Chapter 4
Monsoon over Southern Asia (Comprising Pakistan, India, Bangladesh, Myanmar and Countries of Southeastern Asia) and Adjoining Indian Ocean (Region – I)

4.1 Introduction – Physical Features and Climate

The world's largest and most powerful monsoon circulation develops over the region of Southern Asia and its adjoining Indian Ocean. There are several reasons for this development: The first is a most favorable land-sea distribution. The geographical location of the region (see a physical map in Fig. 4.1) with the Tropic of Cancer passing through almost the middle of the landmass of the region and the Tropic of Capricorn through the middle of the Southern Indian Ocean (indicated by thin dashed lines) appears to provide an ideal setting for a heat source to develop over the land and a heat sink over the ocean during northern summer and vice versa during northern winter.

A second reason is orography. The lofty Himalaya Mountains and the Tibetan Plateau standing along the northern boundary of the region and rising steeply from the plains of northern India to peak heights of 8–9 km above sea level and then an average Plateau height of 4–6 km above mean sea level not only protect it from the icy cold winds of Central Asia during the winter but also help shape the structure of a well-defined summer monsoon circulation over the region and provide a natural barrier to the moisture-laden onshore winds from the Indian Ocean blowing into the region. The high mountain ranges all along the northwestern, northern and the eastern boundaries of the region as well as the coastal mountain ranges, such as the Western Ghats Mountain of peninsular India and the Arakan Yoma and the Tennaserim Ranges of Myanmar (erstwhile Burma) play important roles in shaping the structure of the monsoon circulation and the distribution of monsoon rains over the region.

The geographical location of Southeastern Asia, sandwiched between the Bay of Bengal and the South China Sea, presents a complex array of narrow coastal mountain ranges and broad inland high plateaux with heights ranging from 1.5 to 2.5 km above sea level. Some of these ranges which are mostly north-south oriented rise steeply above the plains of Myanmar, Thailand and Malaysia. It will be shown later in this chapter that the geographical locations of these mountain ranges and high plateaux, both along the northern and the eastern borders of the Indian Subcontinent play crucial roles not only during advance and retreat of monsoon, but also in controlling the distribution of monsoon rains over the subcontinent.

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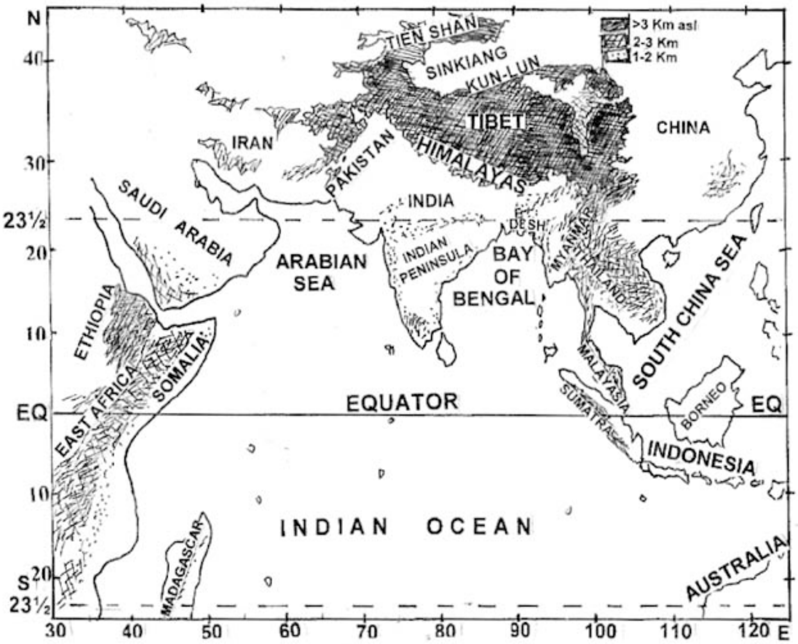

Fig. 4.1 Physical map of Southern Asia and adjoining Indian Ocean and its littoral countries

In the western sector, a chain of semi-arid and desert lands extends from Pakistan and Iran southwestward to Saudi Arabia and beyond to East Africa across Somalia, Kenya, Tanzania, Zimbabwe to Mozambique and South Africa against the backdrop of the high East African Mountains with heights ranging between 2 and 3 km asl. It will be shown in the present text that these varied land features along the western boundary of the Indian Ocean play crucial roles in advance and withdrawal of monsoon as well as in shaping the structure of the circulation and distribution of monsoon rains over the Western Indian Ocean and its littoral countries during both winter and summer.

The seasons over the region which exercise great influence on climatic conditions are as follows: The winter season characterized by extremely cold and dry conditions over the northern parts of the region appears to cover a period of 3 months from December to February. Then comes a period of approximately 3 months, March to May, when the land surface warms up rapidly to very high temperatures and 'heat lows' develop over the land. This is the period of transition from winter to summer monsoon. Summer monsoon proper arrives towards the end of May or early June and covers the whole region including the northernmost parts of the Western Himalayas by end of August. It starts withdrawing from the northwestern part of the region from early September, but the process of withdrawal is not completed till end of November.

Thus, the total period of summer monsoon over the region, from its first onset to final withdrawal, is almost 6 months. It, therefore, stands to reason that in the present text, the period of summer (wet) monsoon be divided into two parts: the onset phase (June to August) and the withdrawal phase (September–November), in order to conform to ground realities. The transition from summer to winter monsoon conditions varies from place to place but is usually about a month. The winter monsoon is of shorter duration.

It is important to bear in mind the great diversity of the seasonal weather and climate over the Indian Subcontinent arising largely from its topography and latitudinal extent from about 5 to 35°N. For example, in winter (December–February), while the northern parts of the subcontinent (north of about 25°N) shiver in near-freezing temperatures at times, the Indian peninsula and a large part of SE Asia enjoys cool, salubrious climate. Conversely, during the SW monsoon season, cool, humid weather of the south and the southeast stands in sharp contrast to the hot, semi-arid desert climate of the NW India and Pakistan.

4.2 The Winter Season (December–February)

4.2.1 General Climatic Conditions

In winter, the surface temperatures are generally low over the subcontinent and decrease with latitude. The lofty Himalayan mountain ranges along the northern boundary separate the region from Central Asia and protect it from the icy cold temperatures of that region. Relatively warmer temperatures prevail over the Indian Ocean, with the warmest temperatures being over the equatorial region where the ITCZ is located along about 15°S over the western Indian Ocean, but is much closer to the equator over the eastern part.

In keeping with the above-mentioned temperature distribution, m.s.l. pressure is generally high over the cold Indian Subcontinent and low over the warm equatorial ocean. An interesting aspect of the pressure distribution is that the three land segments of Southern Asia, viz., the Arabian Peninsula, the Indian Subcontinent, and Southeastern Asia, develop their independent high pressure cells, with low pressure troughs located in between over the somewhat warmer oceans; one over the northern Arabian Sea and the other over the northern Bay of Bengal. The orographic influence of the Himalaya mountain ranges and other coastal mountain systems, such as the Western Ghats of India and the Arakan Yoma of Myanmar is also significant. Generally, across any of these mountain ranges of appreciable height, a trough of low pressure tends to form on the windward as well as the lee sides of the mountain range with a ridge of high pressure on the mountain itself. Accordingly, troughs of low pressure are observed to form along the foothills of the Himalayas; one over Pakistan in Western Himalayas and the other over West Bengal and Bangladesh in the eastern part of the subcontinent.

The prevailing quasi-stationary pressure distribution drives anticyclonic circulation around the high pressure cells and cyclonic circulations around the troughs of low pressure. As part of the anticyclonic circulations, low-level airflow is generally northeasterly over the northern Indian Ocean. However, near the equator, the airflow is cross-equatorial in the western part of the ocean (west of about 70°E), and more or less westerly in the eastern part. The equatorial westerlies in the eastern part of the ocean divide the warm low pressure over the area into two cells, one north and the other south of the equator, each having its own trough, viz., the North Equatorial Trough (NET) and the South Equatorial Trough (SET).

According to rainfall map prepared by the India Meteorological Department (1943) (not shown), there is little rainfall over the subcontinent during the season, except areas affected by traveling disturbances.

