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Interannual Variability of Monsoons

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INTRODUCTION

The basic question addressed in this chapter is why one monsoon season is different from another. This question is not any different than asking why some summers are warmer than others, and some winters are colder than others. This is fundamentally a more difficult question than why summers are warmer than winters and vice versa. For brevity, I first describe the processes that determine the mean seasonal monsoon climate itself and the mechanisms that are responsible for day-to-day weather changes. I then attempt to address the question of the interannual variability of the seasonal averages. This chapter will focus almost exclusively on the Indian summer monsoon for which we have an extensive historical data set.

Let us begin by asking what appears to be a simple question: Why weather on a given date (say, January 1 of one year) is so different from the weather on the same date in another year? We know that the amount and distribution of solar energy, the most important driving force for the atmosphere, remains the same on both dates, as does the rotation of the earth, the composition of the atmosphere, and the distribution of oceans and continents. This would suggest that the causes for different weather on the same date for two different years do not lie outside the atmosphere, but most probably inside the atmosphere. To provide a reasonable answer to the above question, we have to address some additional fundamental questions. For example, what causes weather, and what causes weather to change from one day to another? An understanding of the processes that are responsible for changes in weather from one day to another is a prerequisite to understanding the mechanisms for changes in seasonal mean weather from one year to another. Since seasonal mean weather, which is sometimes referred to as the short term climate, is a consequence of the average of daily weather over the length of a season, it is natural to think that day-to-day changes in weather in a given season

might be responsible, at least in part, for producing changes in the seasonal mean from one year to another.

The basic driving force for atmospheric motions is the uneven solar heating of the earth–atmosphere system due to the near-spherical geometry of the earth's surface and the revolution of the earth around the sun. The actual rates of heating vary with height, latitude and longitude, and the magnitudes are determined primarily by the composition of the earth's atmosphere, and the time of day and day of the year. The earth's equatorial regions receive more energy from the sun than they lose to space, but the reverse is the case for the polar regions. The net heating of the warmer equatorial regions and the net cooling of the colder polar regions is a source of energy for the motion of air particles. Other important quasi-stationary forces are provided by the asymmetric heat sources of land and ocean and by mountains which act as mechanical barriers and produce quasi-stationary circulation features. Solar energy heats the oceans and evaporates the water that later condenses into the atmosphere, often at some distance from where it evaporated, providing another very important energy source, the latent heat of condensation. The three-dimensional structure of this heat source is determined in part by the motion field itself, producing one of the most complex feedback loops (i.e., nonlinearities) of atmospheric dynamics. The mean circulation produced by the forcing functions described above can be considered, hypothetically, as the seasonal mean climate of the earth–atmosphere system. This seasonal mean climate, however, is characterized by horizontal and vertical gradients of wind, temperature, and moisture, that are favorable for the growth of thermodynamic and hydrodynamic instabilities. The day-to-day weather fluctuations are produced by transient weather disturbances which are the manifestations of the growth, decay and propagation of these instabilities. The disturbances can derive their enery from the mean circulation and thereby change the mean circulation itself. The observed, quasi-equilibrium mean circulation, to be referred to as the mean climate, is produced by interactions among the stationary and quasi-stationary forcing functions, and transient disturbances.

Since changes in the weather occur due to instabilities of the atmospheric state, and since the internal atmospheric dynamics are intrinsically nonlinear, atmospheric behavior is aperiodic and, therefore, at long ranges, unpredictable. This inherent aperiodicity of the weather at all time scales also produces aperiodicity of monthly and seasonal averages. Therefore, even if there were no changes in the external forces, it would be reasonable to expect seasonal averages to be different from one year to another. Actually, small changes do occur; for example, although the solar energy input for a given season remains nearly the same from one year to the next, there is some variation due to the interannual variability of seasonal mean cloudiness. The height of the mountains remains constant, but there may be large changes in the circulation because the wind impinging on the mountains may be quite different. Another plausible reason for the interannual variability of seasonal mean climate appears to be the interannual variability of boundary parameters such as seasonal mean sea surface temperature (SST), soil moisture, and sea ice and snow. Based on these considerations, the mechanisms responsible for interannual variability of

seasonal mean climate fall into two categories: "internal dynamics" and "boundary forcing" (1, 2). These will be discussed further in Section 3.

It is reasonable to expect that the nature of the interannual variability of seasonal averages will depend upon the spatial and temporal domains for which the averages are calculated. For example, if we average rainfall over the whole Afro–Asian monsoon region, it may not show large interannual variability, but the average over the Indian subcontinent, or part of it, might show large interannual variability. Similarly, monthly means might display large interannual variability, whereas seasonal means might not. This suggests that in order to get a meaningful description of the interannual variability of any atmospheric phenomenon, we need a good understanding of the dominant space and time scales of atmospheric anomalies. Part of the interannual variability may be due to changes in the intensity of the mean atmospheric circulation systems, and part of it may be simply due to shifts in location and timing of those circulation systems. It would appear, therefore, more appropriate to study the question of interannual variability on a global scale. However, due to insufficient long-term, global data records, we are constrained to study the interannual variability of regional phenomena.

In Section 1, 81 years of monsoon rainfall data is used over different subdivisions of India to describe the observed structure and interannual variability of monsoon rainfall. The interannual variability of rainfall is described here because of its social–economic importance and also because reliable records are available for a long period. Discussion of observed variability will be limited to the summer monsoon season. In Sections 2 and 3, the basic mechanisms for intraseasonal and interannual variability of the monsoon circulation and rainfall are described and the results of some of the key studies which document possible relationships between monsoon and other global circulation features are summarized. It will be shown that the year-to-year changes in the Indian summer monsoon are related to a global scale atmospheric circulation feature called the Southern Oscillation, and ocean temperature anomalies in the tropical Pacific called El Niño. A summary of several numerical experiments with global climate models, and results of observational studies is presented to show that the slowly varying boundary conditions have significant influence on the fluctuations of seasonal averaged monsoons.

1 INTERANNUAL VARIABILITY OF SUMMER MONSOON RAINFALL OVER INDIA

After the great famines of 1877 and 1899, the India Meteorological Department established an extensive and efficient network of rain gauge stations over India. This network has provided a great wealth of monsoon rainfall data for systematic studies of interannual variability which have been carried out by a large number of investigators. Some of the earlier work will be summarized and then an analysis will be presented of the structure and space–time variability of monsoon rainfall over India for the period 1901–1981.

1.1 Previous Studies

There are a large number of scientific publications, especially from Indian scientists, describing various aspects of the interannual variability of Indian monsoon rainfall and its relationship with regional and planetary-scale circulations. It is not possible to summarize here the results of all such papers; I therefore refer only to those works that examine sufficiently long time series. This allows some definitive statements to be made about the nature of variability and its possible causes. Case studies using limited data samples are indeed illuminating but insufficient to distinguish between a genuine signal and random noise because of sampling problems.

Parthasarathy and Dhar (3) studied 60 years (1901–1960) of annual rainfall for 31 subdivisions over India (Fig. 14.1). Their data were derived from approximately 300 rain gauge stations. The analysis showed that rainfall in most of the subdivisions fits a normal distribution. Positive trends (increasing rainfall) were found in many

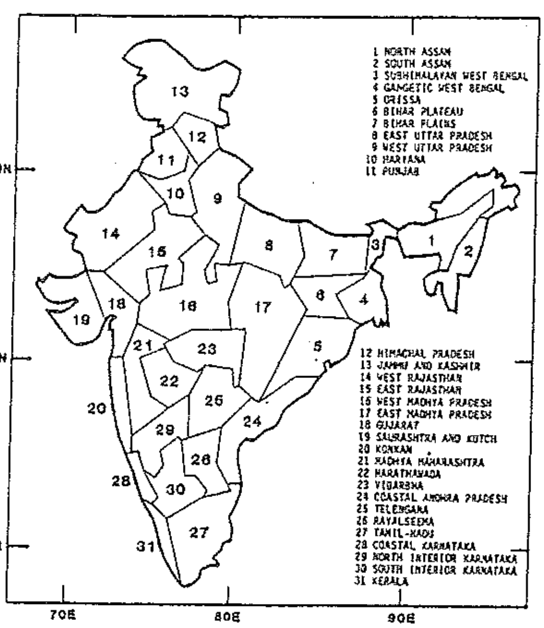

Figure 14.1. Locations and names of 31 subdivisions over India.

subdivisions, but these have been reversed by the large number of years with deficient rainfall in the 1960s and 1970s. Power spectrum analyses did reveal some evidence of a 2- to 3.5-year cycle in some subdivisions. Dhar et al. (4) computed monthly mean rainfall for the whole of India from corresponding subdivisional data for 1901–1960. Annual and seasonal mean rainfall statistics were computed from these monthly values. The authors showed that a direct relationship usually exists between the rainfall amount and the number of cyclonic storms crossing India. Pareek and Ramaswamy (5) analyzed the summer monsoon rainfall over Burma for the period 1907–1938 and concluded that droughts are not a serious problem in Burma. Parthasarathy and Mooley (6) studied seasonal rainfall for the whole of India. The data, covering the period 1841–1977, were derived from rain gauge stations. Two series were constructed; one for prepartition India and Burma covering the period 1841–1935; another for Indian rain only, 1901–1977. From these two series a homogeneous series was constructed for 1866–1970. Some data were discarded because of unreliability and sparse coverage (1971–1977). They found no trend in the rainfall data, but pointed out that the period 1931–1960 had higher than normal rainfall. A power spectrum analysis showed evidence of a possible cycle with a period of between 2 and 3 years.

Bhalme and Mooley (7) used monthly mean percentage departure rainfall data for the period 1891–1975 for 31 subdivisions to construct indices of flood and drought intensity. They found that large-scale floods (heavy rainfalls) and droughts each occur about 15 times per century. They also found a periodicity of about 20 years in the general flood index. Frequent large-scale droughts were found to appear during the periods 1891–1920 and 1961–1975, while there were few droughts between 1921–1960. They also studied composites of circulation patterns for flood and drought years and indicated that departures from normal of the monsoon rain may be foreshadowed by circulation anomalies in the upper troposphere during May. Mooley et al. (8) considered a time series from 1871–1978 of annual rainwater volume falling on the whole of India, derived from a network of 306 rain gauge stations. Time series of the area covered by excess and deficit rain were prepared to show the extreme years. The authors point out that in the period 1921–1950 there was only one excess and one deficit (as per their definition) year of rainfall. They also commented on the economic impact of extreme rainfall. Mooley and Pant (9) studied the historical data for 1771–1977 and classified 32 major droughts. These years did not necessarily correspond to the largest negative departures in the time series for seasonal rain because they considered the annual rainfall.

1.2 Variability of Monsoon Rainfall over India, 1901–1981

The observed interannual variability of monsoon rainfall over India based on 81 years of data is presented here. The percentage departure from normal rainfall for the months of June, July, August, September, and the seasonal mean, for 81 years (1901–1981), for 31 subdivisions of India (Fig. 14.1) was obtained from the India Meteorological Department (the seasonal mean is given in Table 14.1). Various

Table 14.1. Seasonal Rainfall Anomalies (percentage departure from normal) for 31 Subdivisions

Year	Subdivision Number 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
1901	−4	−3	−18	−5	−27	−15	−26	−16	−5	−25	−20	−11	−14	−53	−36	−10
1902	15	2	24	−9	−3	−4	−2	−5	−4	−4	−25	−35	−26	−31	−9	−17
1903	5	4	−3	−14	0	−28	−17	−5	−16	−11	11	−6	45	12	4	−3
1904	−6	−2	−33	0	6	9	−2	6	8	−1	−40	−17	−16	−51	16	−18
1905	2	10	14	5	−15	0	23	−13	−44	−49	−32	−30	3	−59	−62	−23
1906	−4	4	3	−14	−11	−15	1	4	31	21	11	49	24	−16	−3	14
1907	0	9	2	−9	1	8	−9	−40	−39	−32	−22	−53	−20	19	−27	−30
1908	−9	1	−31	12	10	−10	−41	−22	4	45	48	1	34	112	43	1
1909	−7	−7	−1	23	−1	13	20	19	13	36	57	25	52	40	6	−14
1910	1	9	22	−6	4	−1	16	2	8	15	23	12	23	11	3	0
1911	12	−1	9	−8	−2	12	15	1	−20	−15	−51	−34	−31	−63	−35	−23
1912	1	−14	−6	−14	−5	−20	−16	−10	−2	11	−6	−2	−21	−2	5	−12
1913	−16	−3	1	27	5	11	22	−26	−44	−28	−18	−24	−10	−27	−41	−19
1914	−7	−19	−14	−8	12	−15	−11	1	12	31	36	36	35	2	8	−8
1915	3	14	−18	−17	−17	−23	−1	32	−5	−37	−45	−6	−22	−72	−57	−22
1916	−9	−8	27	5	−5	−4	21	27	34	31	18	2	32	45	34	21
1917	8	−6	−5	5	5	11	1	18	30	82	90	48	67	139	82	26
1918	28	26	13	11	−13	−5	21	−35	−43	−57	−49	−53	−40	−76	−58	−40
1919	−4	−13	−6	9	11	17	4	5	5	4	−10	−7	14	3	23	27
1920	−5	−11	17	−8	1	10	−2	−13	−14	−26	−31	−22	−47	−28	−16	−30
1921	5	5	16	−2	−5	1	13	15	48	−5	−33	−7	−2	−20	−13	−9
1922	−7	−3	20	41	10	22	32	42	33	14	0	30	−4	−11	5	−5
1923	−6	1	−11	−5	−23	7	−22	6	−1	−3	22	−15	−9	−4	12	16
1924	1	−3	13	−6	−27	4	25	18	27	27	0	8	−5	4	41	4
1925	−8	−18	5	−25	27	−1	1	19	21	12	26	8	12	−32	−27	−20
1926	−3	−5	3	23	3	7	−6	0	−1	18	8	12	22	56	22	4
1927	10	12	4	−17	4	−14	−17	−11	2	−9	−17	24	−4	16	−1	−18
1928	−1	−3	10	19	4	−6	−12	−37	−35	−43	−29	−26	−9	−4	−30	−14
1929	−2	4	−11	−1	6	0	−7	−4	−25	−43	−27	−11	9	10	−18	−12
1930	1	4	−25	0	−2	5	−6	28	1	10	5	−21	−18	−3	−8	−5
1931	10	−12	12	−11	−10	−8	−7	−1	3	−1	10	−7	−11	40	9	9
1932	10	0	−2	−18	−8	−13	−28	−26	−2	−8	−9	18	−4	−3	−16	3
1933	−8	1	−6	26	27	3	9	−21	20	82	65	24	21	27	43	18
1934	8	4	1	−20	11	−6	2	10	9	0	−22	2	−15	19	20	28
1935	15	13	16	−21	−6	−7	9	−7	−14	−6	−12	−10	−15	−12	0	−4
1936	−5	−4	−2	7	25	17	27	53	49	16	11	14	−1	−10	−15	−6
1937	−9	−2	−17	6	5	0	−9	−1	−17	−16	−15	7	−38	−6	−1	9
1938	7	7	22	−12	−13	−9	26	40	−4	−46	−27	−16	−13	−29	−20	3
1939	−2	−7	−2	23	2	12	5	2	−12	−36	−41	−12	−23	−59	−36	−4
1940	−11	−1	−11	−5	17	−17	−14	−16	−12	−22	−2	−8	−10	−2	−14	8
1941	−7	3	−1	23	8	4	6	−18	−36	−35	−11	−20	−2	−19	−26	−26
1942	−2	4	−23	19	−1	25	7	0	28	63	53	47	28	16	48	24
1943	−1	−6	12	−1	32	13	−6	12	16	−11	−25	47	0	10	7	−3
1944	−6	−1	−2	4	−3	−4	−5	−4	−23	−15	−5	−2	14	87	25	32
1945	−3	9	−6	−24	−3	−11	−16	−13	18	30	49	0	−18	28	30	15
1946	−3	−13	−13	11	19	4	3	−6	3	−4	−20	11	−6	−18	27	15
1947	3	19	−14	−13	−14	−14	−10	3	2	−9	19	9	8	−5	4	15
1948	4	5	1	−8	−13	−2	4	32	17	15	3	−4	15	−23	1	19
1949	7	4	−2	−3	−20	2	10	7	10	5	−4	−22	−31	−15	−17	−2
1950	1	−8	1	13	−7	13	−6	−8	15	11	91	28	46	9	13	−7
1951	−12	−18	−3	−18	−17	−21	−29	−22	−25	−45	−26	−17	−44	−43	−47	−27
1952	−2	−5	3	−8	2	5	2	7	4	2	0	−9	−12	4	7	−7

(Table continues on p. 406.)

