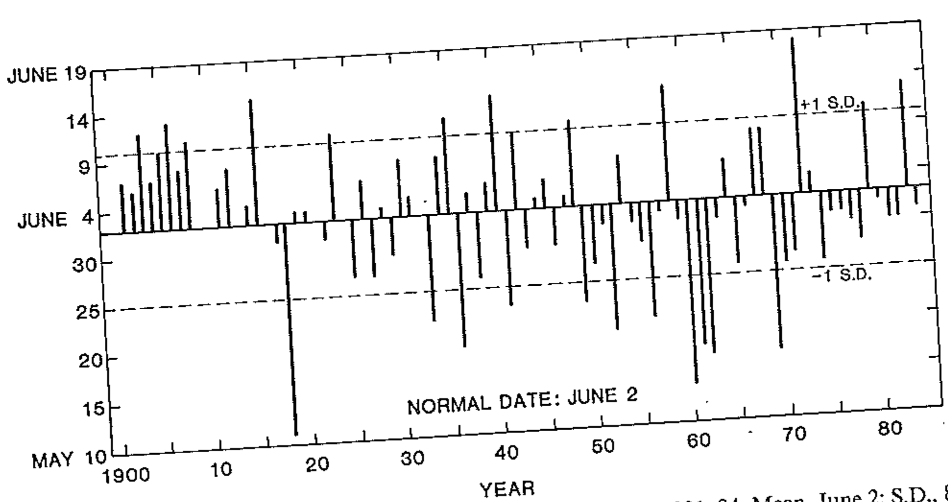

Fig. 2.3. Dates of onset of the summer monsoon over southern Kerala, 1901–84. Mean, June 2; S.D., 8 days.

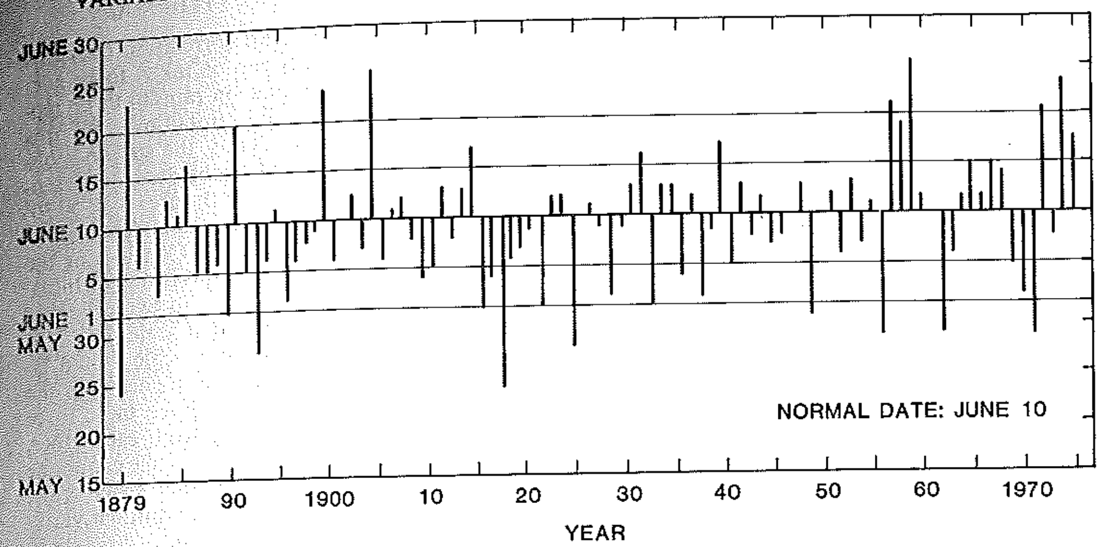

Fig. 2.4. Actual dates of onset of the summer monsoon over Bombay, 1879–1975. (From Rao 1976.)

current withdraws to a lower latitude; in this case the monsoon advance is termed a temporary advance. Over northern India, the westward advance of the monsoon often takes place in association with a low-pressure system moving northwestward from the north Bay of Bengal.

Ramdas et al. (1954) applied the rainfall criterion to the average district rainfall for the period 1901 to 1950 and fixed the dates of establishment of summer monsoon over Travancore-Cochin (presently Kerala State), south Kanara district, Ratnagiri district, and Colaba district (adjoining and south of Bombay), in each of the years.

Figure 2.4 shows the dates of onset of the summer monsoon over Bombay during the years 1879 to 1975, taken from Rao (1976). The mean date is June 10. The onset was later than June 20 during 1880, 1900, 1905, 1957, 1959, 1972, and 1974 and earlier than June 1 during 1879, 1893, 1918, 1925, 1949, 1956, 1962, and 1971.

Bhullar (1952) fixed the dates of onset of the summer monsoon over Delhi in each of the years during the period 1901 to 1950 on the basis of a rainfall criterion, using daily rainfall data of the stations in the districts surrounding Delhi and in Delhi State. The first date of widespread rain in Delhi State and in each of the districts adjoining Delhi State, during the period June 15 to July 20 was taken by Bhullar as the date of onset of monsoon over Delhi. He also examined the synoptic conditions leading to the monsoon onset. These are (1) arrival of the bay monsoon branch in association with a depression from the bay, (2) arrival of the Arabian Sea branch of the monsoon in association with a strengthening of the monsoon over the Arabian Sea, (3) simultaneous strengthening of both the branches of the monsoon, (4) penetration of the bay or Arabian Sea branch of the monsoon under the influence of a western disturbance or a wave in the midlatitude westerlies, (5) simultaneous strengthening of the Arabian Sea and the bay branches and a well-placed western disturbance. The mean date of onset over Delhi is July 3 and the standard deviation is 8 days. The extreme dates are June 17 and July 20. Figure 2.5 shows the dates of onset of the monsoon over Delhi, the mean date, and limits of one standard deviation on either side of the mean. About 75% of the dates lie between June 21 and July 10. The probability of onset before June 21 is only 8%, but the probability of the onset after July 10 is 18%. Except for July 20, 1907, all the dates of onset over Delhi are within two times the standard deviation from the mean date.

By the time monsoon advances over Delhi, the whole country except the extreme northwestern

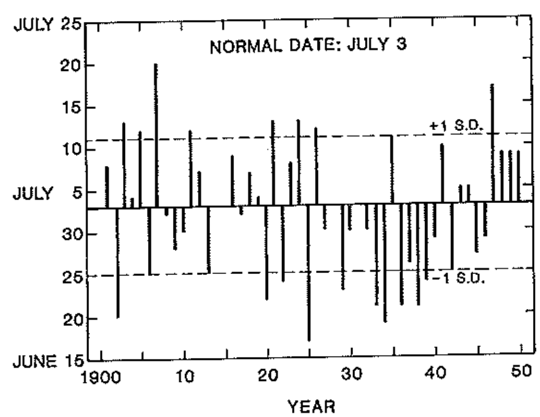

Fig. 2.5. Dates of onset of the summer monsoon over Delhi. Mean, July 3; S.D., 8 days. (From Bhullar 1952.)

part of India is under the influence of the monsoon. An idea of the time taken by the summer monsoon to advance over India can be had from the interval between the dates of onset over southern Kerala and over Delhi. This interval has been computed and its frequency distribution has been examined. The extreme intervals are 11 days in 1934 and 57 days in 1918. The mean interval is 29 days, and the standard deviation of the interval is 9 days. All intervals, except the one in 1918, are within 2 standard deviations from the mean interval. About 75% of the distribution lies between 16 and 40 days. Thus the time taken by the monsoon to advance over the country exhibits high interannual variability. The probability of the monsoon advancing over the whole country in 11 to 15 days is only 10%, and that of its advancing over the whole country in more than 40 days is about 15%.

While the onset of the monsoon over the country is invariably gradual, its withdrawal is relatively rapid. Withdrawal from the northwestern parts of India normally commences by September 1. Cooling of the land masses of northern India and further north and a shift in the activity of the troughs in the westerly wind belt to a relatively lower-latitude belt result in the southward shift of the monsoon trough and withdrawal of the monsoon from northwestern India. As these factors vary from year to year, the withdrawal of the monsoon undergoes interannual variation. In general, the withdrawal of the monsoon from western Rajasthan, Haryana, and Punjab takes place sometime during the first fortnight of September, and withdrawal from over most of the remaining parts of the country occurs during the period mid-September to mid-October.

3 INTERANNUAL VARIABILITY OF THE SUMMER MONSOON RAINFALL

Several studies of the seasonal monsoon rainfall and its interannual variability have been made by research workers. These can be grouped into three distinct space scales for seasonal monsoon rainfall: an all-India scale or a large scale covering practically the whole of India, a regional/subdivisional or medium scale, and a small scale, that is, one covering an area over which the seasonal monsoon rainfall regime can be represented by the rainfall recorded at a rain gauge station. Hourly or daily rainfall shows high variability over short distances, and the rainfall recorded at the station is representative of the rainfall character over an extremely small area. As the time scale of rainfall increases, the areal representation of the station increases. From a study (Mooley and Ismail 1982) of the variation of the correlation coefficient with distance, with respect to seasonal monsoon rainfall over Vidarbha, it is seen that the correlation coefficient falls to 0.8 at a distance of 30 km. The mean correlation over this distance is 0.9, and this can be taken to be adequate for the station monsoon rainfall to be representative of the monsoon rainfall over the area within a 30-km distance. Thus, the station monsoon seasonal rainfall can be considered to be representative of the rainfall over the area $3.14 \times 30^2 \approx 3 \times 10^3$ km^2. Hereafter, this spatial scale will be referred to as the small scale. Considering the sizes of the meteorological subdivisions of India, the medium scale would be one to two orders of magnitude higher than the small scale, i.e., 3×10^4 to 3×10^5 km^2. The large scale would be about three orders of magnitude higher than the small scale, i.e., 3×10^6 km^2. Since these three space scales differ substantially from each other, the results of a study for one space scale cannot be applied to the results from another space scale. We shall review the studies of interannual variability of the monsoon rainfall on these three space scales.

3.1 Large-scale variability

While considering interannual variability of monsoon rainfall on this scale, studies on variability of annual mean rainfall have also been considered, since on this scale, the contribution of the summer monsoon to the annual rainfall is 75 to 80%, and the main features of the variability of the monsoon and annual rainfall are expected to differ little.

The earliest study on fluctuations of Indian rainfall was carried out by Blanford (1886). He computed the annual rainfall of India for the years 1867 to 1885, using a network of about 500 rain gauges over British India. The study had to be based on data for 19 years only since data prior to 1867 were very limited.

Thereafter, Walker (1910a, 1914), Mooley (1975), Banerjee and Raman (1976), Parthasarathy and Dhar (1976), Parthasarathy and Mooley (1978), Mooley and Parthasarathy (1979), Bhalme and Mooley (1980), Paolino and Shukla (1981), Mooley at al. (1982), Joseph (1983), Mooley and Parthasarathy (1982, 1983a, 1983b), Shukla (1987), and Mooley and Parthasarathy (1984a) examined the variability of monsoon rainfall on this spatial scale; and Parthasarathy and Dhar (1975) and Mooley at al. (1981) examined the variability of the annual rainfall on this scale. These studies cover trends/oscillation in rainfall, droughts/floods, indices of dryness/wetness, and monsoon failures and are based on periods of data that vary from 25 to 108 years. The criteria for droughts/floods and dryness/wetness also vary from one study to another. The rain gauge network used in these studies is highly variable over the period of data, and for some stations and for some months or years rainfall values are missing, except in the case of studies by Mooley et al.

(1981, 1982) and Mooley and Parthasarathy (1982, 1983a, 1983b, 1984a), who used the fixed network of 306 rain gauge stations evenly distributed over the country and the longest possible rainfall series that could be constructed for India. All these studies, however, are able to bring out generally the chief features of the variability of the Indian monsoon rainfall and years of well-marked deficiency/excess of the monsoon rainfall. The missing data problem, as we face it in actual practice, is threefold: (1) missing due to nonexistence of rain gauge stations in prior periods, which comes under the category of systematic missing; (2) missing during the periods of existence of the stations, either systematically or randomly over a specific period or a combination of missing of both types; and (3) missing randomly during the periods of existence of the stations, which comes under the category of random missing. A study of the problem covering these three categories of missing data and covering large areas and long periods is very difficult due to the complexity of the problem. It is, however, clear that sampling of different stations during the years covered by the period of data is likely to superpose artificial variability over the natural interannual variability. It is, therefore, advisable to keep to the same network of rain gauge stations while studying the variability and other aspects of rainfall over an area. The network of rain gauge stations used by Mooley et al. (1981, 1982) and Mooley and Parthasarathy (1982, 1983a, 1983b) is fixed, is evenly distributed over the country (one rain gauge station per district), and covers the longest possible period, namely, 1871 to 1978; in view of this, the details of interannual variability will be presented on the basis of rainfall data from this network. Wherever possible, a few more years of rainfall data have been added. The details with respect to the network and the sources from which monthly rainfall data were collected, the method of filling the few data gaps that existed, and the method of constructing all-India and subdivisional rainfall series have been given in Mooley et al. (1981) and Mooley and Parthasarathy (1983b). The network they utilized and the predominately hilly area they excluded are shown in Fig. 2.6. The meteorological subdivisions in which the rain gauge stations are located are shown in Fig. 2.7. The area considered is contiguous India excluding the hilly subdivisions and hilly portions of a few subdivisions shown as hatched in Figs. 2.6 and 2.7, and hereafter this area will be referred to as India or the country, and average monsoon rainfall over this area will be referred to as all-India monsoon rainfall. On application of suitable statistical tests, they found the series to be homogeneous and random.