4.2.2 Disturbances of the Winter Season

4.2.2.1 Western Disturbances (W.D.)

These are large-scale wave disturbances which form in midlatitude-subtropical baroclinic westerlies and usually travel from west to east. During northern winter, they travel from the Eastern-Mediterranean region and enter the Indian region from November onward. Their arrival is heralded by appearance of first high clouds and then medium and low clouds with rain or snowfall in the mountains of western Himalayas. With advance of the season, their track continually shifts equatorward and more and more of the subcontinent are affected by them and experience rain or snowfall. In December, western disturbances cross the mountain ranges of Iran and Afghanistan and arrive over NW Pakistan and adjoining western Himalayas in a diffused state with distorted structure, but on arrival over the plains of Pakistan, some of them interact with the pre-existing orographically-maintained trough of low pressure and regain their frontal structure and identity. The interaction leads to strengthening of the wave disturbance. Those developing large amplitude may draw moisture from the Arabian Sea and strengthen further. The rejuvenated disturbance (which is popularly called a 'Western Disturbance' in the subcontinent) then generally moves eastward, often in an occluded state, affecting southern Himalayas and adjoining plains of northern India. Some of them travel as far as Bangladesh and Assam and the mountains of eastern Himalayas where they usually break up and disappear. While passing over the plains of northern India, they draw moisture from the Bay of Bengal and are often accompanied by thundery rain and squally conditions. The average lifetime of a western disturbance is 4–6 days and there may be 6–8 disturbances per month at the peak of the season. Farmers welcome winter rain, for it helps in the cultivation of winter crops, such as wheat, etc. There is also heavy snowfall in the western Himalayas, the depth of snow occasionally exceeding 15 metres and causing dangerous avalanches. In the wake of some of these disturbances, dense fog may appear over large tracts of northern India and persist for days, interfering with road, rail and aerial communications. During the passage of these disturbances, the subtropical westerly jetstream may often reach a speed of 60–75 m s^{-1} at about 200 hPa over the latitude belt 25–30°N.

4.2.2.2 Easterly Waves and Cyclonic Storms of the Northern Indian Ocean

A surge in the low-level E/NE-ly tradewinds that converge into the equatorial trough of low pressure over South China Sea often give rise to easterly waves which travel westward. Arriving over the Bay of Bengal, some of them develop into westward-propagating depressions and cyclones. Their formation and movement are clearly seen in day-to-day satellite cloud imagery. Large cloud clusters and rainfall associated with them are observed to move westward a few degrees north of the equator. During their movement, they may strike the coast of Sri Lanka and southern Tamilnadu. Once in a while, they may cross the peninsula and emerge into the Arabian Sea. But such occasions are rare.

But not all surges in the tradewind easterlies develop into waves and cyclonic disturbances. Freeman (1948) draws an analogy between these surges and supersonic gas flows and concludes that surges lead to hydraulic jumps in the easterly flow and formation of cloud lines which move downstream over long disturbances.

4.2.2.3 Easterly Waves and Cyclones of the Southern Indian Ocean

Westward-propagating easterly waves also form south of the equator when a surge in the low-level SE-ly tradewinds, or the monsoon northwesterlies, converges into the circulation around the SET and increases its cyclonic vorticity.

The enhanced vorticity gives rise to westward-propagating cyclonic disturbances. After formation, these disturbances usually move westward as lows or depressions. However, a few of them which survive and enter the western Indian Ocean (west of about 80°E) recurve southwestward and develop into tropical storms and cyclones. Several islands of the region including Madagascar and Mauritius are badly hit by these cyclones almost every year. Beyond these islands they recurve further around the subtropical high pressure belt. Some of them are drawn into the circulation of the midlatitude wave disturbances which sweep across South Africa and move eastward across the Southern Indian Ocean towards Australia – New Zealand.

4.3 The Transition Season (March–May)

Several important developments take place in the weather over the Subcontinent and the adjoining Indian Ocean during the transition season. These include:

(1) Passage of WDs across the northern part of the Subcontinent; and
(2) Development of powerful ‘heat lows’ over land and relatively ‘cold highs’ over neighboring oceans.

The equatorial Indian Ocean is, perhaps, one region which becomes most active during this period when summer monsoon withdraws from the southern hemisphere and enters the northern hemisphere. Varieties of interesting phenomena are observed over the equatorial region about this time.

4.3.1 Western Disturbances

Western disturbances, described in Sect. 4.2.2.1, continue to be active over the northern part of the subcontinent during the transition season, though their track gradually shifts poleward with advance of the season. During their eastward passage, they draw additional moisture from the Arabian Sea as well as the Bay of Bengal and cause increased convective rainfall on the plains of northern India and extensive snowfall on the mountains.

4.3.2 'Heat Lows' over Land and 'Cold Highs' over Ocean

During March and April, 'heat lows' form over the Indian Subcontinent as well as other land areas bordering the Northern Indian Ocean, such as Somalia and Saudi Arabia in the west, and Malaysia, Thailand and Myanmar in the east, while the adjoining oceans, the Arabian Sea and the Bay of Bengal continue to be under 'cold highs'. By mid-May, the heat lows over the land areas develop further and move further north to take up their northernmost summer locations. The heat low over the Indian Peninsula also deepens further with its trough at surface oriented in a direction almost paralleling the east coast of the peninsula. While low level convergence and convection occur in the boundary layer of these heat lows, strong subsidence and divergence prevails in the upper troposphere above them. With adiabatic cooling of the rising air and subsidence warming aloft, the result is a stable stratification in the vertical in these heat lows at this time.

The scenario changes when, under the influence of the prevailing pressure distribution, low-level winds diverging from the neighboring oceanic high pressure cell inject cool, moist air into the heat low circulation over the land along the coastal belts of Southeastern India and Bangladesh. The injection of moisture at low levels in this manner makes the atmosphere over the region, especially the coastal belts, conditionally unstable.

4.3.3 Severe Local Storms

Over central and eastern parts of the Subcontinent where a lot of latent instability energy is stored in the atmosphere due to influx of cool, moist air from the Bay of Bengal at low levels, widespread thunderstorms and hailstorms occur whenever a Western Disturbance affects the region. Locally, in Bengal, these storms are known

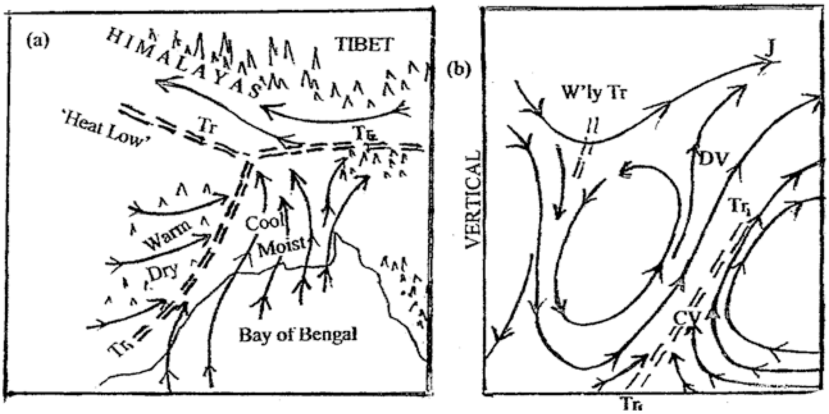

Fig. 4.2 Illustrating the formation of a Kal-Baisakhi: (**a**) A plan view showing locations of low-level troughs (*double-dashed*), (**b**) a vertical section showing superposition of a W'ly jetstream (J) on low-level trough. DV – divergence, CV – convergence

as 'Kal-Baisakhis' or deadly storms of the month of Baisakh (April–May). They are accompanied by thunder and lightning, squally winds, and heavy downpours. Many of them breed deadly tornadoes which inflict heavy loss of life and property.

Figure 4.2 shows schematically a typical synoptic situation favorable for occurrence of a Kal-Baisakhi over eastern India and adjoining Bangladesh.

The left panel of Fig. 4.2 shows the heat low troughs (Tr), to which cool, moist air from the Bay of Bengal converges and rises in strong penetrative convection when a upper-air W'ly trough approaches the region and has its divergent area associated with upper-level Jetstream (J) superimposed upon the pre-existing trough with low-level moisture convergence (right panel).