TABLE 14.1 (Continued)

Year	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31
	Subdivision Number														
1953	−7	23	64	20	4	16	4	7	23	14	23	9	28	26	−15
1954	−11	40	40	42	16	26	10	28	14	−2	−4	21	10	5	−10
1955	1	3	−10	33	4	51	14	6	38	18	−7	−5	20	−18	−13
1956	9	31	61	22	17	16	−3	34	30	23	19	2	31	4	−20
1957	−20	−28	−2	9	−7	6	−6	15	15	−3	−17	−8	17	−12	−2
1958	−3	25	20	49	23	33	4	32	30	3	−17	14	1	15	−7
1959	16	53	102	27	30	43	49	26	40	8	−13	38	26	52	30
1960	−8	−28	−18	11	4	−10	−11	15	−2	2	2	−2	15	−6	−7
1961	38	2	91	31	5	−1	33	11	17	−14	26	62	5	42	50
1962	−26	−13	−19	15	−4	18	−3	21	23	−16	6	13	8	9	2
1963	−10	7	−25	31	−3	47	2	−11	18	−19	3	−6	−11	−16	−15
1964	8	11	35	12	12	15	2	39	10	46	6	−2	54	22	−3
1965	−40	−31	11	2	−2	15	−28	−4	8	2	−8	−9	11	−18	−30
1966	−27	−16	−21	−21	−11	−13	−14	7	−3	17	29	−21	−3	−15	−26
1967	7	−3	40	−1	11	−1	−11	1	7	−3	−7	5	0	−14	−9
1968	−20	−22	−31	−26	−7	−7	−13	−30	−26	−21	−3	5	−9	−4	31
1969	−9	−6	−37	13	31	15	−6	−17	2	−15	−26	4	6	−5	−13
1970	11	53	79	25	9	45	21	15	27	20	−4	26	11	−8	−14
1971	4	−10	13	−4	−21	−22	−29	−18	−39	−29	10	2	−17	19	5
1972	−12	−40	−55	−31	−47	−45	−38	−26	−34	−31	9	−26	−14	−7	−14
1973	−4	38	−18	5	13	20	0	−26	−4	−5	17	0	2	15	−15
1974	−27	−67	−72	10	−5	−34	−34	−13	−25	4	16	14	−5	2	8
1975	15	26	−4	13	24	50	13	27	14	9	43	59	41	24	19
1976	−15	67	−1	9	31	−10	−9	−6	17	−3	15	−7	6	−38	−49
1977	5	43	10	9	−12	−20	−4	−17	−29	3	14	5	−6	0	−16
1978	−2	−4	−5	1	−12	−4	12	47	40	38	2	18	19	−1	−4
1979	−52	−25	60	−38	1	12	−3	−28	−18	−15	13	−19	9	32	1
1980	33	22	61	0	15	24	5	17	−9	−15	−32	−1	2	12	2
1981	−13	4	1	15	24	6	14	10	18	22	41	12	46	16	24
Mean	−3	1	4	3	1	3	0	2	1	−6	−2	2	5	3	−2
S.D.	15	30	39	18	17	25	18	20	21	24	18	17	20	18	19

statistics have been calculated from these data.* The percentage departure from normal is the difference (expressed as a percentage) between the average monthly mean rainfall for all the stations in a subdivision for which data is available for that particular month and year and the normal rainfall (the average, over all years, for that month) for the same stations. The percentage departure of seasonal rainfall is calculated in the same way. The seasonal mean percentage departure is not identical but close to the average of the percentage departures for the four months.

It is important to note that the number of stations varies with month and year, that is, the same stations are not available for each year. Thus a straightforward averaging of rainfall for all the available stations would show some interannual variation solely due to the sampling error that arises from using different stations each year. If the total number of stations is not large, and/or the rainfall in a given subdivision has large geographical variability, the spurious interannual variability arising from the sampling error could be quite large and might mask the real interannual variability. We consider the percentage departure from normal as a more appropriate parameter than the actual rainfall to study interannual variability.

We have also examined the interannual variability of the percentage departure of rainfall for several large regions of India. A regional average is defined as the area weighted mean of all the subdivisions in a given region. The "rainfall anomaly" is defined as the deviation of the percentage departure of rainfall from its long-term mean and the "normalized rainfall anomaly" is defined as the ratio of rainfall anomaly and its standard deviation. The coefficient of variability is defined as the ratio of the standard deviation and the mean rainfall.

Walker (10) suggested that rainfall anomalies over several subdivisions of India should be grouped together to define areal averages for large homogeneous regions. The basis for Walker's groupings was the uniformity of correlation coefficients between the anomalies and geographically distant atmospheric parameters. The two regions (Northwest and Peninsula) which were first defined by Walker and which are still being used by the India Meteorological Department for seasonal forecasts are shown in Figure 14.2 and defined in Table 14.2. The interannual variability of the seasonal rainfall for the subdivisions in the Northeast region was found to be small, and the correlation with the distant parameters was not significant, so Walker did not use these subdivisions for seasonal prediction.

It can be argued that the criteria chosen by Walker to define homogeneous regions of India were not quite appropriate because they were based on correlations with distant parameters. A more appropriate procedure might have been to examine the structure of the dominant mode of variability of the Indian rainfall data itself.

We have attempted to determine spatially homogeneous regions of India by the following three methods: (a) correlation among all the subdivisions, (b) correlation

*Percentage departure from normal, $P = 100(\tilde{R} - \tilde{\bar{R}})/\tilde{\bar{R}}$, where $\tilde{R}$ is the monthly or seasonal mean rainfall, and $\tilde{\bar{R}}$ is the "normal" monthly or seasonal mean rainfall for 50 years (1901–1950).

Rainfall anomaly, $P' = (P - \bar{\bar{P}})$, where $\bar{\bar{P}}$ is time mean of P for 81 years.

Normalized rainfall anomaly, $P^* = P'/\sigma$ where σ is the standard deviation of P' for 81 years.

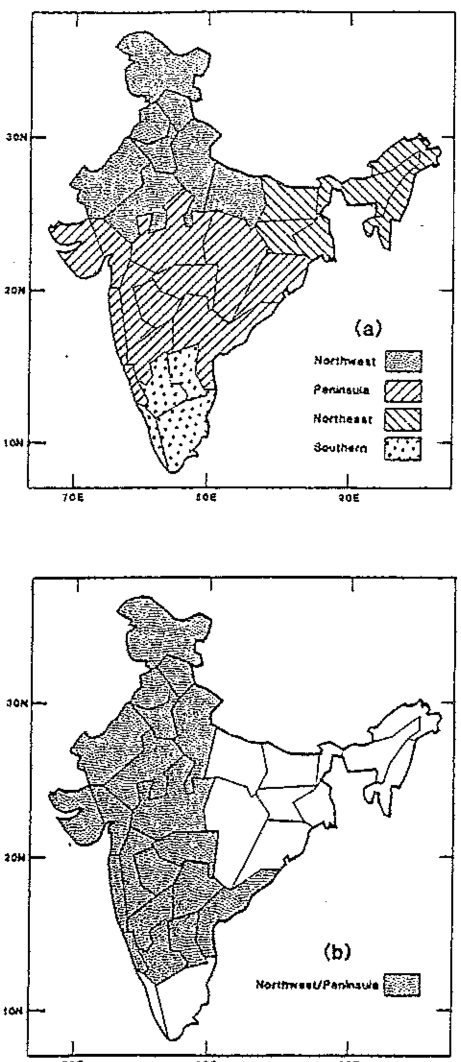

Figure 14.2. Locations and names of regions used for spatial averaging: (*a*) Northwest, Peninsula, Northeast, Southern, (*b*) Northwest-Peninsula. (*Figure 14.2 continues on page 414.*)

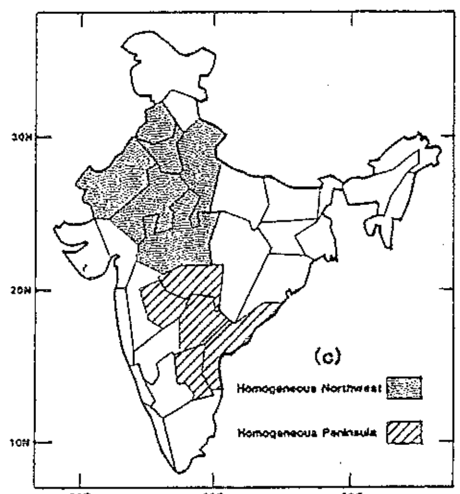

Figure 14.2. (*continued*) (*c*) Homogeneous Northwest, Homogeneous Peninsula.

between the subdivisional rainfall anomaly and the Southern Oscillation, and (c) empirical orthogonal functions (EOFs). Table 14.3 gives the correlation coefficients among rainfall anomalies for all the subdivisions. Correlations of 0.7 or higher are found between adjacent subdivisions as shown in Figure 14.3*a*. Six subdivisions in Northwest India (Western Uttar Pradesh, Haryana, Punjab, Western Rajasthan, Eastern Rajasthan, and Western Madhya Pradesh) and five subdivisions in Peninsula India (Marathawada, Vidarbha, Coastal Andhra Pradesh, Telengana, and Rayalseema) are the only two contiguous groups of highly homogeneous rainfall anomalies. We will refer to them as the Homogeneous Northwest and Homogeneous Peninsula regions. A similar map for the threshold correlation coefficient of 0.5 is shown in Figure 14.3*b*. This defines a much larger number of nearly homogeneous subdivisions which can be used for spatial averaging of Indian summer monsoon rainfall. Figure 14.3*c* shows the correlation coefficients between the normalized winter to spring pressure tendency for Darwin (which can be considered to be an index of the Southern Oscillation phenomenon), and seasonal rainfall anomalies for each subdivision using 81 years of data.

There is a marked similarity between the homogeneous subdivisions of Figure 14.3*b* and subdivisions with large correlation coefficients (significant at the 95% confidence level) in Figure 14.3*c*. This suggests that the Southern Oscillation influences a large number of subdivisions of India and, therefore, the separate groupings of

TABLE 14.2 Names of Subdivisions Constituting the Eight Regions

Region

1. *Northwest India*: East Uttar Pradesh (8), West Uttar Pradesh (9), Haryana (10), Punjab (11), Himachal Pradesh (12), Jammu and Kashmir (13), West Rajasthan (14), East Rajasthan (15)
2. *Peninsula India*: Orissa (5), West Madhya Pradesh (16), East Madhya Pradesh (17), Gujarat (18), Saurashtra and Kutch (19), Konkan (20), Madhya Maharashtra (21), Marathawada (22), Vidarbha (23), Coastal Andhra Pradesh (24), Telengana (25), Coastal Karnataka (28), North Interior Karnataka (29)
3. *Southern India*: Rayalseema (26), Tamil-Nadu (27), South Interior Karnataka (30), Kerala (31)
4. *Northwest India*: North Assam (1), South Assam (2), Subhimalayan West Bengal (3), Gangetic West Bengal (4), Bihar Plateau (6), Bihar Plains (7)
5. *Homogeneous Northwest*: West Uttar Pradesh (9), Haryana (10), Punjab (11), West Rajasthan (14), East Rajasthan (15), West Madhya Pradesh (16)
6. *Homogeneous Peninsula*: Marathawada (22), Vidarbha (23), Coastal Andhra Pradesh (24), Telengana (25), Rayalseema (26)
7. *Northwest–Peninsula*: West Uttar Pradesh (9), Haryana (10), Punjab (11), Himachal Pradesh (12), Jammu and Kashmir (13), West Rajasthan (14), East Rajasthan (15), West Madhya Pradesh (16), Gujarat (18), Saurashtra and Kutch (19), Konkan (20), Madhya Maharashtra (21), Marathawada (22), Vidarbha (23), Coastal Andhra Pradesh (24), Telengana (25), Rayalseema (26), Coastal Karnataka (28), North Interior Karnataka (29), South Interior Karnataka (30)
8. *Whole India*: All 31 Subdivisions

Northwest India and Peninsula India, as suggested by Walker, are not appropriate to study the variability and predictability of Indian summer monsoon rainfall in relation to the Southern Oscillation. This is mainly because large-scale rainfall anomalies over Northwest India and Peninsula India are not independent. This point is further supported by Figure 14.4 which shows the composite rainfall anomaly maps for years with heavy and deficient monsoon rainfall for both Northwest and Peninsula regions. Figures 14.4*a* and 14.4*b* show the composite anomaly maps for the Northwest region for the years with normalized rainfall anomalies equal to or greater than 1.0, and less than or equal to −1.0, respectively. Figures 14.4*c* and 14.4*d* show similar composites for the Peninsula region.* It can be seen that the years with heavy or deficient rainfall over one of the regions is accompanied by anomalies of the same sign (although not of the same magnitude) for a large number of subdivisions over India.

* Composite anomaly for the years 1908, 1909, 1916, 1917, 1933, 1942, 1953, 1956, 1957, 1958, 1961, 1973, 1975, 1976, and 1978 is shown in Figure 14.4*a*; for 1901, 1905, 1911, 1913, 1915, 1918, 1920, 1939, 1951, 1965, 1972, and 1979 in Figure 14.4*b*; for 1914, 1917, 1933, 1955, 1956, 1958, 1959, 1961, 1964, 1970, 1975, and 1980 in Figure 14.4*c*; and for 1901, 1904, 1905, 1911, 1918, 1920, 1941, 1951, 1965, 1966, 1968, 1972, 1974, and 1979 in Figure 14.4*d*.

TABLE 14.3 Correlation Coefficients Among All Subdivisions for Seasonal Mean Rainfall Anomalies

	Subdivision Number														
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
1	1.00	.34	.22	−.23	−.33	−.13	.13	−.00	−.10	−.12	−.07	−.14	−.11	−.12	−.10
2		1.00	.15	−.23	−.24	−.24	.03	−.29	−.26	−.22	−.23	−.23	−.26	−.17	−.31
3			1.00	−.12	−.08	−.06	.43	.07	−.05	−.18	−.21	−.16	−.20	−.12	−.20
4				1.00	.18	.67	.32	.09	−.03	.10	.07	.16	.07	.12	.17
5					1.00	.35	.12	.21	.28	.32	.17	.29	.09	.24	.30
6						1.00	.49	.35	.25	.28	.18	.26	.10	.17	.34
7							1.00	.53	.19	.06	−.05	.01	−.05	−.15	.03
8								1.00	.57	.33	.21	.36	.18	.09	.27
9									1.00	.75	.50	.66	.44	.31	.60
10										1.00	.72	.62	.49	.62	.77
11											1.00	.59	.59	.64	.67
12												1.00	.40	.42	.58
13													1.00	.50	.44
14														1.00	.75
15															1.00
16															
17															
18															
19															
20															
21															
22															
23															
24															
25															
26															
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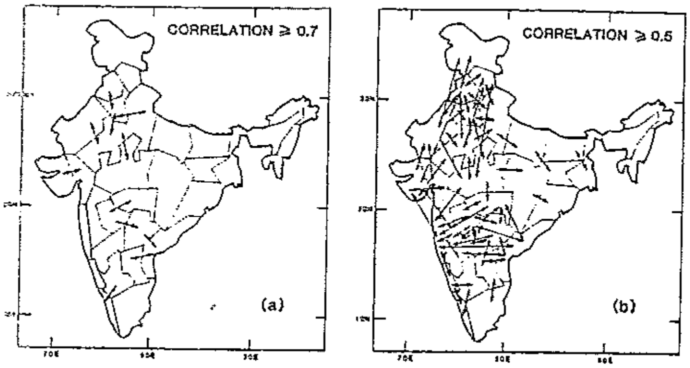
28															
29															
30															
31															

(Table continues on p. 418.)