3.1.1 Statistical properties of all-India monsoon rainfall. Table 2.1 gives the mean monsoon rainfall (in millimeters and as a percentage of

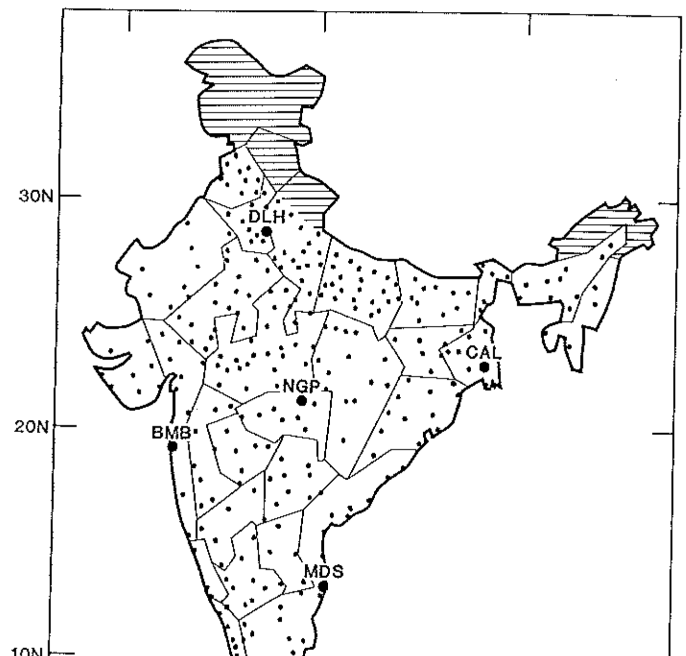

Fig. 2.6. Network of rain gauge stations over the area considered. The area considered is contiguous India excluding the hatched hilly area.

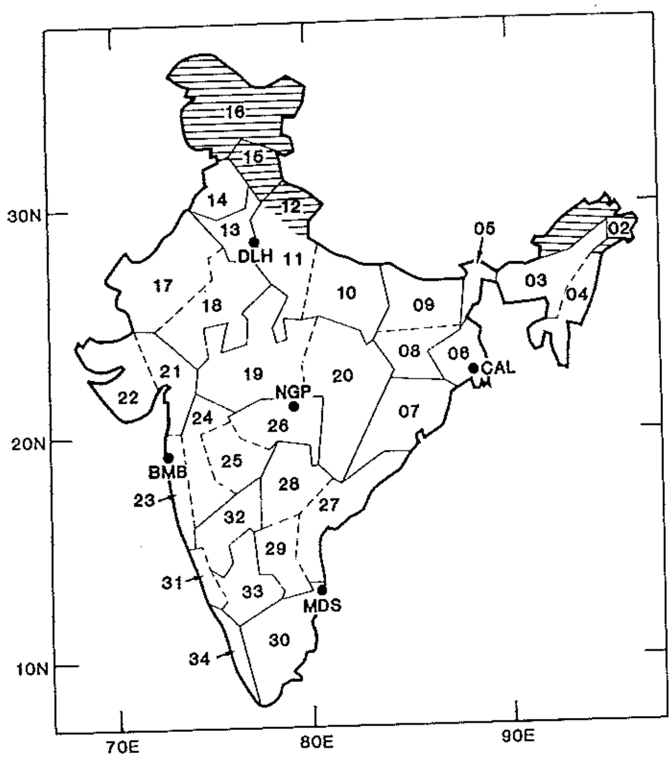

Fig. 2.7. Meteorological subdivisions of contiguous India. (subdivision 1 is Bay Islands and subdivision 35 is Arabian Sea Islands).

2 ARUNACHAL PRADESH
3 NORTH ASSAM
4 SOUTH ASSAM
5 SUB-HIMALAYAN WEST BENGAL
6 GANGETIC WEST BENGAL
7 ORISSA
8 BIHAR PLATEAU
9 BIHAR PLAINS
10 EAST UTTAR PRADESH
11 WEST UTTAR PRADESH PLAINS
12 WEST UTTAR PRADESH HILLS
13 HARYANA
14 PUNJAB
15 HIMACHAL PRADESH
16 JAMMU AND KASHMIR
17 WEST RAJASTHAN
18 EAST RAJASTHAN
19 WEST MADHYA PRADESH
20 EAST MADHYA PRADESH
21 GUJARAT
22 SAURASHTRA & KUTCH
23 KONKAN
24 MADHYA MAHARASHTRA
25 MARATHWADA
26 VIDARBHA
27 COASTAL ANDHRA PRADESH
28 TELANGANA
29 RAYALSEEMA
30 TAMIL NADU
31 COASTAL KARNATAKA
32 NORTH KARNATAKA
33 SOUTH KARNATAKA
34 KERALA

mean annual rainfall), the median, the lower and upper quartiles of the rainfall distribution based on the data for the period 1871 to 1984, and the autocorrelation coefficient with a lag of one for the period 1871 to 1978. It may be noted that the mean is slightly smaller than the median. In 50% of the years, the all-India monsoon rainfall was between 800 and 908 mm. The lag-one autocorrelation coefficient, though not statistically significant, is consistently negative for periods of different length, the values varying from −0.08 to −0.12.

Figure 2.8 shows the monsoon rainfall in standard units, namely, the deviation from the mean divided by the standard deviation, during each year of the period 1871 to 1984. A small deviation of up to 5% on either side of the long-term mean can be considered as normal or average rainfall. With this consideration, the following features are noticed from the rainfall series:

1. The decades 1881 to 1890 and 1941 to 1950 had normal or above-normal rainfall in all years except 1941, when rainfall was below normal.

TABLE 2.1 *Statistical properties of the all-India summer monsoon rainfall series (rainfall on large spatial scale), 1871–1984*

Property	Value	Property	Value
Mean	852 mm	Highest rainfall (1961)	1017 mm
Mean as % of annual mean	78.1	Deviation from mean	19%
Median	864 mm	Range	413 mm
Lower quartile	800 mm	Range as % of mean	48
Upper quartile	908 mm	Mean interannual variation	101.6 mm
Lag-one autocorrelation coefficient[a]	−0.12	Mean interannual variation expressed as % of mean	11.9
Standard deviation	83 mm	Mean interannual variation expressed in terms of standard deviation	1.22
Coefficient of variation	9.7%		
Lowest rainfall (1877)	604 mm	Highest interannual variation (1877–78)	377 mm
Deviation from mean	−29%	Lowest interannual variation (1895–96)	1 mm

[a]Based on data for the period 1871–1978.

2. During the periods 1871 to 1900 and 1921 to 1960 rainfall was normal or above normal in about 80% of the years, but during the period 1901 to 1920 rainfall was normal or above normal only in 45% of the years.

3. Occurrence of above-normal/below-normal monsoon rainfall in three or more consecutive years is very rare.

The different measures of variability of the all-India monsoon rainfall based on data for 1871 to 1984 are also given in Table 2.1. These are the standard deviation (S.D.), coefficient of variation (C.V.)—that is, the ratio of the standard deviation and the mean expressed as a percentage—range and extremes of rainfall, mean interannual variation in absolute and relative units, and extremes of interannual variation. The interannual variation is defined as $\Sigma_{i=1}^{n-1} |x_{i+1} - x_i|/(n - 1)$. The interannual variation is also expressed in relative terms, i.e., as a percentage of the mean and as a ratio of the standard deviation. The coefficient of variation is 10%. The interannual variation is 22% greater than the standard deviation. Landsberg (1951), who studied Oahu (Hawaiian Islands) annual rainfall, found that interannual variation is fairly close to the standard deviation. He has mentioned that for a Gaussian series, the ratio of interannual variation to standard deviation depends on the serial correlation coefficient

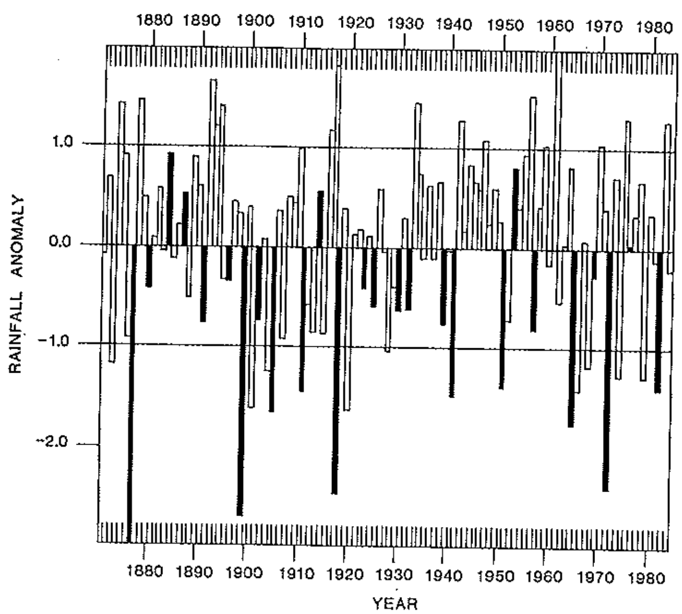

Fig. 2.8. All-India summer monsoon rainfall in standard units (deviation from normal divided by standard deviation), 1871–1984.

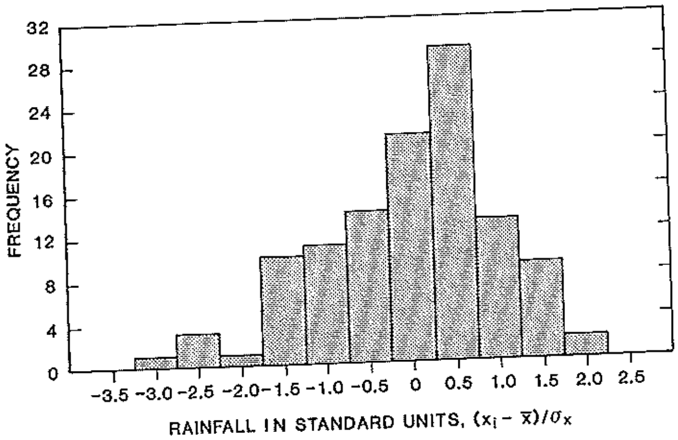

Fig. 2.9. Frequency distribution of the all-India monsoon rainfall (1871–1984). Class interval limits are given in standard units.

with a lag of one. For Indian monsoon rainfall, the ratio should be 1.19. The actual ratio obtained (1.22) is very close to this value.

The frequency distribution of all-India monsoon rainfall is shown in Fig. 2.9. The chi-square test was applied to test the goodness of fit of the Gaussian distribution to the monsoon rainfall distribution (Mooley and Parthasarathy, 1984a). The fit of the Gaussian distribution is found to be good, and monsoon rainfall can be considered to be Gaussian-distributed. In view of this, in 95% of the years all-India monsoon rainfall will be within 2 standard deviations from the mean, that is, between 686 and 1018 mm; and in 67% of the years it will be within 1 standard deviation from the mean, that is, between 769 and 935 mm. It may, however, be noted from Fig. 2.9 that the magnitudes of the highest negative anomalies are relatively much higher than those of the highest positive anomalies.

3.1.2 Trends and periodicities in all-India monsoon rainfall. The all-India monsoon rainfall series for the period 1871 to 1978 was subjected to a binomial low-pass filter, per the procedure given by the World Meteorological Organization (WMO) (1966), by Mooley and Parthasarathy (1984a) to study the fluctuations. A five-term binomial filter was used. The smoothed rainfall RS_i in i^{th} year was obtained from the equation

$$\mathrm{RS}_i = 0.06R_{i-2} + 0.25R_{i-1} + 0.38R_i + 0.25R_{i+1} + 0.06R_{i+2} \quad (2.1)$$

They observed that the filtered rainfall was below normal during the periods 1895 to 1932 and 1965 to 1974, and above normal during the periods 1872 to 1894 and 1933 to 1964.

Power spectrum analysis of the series has been carried out by Mooley and Parthasarathy (1984a), following the procedure given by WMO (1966), to locate periodicities in the series. The power spectrum is given in Fig. 2.10. It is seen from the spectrum that there are two cycles, one of 2.8 years (significant at the 5% level) and the other of 2.3 years (significant at the 10% level), in the Quasi-Biennial-Oscillation (QBO) range. To examine whether the cycles in the QBO range were stable, they applied power spectrum analysis to the two halves of the series. The spectra for the two periods, 1871 to 1924 and 1925 to 1978, are also given in Fig. 2.10. The cycle of 2.8 years is observed in the latter half (1925 to 1978) of the period only, significant at the 10% level. In view of this finding, they applied power spectrum analysis to the monsoon rainfall for the periods 1871 to 1900, 1871 to 1910, . . . , 1871 to 1970, and 1871 to 1978. No QBO was observed during the period 1871 to 1950. Thereafter, QBO was observed for the periods 1871 to 1960, 1871 to 1970, and 1871 to 1978, significant at the 5% level.

3.1.3 Droughts and floods over India. The series of monsoon rainfall has been examined for the occurrence of drought/flood by following the criterion that rainfall is $< -1.28/> +1.28$ in standard units. This criterion enables one to compare droughts/floods over India with those over another area of approximately the same size, since the criterion is based on the value of the deviation from the mean rainfall in terms of the standard deviation. On probability consideration, monsoon rainfall being less than -1.28 or greater than $+1.28$ has a probability of about 10%, in view of the Gaussian nature of the distribution of all-India ranifall. The years of drought and flood during the period 1871 to 1984, as identified by this criterion, are given in Table 2.2, along with rainfall in standard units. It is seen from this table that during the period 1871 to 1900, 2 droughts (i.e., 0.66 per decade) and 3 floods (i.e., 1 per decade) occurred. During the periods 1901 to 1920, 1921 to 1960, and 1961 to 1984, the figures for drought are 5 (2.5 per decade), 2 (0.5 per decade),

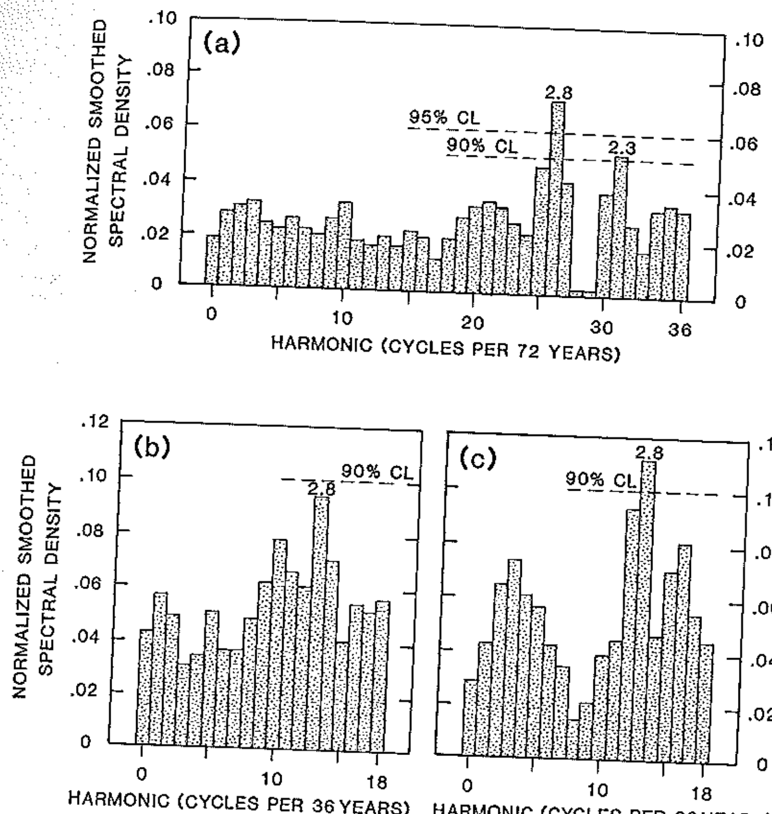

Fig. 2.10. Power spectra of the all-India summer monsoon rainfall. (*a*) For 1871–1978. (*b*) For 1871–1924. (*c*) For 1925–1978. (After Mooley and Parthasarathy 1984a.)