4.3.4 Developments over the Equatorial Indian Ocean

During transition season, summer monsoon withdraws from the southern hemisphere and enters the northern hemisphere. The movement is marked by several interesting events over the equatorial Indian Ocean, including the following:

(a) Cross-equatorial movement of Monsoon Circulation;
(b) Formation of Equatorial Westerlies, Double Equatorial Troughs and Cloud Bands;
(c) Increased Cyclonic and Anticyclonic Activity over the Equatorial Zone

(a) Cross-Equatorial Movement of Monsoon Circulation: With increased warming of the earth's surface and falling of pressure to the north of the equator and cooling and rising of pressure to the south and differential rate of heating between land and ocean, a gradient of pressure tendency develops between the two sides of

the equator, which enables the monsoon current of the southern hemisphere to cross the equator and move into the northern hemisphere to start the process of its advance towards the Indian Subcontinent.

An example of this type of forcing for equatorial crossing and relocation of the monsoon current from the winter to the summer hemisphere may be seen in Fig. 4.3 which shows the distribution of mean sea level pressure at 12 GMT on 7 April 2008 when monsoon appeared to have crossed the equator in 2008 in the Western Indian Ocean. According to the isobaric field shown, a pressure gradient exists in the vicinity of the East African coastline not only between the two sides of the equator but also along the equator from west to east. Also, north of the equator, a pressure gradient exists between the ocean and the land across the coast of Somalia. These pressure gradients appear to provide a safe passage for a parcel of air approaching the equator from the south to move along a path indicated by a thick continuous line with arrow in Fig. 4.3. The new locations of the cold sector of the monsoon and related ITCZ (dashed) north of the equator are also shown.

(b) Formation of Equatorial Westerlies, Double Equatorial Troughs: The low level winds and circulation over the equatorial Indian Ocean during the transition season show a broad band of equatorial westerlies between two troughs of low pressure, one in each hemisphere. Tradewinds diverging from the oceanic high pressures converge into these equatorial troughs forming penetrative convection and Cloud

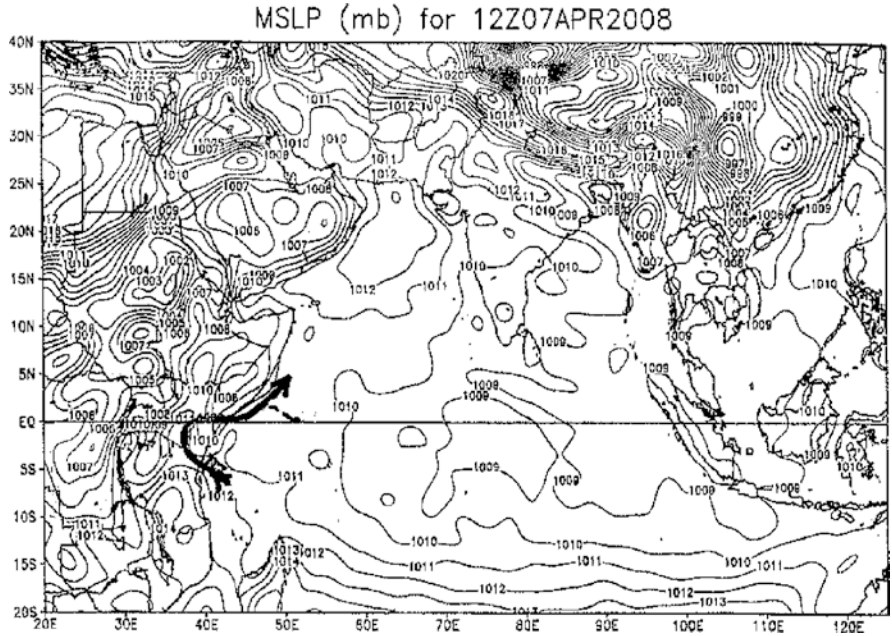

Fig. 4.3 Map showing MSLP (mb) at 12 Z, 7 April 2008 over the Indian Ocean and littoral countries *Thick continuous line* shows how the cool humid monsoon current crosses the equator, with ITCZ (*dashed*) located to its east (NCEP Reanalysis)

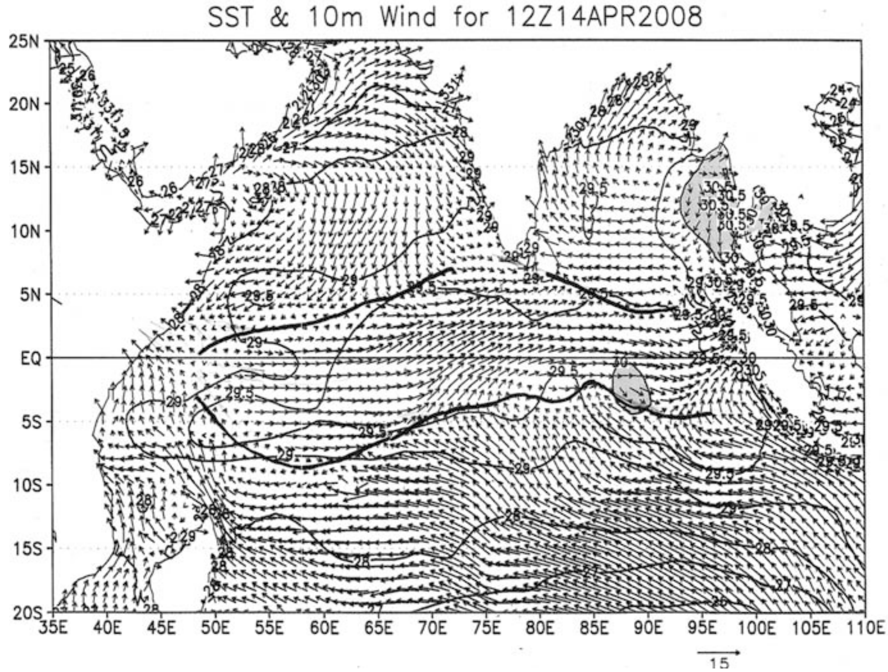

Fig. 4.4 Wind vectors and streamlines over the equatorial Indian Ocean at 10 m above sea surface and ocean surface temperatures (°C) at 12 GMT, 14 April 2008. *Thick continuous lines* show the locations of the troughs (adapted from NCEP Reanalysis)

Bands, one on either side of the equator, during a short period of equatorial crossing of monsoon. Just how this happens is demonstrated by Figs. 4.4 and 4.5 which shows winds and streamlines at 10 m above mean sea level and 850 hPa at 12 GMT on 14 April 2008, approximately a week after equatorial crossing.

The double equatorial convergence zones are characterized by formation of double cloud bands, one on each side of the equator, the presence of which is revealed in satellite cloud imagery (see Fig. 4.6).

Double equatorial troughs and associated cloud bands are persistent features of the circulations over the equatorial Indian Ocean during the transition season. However, they are short-lived and usually observed around the time of equatorial crossing only.

(c) Increased Cyclonic Activity over the Equatorial Eastern Indian Ocean: The spring transition season appears to be the period of maximum cyclonic activity over the equatorial eastern Indian Ocean as well as Arabian Sea (see Fig. 4.7)

The reason why the equatorial eastern Indian Ocean is so cyclogenetic during the transition season is to be sought, inter alia, in the warm ocean surface temperature of the Bay of Bengal. Further, most of the cyclonic storms develop around monsoon troughs which interact with traveling disturbances such as E'ly and W'ly waves. After development, the cyclonic disturbances move northwestward and later recurve