TABLE 14.3 *(Continued)*

	Subdivision Number															
	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31
1	−.05	−.14	−.17	−.33	−.01	−.16	−.13	−.20	.01	−.07	.08	.00	−.11	−.11	−.21	−.13
2	−.24	−.26	−.22	−.17	−.32	−.16	−.14	−.06	−.13	−.10	.06	−.00	−.34	−.20	−.22	−.18
3	−.11	−.19	−.03	.00	.02	−.06	.06	−.08	.14	−.05	.25	−.04	−.16	.01	−.06	−.10
4	.09	.08	.19	.13	−.00	−.02	−.06	.01	−.19	−.14	−.06	.05	.08	−.03	.05	.10
5	.25	.50	.23	.29	.14	.13	.02	.15	.00	.08	−.14	−.16	.16	−.01	.06	.05
6	.23	.37	.17	.07	−.01	−.12	−.21	−.11	−.23	−.28	−.16	−.03	.14	−.21	.10	.17
7	.09	.05	−.03	−.14	−.03	−.06	.04	−.04	−.05	−.09	.10	−.01	−.18	−.20	−.10	−.07
8	.37	.47	.02	−.01	.25	.14	.24	.14	.29	.21	.22	.13	.15	−.00	.07	.09
9	.49	.45	.21	.10	.41	.21	.29	.24	.36	.39	.30	.18	.33	.06	.17	.16
10	.51	.39	.43	.31	.45	.36	.39	.32	.36	.43	.27	.29	.42	.25	.28	.14
11	.48	.30	.54	.37	.47	.37	.42	.35	.39	.48	.18	.22	.46	.22	.27	.08
12	.51	.48	.43	.35	.41	.28	.35	.33	.33	.36	.16	.20	.35	.13	.28	.12
13	.27	.19	.33	.30	.33	.31	.39	.30	.39	.46	.31	.28	.35	.35	.25	.09
14	.52	.35	.65	.50	.42	.41	.37	.44	.34	.43	.30	.31	.43	.32	.25	.04
15	.74	.46	.56	.37	.47	.40	.38	.36	.20	.34	.17	.24	.49	.20	.34	.26
16	1.00	.62	.47	.24	.59	.47	.44	.54	.20	.34	.21	.17	.44	.14	.24	.14
17		1.00	.28	.26	.33	.22	.21	.44	.23	.25	.08	−.09	.38	.02	.21	.14
18			1.00	.72	.54	.58	.33	.47	.27	.36	.30	.16	.33	.25	.22	−.03
19				1.00	.41	.45	.30	.46	.37	.34	.36	.14	.32	.32	.39	.08
20					1.00	.67	.62	.57	.45	.63	.34	.14	.58	.45	.24	.09
21						1.00	.64	.59	.39	.56	.39	.24	.45	.63	.35	.13
22							1.00	.70	.51	.71	.46	.27	.37	.54	.29	.03
23								1.00	.51	.68	.45	.19	.42	.39	.31	.07
24									1.00	.73	.71	.35	.34	.57	.31	.03
25										1.00	.56	.25	.46	.56	.21	.02
26											1.00	.54	.12	.49	.24	−.08
27												1.00	.24	.37	.41	.26
28													1.00	.41	.53	.59
29														1.00	.49	.23
30															1.00	.71
31																1.00



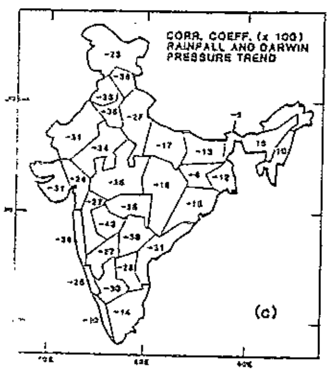


Figure 14.3. Subdivisions with correlation coefficients equal to or greater than (*a*) 0.7, (*b*) 0.5 for seasonal rainfall anomalies are joined by a line with an arrow at each end, and (*c*) correlation coefficient between seasonal rainfall anomaly and normalized winter to spring pressure tendency for Darwin.

Figure 14.5 shows the structure of the empirical orthogonal functions (EOFs) ›r the normalized seasonal rainfall anomaly. EOFs describe large-scale patterns 'hich represent maximum variability. The percentage departure from normal rainfall normalized by dividing it by its standard deviation. The first EOF explains 31.5% f the total variance while the second, third, and fourth functions explain only 1.7%, 7.6%, and 6.3% of the variance, respectively. It is again seen that the large alues for the first function in Figure 14.5 are confined to nearly the same subdivisions hich show large correlation coefficients with the Southern Oscillation in Figure 4.3c. Results shown in Figures 14.3, 14.4, and 14.5 suggest that the most dominant ıttern of the spatial homogeneity of the Indian summer monsoon rainfall is determined

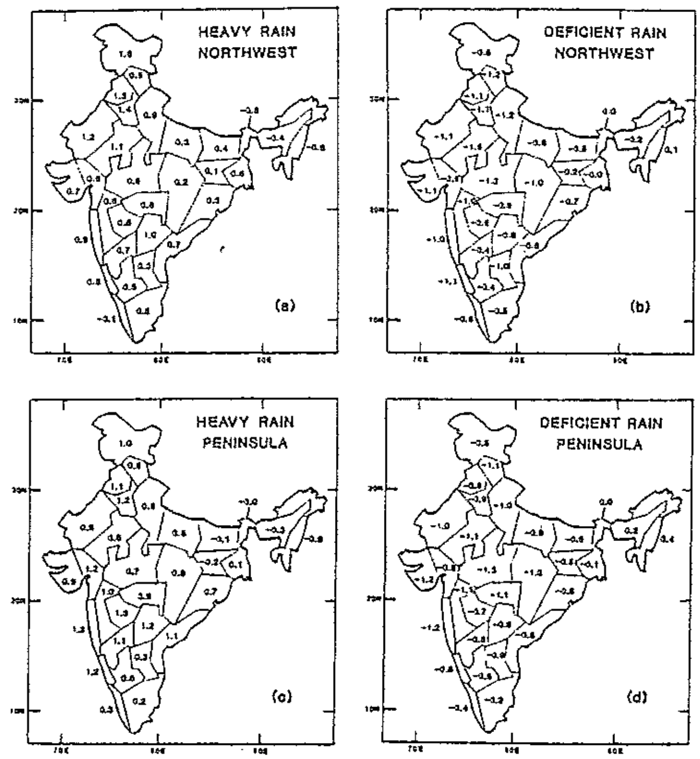


Figure 14.4. Composite maps of normalized seasonal rainfall anomaly for: (*a*) heavy rainfall over Northwest India, (*b*) deficient rainfall over Northwest India, (*c*) heavy rainfall over Peninsular India, (*d*) deficient rainfall over Peninsula India.

by its relationship with the Southern Oscillation, which simultaneously affects a large number of subdivisions over India. We, therefore, propose to examine regional averages over a larger area which would be referred to as the combined Northwest–Peninsula region (see Fig. 14.2*b* and Table 14.2).

As mentioned earlier, the Northwest and Peninsula regions were defined by Walker and are currently used by the India Meteorological Department for long-range forecasting of monsoon rainfall. The Southern and the Northeast regions are simply the remaining subdivisions. Homogeneous Northwest and Homogeneous

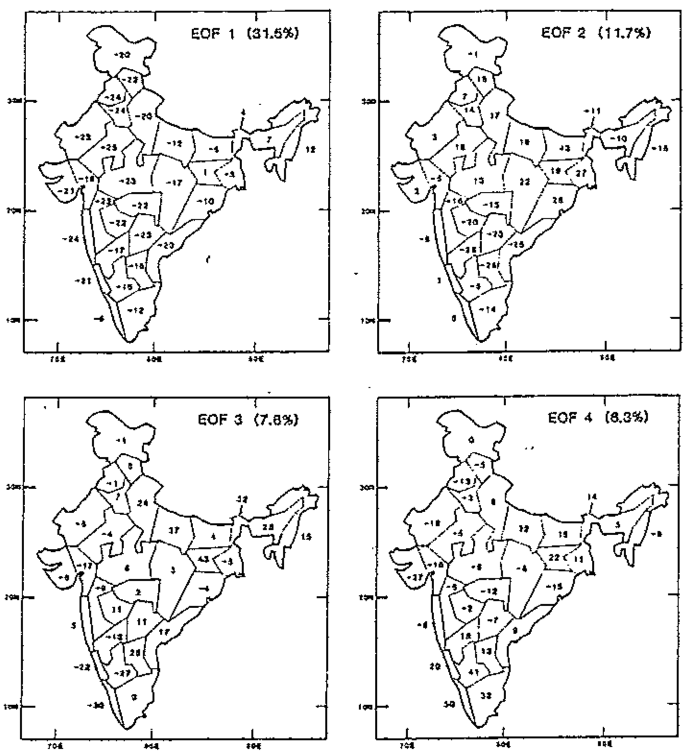


Figure 14.5. First four empirical orthogonal functions (EOFs) for normalized seasonal rainfall anomalies. Percentage of variances explained is indicated at the top.

Peninsula are defined on the basis of high correlation coefficients for seasonal rainfall anomalies (see Fig. 14.3*a*). The combined region of Northwest–Peninsula is defined on the basis of significant correlation with the Southern Oscillation. Table 14.4 gives the rainfall anomalies for the eight regions (listed in Table 14.2) for the period 1901–1981.

Table 14.5 shows the correlation coefficients of the seasonal rainfall anomalies among the eight regions. It is seen that the correlation coefficient between the Northwest and Peninsula regions is 0.71, which is perhaps too high to consider these regions as independent. It should also be noted that the correlation coefficient

TABLE 14.4 Seasonal Rainfall Anomalies (percentage departure from normal) for the Eight Regions in Table 14.2

	Region Number							
Year	1	2	3	4	5	6	7	8
1901	−24	−15	−7	−10	−26	−14	−21	−16
1902	−18	−12	−7	3	−16	−14	−15	−11
1903	10	8	25	−7	0	28	13	8
1904	−12	−21	−24	−2	−16	−28	−22	−15
1905	−33	−21	−24	8	−44	−16	−29	−20
1906	11	6	4	−4	7	14	11	6
1907	−22	−5	−5	1	−19	−14	−10	−9
1908	37	13	−13	−10	41	17	26	14
1909	31	0	9	5	15	6	16	11
1910	12	12	14	5	7	24	15	11
1911	−32	−21	−18	7	−36	−19	−32	−20
1912	−6	0	−3	−11	−3	−11	0	−4
1913	−27	−5	−18	5	−30	−15	−14	−12
1914	16	21	−6	−12	6	27	21	12
1915	−27	−8	8	−3	−40	10	−19	−12
1916	32	13	21	1	32	30	25	18
1917	71	22	14	3	71	28	48	33
1918	−51	−34	−46	18	−54	−33	−48	−32
1919	8	8	7	0	13	−3	7	7
1920	−26	−25	−4	−3	−24	−42	−27	−20
1921	1	2	15	5	−3	14	3	3
1922	11	−8	−20	13	4	−17	−6	0
1923	0	−7	1	−5	7	−11	−3	−4
1924	14	−8	33	4	16	−7	6	5
1925	1	−7	−14	−10	−12	−12	−10	−6
1926	21	9	0	1	20	−12	15	11
1927	1	5	4	−2	−3	1	5	3
1928	−22	1	−22	−1	−21	3	−8	−9
1929	−7	−10	−6	−1	−13	−21	−12	−7
1930	−2	−6	−22	0	−3	−17	−8	−5
1931	7	4	−1	−4	15	7	8	3
1932	−8	0	−5	−7	−4	−6	−1	−4
1933	24	21	11	4	31	16	26	18
1934	5	10	−30	0	16	1	4	3
1935	−10	−1	−4	5	−8	6	−4	−3
1936	13	−1	−7	6	4	−1	−3	4
1937	−13	3	−16	−4	−4	−14	−7	−5
1938	−10	9	14	6	−15	34	0	3
1939	−27	−12	−14	4	−28	−23	−23	−14
1940	−10	2	−3	−9	−4	3	−3	−4
1941	−19	−18	−4	4	−26	−24	−18	−13

(*Table continues on page 424.*)

TABLE 14.4. *(Continued)*

Year	Region Number 1	2	3	4	5	6	7	8
1942	28	12	−6	8	31	7	20	15
1943	8	4	−12	−1	4	−8	1	3
1944	19	13	−2	−3	29	0	18	11
1945	10	11	−15	−7	24	2	12	5
1946	−2	11	3	−2	4	−6	5	4
1947	3	9	14	−1	5	14	8	6
1948	7	−6	−7	2	4	−6	−3	−1
1949	−11	4	12	4	−5	23	0	0
1950	21	−2	−3	1	11	−12	11	6
1951	−36	−15	−13	−18	−35	0	−24	−22
1952	0	−13	−35	−2	1	−30	−12	−9
1953	23	8	18	6	−3	13	17	13
1954	−9	12	−2	−11	−3	15	7	0
1955	19	15	−6	−4	25	25	16	11
1956	23	20	11	−1	13	20	21	16
1957	46	−7	−11	−14	−1	6	21	8
1958	23	15	−3	−11	19	21	22	11
1959	10	29	15	−8	7	35	29	16
1960	0	−1	−1	6	1	−1	−3	0
1961	34	31	25	−7	38	12	32	25
1962	3	−4	2	5	0	10	2	0
1963	−7	−1	−9	−4	−8	7	−4	−4
1964	10	15	18	−1	10	20	15	11
1965	−32	−15	−11	−1	−31	−2	−19	−18
1966	−6	−16	8	−7	−17	−2	−8	−9
1967	8	4	−8	−12	6	−1	5	2
1968	−19	−15	−3	0	−23	−20	−17	−13
1969	−18	1	−16	−4	−9	−4	−8	−7
1970	12	21	−1	0	10	25	19	13
1971	3	−7	4	9	6	−29	−6	0
1972	−26	−27	−6	−16	−23	−34	−29	−23
1973	26	9	8	1	40	−4	20	13
1974	−22	−26	9	21	−25	−21	−22	−14
1975	60	16	28	−8	46	21	39	27
1976	29	5	−11	5	27	0	18	11
1977	19	−1	4	6	24	−15	10	7
1978	24	8	8	3	26	28	18	12
1979	−33	−19	11	−13	−37	−12	−18	−19
1980	11	18	−12	−2	2	3	8	9
1981	−11	4	28	−6	−17	14	0	0
Mean	2	1	−2	−1	0	0	2	1
S.D.	22	14	15	7	23	18	18	12

TABLE 14.5 Correlation Coefficients for Seasonal Mean Rainfall Anomalies Among Eight Regions

Region	1	2	3	4	5	6	7	8
1 Northwest India	1.00	.71	.42	−.06	.93	.55	.93	.93
2 Peninsula India		1.00	.44	−.19	.76	.79	.90	.90
3 Southern India			1.00	−.06	.38	.55	.54	.55
4 Northeast India				1.00	−.05	−.18	−.16	−.03
5 Homogeneous Northwest					1.00	.51	.91	.91
6 Homogeneous Peninsula						1.00	.75	.72
7 Northwest–Peninsula							1.00	.98
8 Whole India								1.00

between the combined Northwest–Peninsula region and whole India is 0.98, which suggests that any statement about the combined Northwest–Peninsula region would generally be valid for the whole of India.