TABLE 2.2 *Droughts/floods over India during the period 1871–1984 (large spatial scale)*

Serial Number	Drought Year	Monsoon Rainfall mm	Monsoon Rainfall Standard Units	Flood Year	Monsoon Rainfall mm	Monsoon Rainfall Standard Units
1	1877	604	−2.99	1874	971	1.43
2	1899	628	−2.70	1878	974	1.47
3	1901	719	−1.60	1892	990	1.66
4	1905	715	−1.65	1894	969	1.41
5	1911	733	−1.43	1917	1003	1.82
6	1918	648	−2.46	1933	973	1.46
7	1920	717	−1.63	1956	980	1.54
8	1941	729	−1.48	1961	1017	1.99
9	1951	737	−1.39	1975	960	1.30
10	1965	707	−1.75	1983	959	1.29
11	1966	735	−1.41			
12	1972	653	−2.40			
13	1979	746	−1.28			
14	1982	736	−1.40			
Mean		701	−1.83		980	+1.54

and 5 (2.1 per decade), respectively, and those for flood are 1 (0.5 per decade), 2 (0.5 per decade), and 3 (1.25 per decade), respectively. Thus, the periods 1901 to 1920 and 1961 to 1984 had relatively much higher drought frequency than did the periods 1871 to 1900 and 1921 to 1960. The period 1921 to 1960 had the lowest frequency of drought as well as the lowest combined frequency of drought and flood; thus it is the best period in the recorded history of the Indian monsoon. This

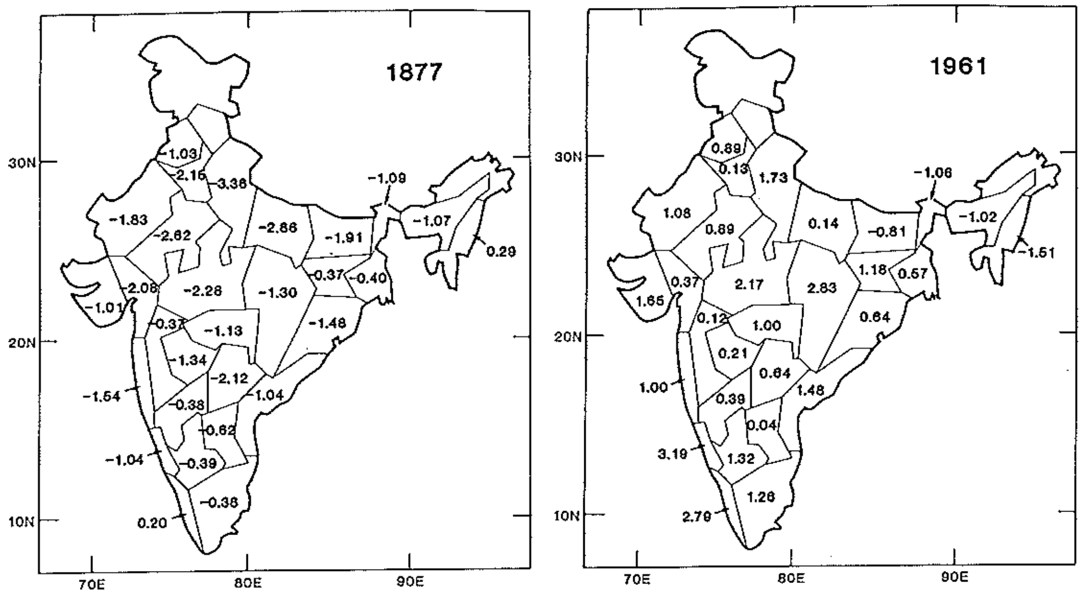

Fig. 2.11. Distribution of subdivisional monsoon rainfall in standard units in 1877 and 1961, the years of lowest and highest all-India monsoon rainfall, respectively.

observed alternation of periods of low and high drought frequency over India—namely, low (during about 40 years prior to 1900) → high (during 1901 to 1920) → low (during 1921 to 1960) → high (during 1961 to 1984)—suggests that there are no systematic trends in the Indian monsoon rainfall.

Figure 2.11 shows the geographical distribution of the monsoon rainfall in standard units for the different meteorological subdivisions of India during 1877, the year with the lowest all-India rainfall on record, and during 1961, the year with the highest all-India rainfall on record. During 1877, it may be mentioned that with the exception of southern Assam and Kerala, the rainfall of all the subdivisions of India was below normal and that the rainfall deficiency was particularly large over northwestern India, the central parts of the country, the Konkan-Karnataka coasts, and northern parts of the peninsula. During 1961, there was a large excess of monsoon rainfall over central India, some parts of northwestern India, and most parts of the southern peninsula; in contrast, Assam, sub-Himalayan western Bengal, and the Bihar Plains suffered well-marked rainfall deficiency.

3.2 Medium/regional-scale variability

The first exhaustive study of the monsoon rainfall on this scale was made by Walker (1910a). He examined the rainfall variation over northwestern India of preindependence days for the years 1841 to 1908, and he concluded that the monsoon rainfall deficiency over this part of the country had not lasted sufficiently long to justify the conclusion that there had been a permanent change in climate.

Monsoon rainfall on this scale has also been studied by Ramdas (1950, 1960), Rao (1958), Rao and Jagannathan (1960, 1963), Raghavendra (1974, 1976, 1980), Shukla (1987a), and Parthasarathy (1984a, 1984b).

Parthasarathy (1984a, 1984b) studied the monsoon rainfall of all the subdivisions of India for variability, trends and cycles, incidence of droughts and floods, and interrelationship. He utilized the monthly rainfall data of the fixed network of stations in each of the subdivisions, one station from each district, and stations evenly distributed over most of the subdivisions for the period 1871 to 1978. The different aspects of the medium- or regional-scale monsoon rainfall that will be discussed here are mostly based on his studies. However, as far as possible, data have been extended, and the results presented here are based on the data for the period 1871 to 1984. It may, however, be mentioned that rainfall data for the period 1981 to 1984 are available for about 80% of the stations, but this is not likely to affect the properties of the subdivisional rainfall when we are considering the long period 1871 to 1984. These regional rainfall series have been found to be homogeneous and free from persistence.

*3.2.1 **Statistical properties of medium-scale monsoon rainfall.*** Table 2.3 gives mean and median monsoon rainfall based on the data for the period 1871 to 1984. The lowest subdivisional mean is 252 mm for West Rajasthan and the high-

TABLE 2.3 *Mean, median, standard deviation (S.D.), coefficient of variation (C.V.), and interannual variation (IAV)/S.D. for medium-scale monsoon rainfall over India, 1871–1984*

Subdivision Number	Subdivision Name	Mean (mm)	Median (mm)	S.D. (mm)	C.V. (%)	IAV/S.D.
03	North Assam	1455	1458	165	11.3	1.07
04	South Assam	1452	1439	181	12.4	1.14
05	Sub-Himalayan West Bengal	1984	1947	299	15.0	1.13
06	Gangetic West Bengal	1149	1135	174	15.1	1.18
07	Orissa	1175	1163	161	13.7	1.10
08	Bihar Plateau	1103	1070	163	14.8	1.18
09	Bihar Plains	1029	1023	194	18.8	1.13
10	East Uttar Pradesh	911	912	206	22.6	1.07
11	West Uttar Pradesh, plains	770	789	179	23.2	1.16
13	Haryana	459	454	137	29.8	1.09
14	Punjab	495	469	168	33.9	1.12
17	West Rajasthan	252	248	99	39.2	1.10
18	East Rajasthan	641	642	168	26.2	1.15
19	West Madhya Pradesh	922	920	166	18.0	1.09
20	East Madhya Pradesh	1207	1228	191	15.8	1.14
21	Gujarat	871	873	267	30.6	1.18
22	Saurashtra and Kutch	438	414	190	43.4	1.16
23	Konkan	2378	2407	469	19.7	1.08
24	Madhya Maharashtra	580	579	122	21.0	1.11
25	Marathwada	639	694	186	26.8	1.08
26	Vidarbha	898	898	179	19.9	1.18
27	Coastal Andhra Pradesh	505	495	111	22.0	1.20
28	Telangana	717	722	163	22.7	1.10
29	Rayalaseema	423	397	121	28.6	1.21
30	Tamilnadu	310	304	72	23.2	1.10
31	Coastal Karnataka	2848	2802	509	17.9	1.19
32	North Karnataka	603	601	117	19.4	1.14
33	South Karnataka	502	497	102	20.3	1.26
34	Kerala	1936	1883	366	18.9	1.14

est is 2848 mm for coastal Karnataka. Thus the spatial variation of the mean is one order of magnitude. Over northern India, the mean rainfall progressively decreases from east to west. The mean is highest over the west coast of peninsular India, decreases eastward over the plateau for some distance, and thereafter increases as the east coast is approached. Over the plateau east of the western Ghats, mean rainfall decreases from north to south. Along the west coast of peninsular India mean rainfall is highest over the central portion of the coast. It is seen from the table that the median monsoon rainfall is generally close to the mean.

Table 2.3 also gives the measures of variability of the regional monsoon rainfall—standard deviation and coefficient of variation. In general, the coefficient of variation is high over areas of low rainfall and low over areas of high rainfall. The lowest coefficient of variation of 11% is found over North Assam, and the highest one of 43% over Saurashtra and Kutch. Over northeastern and central India, the coefficient of variation is low, generally 12 to 18%, but over northwestern India and the Marathwada and Rayalaseema subdivisions in peninsular India, it is relatively high, generally 25 to 40%. Over the remaining parts of India the coefficient of variation is 18 to 25%. It is seen that while the coefficient of variation for monsoon rainfall on a large scale is 10%, that for monsoon rainfall on the medium scale varies from 11 to 44%.

Table 2.3 also gives the ratio of mean interannual variation (IAV) to standard deviation of the monsoon rainfall of the different subdivisions. The interannual variation is the mean absolute variation from one year to the next, namely, $(\Sigma_{i=1}^{n-1} |x_i - x_{i+1}|)/(n - 1)$.

The regional monsoon rainfall series were examined for outliers. The few outliers that were observed in some of the series were removed, and all the series were tested for normality. The subdivisional rainfall series were found to be Gaussian.

3.2.2 Relationship between all-India monsoon rainfall and subdivisional monsoon rainfall. Parthasarathy (1984a, 1984b) computed correlation coefficients (CCs) between all-India monsoon rainfall and monsoon rainfall series for the subdivisions of India, and also the correlation coefficients between each subdivisional series and the remaining subdivisional series. He observed that the rainfall of the subdivisions of northeast India—consisting of North Assam, South Assam, sub-Himalayan West Bengal, Gangetic West Bengal, the Bihar Plains, and the Bihar Plateau—has no significant relationship with all-India rainfall, but the monsoon rainfall of subdivisions that lie west of 84°E is highly and significantly related to all-India monsoon rainfall, the correlation coefficient of 0.24 for Kerala, though not significnt at the 1% level, being close to this significance level. The maximum correlation coefficient is 0.76 with East Rajasthan rainfall. The correlation coefficient with West Madhya Pradesh rainfall is slightly less, 0.74.

3.2.3 Trends and periodicities in subdivisional monsoon rainfall. Parthasarathy (1984a, 1984b) has examined the subdivisional rainfall series for the period 1871 to 1978 for increase or decrease of the mean from the first half to the second half of the period by applying the Student's t test and the Mann-Whitney rank test. The subdivisions for which both the tests show significant change in the mean from the first half to the second half are Konkan, Punjab, and Telangana. The series for Konkan shows an increase from the period 1871 to 1924 to the period 1925 to 1978, significant at the 1% level by the Mann-Whitney test and at the 2% level by the t test. The Punjab and Telangana rainfall series show an increase from the first half to the second half of the period, significant at the 10% level only by both tests. On the whole, it can be inferred that except for Konkan, the subdivisional rainfall series do not show any notable difference in the means for the two halves of the series.

To locate significant periodicities in the subdivisional rainfall, Parthasarathy (1984b) examined the correlogram and power spectrum of these series for the whole period of data as well as for the two halves of the period. He found autocorrelation with lag 14 to be significant (1% level) for the series for Punjab. The periodicities he found to be significant in the spectrum for the whole period as well as for each half of the whole period are in the range of 2.2 to 3.0 years, generally referred to as quasi-biennial oscillation (QBO), in the series for Vidarbha, coastal Andhra Pradesh, Tamil Nadu, South Karnataka, and North Karnataka, and 3.6 years in the series for Kerala. These periodicities generally account for less than 10% of the total variance. It may be noted that the power spectrum analysis of the series for Punjab does not support the 14-year cycle observed in the correlogram. Cycles in the range of 6 to 8 years significant at the 10% level are observed in the series for east Uttar Pradesh, Punjab, and Madhya Maharashtra, and cycles around 12 years significant at 10% are observed in the series for West Rajasthan and West Madhya Pradesh. But these cycles lack consistency; that is, they are observed in only one of the halves of the series or not observed in any of the halves of the series.