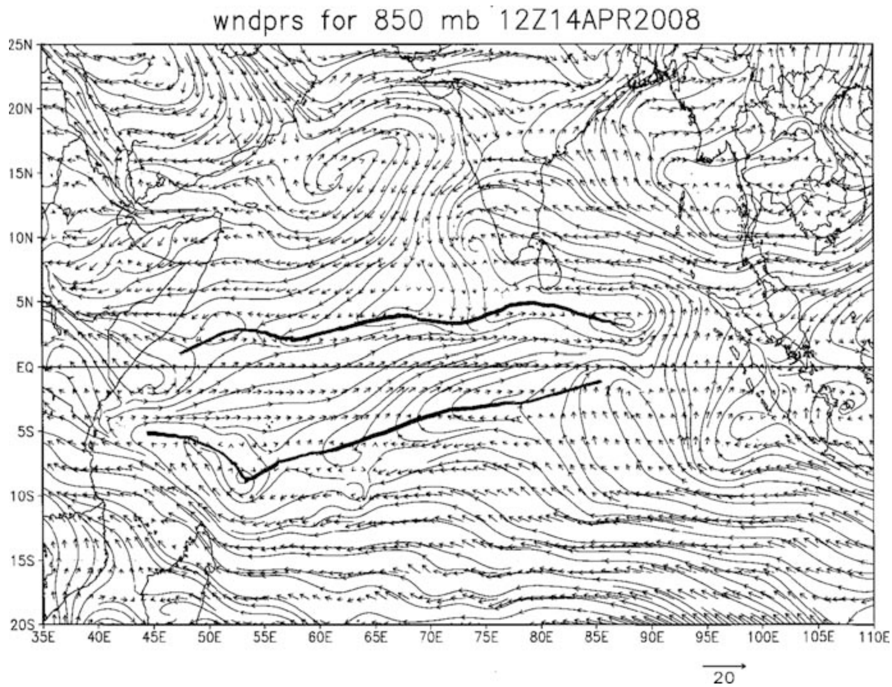

Fig. 4.5 Wind vectors and streamlines at 850 hPa showing the locations of two equatorial troughs, one on either side of the equator at 12 Z, 14 April 2008

northward and then northeastward to strike the coasts of West Bengal, Bangladesh and Myanmar. However, a few of them may continue their northwestward movement to strike the coast of Andhra Pradesh before recurving. A few may even cross the peninsula to emerge over the Arabian Sea to affect the west coast of India. A few cyclones may also form and develop over the Arabian Sea itself.

4.4 Advance of Summer Monsoon to the Indian Subcontinent – General Remarks

The India Meteorological Department (1943) has traditionally used certain criteria associated with rainfall to determine the normal date of onset of summer monsoon at a place.

Ananthakrishnan et al. (1968) who proceeded on that basis and studied the onset of SW monsoon over Kerala found that the synoptic conditions over the Arabian Sea associated with the onset were, in their own words, as follows:

(i) A disturbance in the Arabian Sea/Bay of Bengal. The most common initial form of the disturbance is a trough of low pressure in southeast Arabian Sea;

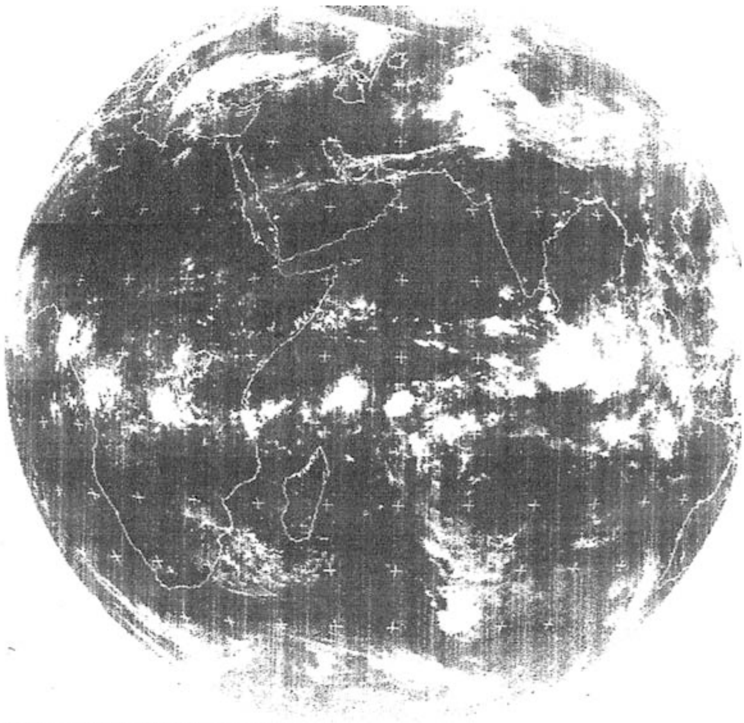

Fig. 4.6 Satellite Cloud Imagery showing double cloud bands, one on each of the equator, over the Equatorial Indian Ocean, at 12 GMT, 14 April 2008

(ii) Reports from ships and island stations in the South Arabian Sea, of heavy convection, squally weather and rough seas or swell from southwest with moderate to strong winds from some southerly to westerly direction;
(iii) The strengthening and deepening of lower tropospheric west winds over extreme south peninsula and Sri Lanka and strengthening of upper tropospheric easterlies to 40 Kts for a few days at 14 to 16 km; at the time of onset, the easterlies reach a maximum speed of about 60 Kts.
(iv) The tendency of the strong westerlies of the upper troposphere over north India to break up or shift northwards.
(v) Persistent moderate to heavy clouding in the south Arabian Sea shown by satellite pictures and its tendency to shift northwards.

In a note added to the criteria, the authors state that all these features may not always be present simultaneously. The organization of the circulation to the monsoon patterns extends over an interval ranging from a few days to 1 or 2 weeks.

The above-mentioned study by Ananthakrishnan et al. (1968) is very significant in the sense that it marked a welcome re-thinking on the nature of monsoon and

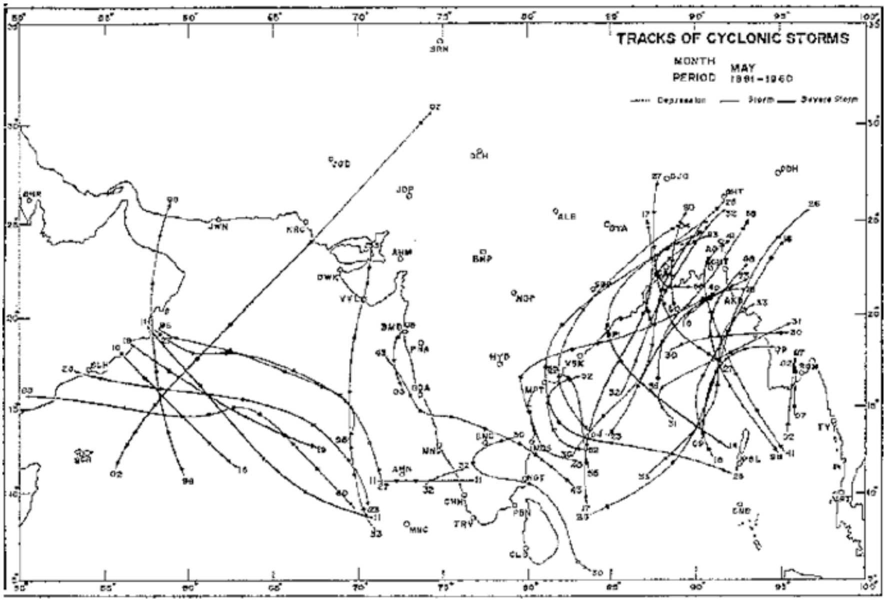

Fig. 4.7 Tracks of cyclonic storms that formed over the Bay of Bengal and the Arabian Sea in May during the 70-year period, 1891–1960 (After Rao, 1981)

its onset. It appeared to recognize for the first time that monsoon was not simply observed rainfall, nor a particular type of wind, but something beyond that, and that we must consider some aspects of the circulation that is associated with the rainfall or the wind to determine the onset. So, the problem of onset of summer monsoon over the Indian Ocean still remains to be addressed.

In this regard, in addition to qualitative analysis of monsoon onset in Chap. 1 and the preceding sections, the author examined data and analyses of several meteorological variables over the Indian Ocean at surface and lower and upper-tropospheric levels at 12 GMT daily over a 5-year period (2004–2008) (January–August), available from NCEP Reanalysis. For tracing the movement of the cool, humid air of the monsoon current, stress was laid on the analysis of temperature, pressure, specific humidity, and wind fields of the different layers extending from the ocean surface to 500 hPa. The results of this examination are depicted in Fig. 4.8.