1.3 The Observed Variability

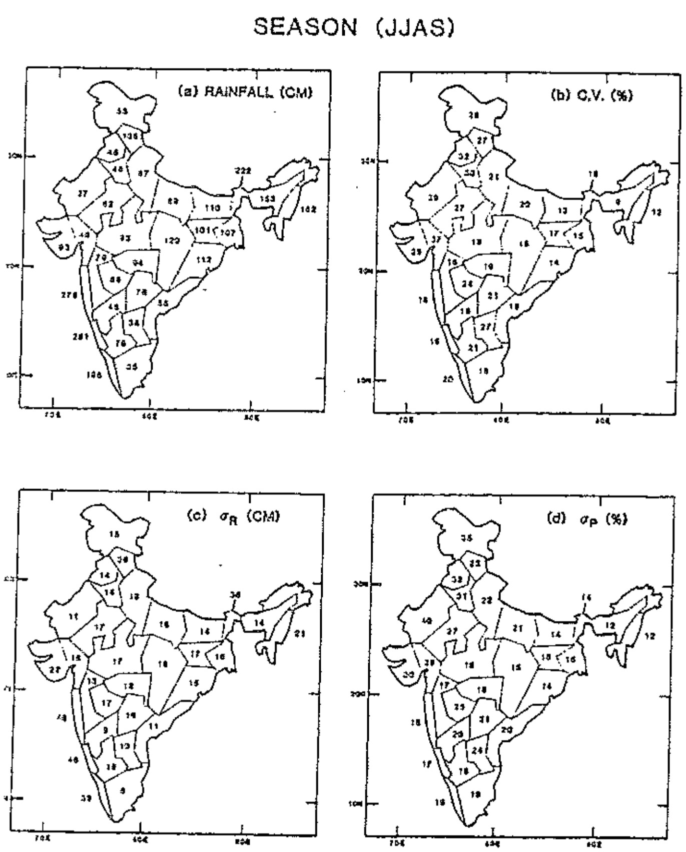
Table 14.6 gives the seasonal mean rainfall ($\bar{R}$) in mm (column 1), standard deviation of seasonal mean rainfall (σ_R) in mm (column 2), and coefficient of variation (CV) calculated from 70 years (1901–1970) of rainfall data for 31 subdivisions and eight regions (column 3). As pointed out earlier, the number and location of stations used for calculating subdivisional mean rainfall is not same for each year and, therefore, part of the variability shown could be due to sampling of different stations. Only 70 years of data were used for these calculations because rainfall reports from numerous stations had not yet been received for the period after 1970. The seasonal normal rainfall given in column 1 is very similar to the rainfall based on 50 years (1901–1950) of data and used as normal by the India Meteorological Department. The differences are less than 5% for most of the subdivisions and less than 10% for a few subdivisions near high mountains. Column 4 gives the standard deviation of seasonal percentage departure from normal rainfall (σ_p) for 81 years (1901–1981). Column 5 gives the standard deviation of monthly percentage departure from normal (σ_m) for all the four months (June, July, August, September) combined, and column 6 also gives the standard deviation of monthly percentage departure from normal (σ_{ms}) for all the four months combined, except that for each year the seasonal percentage departure from normal is subtracted from the monthly percentage departure before calculating the standard deviation of monthly means. The standard deviations σ_m and σ_{ms} are not the same because seasonal mean anomalies contribute towards the interannual variability of monthly means. The area weighted average of seasonal normal monsoon rainfall for 31 subdivisions (whole India) is 890 mm and the coefficient of variation of area weighted seasonal rainfall is 9.5%.

TABLE 14.6. Seasonal Normal Rainfall ($\bar{R}$ in mm), Standard Deviation of Seasonal Mean Rainfall (σ_R) in mm and Coefficient of Variation (CV) Based on 70 Years (1901–1970) of Data.

	MEAN($\bar{R}$)	σ_R(mm)	CV(X100)	σ_p(%)	σ_m(%)	σ_{ms}(%)
1 North Assam	1531.6	139.8	9.1	12.3	27.4	24.7
2 South Assam	1819.5	209.3	11.5	11.7	22.4	19.4
3 Subhimalayan West Bengal	2222.4	358.3	16.1	14.3	30.2	26.8
4 Gangetic West Bengal	1069.8	159.6	14.9	15.9	35.8	31.7
5 Orissa	1123.1	153.4	13.7	13.7	30.3	26.8
6 Bihar Plateau	1101.3	142.2	12.9	13.7	31.7	28.1
7 Bihar Plains	1005.4	166.5	16.6	15.9	35.3	31.1
8 East Uttar Pradesh	886.3	175.1	19.8	21.4	42.9	36.3
9 West Uttar Pradesh	869.9	186.1	21.4	21.9	49.0	43.4
10 Haryana	460.2	138.0	30.0	30.4	66.2	59.3
11 Punjab	451.2	143.5	31.8	32.3	72.6	64.1
12 Himachal Pradesh	1347.5	356.8	26.5	23.2	61.2	54.9
13 Jammu and Kashmir	528.5	152.3	28.8	35.2	62.5	53.1
14 West Rajasthan	270.0	106.2	39.3	40.3	78.7	68.4
15 East Rajasthan	619.0	169.3	27.4	27.0	58.8	52.1
16 West Madhya Pradesh	932.8	173.6	18.6	19.0	43.0	38.4
17 East Madhya Pradesh	1197.3	175.3	14.6	15.3	32.5	28.3
18 Gujarat	929.9	266.0	28.6	30.2	63.9	55.9
19 Saurashtra and Kutch	488.3	179.3	36.7	39.1	86.4	76.8
20 Konkan	2756.2	481.8	17.5	17.9	37.7	32.7
21 Madhya Maharashtra	793.1	127.9	16.1	17.3	35.7	31.2

TABLE 14.6. (Continued) Standard Deviation of Seasonal Percentage Departure from Normal (σ_p), Standard Deviation of Mean Monthly Percentage Departure from Normal (σ_m) for June, July, August, September, and Standard Deviation of Monthly Percentage Departure (σ_{ms}) after Removing Seasonal Means Calculated from 81 Years (1901–1981) of Data.

	MEAN($\bar{R}$)	σ_R(mm)	CV(X100)	σ_p(%)	σ_m(%)	σ_{ms}(%)
22 Marathawada	692.9	167.2	24.1	25.2	48.8	41.7
23 Vidarbha	936.4	174.6	18.6	18.3	37.6	32.3
24 Coastal Andhra Pradesh	583.1	109.8	18.8	20.0	34.9	28.7
25 Telengana	755.0	158.1	20.9	21.3	39.4	33.3
26 Rayalseema	376.2	99.6	26.5	23.7	48.8	42.8
27 Tamil-Nadu	344.7	62.8	18.2	18.5	36.0	31.1
28 Coastal Karnataka	2907.0	455.7	15.7	16.8	36.5	31.6
29 North Interior Karnataka	456.7	84.4	18.5	20.1	39.1	34.8
30 South Interior Karnataka	786.4	163.9	20.8	18.4	34.8	30.1
31 Kerala	1976.0	393.2	19.9	19.4	38.1	32.7
Region						
1 Northwest India	631.9	118.0	18.7	22.4	45.3	39.1
2 Peninsula India	944.3	117.4	12.4	13.5	28.9	24.9
3 Southern India	666.9	98.2	14.7	14.7	29.0	24.8
4 Northeast India	1394.1	90.9	6.5	7.2	18.3	16.9
5 Homogeneous Northwest	649.8	123.5	19.0	22.6	47.8	41.7
6 Homogeneous Peninsula	692.2	120.2	17.4	17.8	30.8	25.0
7 Northwest-Peninsula	729.4	112.7	15.4	17.8	34.3	29.0
8 Whole India	890.2	84.4	9.5	12.5	25.2	21.4



lgure 14.6. (*a*) Seasonal normal rainfall ($\bar{R}$) in cm, (*b*) coefficient of variation (CV) calculated from) years (1901–1971) of data, (*c*) standard deviation of seasonal mean rainfall (σ_R) in cm, and (*d*) andard deviation of seasonal percentage departure from normal (σ_p) calculated from 81 years (1901–)81) of data (see Table 14.6).

Figure 14.6 shows the maps of the first four quantities listed in Table 14.6. The rge values of subdivisional mean rainfall occur along the west coast and over ortheast India. For both of these regions ocal orography plays an important role determining the seasonal mean rainfall. The subdivisions in Northwest and Southeast

India get the least amount of rainfall. The coefficient of variation is largest for the subdivisions in Northwest India, primarily because normally the amount of rainfall is small but it can undergo large changes from year to year. The normal rainfall and coefficient of variation are both small for Southeast India. This can be seen more clearly from values of standard deviation of seasonal mean rainfall which are the smallest for Southeast India. Although the seasonal normal for subdivisions in Northeast India is quite large, the standard deviation of seasonal mean rainfall is relatively small giving rise to rather low values of coefficient of variation. The values of standard deviation of percentage departure of seasonal rainfall (shown in Fig. 14.6*d*) range from 12 to 40%, with the smallest values occurring over the astern part of Northeast India and the largest values over the western part of Northwest India. For most of the subdivisions over Central and Peninsula India, the values range from 15 to 25%. It can also be seen from Table 14.6 that the standard deviation of the percentage departure of seasonal rainfall ranges only from 13.5 to 22.6% for most of the large homogeneous regions of India, and the coefficient of variation is very similar to the standard deviation of percentage departure (Fig. 14.6).

Figures 14.7–14.10 show the mean rainfall, the coefficient of variation, the standard deviation of monthly mean rainfall, and the standard deviation of percentage departure for June, July, August, and September, respectively. In general, normal rainfall for July and August are comparable, but larger than that for June or September. The least rainfall occurs in June. The coefficient of variation is large for June (22 to 86%) and September (26 to 108%) and somewhat less for July and August (19 to 70%). As would be expected, the coefficient of variation for monthly mean rainfall is significantly larger than that for the seasonal mean rainfall. If the coefficients of variation were calculated for the monthly mean rainfall of individual stations, they would be even larger than those for subdivisions. The standard deviation of the percentage departure is comparable for July and August, and relatively less for June or September. The coefficient of variation is very similar to the standard deviation of percentage departure for each month. A more detailed discussion of intraseasonal variability as compared to the interannual variability will be presented later in this chapter. The standard deviation of percentage departure of combined July and August mean rainfall is comparable to that of seasonal mean rainfall for most of the subdivisions of India (Fig. 14.11).

The percentage departure of seasonal rainfall for 81 years for Northwest, Peninsula, Southern, and Northeast regions is shown in Figure 14.12, and for Homogeneous Northwest, Homogeneous Peninsula, combined Northwest–Peninsula, and whole India is shown in Figure 14.13. The years for which the normalized seasonal rainfall anomaly was equal to or greater than 1.0, or less than or equal to -1.0 are shown in Tables 14.7 and 14.8, respectively. It can be seen that during the 81 year period considered here, for the whole of India, there were 12 years of heavy rainfall (normalized anomaly equal to or greater than 1.0) and 14 years of deficient rainfall (normalized anomaly less than or equal to -1.0). A more detailed discussion of possible relationships between heavy and deficient rain and the Southern Oscillation is presented in Chapter 16.

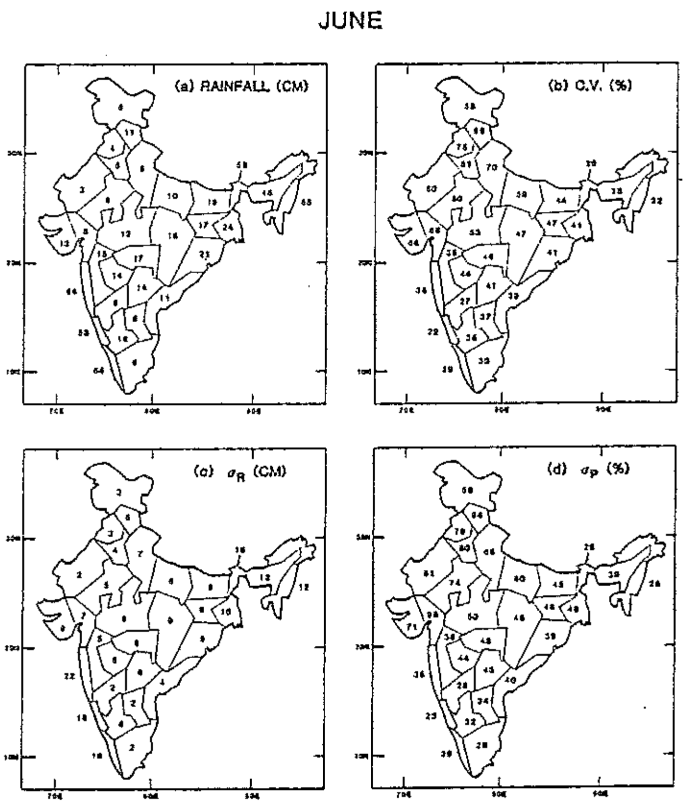


Figure 14.7. Same as Figure 14.6, except for June only.

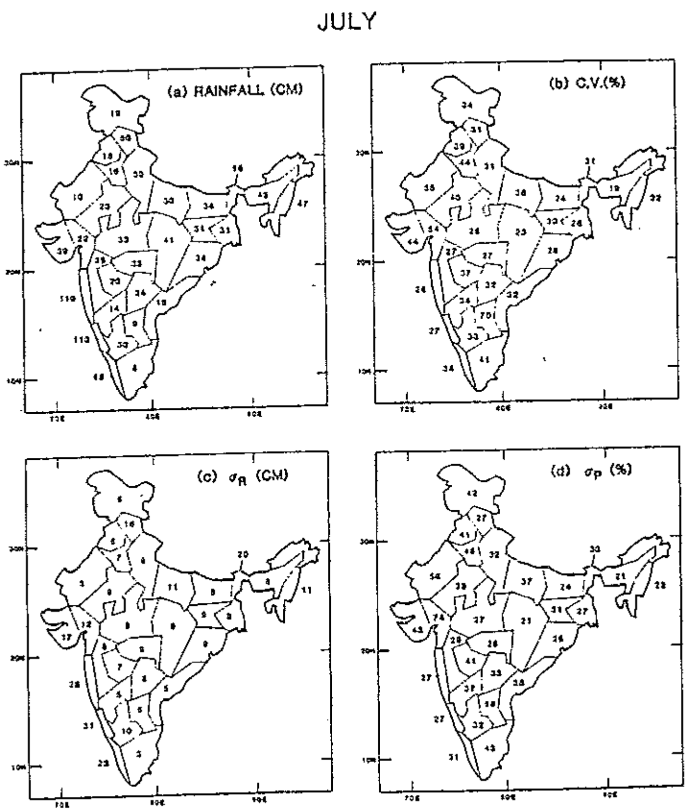


Figure 14.8. Same as Figure 14.6, except for July only.

AUGUST

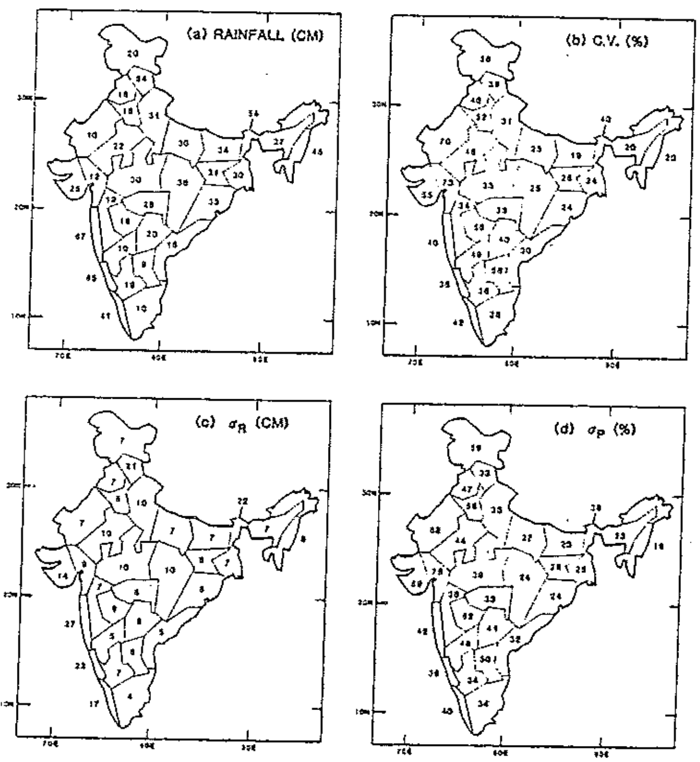


Figure 14.9. Same as Figure 14.6, except for August only.