3.2.4 Incidence of droughts and floods over the subdivisions of India. The criteria adopted for identifying drought and flood over the different sub-divisions are as follows:

$$\text{Drought:} \quad \frac{x_i - \overline{x}}{\sigma_x} < -1.28$$

$$\text{Flood:} \quad \frac{x_i - \overline{x}}{\sigma_x} > 1.28$$

where x_i is the monsoon season rainfall in the i^{th} year, $\overline{x}$ is the long-period mean monsoon rainfall, and σ_x is the standard deviation of the monsoon rainfall. The nondimensional quantity $(x_i - \overline{x})/\sigma_x$ is referred to as the standardized rainfall or normalized rainfall anomaly. Since the regional monsoon rainfall is generally Gaussian, -1.28 and $+1.28$ are 10th and 90th percentiles of the rainfall distribution. Thus the long-term probability of drought/flood is about 10%. By use of the standardized monsoon rainfall series for the different meteorological subdivisions, the years of drought/flood over each subdivision have been identified. Figure 2.12 brings out the incidence of drought/flood over each subdivision during each year of the period 1871 to 1984. The last two columns in this figure give the frequencies of drought/flood for the whole period. It is observed from the figure that a large number of subdivisions experienced drought in 1877, 1899, 1918, and 1972 and flood in 1878, 1892, 1933, 1938, and 1983.

The incidence of drought—that is, the number of subdivisions affected by drought—was low during the decades 1881 to 1890, 1931 to 1940, and 1941 to 1950, and was high during the decades 1871 to 1880, 1891 to 1900, 1901 to 1910, and 1911 to 1920. It will be useful to investigate whether the main features of the atmospheric circulation during the period of high drought incidence (1891 to 1920) were appreciably different from those during the period of low drought incidence (1921 to 1960). The monsoon trough extending from northwestern Rajasthan to the north Bay of Bengal is the most prominent synoptic system with which the Indian monsoon rainfall is associated. The shift of this trough to the foot of the Himalayas profoundly influences the rainfall. Most of the country gets little rain as long as the trough remains over the foot of the Himalayas. Such situations are referred to as a 'break' in the monsoon. The periods of 'break' have been documented by Ramamurthy (1969) for the period 1888 to 1967 for the core monsoon months July

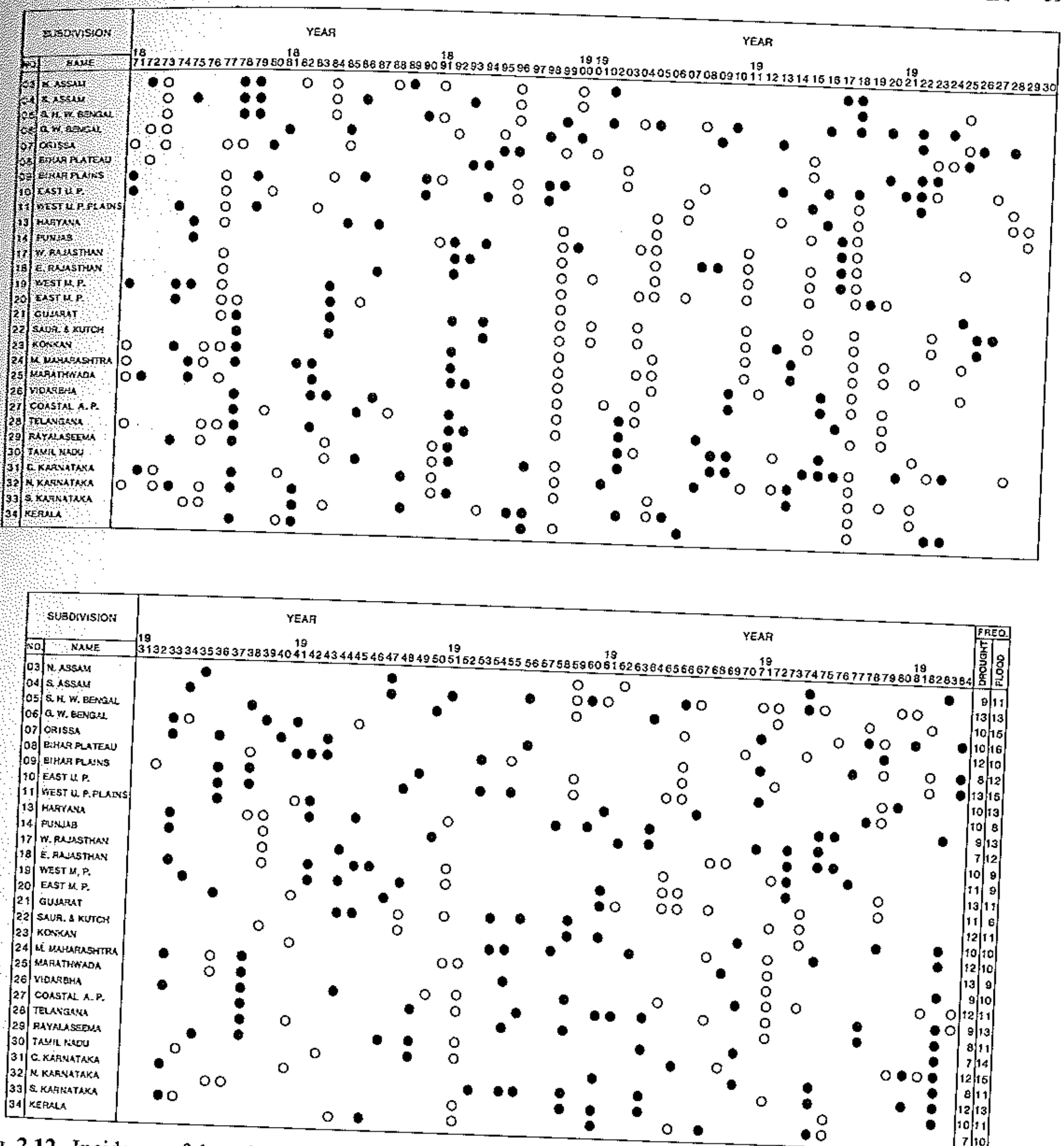

Fig. 2.12. Incidence of drought/flood over the meteorological subdivisions of India during the summer monsoon season, 1871–1984 (drought o; flood •).

and August. For further periods, the information is obtained from the publications of the Indian Meteorological Department. On examination of this information on 'break' it is found the numbers of days of 'break' for the decades 1891 to 1900, 1901 to 1910, and 1911 to 1920 are 97, 109, and 120, respectively; while those for the decades 1921 to 1930, 1931 to 1940, 1941 to 1950, and 1951 to 1960 are 67, 56, 66, and 73, respectively. Thus, the atmospheric circulation features that are favorable for keeping the monsoon trough over the foot of the Himalayas prevailed with much higher frequency during the period 1891 to 1920 than during the period 1921 to 1960. In recent decades, 1961 to 1970 and 1971 to 1980, the frequency of days of 'break' has shown an increase, the values of the past two decades being 95 and 83, respectively.

The incidence of flood, in general, shows little variation from decade to decade. The number of subdivisions experiencing floods is relatively smaller during the decades 1921 to 1930 and 1901 to 1910.

The country's area under drought/flood has been computed for each year. Figure 2.13 shows the frequency distribution. The percentage fre-

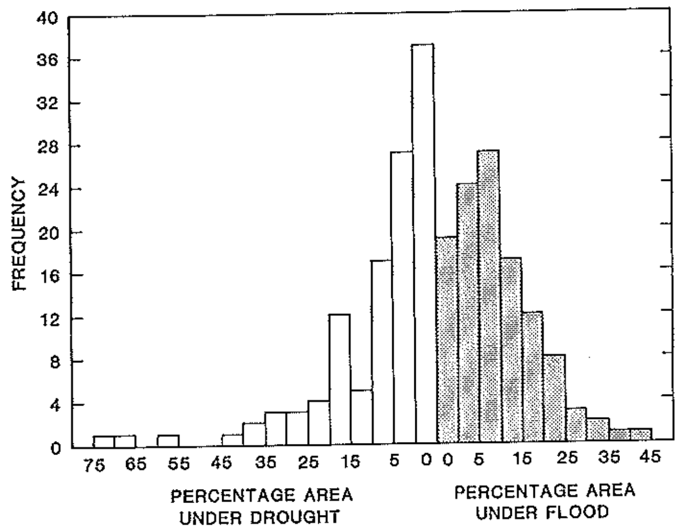

Fig. 2.13. Frequency distribution of the percentage area of India under drought/flood during the summer monsoon season (1871–1984).

quency of years with zero area under drought is 32, and that of years with area under drought of 10% or less is 71. Similar percentage frequencies for flood are 17 and 61, respectively. The percentage frequency of years with drought/flood area exceeding 20% of the country's area is 14/13. The frequency distribution of the percentage area under drought differs from that for the percentage area under flood in three respects: (1) Extreme droughts affect much larger areas than extreme floods (however, if the three extreme and unusual years, 1877, 1899, and 1918, are not considered, both the distributions extend up to 45% area); (2) the percentage frequency of zero area under drought is much higher than that for zero area under flood; (3) from the maximum frequency for zero drought area, the frequencies of different class intervals of drought area generally taper off in the direction of increasing percentage area, but in the case of frequency distribution of flood area, the frequency is maximum for the interval 5 to 10% and tapers off slowly with decreasing area under floods and rather rapidly with increasing area under flood.

The years of large-scale drought/flood over India can be identified on the basis of percentage area under drought/flood. For the criterion of drought/flood area exceeding the 90th percentile of the distribution of the percentage drought/flood area, for large-scale drought/flood over India, such years of large-scale drought/flood have been identified during the period 1871 to 1984. The 90th percentiles of the drought and flood area percentage series are 25.0 and 21.5, respectively. The years of large-scale drought/flood over India in order of decreasing area under drought/flood are given below, along with percentage area under drought/flood in parentheses.

Drought	1899 (70.6), 1918 (65.8), 1877 (56.8), 1905 (40.1), 1972 (37.1), 1915 (36.5), 1911 (34.3), 1966 (30.6), 1904 (30.4), 1965 (29.5), 1979 (29.4), 1920 (26.6)
Flood	1892 (40.9), 1878 (36.7), 1983 (31.6), 1961 (31.5), 1938 (28.4), 1933 (27.4), 1884 (26.0), 1917 (24.9), 1936 (24.9), 1874 (22.8), 1875 (21.9), 1973 (21.7)

These years are not identical with the years of all-India drought/flood as given in Table 2.2. They are, however, not expected to be identical since in one case the criterion is based on percentage area under drought/flood and in another case it is based on standardized all-India rainfall. Standardized all-India rainfall can be low as a result of a few subdivisions having very low rainfall. In view of this, the years of large-scale drought/flood over India based on the area of drought/flood over India are more representative of drought/flood conditions over the country. When the total water deficiency/excess over the country is considered, the years of drought/flood based on all-India standardized rainfall are more useful.

Figure 2.14 gives the rainfall distribution over the subdivisions of India during 1899 and 1892, the years of maximum incidence of drought and flood. During 1899, when most parts of India were suffering from large monsoon rainfall deficiency, Assam, West Bengal, the Bihar Plains, and East Uttar Pradesh were experiencing excess monsoon rainfall. The deficiency over and around the central parts of India was high, rainfall below the long-term mean being more than 2 standard deviations. In 1892, when most parts of the country were experiencing rainfall excess, Assam, Gangetic West Bengal and the Bihar Plateau were experiencing rainfall deficiency. The high excess of rainfall exceeding the mean by more than 2 standard deviations was experienced by the subdivisions of Telangana, Rayalaseema, Marathawada, and North Karnataka in peninsular India. Thus, even in extreme years, the inverse relationship between the rainfall of subdivisions of northeastern India and the rainfall of the subdivisions of the remaining India is observed.

3.3 Small-scale variability

Koteswaram and Alvi (1969, 1970) studied the fluctuations in annual and monsoon rainfall of 20 stations over India, using all available data up to 1967. They found an increasing tendency in the rainfall of west coast stations. Saha and Mooley (1978) examined the monsoon rainfall series of 39 stations in the field of the Asian summer monsoon based on all available data up to 1960, which generally varied from 70 to 100 years. The network contained 18 well-distributed stations from India. They did not find any significant trend or oscillation in the station monsoon rainfall series. Parthasarathy (1984a) has examined the geographical distribution of the mean monsoon rainfall, concentration of rainfall during the summer monsoon season, and coefficient of variation of summer monsoon rainfall of 306 stations evenly distributed over the country for the period 1871 to 1978. These geographical distributions are given in Figs. 2.15, 2.16, and 2.17, respectively.

Figure 2.15 shows the steep variation of monsoon rainfall across the western ghats, the high spatial variation over Tamilnadu and West Rajasthan, and also pockets of monsoon rainfall exceeding 150 cm over Assam and 125 cm over East Madhya Pradesh.

Figure 2.16 shows that while 70 to 95% of the annual rainfall occurs during the monsoon season over most parts of India, only 20% of the annual rainfall occurs during the monsoon season over extreme southeastern Tamilnadu.

Figure 2.17 brings out the high coefficient of variation of small-scale monsoon rainfall, exceeding 60%, over extreme south-eastern Tamilnadu and extreme West Rajasthan. Generally, it is seen that variability of monsoon rainfall on a small scale is much higher than that on the regional scale.

4 INFLUENCE OF GLOBAL SURFACE BOUNDARY CONDITIONS ON MONSOON CIRCULATION

The earth's surface consists of land, water, ice, and snow, and the percentages of these components exhibit seasonal variation. In addition to seasonal variation, these components undergo interannual variation.

Since the global boundary conditions have the potential to influence the monsoon circulation significantly (Charney and Shukla 1981), in this section we review the observational and modeling studies on this topic.

We shall consider the influence of the different boundary forcings, snow cover, sea surface temperature (SST), El Niño, southern oscillation (an oscillation in surface pressure strongly linked to SST distribution), albedo changes, and soil moisture.