For a closer look at the problem of advance, the total period of advance was divided into *three* stages, as follows:

Stage 1. Advance over the Southwestern Indian Ocean, the Arabian Sea, and the Bay of Bengal to the Indian Subcontinent (April–June);
Stage 2. Advance over the Indian Subcontinent (June–July); and
Stage 3. Advance from the Indian Subcontinent to the Western Himalayas and the Tibetan Plateau (July–August)

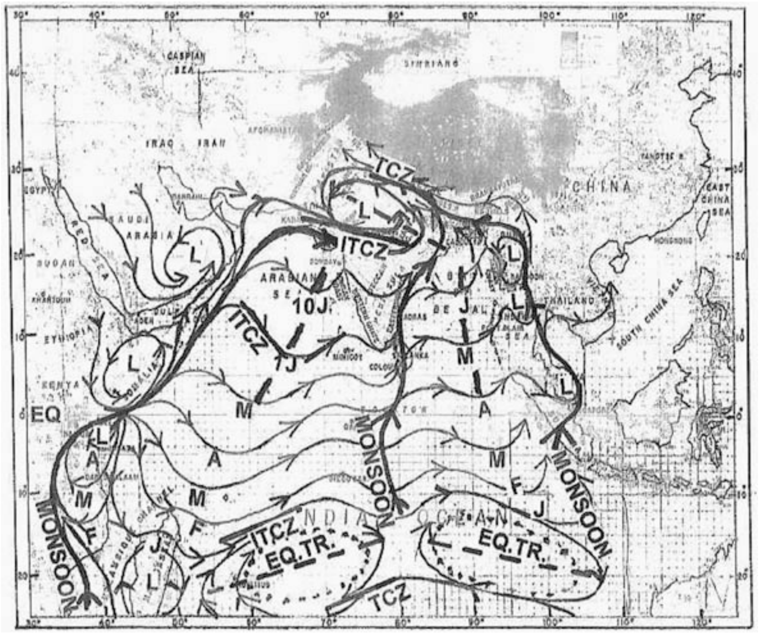

Fig. 4.8 Schematic showing advance of summer monsoon from the Indian Ocean. The *thick continuous lines* with *arrows* show the routes of the major cross-equatorial flows. *Thin continuous lines* with wave structure mark the approximate monthly locations of the advancing monsoon current Troughs are indicated by *thick double-dashed lines*. The monsoon circulation at the beginning of advance showing locations of ITCZ and TCZ (*short thick lines*) over the Southwestern Indian Ocean is indicated by *dotted lines* with *arrows*. ITCZ and TCZ over India are also shown

The following aspects of the advance of monsoon are highlighted by Fig. 4.8:

1. Locations of the main heat lows (L) and their troughs (thick dashed lines) in all the land sectors bordering the Southwestern Indian Ocean, the Arabian Sea, and the Bay of Bengal;
2. The (J-F) location of the equatorial trough of low pressure with cyclonic circulation (dotted lines with arrows) around it over the Southern Indian Ocean with locations of ITCZ and TCZ, *before* it starts its seasonal movement;
3. Three major *monsoon currents* (*thick* **bold** *continuous lines* with *arrow*) transporting *cool*, *humid* air of the winter hemisphere to the summer hemisphere, each being a *divergent* current from a heat sink or a high pressure area. These currents *converge* into the circulations around low pressure areas on either side, one being the heat low over the neighboring land and the other over the oceanic equatorial heat source (Note that the Continents of Africa, Australia and the Maritime Continent play important roles in this regard). A major cross-equatorial flow of monsoon current is influenced by the trough of the heat low over Peninsular India. It appears to cross the equator in the longitudes of Sri Lanka and flow along the western boundary of the Bay of Bengal close to the east coast of the peninsula;

4. Approximate northern boundary of the advancing monsoon current (thin continuous lines with wave structure) in the northern hemisphere at certain epochs of time (months/dates); in the southern hemisphere, the lines refer to the southern boundary of the retreating monsoon;
5. Advance of the monsoon current along the west coast of Peninsular India as well as the Arakan coast of Myanmar appear to be facilitated by movement of a cold sector of the monsoon wave over the respective region;
6. Locations of the *ITCZ* and the *TCZ* (indicated by short thick lines) on either side of the equatorial heat source after its final arrival over the northwestern part of the Indian Subcontinent.

4.4.1 Advance over the Indian Ocean (April–June) – Stage 1

4.4.1.1 Retreat from the Southern Hemisphere

The successive monthly locations of the southern boundary of the monsoon wave during the period, January–April, shown in Fig. 4.8, illustrate how summer monsoon withdraws from the southern hemisphere before it enters the northern hemisphere. Note how the heat low over the land sector jumps from its January location over the island of Madagascar to the African mainland and then travels along the western boundary of the Southwestern Indian Ocean to southern Somalia in April and disappears from the southern hemisphere. The equatorial troughs over the ocean also shift northward. By April they move into the northern hemisphere to start the process of advance of monsoon towards the Indian Subcontinent.

Summer monsoon advances over the Northern Indian Ocean from different parts of the equator, as shown in Fig. 4.8. However, the first move appears to be made from the side of the Eastern Bay of Bengal. So, we start from that side first.

4.4.1.2 Advance over the Bay of Bengal

During late March (M) or early April (A), the increased seasonal warming of the northern hemisphere and cooling of the southern hemisphere in the Southeast Asia sector and differential heating between land and ocean in general deepens a newly-formed heat low over the Malaysia-Sumatra region, which forces the North Equatorial Trough (N.E.T) of low pressure over the Bay of Bengal to move northward along with it. As the heat low crawls slowly northward along the narrow landstrips of Thailand and Myanmar, it carries the W/SW'ly monsoon airstream with it on its southeastern side. Just about this time, under the powerful influence of the heat low over the Indian Peninsula, there is a major cross-equatorial airflow from the western side of the South Equatorial Trough towards the east coast of the Indian Peninsula, a branch of which turns eastward to strengthen the circulation around the northward-moving heat low over the mountain ranges of the Tennaserim coast. Early May, there is another cross-equatorial current from the side of Australia to enter the extreme western part of the South China Sea and flow northward along the eastern coast of the Malaysia-Thailand peninsula, a branch of which turns northward under the influence of a powerful heat low over Central Myanmar, while the other branch turns eastward to flow around a heat low over Thailand. From here, the S-ly current comes under the influences of several mountain ranges and finally turn towards the plains of northern India to flow as a ESE-ly current around the heat low over India along the foothills of the Himalayas. Needless to state, the cool, moist winds converging at the mountain slopes all along the routes produce heavy rainfall over the mountain ranges.

Thus, on the Bay of Bengal side, summer monsoon advances to the Indian Subcontinent via two main routes: (i) A strong cross-equatorial airflow by the side of Sri Lanka and parallel to the east coast of the Indian Peninsula, and (ii) a cross-equatorial airflow from the side of Australia and Indonesia.

4.4.1.3 Advance over the Arabian Sea

The Arabian Sea during early summer when the monsoon current crosses the equator is relatively cold compared to surrounding land areas where heat lows form, and remains under a strong temperature inversion with a strong anticyclonic circulation prevailing at low levels. The conditions inhibit further northward advance of the monsoon current. Further, part of the southern-hemispheric tradewinds that cross the equator near the Somali coast at this time appears to be diverted westward by the heat low circulation over the Congo region of Equatorial Africa, leaving only a feeble narrow coastal current to circulate around the heat low over Somalia.

But, the situation changes, though gradually at first, when a warming ocean surface and deepening heat lows over land allows a further northward movement of the cool monsoon current towards the Indian Subcontinent. The land-sea thermal contrast across the Somali coast intensifies resulting in development of the so-called *Somali jet* and intense *upwelling* along the Somali coast. Similar developments also occur along the coast of the Arabian Peninsula, further north. These developments help *the cold monsoon current which binds the heat low over the land to the oceanic heat low* to move further northeastward and cover more than half of the Arabian Sea by end of May. So, on or around 1 June, the equatorial trough of low pressure over the northeastern Arabian Sea is so placed as to have its associated *ITCZ* oriented in a more or less NW-SE direction over mid-ocean and *TCZ* near the coast of Kerala, the southernmost Indian State, where a S/SW-ly flow is enforced partly by the movement of the cold sector of the monsoon wave and partly by the local orography of the Western Ghats Mountains. From this stage onward, monsoon current moves rapidly northward under the influence of the heat low over India-Pakistan to cover most of the northern Arabian Sea and the Indian Peninsula by June 15. The progress of the monsoon over the ocean and the land during this period is indicated in Fig. 4.9.