SEPTEMBER

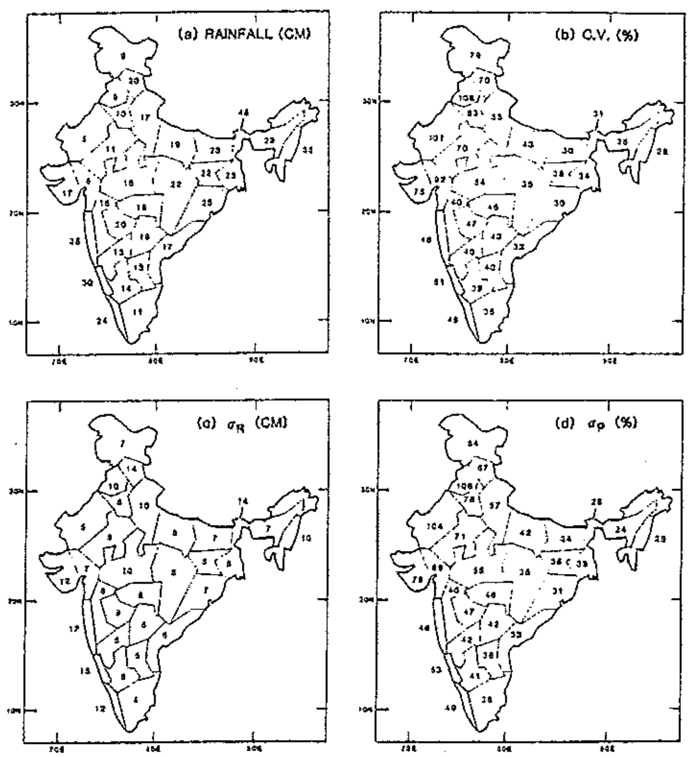


Figure 14.10. Same as Figure 14.6, except for September only.

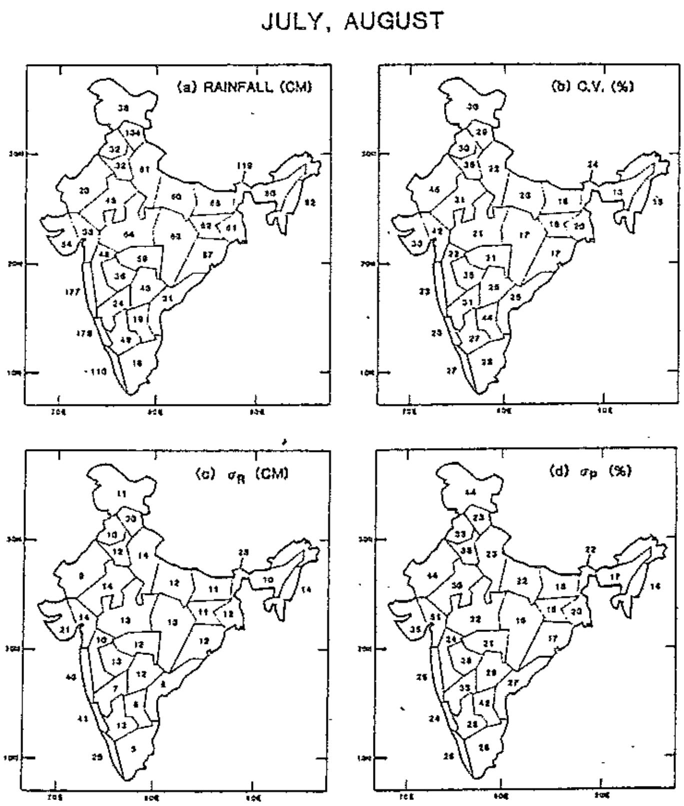


Figure 14.11. Same as Figure 14.6, except for average of July and August.

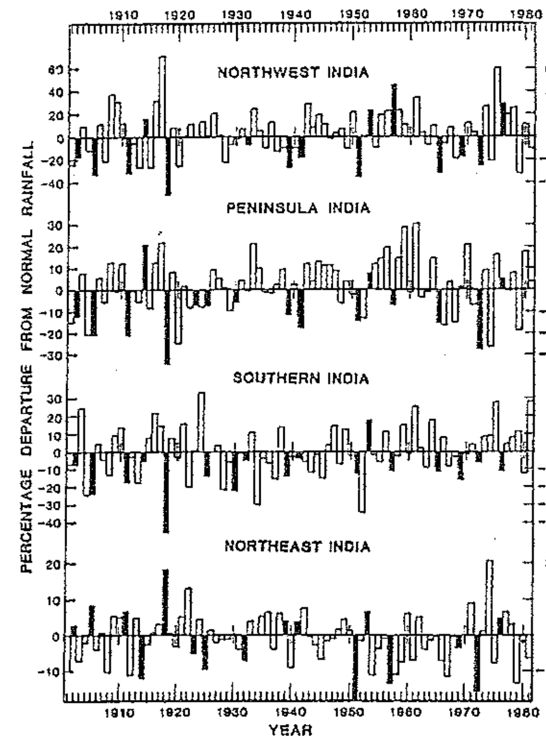
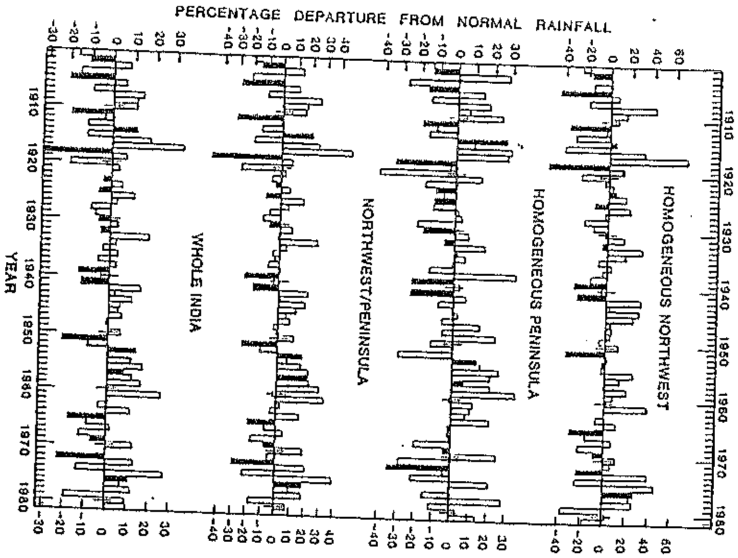


Figure 14.12. Normalized seasonal rainfall anomalies for Northwest, Peninsula, Southern, and Northeast regions of India (in cm). Solid bars denote the El Niño years (11).



...re 14.13. Same as Figure 14.12 for Homogeneous Northwest, Homogeneous Peninsula, Northwest–...sula, and Whole India (11).

TABLE 14.7 Years with Normalized Rainfall Anomalies Greater Than or Equal To 1.0 for the Eight Regions

Region																	
1 Northwest India	08[a]	09	16	17	33	42	53	56	57	58	61	73	75	76	78		
2 Peninsula India	14	17	33	55	56	58	59	61	64	70	75	80					
3 Southern India	03	16	21	24	53	61	64	75	81								
4 Northeast India	05	18	22	42	71	74											
5 Homogeneous Northwest	08	16	17	33	42	44	45	55	61	73	75	76	77	78			
6 Homogeneous Peninsula	03	10	14	16	17	38	49	55	56	58	59	64	70	75	78		
7 Northwest–Peninsula	08	14	15	17	33	42	44	56	57	58	59	61	70	73	75	76	78
8 Whole India	08	16	17	33	42	53	56	59	61	70	73	75					

[a]Year 1908 is written as 08, etc.

TABLE 14.8 Years with Normalized Rainfall Anomalies Less Than or Equal To −1.0 for the Eight Regions

Region																	
1 Northwest India	01[a]	05	11	13	15	18	20	39	51	65	72	79					
2 Peninsula India	01	04	05	11	18	20	41	51	65	66	68	72	74	79			
3 Southern India	04	05	11	13	18	22	28	30	34	37	45	52	69				
4 Northeast India	01	03	08	12	14	25	40	51	54	57	58	59	66	67	72	75	79
5 Homogeneous Northwest	01	05	11	13	15	18	20	39	41	51	65	68	72	74	79		
6 Homogeneous Peninsula	04	11	18	20	29	39	41	52	68	71	72	74					
7 Northwest–Peninsula	01	04	05	11	15	18	20	39	41	51	65	72	74				
8 Whole India	01	04	05	11	18	20	39	41	51	65	68	72	74	79			

[a]Year 1901 is written as 01, etc.

There is considerable intraseasonal variation in the data. The standard deviation (σ_m) of monthly mean anomalies is twice that of the standard deviation of the seasonal mean anomalies (σ_p); see columns 4 and 5 of Table 14.6. It is of interest to note also that during the heavy or deficient rain seasons, the individual months also show heavy or deficient rain. In the next section we describe the mechanisms of intraseasonal variability.

Table 14.9 gives the normalized rainfall anomalies for the summer season and for the individual months of June, July, August, and September for the years of heavy rain and deficient rain for four regions. For most of the years, at least three of the four months have anomalies of the same sign. This indicates that, in spite of the large intraseasonal variability in general, the particular seasons of heavy and deficient rain have significant temporal and spatial coherence (Fig. 14.4). The temporal coherence is higher for deficient years compared to heavy rainfall years.

2 INTRASEASONAL VARIABILITY OF THE INDIAN SUMMER MONSOON

In spite of the highly periodic nature of the planetary-scale monsoon circulation, there are large variations in the circulation and rainfall within the monsoon season. The intensity of the seasonal mean monsoon is influenced by the nature of variability within a monsoon season. Dhar et al. (12) have examined the possible association between the monsoon rainfall over three west coast subdivisions of India (Kerala, Coastal Karnataka, and Konkan) and dates of onset of the monsoon over the respective regions. Despite the fact that the dates of onset can fluctuate by more than 30 days (see Table 14.10), they found the interesting but counterintuitive result that the rainfall anomaly for the month of June as well as for the whole monsoon season is not related to the date of onset. This also indicates that the intraseasonal variability of monsoon rainfall is quite large. This result may, at least in part, be due to the arbitrariness in defining the dates of onset. A season of highly deficient monsoon rainfall and severe drought does not imply an absence of rainfall for the whole season; rather, it can occur due to prolonged periods of reduced rainfall. These periods are usually referred to as the break monsoon conditions by the India Meteorological Department.

2.1 Break Monsoons

Ramamurthy (13) has studied the climatology of break monsoon conditions over India. The phrase "break in the rains" can be traced back as far as the Indian Daily Weather Reports of 1888, and even now it refers only to the situations with reduced or no rainfall. The systematic study of break conditions has shown that although a reduction in rainfall occurs over most of the Indian subcontinent, it is accompanied by increased rainfall over extreme northern India (near the Himalayan foothills) and extreme southern India. This suggests that the break monsoon is actually a spatial redistribution of the monsoon rainfall. During break monsoon conditions, the monsoon trough shifts to its extreme northern position, the surface pressure

TABLE 14.9 Normalized Monthly and Seasonal Anomalies for Heavy and Deficient Monsoon Years

	Northwest (Heavy Rain)						Northwest (Deficient Rain)				
Year	Season	June	July	Aug	Sept	Year	Season	June	July	Aug	Sept
1908	1.7	−0.7	1.6	2.4	−0.4	1901	−1.1	−1.2	−0.7	−0.1	−0.9
1909	1.4	1.1	1.0	−0.7	1.1	1905	−1.5	−1.2	−1.2	−1.7	0.7
1916	1.4	1.2	0.2	1.6	0.5	1911	−1.4	0.3	−2.6	−1.3	0.9
1917	3.2	2.2	0.6	1.4	3.5	1913	−1.2	1.5	−1.0	−1.2	−1.1
1933	1.1	1.7	−0.5	1.1	0.6	1915	−1.2	−0.4	−1.5	−0.6	−0.2
1942	1.3	−0.2	1.2	0.9	0.7	1918	−2.3	−0.3	−2.4	−1.0	−1.3
1953	1.0	0.6	1.1	0.8	−0.2	1920	−1.2	0.6	0.4	−1.7	−1.2
1956	1.0	0.2	2.2	0.3	−0.6	1939	−1.2	0.2	−1.0	−1.4	−0.2
1957	2.0	−0.1	1.6	2.0	0.6	1951	−1.6	−0.5	−1.7	−0.6	−0.9
1958	1.0	−0.9	0.6	0.1	2.2	1965	−1.4	−1.3	−0.1	−1.3	−0.9
1961	1.5	0.7	0.4	1.0	1.8	1972	−1.1	−0.5	−1.2	−0.3	−0.6
1973	1.2	0.1	−0.1	3.0	−0.2	1979	−1.5	−0.2	−0.7	−1.1	−1.1
1975	2.7	1.2	2.4	1.8	1.1						
1976	1.3	0.4	−0.1	2.2	0.4						
1978	1.1	1.3	1.7	0.3	−0.3						

	Peninsula (Heavy Rain)						Peninsula (Deficient Rain)				
Year	Season	June	July	Aug	Sept	Year	Season	June	July	Aug	Sept
1914	1.5	1.0	1.0	0.1	1.4	1901	−1.1	−1.1	−0.8	0.6	−1.3
1917	1.6	1.2	−0.6	1.2	1.7	1904	−1.5	−0.0	−1.3	−1.5	−0.4
1933	1.6	0.9	−0.2	1.3	1.5	1905	−1.5	−1.5	−0.7	−0.9	−0.1
1955	1.1	0.6	−1.5	1.8	1.8	1911	−1.5	0.5	−2.2	−0.5	−0.6
1956	1.5	0.5	1.8	0.5	0.2	1918	−2.5	0.1	−2.7	−0.9	−1.6

(Table continues on p. 440.)