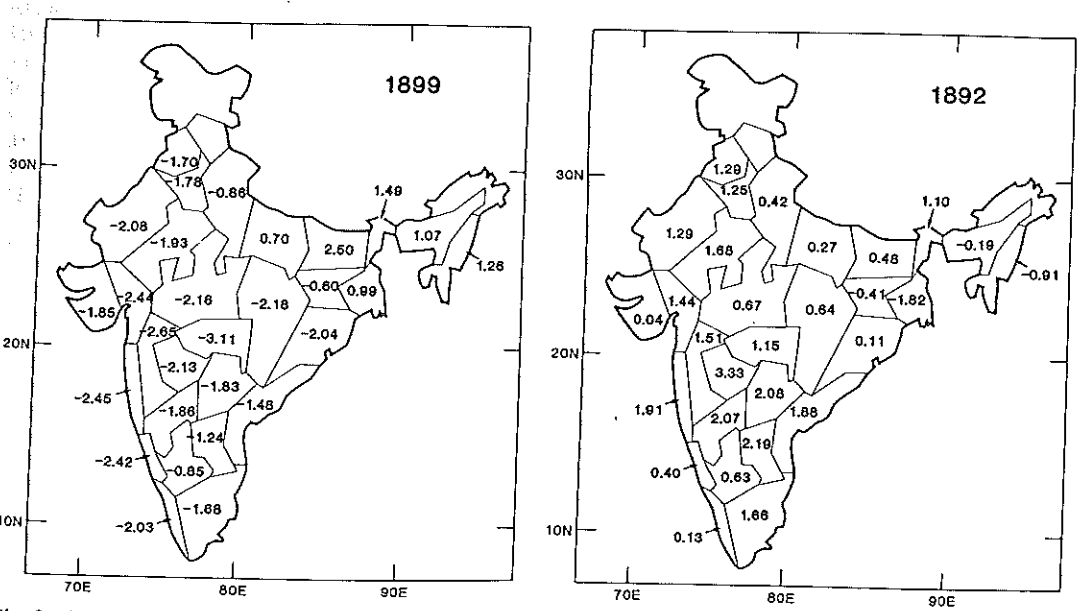

Fig. 2.14. Distribution of subdivisional monsoon rainfall in standard units in 1899 and 1892, the years of maximum incidence of drought and flood, respectively, over India.

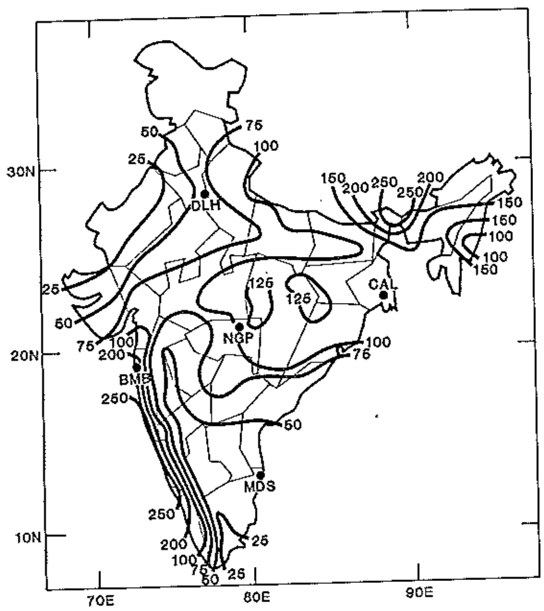

Fig. 2.15. Geographical distribution of the mean monsoon rainfall (cm) over India, 1871–1978. (After Parthasarathy 1984b.)

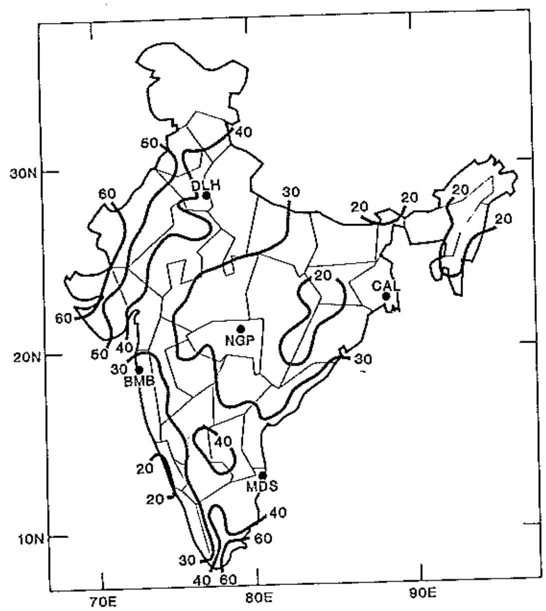

Fig. 2.17. Geographical distribution of the coefficient of variation (%) of the monsoon rainfall over India, 1871–1978. (After Parthasarathy 1984b.)

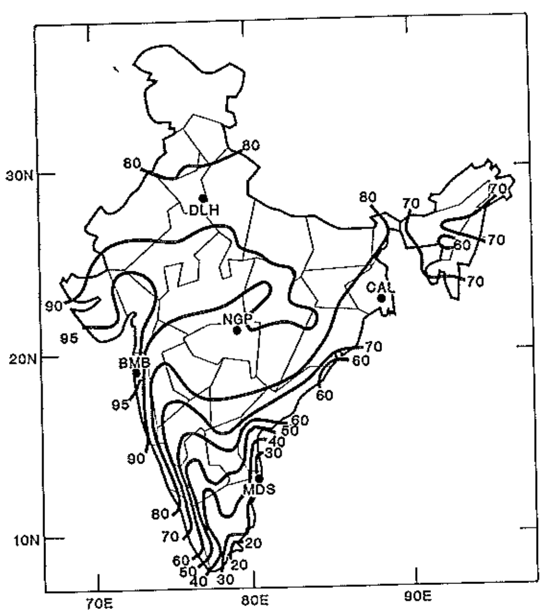

Fig. 2.16. Geographical distribution of the concentration of rainfall during the summer monsoon season over India (percentage of annual), 1871–1978. (After Parthasarathy 1984b.)

4.1 Snow cover

Blanford (1884) observed that "the excessive winter and spring snowfall in the Himalayas is prejudicial to the subsequent monsoon rainfall in India." This observation was later substantiated by Walker (1910b). According to Jagannathan (1960), the amount and time of occurrence of cold weather (October through May) snowfall in the mountain districts adjacent to northern India was one of the important factors used by Blanford for monsoon rainfall forecasts, which he started issuing from 1882. Higher winter snowfall was found to be related to deficient monsoon rainfall during the period 1880 to 1920. However, for the subsequent 30-year period reported snow accumulations showed very large variability, and the relationship with the monsoon rainfall was opposite to that in the prior four decades. In view of this change in the character of this parameter, the use of this parameter was discontinued after 1950.

This parameter went into oblivion until the earth-orbiting satellites made snow cover data available. Then Wiesnet and Matson (1976) showed on the basis of these data that December snow cover for the Northern Hemisphere is a very good predictor of the following January-through-March snow cover.

Hahn and Shukla (1976) found an apparent inverse relationship between Eurasian snow and the Indian monsoon rainfall during the period 1967 to 1977. Large and persistent snow cover anomalies over Eurasia can produce a colder midlatitude troposphere in the following spring, which can strengthen the upper-level anticyclone, slow its northward movement over India, and lead to delayed and weaker monsoon rainfall.

Dickson (1983, 1984) extended this study up to 1980. On deleting the erroneous data for 1969 and adjusting the data for 1967 to 1974 for the bias resulting from noninclusion of Himalayan snow cover, he found that the correlation coefficient improved from −0.44 to −0.59.

Figure 2.18 shows the normalized anomaly (i.e., anomaly divided by S.D.) of Indian monsoon rainfall and Eurasian December-through-March snow cover from 1967 to 1984. The snow cover data from 1967 to 1982 (adjusted for bias for the period 1967 to 1974) were obtained from Dickson (1986), and for 1983 and 1984, from Ropelewski (1986). As mentioned by Dickson (1984), snow cover for 1969 is not reliable and has not been considered for computation of the correlation coefficient. Figure 2.18 brings out the inverse relationship between rainfall and Eurasian snow cover. The correlation coefficient for the period 1967 to 1979 is −0.57, which is just significant at the 5% level. However, for the period 1967 to 1984, the correlation coefficient is −0.45, which fails to attain significance, the coefficient significant at 5% being 0.48.

Dey and Bhanukumar (1982) have found a direct relationship between the anomaly of Eurasian snow cover during spring and the date of onset of monsoon over the southern tip of India as well as the time taken by the monsoon to advance over the whole country on the basis of data for the period 1967 to 1978. They obtained an inverse relationship between snowmelt during spring (snow cover in March minus snow cover in May) and the period of monsoon advance over the whole of India. However, Eurasian snow cover data used by them did not include snow cover over the Himalayan region during the period 1967 to 1974, and they did not apply any correction to the data for this period. Ropelewski et al. (1984) computed correlation coefficients between Eurasian spring snow cover and monsoon advance date as well as between spring snowmelt and monsoon advance period, for each of the two periods 1966 to 1972 and 1973 to 1978 and found that these coefficients were not significant at the 5% level.

Chen and Yan (1978) and Yeh et al. (1981) have shown a possible relationship between winter-spring snow cover over Eurasia and variability of atmospheric circulation and rainfall over India and China.

An inverse relationship between the Eurasian snow cover and summer monsoon rainfall is understandable, since large and persistent Eurasian snow cover would substantially reduce the rate of heating of the concerned land masses during spring and summer and thus delay the onset of the monsoon and prevent normal monsoon activity. It may be mentioned that the snow cover area as measured by the satellite is not quite a representative parameter to assess the amount of snow. The snow cover area in two cases may be the same but depths may be quite different, and the impact on monsoon would be quite different in the two cases.

4.2 Sea surface temperature (SST), El Niño, and southern oscillation

The northward flow of cold water off the Peru coast is due to the upwelling caused by the strong southeastern trade winds. The onset of a southward coastal current heralds the commencement of annual warming of the surface off the Peru coast. This warming, which marked the end of the fishing season, was named El Niño (the child) by coastal inhabitants, since it appears at about the Christmas season. However, an unusually large rise in SST occures every few years. Such a large rise is referred to as the El Niño phenomenon.

The southern oscillation was suggested by Hildebrandsson (1897) and was confirmed by Lockyer and Lockyer (1904) as a pressure seesaw between the Indian Ocean and Argentina. Walker (1924a), while trying to develop techniques for

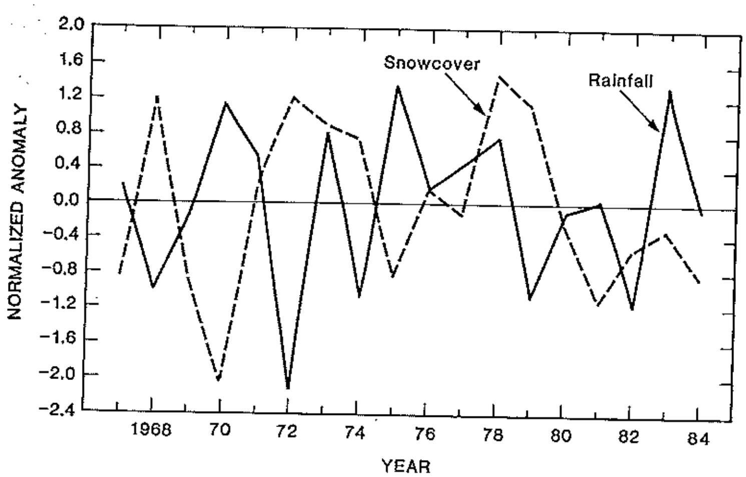

Fig. 2.18. Normalized anomaly of Indian monsoon rainfall and of Eurasian December-to-March snow cover, 1967–84.

monsoon prediction, was able to establish three pressure oscillations—North Atlantic, North Pacific, and southern oscillation. He showed that the third one, the southern oscillation, is more extensive, has large year-to-year fluctuations, and has its centers of action in the lower latitudes of the Southern Hemisphere. Walker and Bliss (1932) described it as a high pressure in the Pacific Ocean associated with a low pressure in the Indian Ocean from Africa to Australia. Studies during the past two decades have brought out that the southern oscillation is the oscillation between the pressure over the Indian Ocean–Australian region and the Southeast Pacific. When pressure is high over one area, it is low over the other area, Tahiti (17.5°S, 149.5°W) and Darwin (12.4°S, 130.9°E) being the opposite poles of the southern oscillation. Recent studies have clearly shown that the El Niño and southern oscillation are strongly coupled, and these are now referred to as the El Niño–southern oscillation (ENSO) phenomenon.

4.2.1 Sea surface temperature. About 70% of earth's surface is covered with water, and changes in SST take place slowly. In view of this position, it is reasonable to expect that interannual variability of the SST may account for part of the interannual variability of the atmospheric circulation and rainfall.

Bjerknes (1966, 1969, 1972), Rowntree (1972), Shukla (1975), and WMO (1977) have shown that the atmosphere is closely coupled to the underlying ocean in the tropics. Studies by Shukla and Bangaru (1978), Khandekar (1979), Weare (1979), Anjaneyulu (1980), Joseph (1981, 1983), Pisharoty (1981), Shukla (1982), Webster (1981), Newell et al. (1982), and Rasmusson and Carpenter (1982) have brought out the role played by the SST of tropical oceans in modifying atmospheric circulation and in the distribution of cloudiness and precipitation.

It has been pointed out by Shukla and Bangaru (1978) and Shukla (1982) that the influence of SST anomalies on the atmosphere depends on the magnitude and structure of the anomalies, normal SST, the latitudinal location of the anomalies (Hoskins and Karoly 1981; Webster 1981), the prevailing circulation regime, the instability mechanism, and the structure of the zonal flow.

Sometimes, persistent positive SST anomaly may result in anomaly in the location of the intertropical convergence zone (ITCZ) and anomalous weather. For example, the ITCZ remaining over a positive SST anomaly area for a long time leads to copious precipitation over the area and little or no precipitation over the area where the ITCZ should have been located according to its seasonal march. In the same way, the locations of Hadley and Walker cells are affected by SST anomalies, leading to anomalies in weather.

Most of the observational and numerical studies that have been made so far have considered the influence of regional SST anomalies only. We shall present briefly the results of these studies. It may, however, be mentioned that for a better understanding of the ocean-atmosphere interaction, the studies covering the influence of hemispheric and global-scale SST on the atmosphere are necessary.