4.4.1.4 Weather over the Northern Indian Ocean During Advance of Monsoon

Summing up the preceding paragraphs, it may be stated that the conceptual or idealized model of the advance of summer monsoon over the Indian Ocean suggested in

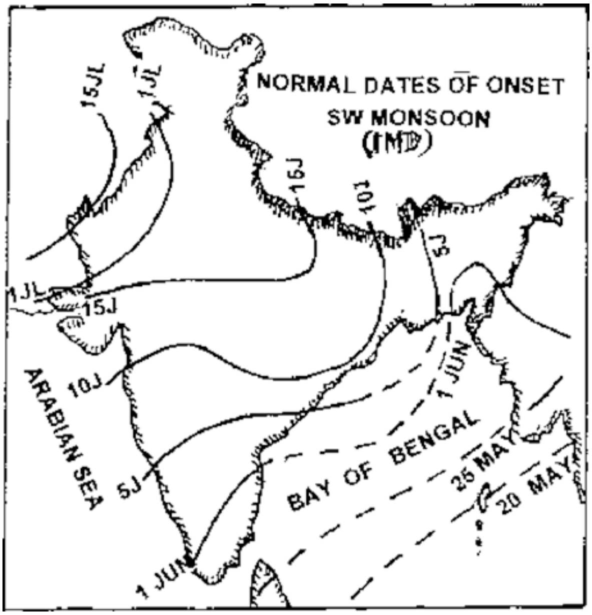

Fig. 4.9 Dates of onset of summer monsoon over the Bay of Bengal and the Indian Subcontinent, as determined by India Met. Dept (Rao, 1981)

the Text appears to account qualitatively for several observed features of the weather over the Bay of Bengal and the Arabian Sea. These include cold SST anomaly, strong temperature inversion, low level jets with highly stormy seas and unusually strong ocean currents, little cloud and rain, and few cyclonic disturbances along and to the west of the cold monsoon current, as against warm SST anomaly, weak temperature inversion, light to moderate winds, heavy clouding and precipitation and a high frequency of cyclonic disturbances, such as depressions and cyclones, to the east. Both ITCZ and TCZ are characterized as zones of cloudy and rainy weather with relatively clear skies in between, confirming the wave structure of the monsoon circulation.

4.4.2 Onset over the Indian Subcontinent (June–July) – Stage 2

The India Meteorological Department (1943) worked out the normal dates of onset of summer monsoon over the Indian Subcontinent from climatological records of observed rainfall from coastal and inland stations. Figure 4.9 shows these dates by isolines over land and dashed lines over the Bay of Bengal. No isolines or dashed lines are drawn over the Arabian Sea.

As stated in the preceding subsections, summer monsoon advances over the Indian Subcontinent in two broad airstreams; the Bay of Bengal stream from the southeast and the Arabian Sea stream from the SW, as shown in Fig. 4.8. In their final position by end of June, these airstreams converge into the circulation around the heat low over northern India, forming the ITCZ on the equatorial side and the TCZ on the poleward side of the so-called monsoon trough. The progressive advance of the two streams to their final destinations over the Subcontinent is given in the map. From SE to NW, the Bay of Bengal branch of the monsoon advances to Bangladesh by 1 June, Bihar by 10 June, East Uttar Pradesh by 15 June, Punjab, Rajasthan, Pakistan and some parts of the State of Jammu and Kashmir by early July. The Arabian Sea branch approaching from the SW crosses the Indian Peninsula by 10 June, Madhya Pradesh by 15 June, and Rajasthan by 30 June. Monsoon remains established over the subcontinent till the end of August. During this period, moderate to heavy rain falls along the ITCZ to the S/SW of the monsoon trough and along the TCZ which runs along the foothills of the Himalayas to the N/NE. Rainfall appears to be deficient over the trough zone in between. However, as shown in the following subsection, the monsoon wave does not remain in this position for long, since it moves further north to Western Himalayas during late July or early August.

4.4.3 Advance to Western Himalayas (July–August) – Stage 3

After summer monsoon gets fully established over the Indian Subcontinent, an extraordinary development takes place further north, which shifts monsoon wave from the plains of northern India to the top of the Himalayan Mountain complex on account of the development of a series of heat lows to the north.

It is well-known that during northern summer a heat low develops over the western part of the elevated Tibetan Plateau (e.g., Flohn, 1968; Yeh and Gao, 1979; Murakami and Ding, 1982; Luo and Yanai, 1984; Feng et al., 1984; Murakami, 1987a).

Almost simultaneously, heat lows also develop to the north of the mountain complex over the extensive lowlands of Uzbekistan and Kazakhstan to the northwest, the extensive desert lands of the Sinkiang province of China to the north, and the Inner Mongolian region of Northeastern China to the northeast. All these heat lows have their warm anticyclonic circulations above them in the upper troposphere. But on a larger scale, they combine to form a powerful anticyclonic circulation centered over the Tibetan Plateau and extending from the Mediterranean Sea in the west to the Pacific Ocean in the east. It is the development of this giant anticyclonic circulation and its sudden poleward movement, which appears to draw the *Monsoon* and related *Hadley circulations* over the Indian Subcontinent within its fold.

The poleward movement of the monsoon trough zone to the Himalayas in July–August causes a total re-organization of the associated Monsoon and Hadley circulation cells associated with it over the region. This movement implies a temporary bifurcation of the monsoon wave, one part remaining over the plains of Northern India with somewhat subdued activity, while the other part jumps over to the mountains to the north. The movement simply means a northward shift of the associated Monsoon and Hadley circulation cells, as shown schematically in Fig. 4.10.

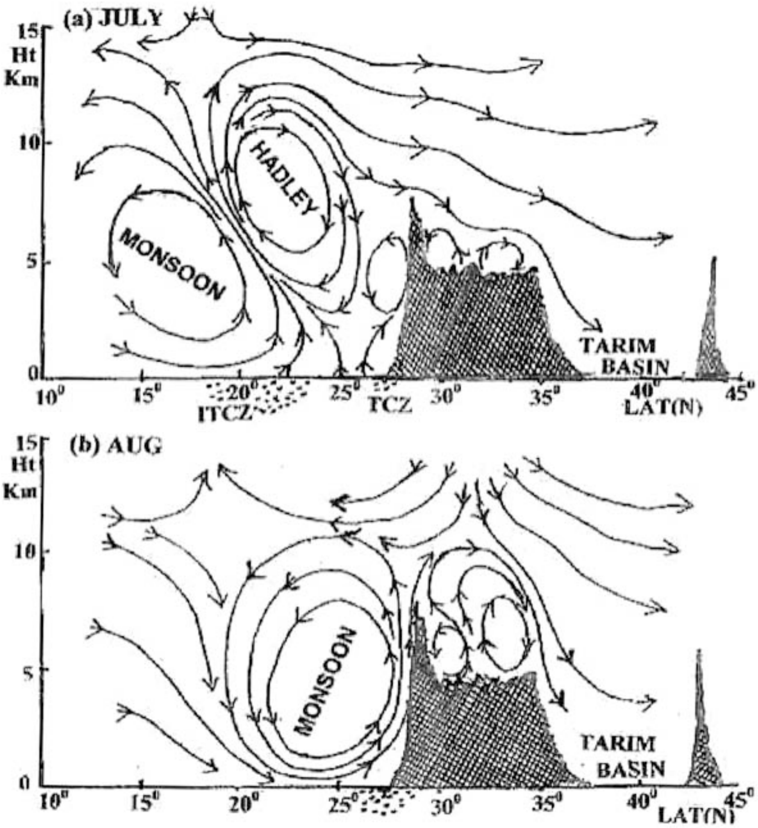

Fig. 4.10 Schematic showing mean meridional-vertical circulations and their associated rainbelts (*dotted*) over the Indian Subcontinent before (*upper panel*) and after (*lower panel*) development of the elevated heat source over Western Tibet

4.4.4 Source of Moisture for Monsoon Rainfall

Over most parts of the globe, the main source of moisture for monsoon rainfall in the summer hemisphere is the cool, moist tradewinds of the winter hemisphere which after crossing the equator in narrow longitudinal segments converge into the ITCZ of the summer hemisphere (e.g., Simpson, 1921; Findlater, 1969a,b; Saha, 1970).

While traveling over the ocean, the tradewinds pick up additional moisture from the underlying ocean surface. A limited amount of moisture is also injected into the TCZ.

The cross-equatorial origin of moisture-bearing tradewinds ushering in moisture for monsoon rainfall in the summer hemisphere is evident over several other parts of the globe as well, such as Eastern Asia, Australia, Africa and South America. For details, see the respective chapter on regional monsoon in the present text.

4.5 Disturbances of the Summer Monsoon during the Onset Phase

4.5.1 Onset Vortex over the Arabian Sea and the Bay of Bengal

During advance of the summer monsoon wave over the Arabian Sea as well as the Bay of Bengal, the atmosphere becomes dynamically unstable, often resulting in the formation of a cyclonic vortex, whenever it is disturbed by a traveling wave. Such a vortex has come to be known as an ‘Onset Vortex’.