TABLE 14.9. *(Continued)*

	Peninsula (Heavy Rain)						Peninsula (Deficient Rain)				
Year	Season	June	July	Aug	Sept	Year	Season	June	July	Aug	Sept
1958	1.1	−1.0	1.3	1.1	1.3	1920	−1.8	−0.1	−0.6	−1.5	−1.4
1959	2.1	0.0	2.0	0.6	1.6	1941	−1.3	−0.7	−0.5	−0.8	−1.0
1961	2.3	0.2	1.8	0.4	2.7	1951	−1.1	−0.5	0.0	−0.4	−1.3
1964	1.1	0.1	−0.0	1.4	0.7	1965	−1.1	−1.3	0.4	−0.7	−0.7
1970	1.5	1.4	−1.1	2.5	0.9	1966	−1.2	−0.4	−0.1	−1.5	−0.4
1975	1.2	0.8	−0.4	1.0	1.2	1968	−1.1	−1.0	−0.3	−0.7	−0.4
1980	1.3	2.9	−0.8	0.7	−1.1	1972	−2.0	−0.6	−1.9	−0.6	−1.2
						1974	−1.9	−1.0	−1.7	−0.3	−1.2
						1979	−1.4	−0.7	0.9	−1.3	−1.1

	Whole India (Heavy Rain)						Whole India (Deficient Rain)				
Year	Season	June	July	Aug	Sept	Year	Season	June	July	Aug	Sept
1908	1.1	−0.8	1.1	1.6	−0.2	1901	−1.3	−1.3	−1.1	0.2	−1.1
1916	1.4	1.2	0.3	1.4	0.7	1904	−1.2	−0.3	−0.5	−1.2	−0.9
1917	2.7	1.9	−0.2	1.4	3.1	1905	−1.6	−1.6	−1.2	−1.1	0.3
1933	1.5	1.3	−0.5	1.7	1.0	1911	−1.6	0.6	−2.9	−1.2	0.3
1942	1.2	−0.1	1.2	1.0	0.5	1918	−2.6	0.0	−3.3	−0.9	−1.7
1953	1.1	0.3	1.0	1.3	−0.1	1920	−1.6	0.2	−0.2	−1.9	−1.3
1956	1.3	0.7	2.1	0.3	−0.2	1939	−1.1	−0.3	−1.0	−0.5	−0.6
1959	1.3	−0.4	1.5	0.1	1.5	1941	−1.1	−0.1	−1.3	−0.7	−0.5
1961	2.0	0.6	1.4	0.8	2.3	1951	−1.7	−0.6	−1.1	−0.8	−1.3

1970	1.0	1.3	−1.4	1.7	0.9	1965	−1.4	−1.4	−0.0	−1.0	−0.9
1973	1.0	−0.2	−0.1	2.4	0.2	1968	−1.0	−0.8	0.3	−1.0	−1.1
1975	2.2	0.8	1.8	1.4	1.4	1972	−1.8	−0.6	−2.1	−0.9	−0.8
						1974	−1.1	−0.4	−0.3	−0.8	−1.0
						1979	−1.5	−0.5	0.1	−1.6	−1.0

	Northwest–Peninsula (Heavy Rain)						Northwest–Peninsula (Deficient Rain)				
Year	Season	June	July	Aug	Sept	Year	Season	June	July	Aug	Sept
1908	1.4	−0.8	1.4	1.7	−0.1	1901	−1.2	−1.1	−0.8	0.1	−1.2
1914	1.2	1.0	1.6	−0.3	1.0	1904	−1.2	−0.6	−0.7	−1.2	−0.6
1916	1.4	0.9	0.4	1.5	0.7	1905	−1.6	−1.3	−1.3	−1.2	0.1
1917	2.7	2.0	−0.1	1.6	3.1	1911	−1.8	0.3	−2.4	−1.3	−0.0
1933	1.4	1.5	−0.4	1.5	1.1	1915	−1.1	0.1	−1.5	−0.9	−0.3
1942	1.1	0.1	1.3	1.0	0.1	1918	−2.7	−0.6	−2.9	−1.1	−1.5
1944	1.0	−0.4	1.6	1.2	−0.8	1920	−1.5	0.6	−0.6	−1.7	−1.3
1956	1.2	0.3	2.6	0.4	−0.4	1939	−1.3	−0.5	−1.2	−0.4	−0.6
1957	1.2	0.2	0.7	1.7	−0.2	1941	−1.0	−0.3	−1.1	−0.6	−0.4
1958	1.2	−0.9	1.1	0.9	1.9	1951	−1.4	−0.5	−0.7	−0.7	−1.1
1959	1.6	−0.3	1.9	0.3	1.5	1965	−1.1	−1.3	0.4	−0.8	−1.0
1961	1.8	0.6	1.5	0.6	2.0	1972	−1.6	−0.3	−1.8	−0.7	−1.0
1970	1.1	1.3	−1.3	2.0	0.8	1974	−1.3	−0.1	−0.8	−0.9	−1.0
1973	1.1	−0.1	0.2	2.5	−0.1						
1975	2.2	1.0	1.4	1.4	1.4						
1976	1.0	0.6	0.4	1.9	−0.3						
1978	1.0	1.5	1.1	0.9	−0.6						

TABLE 14.10 Dates of Onset of the Monsoon over Kerala, the Southern Tip of India (Range: May 11–June 18)

Year	Date	Year	Date	Year	Date	Year	Date
1901	June 7	1921	June 2	1941	May 23	1961	May 18
1902	June 6	1922	May 31	1942	June 10	1962	May 17
1903	June 12	1923	June 11	1943	May 29	1963	May 31
1904	June 7	1924	June 2	1944	June 3	1964	June 6
1905	June 10	1925	May 27	1945	June 5	1965	May 26
1906	June 13	1926	June 6	1946	May 29	1966	June 1
1907	June 8	1927	May 27	1947	June 3	1967	June 9
1908	June 11	1928	June 3	1948	June 11	1968	June 8
1909	June 2	1929	May 29	1949	May 23	1969	May 17
1910	June 2	1930	June 8	1950	May 27	1970	May 26
1911	June 6	1931	June 4	1951	May 31	1971	May 27
1912	June 8	1932	June 2	1952	May 20	1972	June 18
1913	June 2	1933	May 22	1953	June 7	1973	June 4
1914	June 4	1934	June 8	1954	May 31	1974	May 26
1915	June 15	1935	June 12	1955	May 29	1975	May 30
1916	June 2	1936	May 19	1956	May 21	1976	May 31
1917	May 31	1937	June 4	1957	June 1	1977	May 30
1918	May 11	1938	May 26	1958	June 14	1978	May 29
1919	June 3	1939	June 5	1959	May 31		
1920	June 3	1940	June 14	1960	May 14		

departure from normal is positive over most of the country and maximum over central India, and negative over extreme northern and southern parts of the country. The easterly jet at the upper levels is stronger and shifts northward. The meridional component of the wind over the northern parts of India is from the north in the middle troposphere and from the south in the upper troposphere, just the opposite of the meridional wind directions during active monsoon conditions.

The duration of break conditions ranges from 3 to 21 days. During the 80 year (1888–1969) period examined by Ramamurthy, there were 56 cases of break monsoon conditions in July of which 38 lasted for 3–5 days, 14 for 6–10 days and 4 for 11–20 days; of 57 cases in August, 31 lasted for 3–5 days, 19 for 6–10 days and 7 for 11–20 days. The total number of the break days were 306 for July and 380 for August. Although the number of break periods are about the same in July and August, they tend to last longer in August.

The observed rainfall distribution and the circulation patterns support the idea that break monsoon conditions are associated with a northward shift of the normal meridional circulation and the occurrence of enhanced convective activity near the southern tip of India. Most of India is therefore under the descending branches of the two thermally forced meridional circulations with ascending branches near the foothills of the Himalayas and the extreme southern tip of India. The northward shift and the increase in the speed of the easterly jet stream during the monsoon breaks are consistent with the possibility that an enhanced meridional circulation accelerates the zonal flow by deflection of the winds by the Coriolis force.

2.2 Factors Causing Intraseasonal Variability

It is possible to identify certain phenomenological factors that produce variability within a monsoon season on time scales of a few days to a few weeks. They fall into four broad categories: synoptic-scale disturbances, monsoon troughs, quasi-periodic oscillations, and mid-latitude effects.

2.2.1 Synoptic-Scale Disturbances (Lows, Depressions, Storms). An examination of daily weather charts and daily rainfall amounts over India suggests that the rainfall distribution on any given day is generally related to the presence of synoptic-scale disturbances, quasi-symmetric zones of convergence, and the interaction of strong monsoon flow with orographic barriers. In particular, rainfall depends upon the frequency, intensity, life cycle, and propagation characteristics of the synoptic disturbances that influence a particular region. Dhar and Rakhecha (14) found that in the absence of tropical disturbances affecting the Indian subcontinent, the July and August rainfall over the northern Indian plains was reduced by 19 and 14%, respectively. This study did not include the effects of low-pressure areas which can also produce large amounts of rainfall. Sikka (15) examined changes in the monsoon rainfall for five years of high rainfall and deficient rainfall. He also found that the number of monsoon depressions as well as the number of depression days were quite similar for both heavy and deficient rain years. The most striking difference was found in the number of low-pressure areas and the number of days with low-pressure areas. The ratio of the number of lows for heavy and deficient rain years was 1.6. Based on these results, Sikka concluded that the higher number of monsoon lows is a manifestation of greater instability of the monsoon trough.

2.2.2 The Monsoon Trough. As mentioned earlier, the northward shift of the monsoon trough is accompanied by break monsoon conditions over most of the central Indian regions, and enhanced rainfall near the southern tip of India. A weakening of the monsoon trough over North India is associated with the strengthening of the convergence zone near the southern tip of India. We are not aware of any physical explanation for this behavior of the monsoon trough. The northward shift of the monsoon trough could be related to changes in the large-scale circulation in middle latitudes, and to changes in the intensity of the near-equatorial trough. The latter could be due either to air–sea and/or interhemispheric interaction over the Indian Ocean or the formation of tropical disturbances. A partial explanation for the observed fluctuations of the monsoon trough and the equatorial convergence zone is provided by the recently documented northward propagation of cloudiness described in the following text.

2.2.3 Quasi-Periodic Oscillations. Yasunari (16), Sikka and Gadgil (17), and Krishnamurti and Subrahmanyam (18) have presented observational evidence for a possible northward propagation of the convergence zone and cloudiness. Fluctuations in the intensity of the monsoon trough could be interpreted in terms of the phase of this northward propagating convergence zone. There is a large body of observational evidence for 15-day oscillations during the monsoon season [for references see the paper by Krishnamurti and Ardanuy (19)]. From spectral analysis of digital cloud data, Yasunari (16) showed dominant periodicities near 15- and 40-day periods. Both oscillations showed a tendency for northward phase propagation. From the analyses by Yasunari (20, 21) and Krishnamurti and Bhalme (22), there appears to be a significant relationship between the phase of 15-day oscillations and the formation of monsoon depressions. However, this relationship by itself does not clarify whether these oscillations are responsible for the formation of monsoon depressions by creating a favorable large-scale environment, or whether the oscillations are merely a consequence of a regular formation of depressions at 15-day intervals. The 15-day oscillations may also be due to instabilities produced by interaction of the zonal flow with mountains and diabatic heat sources (23). The 40-day oscillation is of much larger scale and not necessarily unique to the monsoon flow (24). Physical mechanisms responsible for these fluctuations are not yet understood, and more observational studies are needed to describe their structure and origin. Webster and Chou (25) conducted numerical experiments with a simple model to study low frequency transitions of a monsoon system and found that the model fluctuations are very sensitive to the parameterizations of land surface processes and treatment of soil moisture. Goswami and Shukla (26) showed that a zonally symmetric general circulation model of the atmosphere exhibited quasi-periodic oscillations with periods of 15–40 days due to the interaction of the motion field and moist convection. If this interaction was eliminated by prescribing the diabatic heating, the oscillations disappeared. Webster (see Chapter 11) has shown that the feedback mechanisms between soil moisture changes and atmospheric circulation play an important role in northward propagation of the monsoon trough.

2.2.4 Mid-Latitude Effects. A large number of observational studies indicate a possible relationship between the mid-latitude circulations of both hemispheres and the summer monsoon circulation and rainfall (see Chapter 11 for an extended discussion). Ramaswamy (27) has suggested that intrusions of large-amplitude troughs from the mid-latitudes of the Northern Hemisphere are associated with break monsoon conditions over India. The preponderance of westerly winds over northern India is favorable for the propagation of mid-latitude influences to the monsoon region. Due to lack of upper air data over the southern Indian Ocean, it has not been possible to examine the upper air circulation over the Southern Hemisphere. Future studies with the special 1979 Global Weather Experiment data and the global analyses produced from these data will be required to gain a better understanding of such relationships.

3 MECHANISMS OF INTERANNUAL VARIABILITY OF MONSOONS

As mentioned earlier, a convenient framework for understanding and describing the mechanisms of interannual variability is to isolate the factors associated with

the atmosphere's internal dynamics and with its lower boundary conditions. It is important to understand the relative contributions of these two factors to the observed interannual variability of the Indian monsoon. This separation is only an idealization of the real atmosphere where the internal dynamics and boundary conditions continuously interact. The recent research has been summarized under these two categories.

3.1 Internal Dynamics

Even if the external forcing by solar radiation and boundary conditions at the earth's surface were constant in time, the atmospheric circulation would exhibit interannual variability due to the inherent aperiodic nature of the system. The combined effects of dynamical instabilities (manifested as synoptic-scale disturbances—see Section 2.2.1), nonlinear interactions among various scales of motion, thermal and orographic forcing, tropical–extratropical interactions, and so on, can be considered as examples of internal dynamical processes that can produce interannual variability. The orography and the land–sea distribution at the earth's surface are fixed with time; however, their interactions with fluctuating winds can produce large changes and, therefore, they are to be considered as part of the variability associated with internal dynamics. It should be noted that if all the observed interannual variability were due to internal dynamical processes alone, the prospects for long-range forecasting would be rather limited because of the inherent limits of predictability of internal dynamics (1). A number of observational studies suggest possible relationships between the monsoon circulation and other features of the global circulation.

Bannerjee et al. (28) found a significant correlation between the average position of the ridge at 500 mb over India during April and summer monsoon rainfall. If the ridge position is to the south (north) of its climatological position, the summer monsoon rainfall is deficient (excessive). This result is substantiated by Rao (29). Parthasarathy and Mooley (6) concluded from a time series of Indian summer monsoon rainfall that the rainfall is random and normally distributed. This does not necessarily imply that it is unpredictable; realizations of a nonlinear deterministic system can also have certain statistical properties which are similar to that of a random process. Verma (30) examined the monthly mean anomalies of 300–100 mb thickness, which is a measure of the upper-tropospheric temperature anomaly, for 10 years (1968–1977) for selected stations over India. He found that anomalies in April and May tend to persist for the whole monsoon season, and that negative (positive) thickness anomalies in the pre-monsoon months are associated with negative (positive) anomalies of Indian summer monsoon rainfall. It is difficult to explain this long persistence as an internal dynamical process of the atmosphere; it seems likely that it is related to some slowly varying boundary forcing at the earth's surface (for example, snow over Eurasia or SST anomalies over the tropical oceans). Joseph et al. (31) found highly significant correlation between monthly mean meridional wind during May at selected Indian stations and summer monsoon rainfall over India for 15 years (1964–1978) of data. However, the correlation coefficients drop abruptly for the meridional winds in both April and June.

Tanaka (32) examined the monthly mean rainfall and height fields for June, July, August and September over the Asian monsoon region, and the wind speed of the tropical easterly jet above 10°N at 150 mb for 17 years (1964–1980). He found that a strong tropical easterly jet at 10°N is associated with a strong low-level monsoon circulation and heavy monsoon rain over India and predominantly zonal circulation near 50°N. In contrast, a weaker tropical easterly jet at 10°N is associated with deficient monsoon rainfall over India and a blocking high to the north of the Caspian Sea. Based on these observations, Tanaka concluded that the interannual fluctuations of the summer monsoon are strongly influenced by the middle latitude circulation of the Northern Hemisphere. This conclusion seems tenuous; the causal mechanism is not clear. He has not considered the possibility that the changes in diabatic forcing due to changes in the rainfall could have produced anomalous tropical and middle latitude circulations. Moreover, in examining the wind speed only at 10°N, Tanaka may have missed the jet core since the latitudinal position of the jet maximum varies from one year to the next. In fact, Ramamurthy (13) has shown that during the break-monsoon situations, when most of India experiences deficient rainfall, the tropical easterly jet shifts northward and is stronger than normal. Raman and Rao (33) suggested that blocking ridges over East Asia are also associated with prolonged breaks in monsoon rainfall.

3.2 Influence of Global Surface Boundary Conditions

There is a growing body of modeling and observational evidence that suggests that the slowly varying boundary conditions of sea surface temperature (SST), soil moisture, and sea ice and snow at the earth's surface can influence the interannual variability of the atmospheric circulation.

The variations in these surface boundary conditions can influence the location and intensity of diabatic heat sources that drive the atmospheric circulation. Anomalous boundary conditions can be more effective in producing circulation anomalies in the tropics than in the mid-latitudes because the tropical circulation is dominated by the planetary-scale Hadley, Walker, and monsoon circulations, and changes in the boundary conditions can alter the locations and intensity of these systems. They can also influence the amplitude and phase of planetary waves in mid-latitudes which, in turn, can influence the tracks and intensity of cyclone-scale disturbances. The physical mechanisms responsible for such influence are rather complex (see Chapter 16) and depend upon the nature of the boundary forcing. Charney and Shukla (2) suggested that the Asiatic monsoon is a dynamically stable circulation system, and, its interannual variability is largely determined by the slowly varying boundary conditions; therefore, monsoons are potentially more predictable than the mid-latitude circulations.

The following is a summary of selected observational and general circulation model studies of the relationships between these boundary conditions and monsoon circulation and rainfall.

3.2.1 *Snow Cover.*

As shown by Wiesnet and Matson (34), December snow cover for the Northern Hemisphere is a very good predictor of the snow cover for the

following January through March. Snow cover, therefore, is considered to be a slowly varying boundary condition useful for prediction.