4.2.1.1 SST over the Arabian Sea. Pisharoty (1965) computed the water-vapor flux across the equator into the Arabian Sea and across the Indian west coast from the Arabian Sea, and he found that the latter was more than twice the former. From this, he concluded that evaporation over the Arabian Sea is an important source of moisture that crosses the west coast of India and is available for precipitation.

Saha and Bavadekar (1973) used data for September 1963 and June to September 1964 and also used data for additional stations in the western Arabian Sea to cover the area of strongest northward flow, repeated the calculation of Pisharoty, and found that the flux of water vapor across the equator is about 30% larger than the evaporation over the Arabian Sea, a result contrary to the result of Pisharoty.

Ghosh et al. (1978) calculated the water vapor budget over the Arabian Sea on the basis of the Monsoon Experiment (MONEX) 1973 data. They got results similar to the results obtained by Pisharoty and they emphasized that the Arabian Sea plays a dominant role in the monsoon activity over the west coast of India.

Later calculations by Cadet and Reverdin (1981) support the results obtained by Saha.

As mentioned by Shukla (1987a), widely different results obtained by these investigators are largely due to the differences in the quality and density of data and in the techniques of analysis of data.

Shukla and Misra (1977), using SST data from Fieux and Stommel (1976), examined the relationship between SST along 10°N and between 60° and 70°E, and Weare (1979) calculated the empirical orthogonal functions (EOFs) for the mean monsoon rainfall of 53 Indian stations and for SST anomalies over the Arabian Sea and Indian Ocean, and the correlation coefficients between the EOFs of monsoon rainfall and of SST. Shukla (1983a), on the basis of critical analysis of SST anomaly along ships tracks over the Arabian Sea and Indian Ocean, found that the SST data used by Shukla and Misra (1977) and Weare (1979) suffered from the defect of significant negative bias before 1940 and positive bias thereafter. After correcting SST data for the bias, Shukla (1986a) composited SST anomaly for heavy/deficient monsoon rainfall years. The main features of these composites are (1) the ocean surface, at least along the ship tracks, is relatively warmer during April, May, and June for heavy rainfall years but is relatively colder during August, September, and October, (2) the differences in SST anomalies between heavy rainfall years and deficient rainfall years for August, September, and October are larger than those for April, May, June. These results suggest that heavy rainfall is associated with high wind speeds, which cool the Arabian Sea more than the weaker winds associated with deficient rainfall. These results are consistent with the earlier results that suggested a negative correlation between surface wind and SST anomaly, but they have little predictive value since premonsoon SST anomaly is small enough to be within the range of observational noise.

Sensitivity of model-simulated rainfall over India during monsoon to SST anomalies over the Arabian Sea and Indian Ocean has been studied by Shukla (1975, 1976) with the Geophysical Fluid Dynamics Laboratory (GFDL) global circulation model, by Washington et al. (1977) with National Center for Atmospheric Research (NCAR) model, and again by Shukla (1981a) with the Goddard Laboratory for Atmospheric Sciences (GLAS) model. However, the NCAR model was later found to be unable to simulate the mean monsoon circulation, and the sea surface anomalies used by Shukla in his experiments were later found (Shukla, 1986a) to be not realistic, being too high and lacking the change of sign from the earlier part to the latter part of the monsoon season as observed with the actual anomalies. Experiments with realistic anomalies need to be performed.

4.2.1.2 SST over the Pacific Ocean. Bjerknes (1969) proposed a thermally direct Walker circulation along the equator, with an ascending branch over the western Pacific Ocean and Indonesia and a descending branch over the East Pacific Ocean near the west coast of South America. In addition, he postulated that this equatorial circulation also exchanges absolute angular momentum with circulations to the north and south. Studies by Krishnamurti (1971) and Krishnamurti et al. (1973) have established east-west circulations for the Northern Hemisphere summer as well as winter. They have shown that the intensity of the Walker circulation is comparable to that of the Hadley circulation and that the east-west circulation consists of several cells in vertical zonal planes. This would suggest that changes in the Walker circulation caused by changes in the SST distribution over the Pacific would affect these cells in vertical zonal planes, which in turn would influence the Hadley and monsoon circulations.

From his theoretical studies related to response of tropical atmosphere to local steady forcing, Webster (1972, 1973) has shown that the correlations of Walker and others are the variations in the magnitude of the response of ultralong slow-moving waves of low latitudes.

Kidson (1975), Rao and Theon (1977), Quinn et al. (1978), Heddinghaus and Krueger (1981), and Liebmann and Hartmann (1982) have shown the relationship between the occurrence of positive SST anomalies and a shift of heavy precipitation from the extreme western Pacific to the central Pacific.

Khandekar (1979) hypothesized that the Walker circulation propagates over the equatorial Indian Ocean region during the early stages of the Indian southwest monsoon and that this circulation pattern and the associated pressure distribution may determine the subsequent monsoon activity. He has postulated a lag of at least one month or more between the establishment of the pressure gradient over the equatorial Pacific and the subsequent monsoon rainfall over India.

Angell (1981) correlated Indian summer monsoon rainfall to eastern equatorial Pacific (0–10°S and 90–180°W) SST anomaly over the period 1868 to 1977. He obtained a correlation coefficient around -0.6 between all-India monsoon rainfall and SST one to two seasons later. This relationship is highly significant. He also computed correlation coefficients for 22-year component periods with SST one season later, and he obtained coefficients from -0.51 to -0.67, which are highly significant.

Fu and Fletcher (1985) examined large-scale thermal contrast between the soil temperature over the Tibetan Plateau and the SST over the eastern equatorial Pacific (5°N–10°S and 120°–160°W) and found that higher (lower) values of land-ocean contrast (i.e., Tibetan Plateau temperature minus SST) are associated with higher (lower) monsoon rainfall. The correlation coefficient with land minus sea temperature is higher than that with either land or SST alone.

Mooley and Parthasarathy (1984b) examined the relationship between all-India summer monsoon rainfall and SST anomaly over the eastern equatorial Pacific (0–10°S and 90–180°W) for the seasons December-January-February (DJF) (lag of -2), March-April-May (MAM) (-1), June-July-August (JJA) (0), September-October-November (SON) ($+1$), and DJF ($+2$) over the period 1871 to 1978. They obtained negative correlation coefficients (inverse relationships) that were significant at the 0.1% level for JJA (0), SON ($+1$), and DJF ($+2$), and at the 5% level for MAM (-1). They found the relationship to be consistent and stable.

4.2.2 El Niño. Sikka (1980a) showed a general association between El Niño events and deficient rainfall. He used the Line Islands precipitation index (LIPI) as the indicator of El Niño occurrences during the period 1922 to 1974. However, as mentioned by Rasmusson and Carpenter

(1983), because of the tendency of LIPI to peak near the end of an El Niño year, he listed all El Niños as occurring during pairs of consecutive years; and on account of this ambiguity in the identification of El Niño years, the degree of correspondence and the lead-lag relationship were left in doubt.

Rasmusson and Carpenter (1983) examined the relationship between rainfall over India and El Niño events. They found that during 21 out of 25 El Niño years, all-India summer monsoon rainfall was below the median rainfall, and in 19 years it was below the mean. According to them, the association between the Indian monsoon and the warm episode has some predictive value.

Mooley and Parthasarathy (1983c) found a signficant association between monsoon rainfall over India and El Niño events on the basis of data for the period 1871 to 1978. They considered 22 moderate and severe El Niño events, as defined by Quinn et al. (1978). In all the severe El Niño events the all-India monsoon rainfall in standard units was less than -0.60, with the exception of 1884, which had an all-India monsoon rainfall of $+0.92$ in standard units. On a careful examination of the rainfall in all the subdivisions during 1884, it is observed that 20.4% of country's area had drought (rainfall < -1.28 in standard units) and 26% of country's area had flood (rainfall $>$ 1.28 in standard units). Thus the effect of an El Niño event is clearly seen, but it has been overshadowed by the effect of some factors that enhance rainfall. This example clearly brings out that two or more factors influence monsoon rainfall over India. Whether or not an El Niño occurs, other factors continue to influence monsoon rainfall in their own way. Mooley and Parthasarathy have shown that the difference between the means of all-India rainfall in El Niño years and the non–El Niño years that is negative is highly significant. This also clearly brings out the inhibiting influence of an El Niño event on monsoon rainfall.

The El Niño years are indicated by black bars in Fig. 2.8, which shows the year-by-year, all-India summer monsoon rainfall.

4.2.3 Southern oscillation. Utilizing the southern oscillation index of Wright (1975), Pant and Parthasarathy (1981) computed the correlation coefficient between all-India summer monsoon rainfall and the southern oscillation index for the JJA season over the period 1871 to 1976 and obtained a value of 0.34, which is highly signifcant.

Bhalme et al. (1983) obtained a correlation coefficient of $-0.28(+0.29)$ between the April pressure index (based on pressures at Santiago, Perth, Jakarta, and Bombay) and the drought area index (flood area index) for India over the period 1891 to 1979.

Mooley and Parthasarathy (1983b) computed the correlation coefficients between the indices of dryness/wetness over India and Wright's (1975) southern oscillation index over the period 1871 to 1974. They found that the coefficients were significant at the 0.1% level for the JJA and SON seasons and significant at the 1% level for the MAM season, the relationship being inverse with dryness index and direct with wetness index.

Shukla and Paolino (1983) examined the relationship between all-India summer monsoon rainfall and Darwin pressure anomalies over the period 1901 to 1981, and found that the tendency of Darwin pressure anomaly before the monsoon season is a good indicator of the monsoon rainfall anomaly. They obtained a correlation coefficient of -0.42 between the monsoon rainfall anomaly and the normalized Darwin pressure trend from the DJF season to the MAM season. They prepared a composite of the normalized Darwin pressure anomaly (three-month running mean) for heavy/deficient monsoon rainfall years, which is given in Fig. 2.19. According to them, if the Darwin pressure anomaly has fallen from season DJF to season MAM, nonoccurrence of drought over India can be predicted with a very high degree of confidence.

4.3 Changes in surface albedo

Charney et al. (1977) have suggested that changes in the surface albedo can produce significant changes in local rainfall and atmospheric circulation, particularly over the desert margins of the subtropics. Increase in albedo leads to a net reduction in the total radiative energy at the ground, leading to a net reduction in evaporation, cloudiness, and precipitation. Since these subtropical regions are not affected by large advective effects, local changes in radiative and latent heating are accompanied by dynamical circulations that produce descending motion over the regions of albedo anomaly.

An experiment to study the effect of albedo changes has been conducted by Sud and Fennessy (1982), using the recent version of the GLAS climate model, which has a better parameterization for evaporation and sensible heat fluxes. They found that their results support all the conclusions of the earlier studies by Charney et al. (1977).

Changes in albedo in desert margins can result from excessive grazing. In a recent numerical study, Sud and Smith (1985) investigated the influence of (1) increased land surface albedo, (2) increased land surface albedo and reduced land surface roughness, (3) increased land surface albedo, reduced surface roughness, and no evapotranspiration on the Indian monsoon. They found that (1) higher albedo (from 0.14 to 0.20) reduced the rainfall, (2) reduction in surface roughness also reduced rainfall, (3) absence of evapotranspiration did not materially alter the rainfall. This study

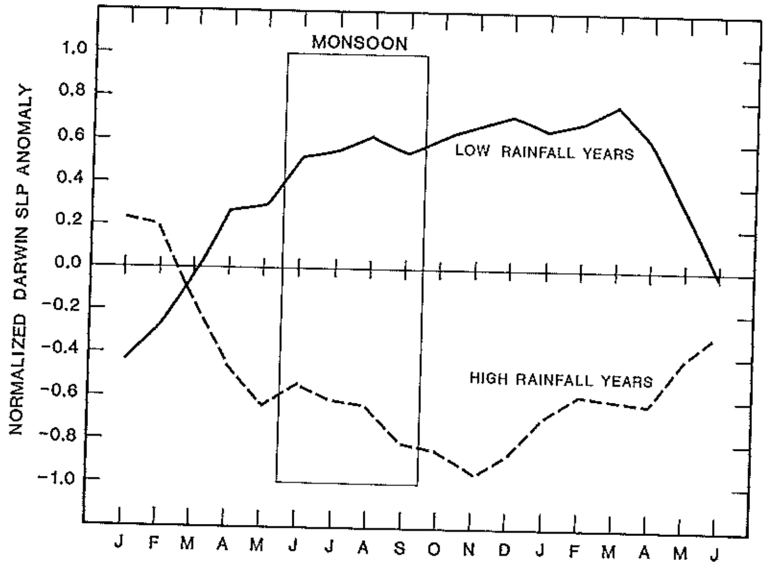

Fig. 2.19. Composite of the normalized Darwin pressure anomaly (three-month running mean) for heavy monsoon rainfall years and deficient monsoon rainfall years. (After Shukla and Paolino 1983.)

thus suggests that an increase in forest area over India would decrease the albedo and increase the roughness of the surface, both of which enhance rainfall, and the net effect would be a good increase in rainfall.

4.4 Soil moisture

Soil mositure is an important component in the global water budget and the hydrological cycle. The amount of moisture in the soil determines the rate of evaporation and hence the rate of moisture and latent heat supply to the atmosphere. Soil moisture also influences the ground heating, which in turn determines the sensible heat flux and affects dynamical circulation by generation and dissipation of heat lows. The net diabatic heating of a vertical atmospheric column is maximum over the tropical land masses (Shukla 1982). Hence even small fractional changes in these tropical asymmetric heat sources can produce considerable changes in the planetary scale circulations of the tropical and the extratropical atmosphere. Thus, even though the land surface is less than half the water surface on earth, the soil moisture effects can be as important as SST anomaly effects. Soil moisture effects strongly depend on the season and the latitude.

Shukla and Mintz (1982) have brought out the importance of soil moisture and vegetation for rainfall over an area through two numerical experiments with extreme conditions: (1) no evaporation (or dry soil or no vegetation) and (2) evaporation from land equaling model-calculated potential evapotranspiration (wet soil or completely vegetated land surface). In each case, 60-day integration was carried out with the GLAS climate model. Over the monsoon lands solar heating of the dry soil produced intense low pressure and convergence, leading to increased moisture flux convergence from the neighboring ocean, which in turn contributed to increased heating of the vertical air column due to latent heat that maintained and intensified the surface low. This happened prominently for the unique monsoon circulation over India for which the oceanic moisture was brought in by the monsoon current.