In the Arabian Sea, a few studies (e.g., Krishnamurti et al., 1981; Saha and Saha, 1993a) have shown that both barotropic and baroclinic instability may be involved in the genesis of these vortices. After formation, these disturbances usually move in a northerly direction and accelerate the advance of the monsoon current along the West Coast of India. However, a few of them may move westnorthwestward and develop into cyclonic storms over mid-ocean before they hit the coast of Oman and then fizzle out over the sandy Arabian Desert.

In the Bay of Bengal, traveling E’ly as well as W’ly waves by their interaction with the quasi-stationary monsoon wave play an important role in the formation of an onset vortex. The formation of such a vortex may upset the normal schedule of advance of monsoon by either accelerating or delaying it by a few days.

4.5.2 Monsoon Depressions and Cyclonic Storms

During the monsoon onset phase, June to August, a number of depressions and cyclonic storms form in the monsoon trough zone over the Bay of Bengal and the Arabian Sea. A few also form over the land area adjoining the Head Bay of Bengal. Table 4.1 gives the total number of depressions and storms that formed over these areas during an 80-year period, 1891–1970 (After Rao, 1976).

The tracks of these disturbances are shown in Fig. 4.11.

After formation, most of the disturbances move in a WNW direction and yield heavy precipitation over their SW quadrant. Several States in India, such as Orissa, Andhra Pradesh, Southern Bihar, Jharkhand, Chhatishgahr, and Madhya Pradesh, receive a significant proportion of their annual rainfall during the passage of these

Table 4.1 Number of monsoon depressions (D) and cyclonic storms (S) during the 80-year period (1891–1970) in June, July and August over different areas

Area	June		July		August	
	D	S	D	S	D	S
Bay of Bengal	71	35	107	38	132	26
Arabian Sea	18	15	9	3	2	2
Land area	12	1	39	1	42	0

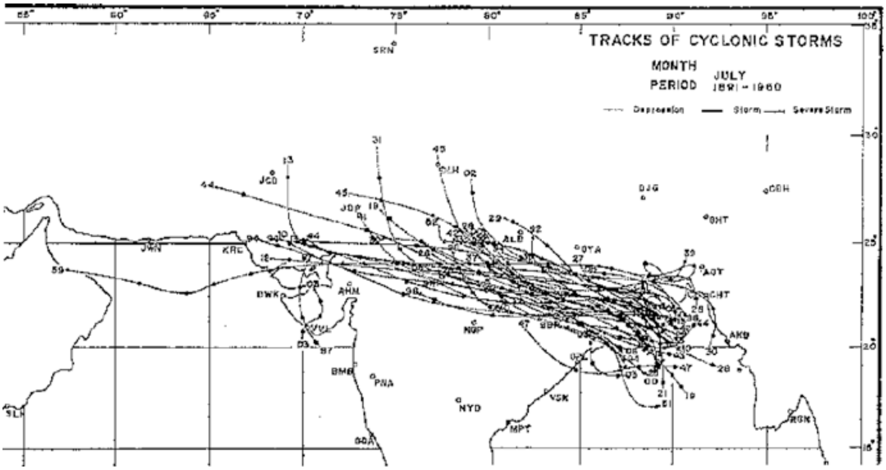

Fig. 4.11 Tracks of depressions and cyclonic storms that formed over the Bay of Bengal and the Arabian Sea in July during the 70-year period, 1891–1960 (after Rao, 1981)

disturbances. By contrast, little rain falls to the northeast of the trough axis. In the Arabian Sea sector, few lows and depressions form in this period. Those which form over the northeastern corner of the sea usually move in a northerly direction. Their contribution to annual rainfall is normally negligible. However, occasionally, midtropospheric cyclones form over the northeastern corner of the Arabian Sea and add significantly to coastal rainfall. More than 80% of all Bay disturbances during the onset phase (June–August) formed over the latitude belt, 17.5–22.5°N, and moved westnorthwestward.

4.5.3 Interaction of Monsoon with W'ly Waves

During summer, large-amplitude W'ly waves moving across the Himalayan region, between about 30 and 50°N, interact with the circulation around the quasi-stationary monsoon trough over the Indian Subcontinent as well as with traveling monsoon disturbances, forming an extended trough between the two regions. On such occasions, a ridge of high pressure with anticyclonic circulations prevails over northwestern and Central India and weather remains dry over these areas (see Fig. 4.12).

During the period of its eastward movement, the extended trough causes a relocation of the east-west oriented monsoon trough and its associated rainbelt which now lies across southern India. Thus, the interaction simply causes a redistribution of monsoon rainfall with two belts of heavy rain, one in the mountains in the north and other over Southern India in the south and a wide area of little or no rain in between over Central India. As the W'ly wave moves further eastward across the Himalayas taking the extended trough along with it, the belt of heavy rain also shifts eastward. A period of 3–5 days may be taken for the rainbelt to move across the mountains from west to east.

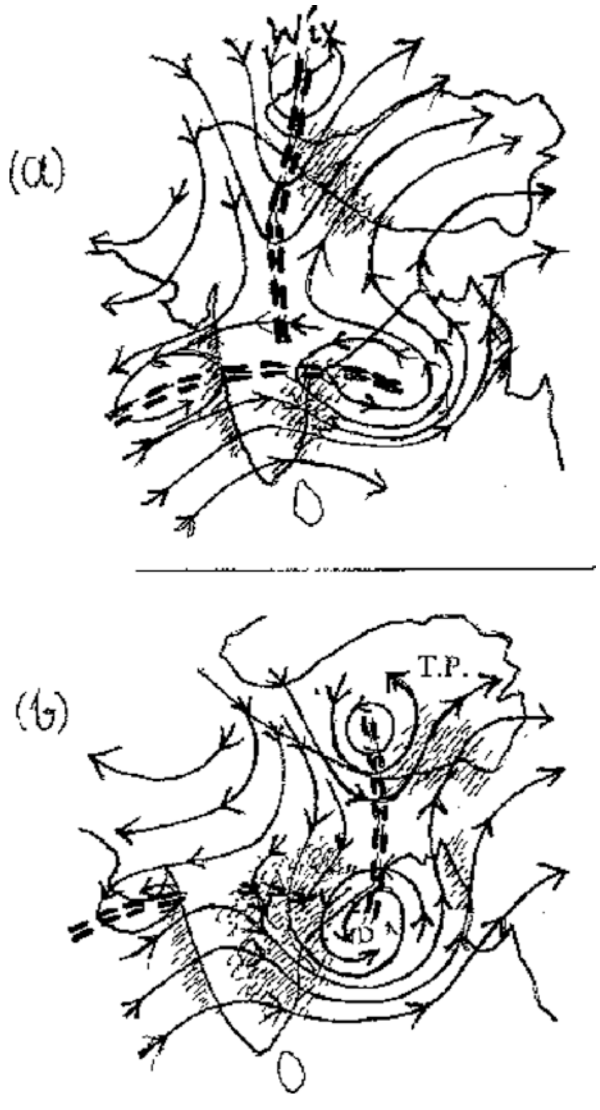

Fig. 4.12 Schematic showing interaction of a W'ly wave trough with: (**a**) the monsoon trough over the Indian Subcontinent, and (**b**) a Depression (D) over the Bay of Bengal. Heavy rainfall areas are hatched. Troughs are *double-dashed*

The W'ly wave troughs moving across the eastern Himalayas also interact with westward-propagating depressions over the Bay of Bengal forming an extended trough with it. In this case also, we have two areas of heavy rainfall, one in the north over the eastern Himalayas and the other over an extensive area of central and southern India. In fact, the 'break monsoon' situation over Central India almost disappears in this case. The windward slopes of the Western Ghats Mountain as well as the Arakan Yoma experience heavy rainfall on these occasions.

4.6 Rainfall over the Indian Subcontinent during the Onset Phase

Several factors appear to contribute to the distribution of monsoon rain over the Indian Subcontinent during the onset phase. They include: (1) Orography, (2) Heat Lows, (3) Location of the monsoon trough, (4) Depressions and cyclones, and (5) Travelling wave disturbances.