Blanford (35) found that excessive winter and spring snowfall in the Himalayas was an indicator of the subsequent monsoon rainfall in India. The amount and time of occurrence of the cold weather (October–May) snowfall in the mountain districts adjacent to northern India was one of the important factors used by Blanford for monsoon rain forecasts which he started issuing in 1882. For the period 1880–1920, greater winter snowfall was found to be related to deficient monsoon rainfall, but for the subsequent 30-year period, reported snow accumulation showed very large variability and the relationship with the monsoon rainfall was opposite to what it was in the earlier four decades. After 1950 the India Meteorological Department dropped this factor as one of the predictors of monsoon rain.

Hahn and Shukla (36) examined 11 years of satellite-derived snow cover over Eurasia and summer monsoon rainfall over India, and found (Fig. 14.14) an apparent inverse relationship between the area extent of winter snow cover and Indian monsoon rainfall. This result, although based on a rather limited sample, supported the earlier

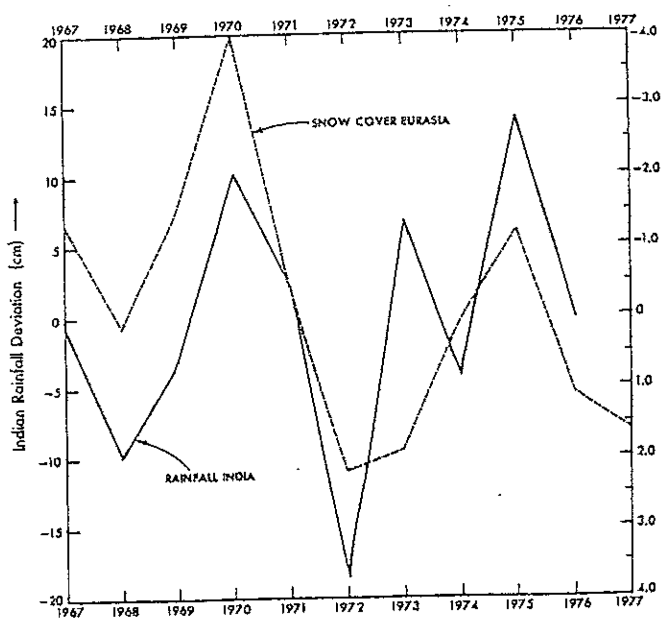


Figure 14.14. Area weighted summer monsoon rainfall anomaly over India (solid line) and winter snow cover departure over Eurasia (dashed line). Note the inverted scale for the latter.

findings of Blanford. Dickson (37) has extended this study to include data up to 1980 and found that the relationship still holds for the extended data period; however, the magnitude of the correlation between the summer monsoon rainfall and winter snow cover is less for the period 1967–1980, compared to 1967–1975. Dickson also found that the snow cover data for the period 1966–1974 did not include information on Himalayan snow cover and, in particular, the data for 1969 were considered to have a large error. He showed that if the data for 1969 are not included and the data for 1967–1974 are adjusted for bias, the correlation coefficient between snow cover and rainfall for the period 1967–1980 changes from −0.44 to −0.59, and for the period 1967–1975, it changes from −0.62 to −0.74. The statistical significance and practical utility of this relationship for predicting summer monsoon rainfall can be examined more systematically when a large sample size is available. In light of these recent results, we speculate that a lack of systematic relationship between Himalayan snowfall and monsoon rainfall during the period 1920–1950 (35) could have been, at least in part, due to the changes in the quality of reporting of the snow cover and snow depth over the Himalayas.

Dey and Bhanukumar (38) have examined the relationship between spring snow cover over Eurasia and the time taken by the Indian summer monsoon rainfall to advance from the southern tip of India to the northern border of India. They found that, for the period 1967–1978, when snow cover during the spring was greater than normal, the time taken by the monsoon to advance from south to north was also greater than normal, and vice versa. They also found a negative correlation between the amount of snow melt during spring and the advance period for the summer monsoon. When the difference of snow cover from March to May (snow melt) was above normal, the speed of the advance of the monsoon was slower than normal and vice versa. These results are not inconsistent with the earlier results of Hahn and Shukla because high snow melt is indicative of enhanced solar heating or shallow snow depth.

An inverse relationship between Eurasian snow cover and the summer monsoon is not implausible because large and persistent winter snow cover over Eurasia can delay and weaken the spring and summer heating of the land masses that is necessary for the establishment of the large-scale monsoon flow. During the spring and summer seasons following winters with excessive snow, most of the solar energy is used for melting the snow or evaporating from the wet soil. Systematic numerical experiments using global general circulation models (GCMs) with adequate treatment of the effects of albedo and ground hydrology can be used to understand the physical mechanisms that influence the atmospheric circulation due to excessive snowfall (39). Analysis of data has also shown (40) that for large (small) snow cover over the Tibetan Plateau, the arrival of the summer monsoon circulation over the plateau is late (early).

3.2.2 *Sea Surface Temperature.* Since 70% of the earth's surface is covered with water, and since changes in the sea surface temperature (SST) are much slower compared to the atmospheric fluctuations, it is natural to think that interannual variability of SST might contribute to the interannual variability of the atmospheric

circulation and rainfall. During the last 30 years there have been numerous studies suggesting possible relationships between SST anomalies and atmospheric anomalies (see e.g., 41 and 42).

Arabian Sea Surface Temperature Anomalies. During the last 20 years, there have been several observational and GCM sensitivity studies that have suggested a possible relationship between anomalies of SST over the Arabian Sea and summer monsoon rainfall over India. These studies have produced, at times, conflicting results, and they have used different data sets and different analysis schemes. The first observational study on this topic was by Ellis (43) who showed that SST over the Arabian Sea was warm during 1920 when several parts of India experienced floods, and was cold during 1923 when several parts of India experienced droughts. In this study, data were examined for only these two years and therefore the results are of limited statistical significance. The characterization of 1920 as a flood year and 1923 as a drought year is also questionable. The concept, however, was physically appealing: warm SST can produce high evaporation and possibly larger rainfall.

In another observational study, Pisharoty (44) calculated the flux of water vapor across the equator into the Arabian Sea, and across the west coast of India from the Arabian Sea, and found that the latter was more than twice the former, and concluded that evaporation over the Arabian Sea is an important source of moisture for precipitation over India. Saha (45) and Saha and Bavadekar (46) repeated the calculations of Pisharoty but used additional stations in the western Arabian Sea where the northward flow is the strongest. They concluded that, contrary to the results of Pisharoty, the flux of water vapor across the equator is about 30% greater than evaporation over the Arabian Sea. They pointed out that owing to scarcity of data over the Arabian Sea, Pisharoty may have underestimated the cross-equatorial moisture flux and, consequently, overemphasized the role of evaporation over the Arabian Sea. Ghosh et al. (47) used observations taken during a monsoon field program called Monsoon-77 to calculate the water vapor budget over the Arabian Sea. They, like Pisharoty, found that the moisture flux across the west coast of India was more than twice the moisture flux across the equator from the Southern Hemisphere, thus supporting the idea that evaporation over the Arabian Sea is a significant moisture source for monsoon rainfall over India. A recent calculation by Cadet and Reverdin (48) has shown that most of the water vapor (about 70%) crossing the west coast of India comes from the Southern Hemisphere. This is in agreement with Saha. In summary, there is a large discrepancy among the calculations of different investigators. Part of this discrepancy can be attributed to the interannual variability of SST, evaporation, and wind speed over the Arabian Sea, but perhaps a much larger part is due to differences in the quality and density of data and the techniques used for analyzing the data.

Motivated by the earlier observational studies, Shukla (49) used a GCM to investigate the sensitivity of the model simulated monsoon rainfall over India to SST anomalies over the Arabian Sea. He ran the model for a period representing 60 days with and without prescribed negative (cold) SST anomalies (which were −3°C near the African coast and −1°C in the central Arabian Sea). He found that the monsoon rainfall over India was significantly reduced. Sikka and Raghavan (50) pointed out that the drastic reduction of rainfall over India could be due to limitations of the model parameterizations or to the choice of a large verification area which included the oceanic regions of the negative SST anomaly. Shukla (51) recomputed the rainfall over the Indian subcontinent for the two model runs and found that the earlier results did not change when the verification domain was changed. A similar GCM sensitivity study was carried out by Washington et al. (52) who also found a significant reduction of rainfall but only over the region of colder SST. Rainfall over India and adjacent regions also decreased but the decrease was not statistically significant. Washington et al. also studied the sensitivity of a warm SST anomaly over the eastern Arabian Sea and the central (equatorial) Indian Ocean. Only in the latter did they find any evidence of remote reponse. In order to resolve the conflicting results of the two GCMs, Shukla (41) repeated the same numerical experiment with a third climate model and found that the results were very similar to his earlier results. He pointed out that the primary reason for the differences in the results was the ability of each model to simulate the mean monsoon circulation in the control experiments. In the July mean simulation of the low-level monsoon flow by the model used by Washington et al., the southwesterly monsoon current did not reach the western Ghats of India.

Considering the large differences in the basic physical parameterizations of different GCMs, it is rather remarkable that the model simulated response of monsoon rainfall to SST anomalies over the Arabian Sea is so similar. It appears reasonable to conclude that any model with a good parameterization of moist convection and the boundary layer would produce a somewhat similar response. However, recent analyses of the observed SST anomalies over the Arabian Sea have suggested that the magnitudes of the SST anomalies used in these GCM experiments were too high to be realistic. Analyses of data along the ship tracks suggest that rarely does such a large anomaly occupy so broad an area in the Arabian Sea. In addition, SST anomalies change sign from the early part of the summer monsoon season to the later part. Thus it is not reasonable to assume that the anomalies persist for the entire monsoon season as was done in the modeling experiments. Following the earlier work by Ellis (43), observational studies of possible relationships between SST over the Arabian Sea and Indian monsoon rainfall have been carried out by several investigators. Shukla and Misra (53) calculated the correlation coefficients between SST anomalies along a ship track at about 10°N between 60°E and 70°E using the data from Fieux and Stommel (54) and seasonal mean rainfall over Indian subdivisions. They found weak positive correlations for subdivisions along the west coast of India. Raghavan et al. (55) showed that during the 1964 monsoon season, the western Arabian Sea was 2–3°C colder during a short period of reduced rainfall over India compared to another short period of enhanced rainfall. Weare (56) calculated the empirical orthogonal functions (EOF) for seasonal mean rainfall of 53 stations over India, and sea surface temperature anomalies on a 5° × 5° grid over the Arabian Sea and Indian Ocean for six months (May–October). Correlation coefficients between the EOFs of monsoon rainfall and SST were of different sign. A recent analysis of SST anomalies along the ship tracks over the Arabian Sea

suggests that the results of Shukla and Misra and Weare should be interpreted with caution. In Shukla and Misra's study the SST anomaly data have a significant negative bias for the period before 1940 and a positive bias after that (as will be discussed further on in this section). This bias was not removed before the calculation of correlation coefficients. The problem with Weare's calculations is that he grouped together the SST anomalies of pre-monsoon months and post-monsoon months. SST anomalies during April, May, and June are generally of opposite sign to those of August, September, and October and it is not meaningful to combine SST anomalies for all the six months together.

Analysis of Observed Sea Surface Temperature Anomalies. The monthly mean SST data for 75 years along the ship track shown in Figure 14.15 has been analyzed. The climatological monthly mean and the standard deviation of the SST were first calculated for 2° × 2° regions along the ship tracks. Anomalies with absolute value greater than four times the standard deviation or 3°C, whichever was larger, were not considered. (Such cases were rare and the discarded values were generally about 10°C or more). Spatial averages over the regions R1, R2, R3, R4, and R5 shown in Figure 14.15 were calculated by using appropriate weights based upon data density in each 2° × 2° box. Figure 14.16 shows the seasonal cycle which was subtracted to obtain the anomalies. Figure 14.17 shows the 10-year running mean for average SST anomalies for the regions R1, R2, R3, R4, and R5. For all the regions the anomaly was colder by about 0.25°C before the 1940s and warmer by about the same amount after the 1940s. This bias was seen for several other areas in the Indian Ocean and equatorial Pacific. Similar bias was also noted by other investigators. We have removed this bias from the observed SST anomalies for further analysis of data.

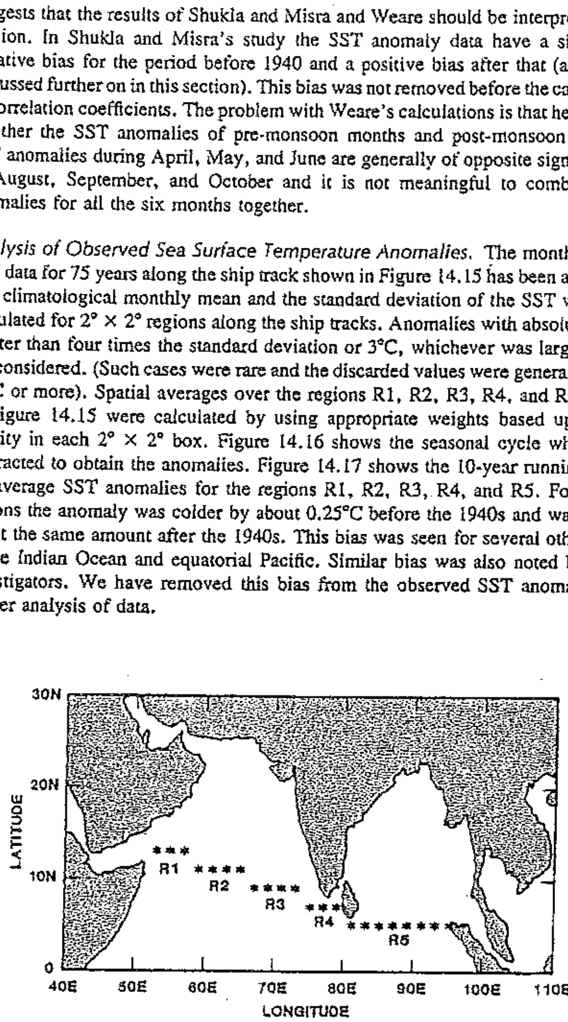


Figure 14.15. Ship track with highest sea surface temperature (SST) data density over the Arabian Sea and regions R1 through R5. The width of each region is 2° latitude.

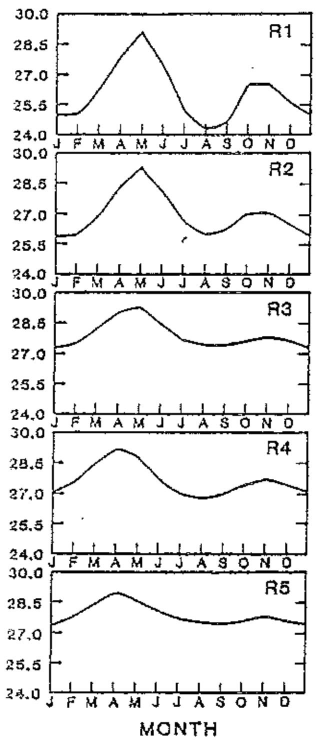


Figure 14.16. Long-term monthly mean sea surface temperature (SST) for the regions R1 through R5.

Figure 14.18 shows the time series of bimonthly [February, March (FM); April, ay (AM); June, July (JJ); August, September (AS); October, November (ON); d December, January (DJ)] SST anomalies for 75 years for region R2. Figure 19 shows the June–July bimonthly SST anomalies for the five regions R1 through . These two figures show that most of the observed SST anomalies are within °C. There is reasonable spatial and temporal consistency in the values of SST malies among the five locations.

We have averaged the SST anomalies for five heavy rainfall years (1916, 1917, 3, 1961, 1975) for which the standardized rainfall anomaly (ratio of anomaly its standard deviation) over India was 1.4, 2.7, 1.5, 2.0, 2.2, respectively, for five deficient rainfall years (1905, 1911, 1918, 1951, 1972) for which the adardized rainfall anomaly was −1.6, −1.6, −2.6, −1.7, −1.8, respectively.