5 INFLUENCE OF OTHER FACTORS

Here we shall consider extraterrestrial factors (solar activity) and such atmospheric factors as undergo year-to-year variation and that influence the monsoon activity on an interannual scale. These can be classified extraterrestrial, global, and regional.

5.1 Extraterrestrial factors

We shall consider only solar activity as judged by the sunspot numbers. Walker (1915a, 1915b, 1915c) examined the relationships between sunspots and the meteorological elements of rainfall, temperature, and pressure at many stations over the world. He used the annual values of the meteorological elements and the mean sunspot number of the contemporary year. The number of years of data varied largely with stations but was generally more than 30. From his studies, Walker concluded that sunspot numbers play a definite but minor direct role in yearly or seasonal weather. After Walker, several research workers examined the relationship between solar activity and Indian weather. The results of these studies are found to be conflicting and lacking in consistency. Critical reviews of papers on the long-term sun-weather relationship have been made by Pittock (1978, 1983).

Bhalme and Mooley (1981) found a highly significant, approximately 22-year cycle in the flood area index and a weak quasi periodicity of 2.7 to 3.0 years (QBO) in the drought area index for India for the monsoon season. From cross-spectrum analysis they showed that the approximate 22-year cycle in the flood area index is related to the double (Hale) sunspot cycle and that it is nearly in phase with the double sunspot cycle. From harmonic dial analysis, they brought out that all the large-scale flood events over India occurred in the major sunspot cycle. According to them, the approximate 22-year cycle in the flood area index explains 16% of the total variance.

5.2 Global factors

5.2.1 Sea level pressure. In the global sea level pressure distribution, there are four pressure systems, referred to as centers of action: Icelandic low, Aleutian low, Atlantic high, and Pacific high. These are semipermanent in character and are well marked in certain seasons. However, their seasonal locations and intensity vary from one year to the next. Parthasarathy (1984a) has examined the relationship between the Indian monsoon and the seasonal mean sea level pressures at these four centers of action. He found that only the correlation coefficient between monsoon rainfall and pressure at the center of the Pacific high in the MAM season that is positive is consistently significant. This would perhaps suggest that when the intensity of the Pacific high is higher than normal during the MAM season, more low-pressure systems move during the monsoon season across Vietnam in a westerly direction and emerge into the north Bay of Bengal as remnant lows to be revived into well-marked lows/depressions under favorable conditions over the bay. Moving west-northwest from the north bay, these systems maintain the normal position of the monsoon trough and a good activity of the monsoon over India. However, the relationship between the pressure at the Pacific high during the MAM season and the number of low-pressure systems over the north bay during the monsoon season needs to be examined further.

5.2.2 Troughs in midlatitude westerlies. During the monsoon season, these troughs normally move north of India in an easterly direction. However, occasionally they move across northern parts of India and profoundly influence the circulation and rainfall within the season. Their influence has been studied extensively by Pisharoty and Desai (1956), Mooley (1957), and Ramaswamy (1958, 1962). Raman and Rao (1981a, 1981b) have examined the influence of the persistence of a blocking ridge over East Asia on the rainfall over India during the monsoon season. Joseph (1978) found that during years of monsoon failure, subtropical westerlies of the upper troposphere protrude more southward over areas immediately west of India during the monsoon season, and that the southward intrusion of the westerlies has a persistence of a few months prior to monsoon and on many occasions from winter or postmonsoon season. In good monsoon years, he found northerlies in the upper troposphere over the whole of India during June, July, and August. According to Sikka (1980b) and Sikka and Gray (1981) the baroclinic perturbations passing eastward across the Mozambique channel probably result in the fluctuations of the low-level equatorial jet off the African coast, which in turn influences the Indian monsoon. These connections are yet to be explored on the basis of upper-air data.

5.2.3 Stratospheric circulation. Mukherjee et al. (1979) examined the stratospheric wind data of Thumba (near Trivandrum) for five summer seasons (1971 to 1973, 1975 to 1976) with reference to the monsoon activity over India. They found a high positive correlation between monsoon activity (departure of monsoon season precipitation from normal) over India and the 25-km mean zonal wind. According to Raja Rao and Lakhole (1978), the appearance of easterlies in the lower stratosphere (16 to 24 km) over Gan Island in May is an indication of the onset of the monsoon over Kerala a month later.

Thapliyal (1979) has shown that the features of circulation at the 50-mb level during westerly and easterly QBO years are very different and that these can be used as guidance material in the issue of seasonal forecasts of monsoon rainfall. However, since drought does not follow only the westerly phase of the QBO in January but is also seen to follow the easterly phase of the QBO in January, these circulation features cannot be used as predictors for forecasting monsoon seasonal rainfall in any individual year. The study brings out that there are other factors that affect the monsoon rainfall.

5.3 Regional factors

5.3.1 Temperature in the upper troposphere. Bhalme and Mooley (1980) found that the mean heights of the 200-mb surface during May were much below (above) normal over the latitudinal belt 15 to 30°N along 70°E during drought (good monsoon), suggesting colder (warmer) upper troposphere. Verma (1980, 1982) found a good association between cooler (warmer) upper troposphere over north and northwest India and poor (good) summer monsoon. He also found that the cooler (warmer) upper troposphere in May generally persists during the monsoon season. However, Paolino (1984) did not find good association between the upper atmospheric temperatures and performance of the summer monsoon for a later period (i.e., after 1977).

5.3.2 Monsoon trough location. Monsoon trough location exercises a profound influence on the rainfall during the monsoon season. In the normal position, this trough over India extends from northwestern India to western Bengal. The trough, particularly the western half, has a tendency to move north to the foot of the Himalayas under the influence of a westerly trough moving eastward across northwest India. On the other hand, monsoon lows/depressions periodically forming over the north Bay of Bengal and moving west-northwestward across the country maintain the normal position and activity of the trough. Sometimes, the low-pressure systems do not form over the bay, and in this situation a westerly trough affecting northwest India results in the shift of the whole trough to the foot of the Himalayas. This situation of the monsoon trough over the foot of the Himalayas is referred to as a 'break' in the monsoon, since except for the sub-Himalayan area and Tamilnadu, the whole country gets very little rain. The sub-Himalayan area receives heavy rainfall during the 'break' period. July and August are the core monsoon months contributing largely to the monsoon seasonal rainfall. Days of such 'breaks' during July and August have been documented for each of the years during the period 1888 to 1967 by Ramamurty (1969). Similar information for the period 1968 to 1983 has been collected from the publications of the Indian Meteorological Department. From this information, the frequency distribution of the number of days of 'break' in the monsoon during July and August over the period 1888 to 1983 has been constructed. The mean of the distribution is 8.1. The frequency distribution is skewed. The frequency of 31 for the class interval 6 to 10 days is highest and falls off successively to 17, 7, 3, and 1 for the higher class intervals of 11 to 15, 16 to 20, 21 to 25, and 26 to 30 days, and it falls to 22 for the lower class interval 1 to 5 days. It is observed that in 15 years of the period 1888 to 1983, there was no 'break' in the monsoon.

The number of days of 'break' and the length of the longest 'break' spell are tabulated in Table 2.4 for all the drought and flood years for which this information is available. The drought and the flood years considered here have been identified by the criterion of the tenth deciles of the percentage drought and flood area series. Table 2.4 clearly brings out that at least in extreme years—that is, drought/flood years—a good part of the interannual variability of the monsoon seems to be accounted for by the interannual variability in the days of 'break' in the monsoon.

5.3.3 Westward-moving low-pressure systems during the monsoon season. Mooley (1976) has shown that during large-scale drought years, the monsoon depressions either dissipate or recurve north/northeast before reaching longitude 80°E.

Sikka (1980a) has shown that the main features that distinguish years of heavy and deficient monsoon rainfall are the number of monsoon low-pressure systems and the number of rainy days in a season.

Bhalme and Mooley (1980) found that during the monsoon season (1) the mean frequency of storms/depressions during drought years is

TABLE 2.4 *Number of days of 'break' in the monsoon and length of the longest 'break' spell during drought/flood years*

Drought Years			Flood Years		
Year	No. of Days of 'Break' During July and August	Length of Longest 'Break' Spell During July and August	Year	No. of Days of 'Break' During July and August	Length of Longest 'Break' Spell During July and August
1877	—	—	1874	—	—
1899	23	8	1875	—	—
1904	8	5	1878	—	—
1905	13	9	1884	—	—
1911	11	11	1892	0	0
1915	16	9	1917	10	7
1918	23	17	1933	8	8
1920	12	6	1936	0	0
1965	15	12	1938	0	0
1966	15	10	1961	0	0
1972	14	14	1973	7	7
1979	24	17	1983	0	0
Mean	15.8	10.7	Mean	3.1	2.7

smaller than that during flood years, (2) the westward extent of their track is shorter during drought years than during flood years; in particular, for June the storms/depressions are confined to east of 80°E during drought years.

Mooley and Parthasarathy (1983a), who examined the tracks of only depressions during monsoon months, found that (1) westward penetration of depressions was more in flood years than in drought years for all the monsoon months, being relatively much more in July, and (2) of the total depressions, a larger proportion moved west of longitude 80°E during flood years, the proportion being much larger for August and September. In brief, the depressions during drought years are characterized by lesser westward activity.

It may be mentioned that a depression is a low-pressure system of specified intensity, that is, system wherein the associated wind speed is 17 to 33 knots (30 to 60 km h^{-1}). While systems of higher intensity (i.e., cyclonic storms) generally weaken into depressions shortly after landfall, those of lower intensity (generally referred to as lows) move across the country, keep the monsoon trough in near-normal position, and make a good contribution to rainfall. In view of this position, investigation of monsoon rainfall in relation to the total number of low-pressure systems, their tracks, and their westward penetration is needed.

5.3.4 Mean 500-mb ridge over India during April. The tropospheric circulation over India and neighboring countries is characterized by a subtropical ridge that undergoes seasonal north-south migration. During winter, the ridge is located in the southernmost position, and with the advance of the season, it gradually shifts northward. Over the peninsula, the summer conditions begin to get established by March in the lower troposphere. However, over northern India, summer conditions commence getting established in the upper troposphere by the beginning of May. April is the month just before the month during which transition to summer circulation takes place over India. The mean (based on data for 1951 to 1965) ridge at 500 mb along longitude 75°E is located at 11.5°N in January, 15.0°N in April, 28.5°N in July, and 20.0°N in October. The largest shift occurs from April to July, the period during which the transition from summer to monsoon takes place.

Banerjee et al. (1978) showed for the first time that the percentage of subdivisions with normal or above-normal rainfall is related to the position of the mean subtropical ridge at 500 mb along 75°E during April, suggesting indirectly a relationship between monsoon activity and ridge location. Thapliyal (1981, 1982) brought out that the autoregressive integrated moving average (ARIMA) model with 500-mb ridge position along 75°E during April as a leading indicator can provide more accurate forecasts for peninsular India than other models.

Mooley et al. (1986) examined in detail the relationship between the 500-mb ridge location along 75°E during April and the all-India monsoon rainfall as well as subdivisional monsoon rainfall for the period 1939 to 1984 and the stability of the relationship for the period 1939 to 1984. They found that (1) the correlation coefficient between all-India monsoon rainfall and ridge location is 0.71 (significant at the 0.1% level and above); (2) the lowest period for which the correlation coefficient is consistently significant at the 1% level or above is a 20-year period, thus exhibiting the high stability of the relationship; (3) the mean ridge position is 3.0° south of the normal position during years of deficient all-India monsoon rainfall and 1.1° north of the normal position during years of excess all-India monsoon rainfall; (4) the correlation coefficients between the ridge position and the subdivisional monsoon rainfall are significant at the 5% level or above for 19 contiguous subdivisions of India covering mostly the Indian area north of 12°N and west of 84°E.

From the good relationship observed between the ridge and the all-India monsoon rainfall, it can be inferred that the location of the ridge in April along 75°E at the 500-mb level is a good indicator of whether the transition to the summer conditions in the troposphere over the Indian subcontinent will be behind schedule, on schedule, or ahead of schedule.

It has been found that the 500-mb May ridge along 75°E is also significantly (at the 5% level) related to all-India rainfall.

6 FORECASTING OF MONSOON SEASONAL RAINFALL AND MONSOON ONSET

Earlier studies on a search for suitable predictor parameters and on forecasting monsoon seasonal rainfall by Blanford (1884) and Walker (1910b, 1914, 1915a, 1915b, 1915c, 1922, 1924a, 1924b, 1928, 1933) have been reviewed by Montgomery (1940a), Jagannathan (1960), Normand (1953), and Rao (1964). Montgomery (1940b) has carried out verification of three of Walker's forecasting formulas for Indian monsoon rainfall. Normand (1953) has verified seasonal rainfall forecasts for a later period. Jagannathan (1960) has also examined the variation in the relationship between monsoon rainfall and Walker's predictor parameters over the period 1881 to 1960. For forecasting the date of establishment of monsoon over Kerala, the south Kanara district, and the Colaba district on the west coast of the India peninsula, Ramdas et al. (1954) evolved regression equations, using all available data up to 1950. They have not given any verification of these regression equations for the independent years 1951 to 1954.

6.1 Recent efforts in forecasting monsoon rainfall

Search for new and useful predictor parameters was continued in the Indian Meteorological Department. Jagannathan and Khandekar (1962) showed that the contour height and thickess between two pressure levels for some locations are useful predictor parameters. In 1975, the mean minimum temperature of Jaisalmer, Jaipur, and Calcutta for March was identified as a predictor parameter for the first time. The correlation coefficient between peninsula monsoon rainfall and this parameter over the period 1944 to 1973 is 0.66 (Thapliyal 1981). This factor was used for the first time for preparing experimental forecasts of peninsula monsoon rain for 1976. Later, from 1977, it was used by the India Meteorological Department in preparing monsoon rainfall forecasts for the peninsula. Banerjee et al. (1978) have brought out that the regression equation between the number of subdivisions with normal or above-normal monsoon rainfall (i.e., greater than 81% of normal) and the location of the April 500-mb ridge along 75°E can be used to estimate the number of subdivisions with normal or above-normal monsoon rainfall in subsequent monsoon seasons. Since 1977, the Indian Meteorological Department has been using three predictor parameters—the South American pressure, the March mean minimum temperature of Jaisalmer, Jaipur, and Calcutta, and the April 500-mb ridge along 75°E—for forecasting peninsula monsoon rainfall. In subsequent years, these three predictor parameters have maintained the same sign of association with peninsula monsoon rainfall.