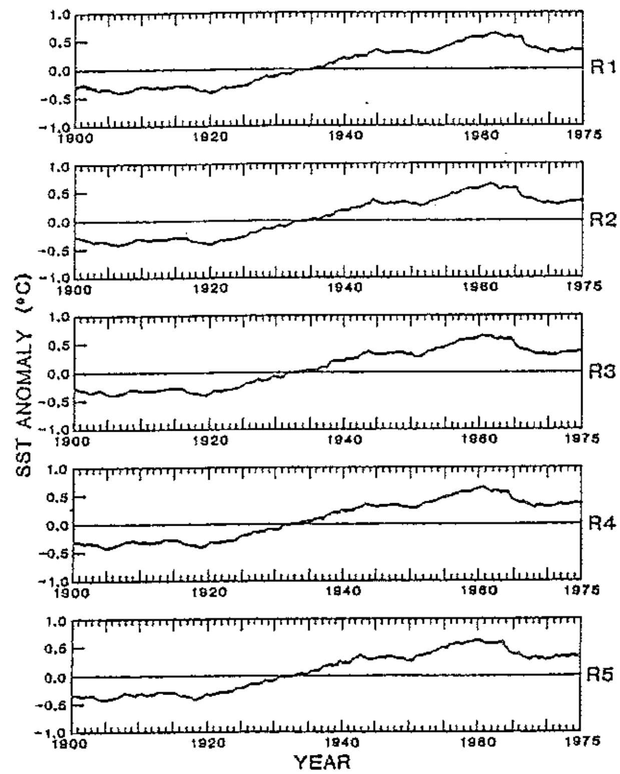


Figure 14.17. Ten-year running mean of sea surface temperature (SST) anomalies for the regions R1 through R5.

The composite time series for SST anomalies for these heavy rainfall and deficient rainfall years are shown in Figure 14.20. The SST anomaly for the heavy rain years is relatively warmer than deficient rain years during March, April, and May, and is colder during August, September, and October. This distribution is consistent with the inverse relationship between wind speed and SST anomaly earlier noted by Shukla and Misra (52) because stronger winds associated with above average monsoon rainfall tend to cool down the ocean surface in the post-monsoon months. This anomalous cooling effect is superimposed upon the annual cycle for which

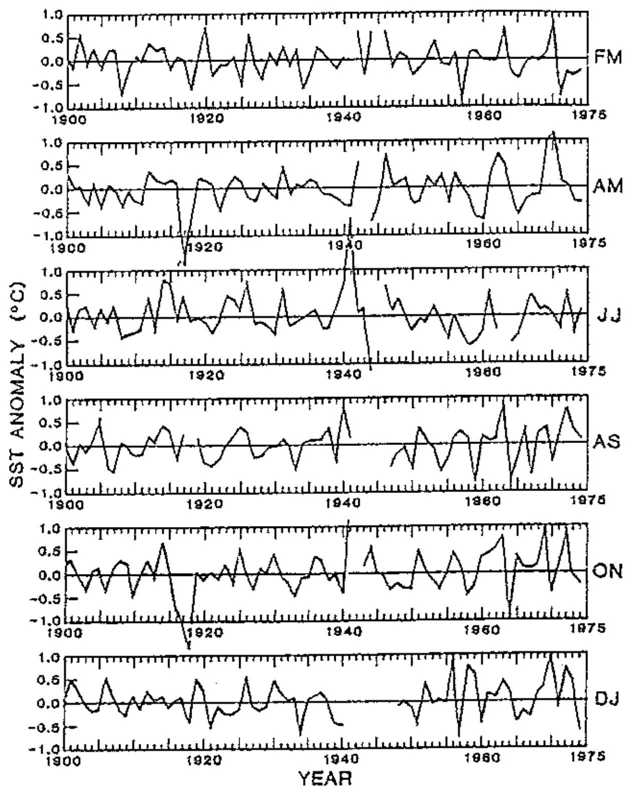


Figure 14.18. Time series of bimonthly [February, March (FM); April, May (AM); June, July (JJ); August, September (AS); October, November (ON); and December, January (DJ)] sea surface temperature (SST) anomalies for the region R2.

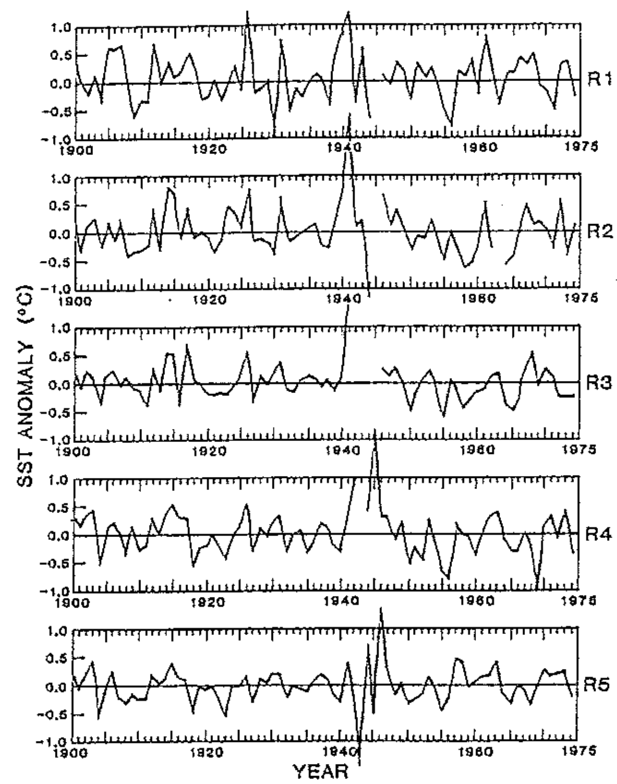
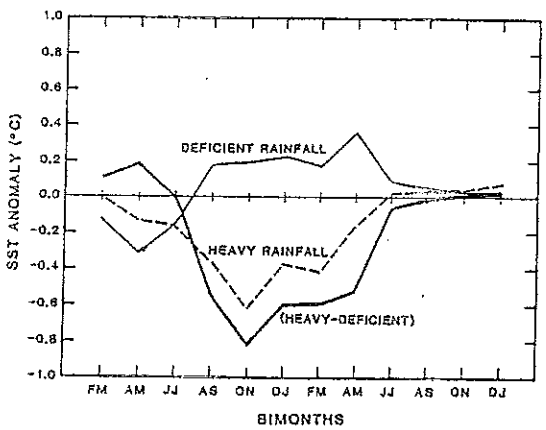


Figure 14.19. Time series of June, July (JJ) SST anomalies for the regions R1 through R5.



-igure 14.20. Composite sea surface temperature (SST) anomalies for heavy (dashed line) and deficient -olid line) monsoon rainfall. The thick solid line shows the difference (heavy-deficient) of SST anomalies.

:ooling is primarily due to advective effects (see Chapter 13, Section 2.4). The nagnitudes of the SST anomalies before the peak monsoon months are smaller than hose for the post-monsoon months and therefore an average for the whole monsoon eason is dominated by the post-monsoon months (56).

The general picture that emerges is as follows: a warm (cool) SST anomaly uring April, May, and June is not necessarily indicative of above (below) average ionsoon rainfall; however, heavy monsoon rainfall is followed by negative SST nomalies. The predictive value of this relationship is rather limited because the iagnitude of SST anomalies during the pre-monsoon months is within the range f observational error (−0.1°C to −0.5°C).

quatorial Pacific Sea Surface Temperature Anomalies. Subsequent to the pi-ieering works of Walker [(10), who discovered the "Southern Oscillation" while oking for correlations between the Indian monsoon rainfall and other atmospheric irameters over the globe] and Walker and Bliss (57), and the works of Bjerknes 58), who coined the term "Walker circulation"], several observational papers ive documented the relationship between the occurrence of warm SST anomalies the equatorial Pacific and a shift of the heavy precipitation regime from the treme western Pacific to the central Pacific near the international dateline. Aperiodic currences of warm equatorial Pacific SST anomalies, referred to as El Niño events, ve been found to be associated with below normal summer monsoon rainfall over dia. Sikka (15) showed a general association between El Niño events and deficient mmer monsoon rainfall. Angell (59) showed that SST anomalies in the equatorial

Pacific were highly correlated (correlation coefficient 0.62) with rainfall over India during the preceding summer monsoon season. Neither Sikka nor Angell recognized the predictive value of their findings because of incomplete knowledge of the life cycle of the warm SST events later described by Rasmusson and Carpenter (60). The positive (warm) El Niño SST anomalies usually appear along the Ecuador–Peru coast several months before the monsoon season, so the association between El Niño events and deficient monsoon rainfall can be a useful forecasting tool. Rasmusson and Carpenter (11) have identified 25 El Niño events during the period 1875–1979, and have shown that the area averaged summer monsoon rainfall over India was below the median value in 21 of the 25 events. The solid black bars in Figures 14.12 and 14.13 denote these El Niño years. This relationship will be discussed in Chapter 16.

Fu and Fletcher (61) have examined the large-scale thermal contrast between the soil temperature over the Tibetan Plateau (given by the mean value for five representative stations), and sea surface temperature over the eastern equatorial Pacific averaged between 5°N–10°S and 120°W–160°W, and found that the higher (lower) values of the land–ocean contrast (i.e., warmer Tibetan Plateau and colder SST) are associated with higher (lower) monsoon rainfall. The correlation between an index of the Indian summer monsoon rainfall and the land–ocean temperature contrast is higher than that with either land or ocean temperature alone.

These results pertain only to a possible relationship between the summer monsoon rainfall and equatorial Pacific SST. Owing to the lack of data, relationships to Southern Hemisphere SST have not been investigated.

3.2.3 Soil Moisture. The annual net rainfall for the global continents is estimated to be about 764 mm, and runoff to the oceans about 266 mm (62). If there were no secular trends in the annual mean global soil moisture this would suggest that the annual global mean evaporation from the land surface alone is more than 60% of the annual and global mean precipitation over the land and is a very important component of the global water budget and hydrologic cycle. It does not necessarily follow that water evaporated locally from the land is important in determining the local rainfall over the land because the total rainfall is determined by the combined effects of available and precipitable moisture and the nature and intensity of the dynamical circulation.

The role of soil moisture in determining the interannual variability of atmospheric circulations is two-fold. First, it strongly influences the rate of evaporation and therefore, the moisture supply to the atmosphere. Second, it influences the heating of the ground which affects the sensible heat flux and ground temperature. In a set of idealized numerical experiments with a GCM, Shukla and Mintz (63) examined the role of soil moisture. They found that when the land surface is dry, and no evaporation is allowed from it, a very intense surface low develops over India during the summer monsoon season. Despite the lack of evaporation from the land surface in the model experiments, monsoon rainfall is greater than when the soil is wet and evaporation over land can take place. This is because the reduction of moisture caused by the lack of evaporation is more than compensated by moisture ux convergence from the surrounding oceans caused by the intensified monsoon ow. If this result were valid for the real atmosphere, it implies that a very dry pre-onsoon season would be followed by enhanced rainfall during the monsoon season. o observational study has been carried out to verify this hypothesis.

Charney et al. (64) suggested that significant changes in precipitation over subtropical esert margin regions can occur by changes in albedo at the earth's surface. An ncrease in albedo reduces the absorption of the incoming solar radiation, and hence vaporation and cloudiness. The increase in the solar radiation reaching the ground lue to reduced cloudiness is more than compensated by the reduction in the long wave radiation reaching the ground from the cloud base; therefore, there is a net eduction in solar radiative heating of the ground, evaporation, and precipitation. hese factors appear to be of some importance in producing changes over northwest dia near the desert region, but a more quantitative evaluation of their influence observed interannual variability of monsoons has not been carried out.

Bavadekar and Mooley (65) computed the evapotranspiration and the moisture ux convergence of the atmosphere for a triangular volume of Peninsula India and ound that the interannual variability as well as the intraseasonal variability during e monsoon season is negligible. This suggests that the land surface processes are ot important in determining the rainfall variability over this region. However, the curacy of evapotranspiration data used in this study is not known. From the nited number of observational and modeling studies, it is not possible to determine e contribution of soil moisture toward the observed interannual variability of onsoon circulation and rainfall.

CONCLUDING REMARKS

n the context of the annual variations of the global circulation the Asiatic monsoon ppears to be a highly periodic phenomenon primarily determined by the seasonal riations of solar radiation and asymmetric continentality with respect to the equator. owever, although the onset and duration of the monsoon is quite regular, the recise date of onset at a given location is highly variable (up to 1 month). Once e onset has taken place, the nature of the day-to-day fluctuations and seasonal ean for one year can be very different from another. The day-to-day changes uring a monsoon season seem to be associated with a variety of factors, the most otable of which are:

1. the formation, growth, decay and propagation of weather-scale disturbances (lows and depressions, etc.) which are manifestations of the hydrodynamic and moist-convective instabilities of the large-scale flow; and
2. changes in the locations and intensity of east–west oriented convergence zones (the near-equatorial trough, the monsoon trough, etc.) which can occur either due to quasi-periodic (15–40 days) fluctuations in the tropics or due to influences of the traveling and quasi-stationary waves in the mid-latitudes of the Northern and Southern Hemispheres.

The role of these short period (i.e., shorter than a season) fluctuations in determining the seasonal mean is not clearly understood. It is also not clear how these transient (high frequency) fluctuations are influenced by global- or planetary-scale low frequency changes.

Since space and time averaged precipitation is determined largely by the amount of moisture that converges in a region, it is reasonable to suggest that large-scale—larger than the scale of lows and depressions—convergence is the primary determinant of the rainfall intensity. Monsoon lows and depressions are manifestations of dynamical instabilities which organize precipitation at preferred scales and, therefore, it may be more profitable to study the large and planetary-scale circulation features that produce a suitable environment for these instabilities to grow. If intraseasonal and interannual variability of rainfall were determined solely by the intensity and frequency of disturbances, the prospects for long-range prediction of time averaged rainfall would be hopeless, but there seems to be sufficient reason to believe that these instabilities are strongly controlled by planetary-scale circulations which are perhaps more predictable than the instabilities themselves. A part of the intraseasonal variability of rainfall is indeed accounted for by the fluctuations associated with these instabilities, and that part would be difficult to predict. However, the remaining part could be more predictable, at least in principle, if the planetary-scale circulations responsible for the rainfall variability were forced by slowing varying boundary conditions at the earth's surface.

We can pose the following questions:

1. Is the behavior of the seasonal mean monsoon primarily determined by a statistical average of a variety of independent short period fluctuations that are not related to any seasonal or other low frequency forcings?
2. Are there global- and planetary-scale "forcing functions" (either due to slowly varying boundary conditions at the earth's surface or due to very low frequency changes like the Southern Oscillation) that determine the interannual behavior of the seasonal mean monsoon circulation and rainfall, and is the interannual variability of the short period fluctuations controlled by such large-scale low frequency forcings?

These questions are similar to those raised in the beginning of the chapter about the relative importance of the internal dynamics and boundary conditions. Clearly the questions represent the two extreme possibilites and the reality must lie between. This author is reluctant to accept the premise of the first question, and is inclined to accept the hypothesis implicit in the second question. The role of the unpredictable day-to-day changes due to the instabilities and the nonlinear interactions can not be ignored; however, it is unlikely that they can account for the total observed variability without considering the influence of the boundary conditions. The significant correlations between the Southern Oscillation index and seasonal mean monsoon rainfall clearly suggest that the behavior of the seasonal mean monsoon is strongly influenced by planetary-scale low frequency changes. It is shown in Chapter 16 that 9 out of 12 heavy monsoon rainfall seasons were preceded by a negative Darwin

ure trend (a measure of the Southern Oscillation), and 12 out of 14 deficient
on rainfall seasons were preceded by a positive Darwin pressure trend. Moreover,
re significant correlations between the monsoon rainfall and the equatorial
c SST anomalies, and snow cover over Eurasia and the Himalayas. It would
easonable to discard these relationships, attributing them to random chance.
ese observational facts combined with their physical plausability support the
nesis that the interannual variability of monsoons is significantly influenced
owly varying boundary conditions and very low frequency planetary-scale
pheric fluctuations. In order to advance our understanding of these phenomena
l be necessary to examine the behavior of the three-dimensional flow at plan-
and global scales rather than over limited regions as has been mostly the case
g the past.

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