Based on the work by Box and Jenkins (1968, 1970), Thapliyal (1981, 1982) has developed an ARIMA model for long-range prediction of monsoon rainfall in the peninsula. In this model, he has used the April 500-mb ridge on 75°E as the leading indicator. Models using such associated time series are called leading-indicator ARIMA models. Chatfield (1975) also found that the leading-indicator ARIMA model is able to give reasonable accurate forecasts in many fields. The basic concept involved is that specific operations on the input that is the associated series leads to the output of the predictand series. Thapliyal (1981) uses the following equation for forecasting monsoon rainfall for the peninsula, R_t for the tth year. Thus, R_t

$$= (R_{t-1})\left[\left(\frac{R_{t-1}}{R_{t-2}}\right)^{\delta_1}\left(\frac{R_{t-2}}{R_{t-3}}\right)^{\delta_2}\left(\frac{R_{t-3}}{R_{t-4}}\right)^{\delta_3}\left(\frac{R_{t-4}}{R_{t-5}}\right)^{\delta_4}\right]$$
$$\times \exp\left[\omega_0(x_t - x_{t-1}) + \sum_{i=1}^{4}(-1)\omega_i(x_{t-i} - x_{t-i-1}) - (0.52a_{t-1} + 0.15a_{t-4} - 0.28a_{t-13})\right] \tag{2.2}$$

where x_t is the location of the 500-mb ridge in the tth year. He has tabulated the final values of δ_1 to δ_4 and ω_0 to ω_4. Utilizing Eq. (2.2), he obtained forecast peninsula monsoon rainfall for the independent years 1973 to 1980. He used the regression equation between peninsula monsoon rain and location of the April 500-mb ridge for the period 1939 to 1976 and computed peninsula monsoon rainfall for each of the independent years 1977 to 1980. According to him, the error in the peninsula monsoon rainfall forecast on the basis of the regression equation is higher than that for the forecast on the basis of the ARIMA leading-indicator model for each of the independent years. This verification, however, is based on a very small sample (four years).

Joseph et al. (1981) have developed a method of forecasting Indian summer monsoon rainfall by using a meridional wind index, based on the meridional wind components at Srinagar, Delhi, Nagpur, Bombay, and Madras during May at the 200-mb level. The regression equation obtained by him between all-India monsoon rainfall (Parthasarathy and Mooley 1978, rainfall series) and the meridional wind index, V_m, for the period 1964 to 1978 is given by (R, rainfall in centimeters)

$$R = 92.55 - 3.15V_m \tag{2.3}$$

where V_m northerly is taken as positive and V_m southerly as negative. Thus northerly meridional wind—which means larger activity of westerly troughs and consequently colder conditions persisting—results in deficient rainfall, and southerly meridional wind results in higher rainfall. The correlation coefficient obtained by him between Indian monsoon rainfall and V_m is −0.89. Joseph (1983) (see below) has verified this regression

Year	V_m (m s)$^{-1}$	Estimated Rainfall, R_e	Actual Rainfall, R_0 (cm)	Error of the Estimate, $R_e - R_0$	Percentage Error
1979	8.6	65.5	69.7	−4.2	−6.0
1980	3.0	83.1	92.0	−8.9	−9.6
1981	0.7	90.3	87.4	+2.9	+3.3
1982	4.0	79.8	75.8	+4.0	+5.3

equation for the independent years 1979 to 1982.

The relationship between the April 500-mb ridge over 75°E and the meridional wind index (V_m during May) of Joseph et al. (1981) has been examined for the period 1964 to 1978. The correlation coefficient between these two parameters is −0.73. The relationship is thus inverse. Thus when the April 500-mb ridge is north of the average position, the May 200-mb meridional wind over India is smaller than the mean, and vice versa. Thus the April 500-mb ridge location and the May mean meridional wind at 200 mb are not independent parameters. With a 500-mb ridge location, the forecast can be prepared one month earlier, and hence it is more useful than the mean meridional wind. It may, however, be mentioned that fixing ridge location is subjective, but fixing mean meridional wind is objective. If we take this advantage of the meridional wind into account, it may be worthwhile to examine the correlation coefficient between mean meridional wind at 200 mb in April and mean meridional wind at 200 mb in May. Joseph et al. (1981), however, found that the correlation coefficient between Indian summer monsoon rainfall and the 200-mb April meridional wind was only −0.16. In view of this, the relationship between April and May meridional wind at 200 mb is not expected to be signficant.

Shukla and Paolino (1983), on the basis of data for the period 1901 to 1981, have shown that the Port Darwin pressure tendency from the winter (DJF) to the spring (MAM) season is significantly related to all-India monsoon rainfall. Their scatter diagram between the normalized Darwin pressure trend and the normalized Indian monsoon rainfall brings out that most of the severe drought years are in the lower-right quadrant wherein the Darwin pressure trend is positive, and most of the excessive rainfall years are in the upper-left quadrant wherein the Darwin pressure trend is negative. They have argued that since there is a physical basis to assume that lower pressure over Darwin would be favorable for above-average rainfall over India, a negative trend in the winter-to-spring pressure of Darwin is an indicator of below-normal pressure over India during the monsoon season, and hence a combination of Darwin pressure anomaly and its trend would provide a better guidance for forecasting monsoon rainfall anomaly over India. They have brought out that nonoccurrence of deficient/heavy monsoon rainfall can be forecast with high confidence when both Darwin pressure tendency and Darwin pressure anomaly are negative/positive.

On the basis of data for the period 1939 to 1984, Mooley et al. (1986) examined in detail the behavior and stability of the location of the April 500-mb ridge on 75°E and the relationship between the location of this ridge and all-India monsoon rainfall. The contingency table for three classes of all-India monsoon rainfall and ridge location, namely, less than or equal to −1.0, greater than −1.0 but less than 1.0, and greater than or equal to 1.0 in standard units, as prepared by them for the period 1939 to 1984, is given in Table 2.5. This table brings out clearly that on 80% of the occasions when ridge location is south of the mean location by 1 standard deviation or more, all-India monsoon rainfall is deficient; and on 67% of the occasions when ridge location is north of the mean location by 1 standard deviation or more, all-India monsoon rainfall is in excess. It is also seen that in 87% of the occasions when the deviation of the ridge location from the mean is within 1 standard deviation, all-India monsoon rainfall lies within 1 standard deviation of the mean, and that nonoccurrence of deficient/excess all-India monsoon rainfall can be forecast with high confidence when the ridge is north/south of the mean location by 1 standard deviation or more. Figure 2.20 gives the scatter diagram between the all-India monsoon rainfall and the location of ridge, and it also gives the regression equation. Mooley et al. (1986) also examined the relationship between subdivisional monsoon rainfall and ridge location. The subdivisions for which the relationship is significant are located north of 12°N and west of 84°E.

Kung and Sharif (1980) developed a regression model for prediction of the date of onset of the summer monsoon over the Kerala coast. The predictors they considered are geopotential heights, kinetic energy, and zonal and meridional wind components, each at 700 and 100 mb (in all, ten predictors) over the point 12.4°N and 76.5°E on the Kerala coast for April. These predictor parameters represent the circulation in the lower and upper troposphere over the point. They used data for the period 1958 to 1978. They obtained data relating to the predictand (monsoon onset date over Kerala) from the Indian Meteorological Department and data relating to geopotential heights and temperature at 700-mb and 100-mb levels from the National Meteorological Center's

TABLE 2.5 *Contingency table showing the frequencies of each of the three class intervals for the all-India summer monsoon rainfall and the April 500-mb ridge; period of data, 1939–84*

	Ridge Position (Standard Units)		
Rainfall (Standard Units)	South of Normal, ≤ -1.00	Normal between -1.00 to $+1.00$	North of normal, $\geq +1.00$
Deficient, ≤ -1.00	8	1	0
Normal, between -1.00 to 1.00	2	26	2
Excess, $\geq +1.00$	0	3	4

Source: From Mooley et al. (1986).

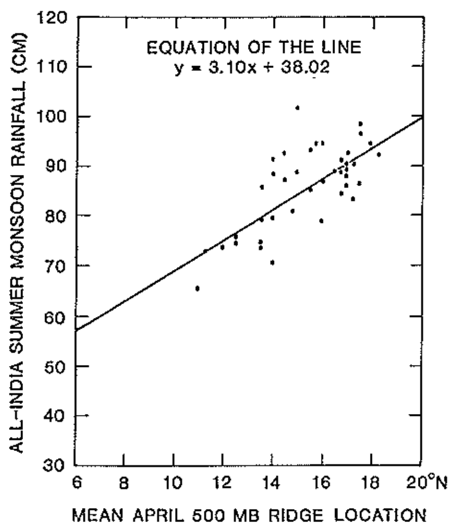

Fig. 2.20. Scatter diagram between the all-India monsoon rainfall and the April 500-mb ridge axis location along 75°E. (After Mooley et al. 1986.)

(NMC) octagonal-grid data archives through the National Center for Atmospheric Research (NCAR) in Boulder. They obtained a forecast for each of the years 1958 to 1978 by utilizing the regression equations based on all data except the data for the forecast year. According to these authors, the recorded onset dates of the summer monsoon in southwestern India can be closely related functionally to the upper-air conditions at the selected point on the Kerala coast. But they have advised constant updating of the regression equation before using it for forecasting the onset date because of the time variations of the general circulation pattern. The success of the regression equations in forecasting monsoon onset date over Kerala is seen to be moderate.

In another study, Kung and Sharif (1981) considered additional predictor parameters—700-mb January zonal wind and 700-mb March temperature over Australia, April Indian Ocean temperature over the area 10–20°N and 60–70°E—and developed regression equations for predicting the monsoon onset date over Kerala and for predicting monsoon rainfall over central India on the basis of data for 1958 to 1977. According to them, the regression equations predict the monsoon onset date with an average of 5% accuracy and summer monsoon rainfall over central India with an average of 6% accuracy.

6.5 Prospects for the future

There are several observational and General Circulation Modeling (GCM) studies (Madden 1976; Charney and Shukla 1981; Manabe and Hahn 1981; Shukla 1981a, 1981b, 1981c, 1982; Shukla and Gutzler 1983) that provide strong but indirect evidence that boundary anomalies in the tropics contribute significantly to the interannual variability of the tropical and monsoon circulation. Air-sea and air-land interactions are the most important mechanisms for global and regional variability.

The cryosphere, the ocean, and the atmosphere are strongly coupled, and the need to develop coupled atmospheric models for a more realistic simulation of the observed state of the atmosphere has been recognized. The World Meteorological Organization has recently commenced a program called tropical ocean and global atmosphere (TOGA) for development of coupled models. A land-ocean-atmospheric model which incorporates the physical processes in the ocean and in the atmosphere as well as the interactive processes at the interface between the earth and the atmosphere is required to investigate the question of predictability.

In recent years, considerable knowledge has been acquired with respect to the El Niño and the southern oscillation, their global repercussions, and their relationship with the Indian monsoon. Efforts aimed at greater understanding of the factors responsible for interannual variability of the Indian monsoon are necessary. For this purpose we recommend detailed studies based on long records covering the following aspects (Shukla, 1987b):

1. Special features of the sea level pressure distribution over the Asian monsoon region and Australia during the months March to May.
2. Snow cover and snow depths over Eurasia. As the physical basis for the influence of winter/spring snow cover on the Indian monsoon appears to be rational, it is felt that the inconsistent behavior of this predictor parameter after 1920, as mentioned by Jagannathan (1960), could be due to defective snow cover observations. Satellite snow cover data need to be assessed carefully for any defects. If there are any remediable defects, these should be removed, and a homogeneous snow cover series may be used (Hahn and Shukla 1976; Dickson 1984).
3. A detailed examination of the SST over different sectors of the eastern Pacific with reference to monsoon performance. These studies may be made to find out whether SST over any particular section has a signal about the monsoon behavior (Rasmusson and Carpenter 1982, 1983).
4. Monsoon behavior in relation to the intensity of the ITCZ over the Indian Ocean in premonsoon months.
5. The different features of the upper-air circulation over and near India during premonsoon months. These features may be studied with reference to monsoon rainfall to obtain useful predictor parameters. The parameters that can be examined are the April ridge in the upper troposphere (Banerjee et al. 1978), heights at stan-

dard-pressure surfaces (Jagannathan and Khandekar 1962), the meridional wind component in the upper troposphere (Joseph et al 1981), upper tropospheric thickness/temperature anomalies (Verma 1980, 1982), and the location of the westerly jet stream, trough and ridge positions at 50 mb (Thapliyal 1979).

6. The influence of blocking highs in the midlatitudes in the premonsoon season on the monsoon (Raman and Rao 1981a, 1981b; Tanaka 1982).

7. The behavior of the monsoon in relation to the phase of the stratospheric quasi-biennial oscillation over the equator on the basis of adequate rocketsonde/high-level radio-wind data (Raja Rao and Lakhole 1978).

8. Monsoon rainfall in relation to the activity of tropical storms/depressions/lows in the western Pacific in the premonsoon season (Kanamitsu and Krishnamurti 1978).

9. The influence of boundary conditions and circulation in the Southern Hemisphere.

These studies are likely to give us a useful insight into the causes of the interannual variability of the monsoon as well as provide us with useful predictor parameters. The predictor parameters arrived at may be examined for an interrelationship among themselves. If two predictor parameters are related among themselves, the one that is better related to the predictand may be used. In this way, the predictor parameters may be so chosen so that among themselves the relationship is not high. With such predictor parameters a regression equation may be obtained for forecasting Indian monsoon rainfall. It is also necessary to examine the performance of a regression equation on independent data.

For the practical problem of long-range forecasting it may be useful to form a regression equation on the basis of the predictor parameters that have retained significant relationship with the predictand during the preceding 25-to-30-year period and estimate the Indian monsoon rainfall in the coming season from this equation.

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