More than any other factor, orography appears to contribute the most to the distribution of monsoon rains over the Indian Subcontinent, as would be evident from Fig. 4.13, which records the highest concentrations of rainfall on the windward slopes of the mountains, and scanty rainfall on the leeside.

According to Fig. 4.13, the Western Ghats Mountains of the Indian Peninsula, the Arakan Yoma and the Shan States of Myanmar, and the Mountains of Eastern Himalayas, especially in Bangladesh, Assam and Arunachal Pradesh experience heavy rainfall, while the leesides of these mountains have deficiency of rainfall.

Along the foothills of the Himalaya Mountains, there appears to be a general decrease of rainfall from east to west. A rainfall maximum appears over Orissa and adjoining Central India to the southwest of the monsoon trough zone.

While the eastern and southern parts of the Indian Subcontinent receive substantial rainfall during the onset phase of the summer monsoon, the northwestern part especially the Thar Desert area of Pakistan and India suffers from shortage of rain because of the presence of the heat low circulation over the region.

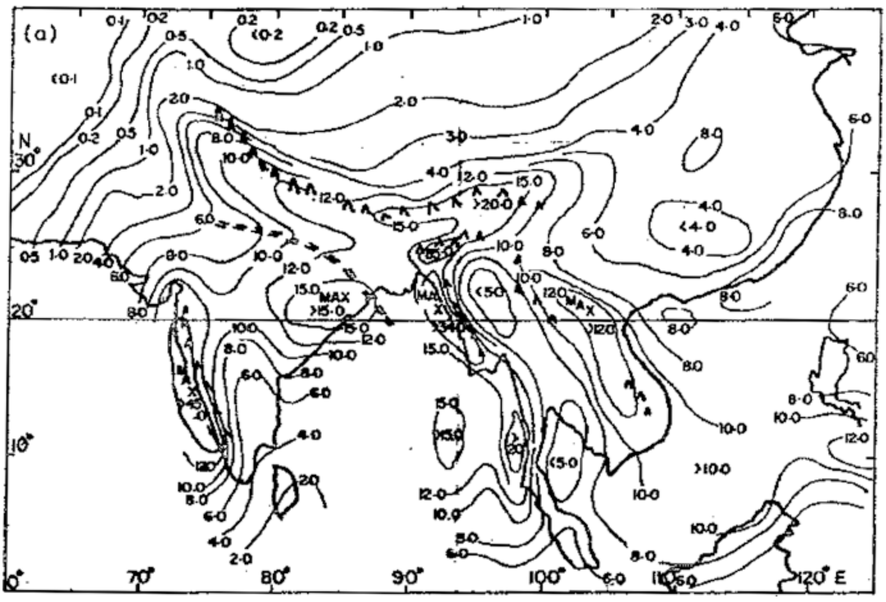

Fig. 4.13 Ten-year (1976–1985) mean July rainfall (unit: 10^{-5} kg m^{-2} s^{-1}) over the Indian Subcontinent. The *double-dashed line* shows the location of the monsoon trough (after Saha and Saha, 1996)

The distribution of rainfall shown in Fig. 4.13 agrees substantially with that of normal summer monsoon rainfall during the period, June–September, as determined by the India Meteorological Department (not shown).

4.7 Summer Monsoon – Withdrawal Phase (September–November)

4.7.1 Dates of Withdrawal of Monsoon

By and large, summer monsoon withdraws from the Indian Subcontinent and the Northern Indian Ocean by following the same route as it did during advance (Fig. 4.9), but in the reverse direction.

After reaching its peak intensity during July–August, monsoon starts withdrawing from the Western Himalayas in early September. The process begins with the filling up of the heat low over the elevated Tibetan Plateau and the re-establishment of the heat low over the northwestern part of the Indian Subcontinent. However, the transition takes place very gradually and almost imperceptibly for a while. The withdrawal from northwestern India is marked by a weakening of the heat low and associated monsoon trough, reversal of the low-level wind from southerly to northerly and a decrease of rainfall. The change ushers in a regime of somewhat cooler and drier air from the north.

Figure 4.14 gives isolines of the dates of withdrawal of the summer monsoon from India, as worked out by the India Meteorological Department on the basis of the climatological distribution of rainfall.

According to Fig. 4.14, the SW monsoon pulls out of nearly half the subcontinent by 1 October when its northern boundary runs from the hills of Uttar Pradesh to the middle of the West coast of India. By 15th October, it moves further southeastward so as to have its western end over the middle of the Indian peninsula and the eastern end over central Myanmar or even further south. From this stage onward, its southward movement over land and ocean is very slow, and it is not until the end of November that monsoon totally withdraws from the Indian peninsula and reaches the latitude of Sri Lanka. The Myanmar branch of the monsoon continues to move south/southeastward to reach eventually its winter location as NET within about 5° of the equator in the eastern Indian Ocean.

4.7.2 Retreating Monsoon Rain over Tamil Nadu

During the stage of withdrawal of summer monsoon from India, the area along the east coast of the Indian Peninsula, especially the state of Tamil Nadu, exposed to the moisture-laden ENE-ly winds at low levels, experiences retreating monsoon rains. In literature, this rain is often projected as winter monsoon rain. However, in reality, it is summer monsoon rain during its withdrawal phase. The mechanism of this rain is illustrated schematically in Fig. 4.15.

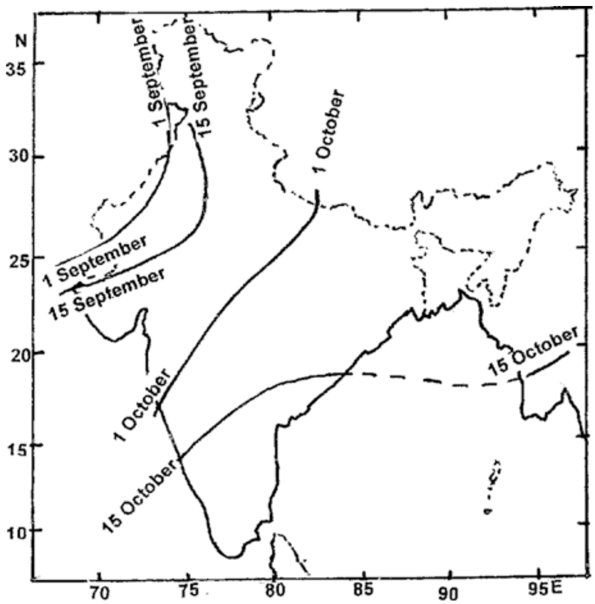

Fig. 4.14 Normal dates of withdrawal of the SW monsoon from the Indian subcontinent (after Rao, 1981)

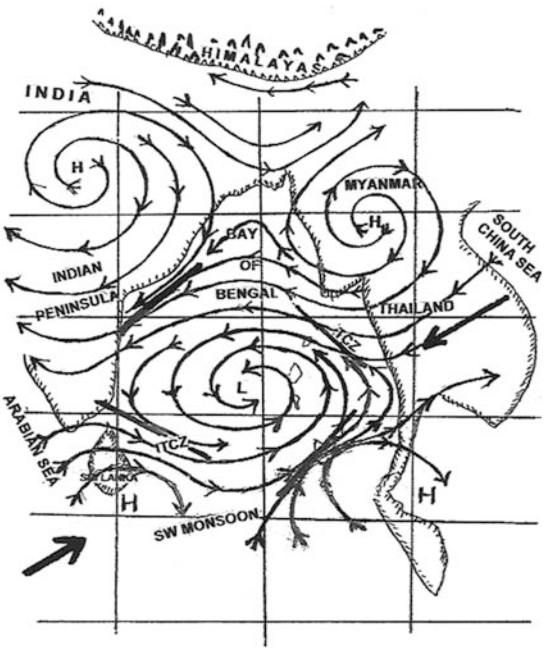

Fig. 4.15 Illustrating the mechanism for formation of a low level moisture convergence zone and rain in Tamil Nadu during monsoon withdrawal phase

Following the withdrawal of the monsoon trough from northern India, two high pressure cells with anticyclonic circulations develop over land, one over Central India and the other over Myanmar. Between their anticyclonic circulations, a trough of low pressure forms over the North Bay of Bengal (see Fig. 4.15).

Now, the wind diverging from the high pressure cell over central India has a limited sea travel before it approaches the coast of Andhra Pradesh and Tamil Nadu from N/NE-ly direction and its moisture content is low. On the other hand, the wind diverging from the high pressure cell over Myanmar has first to flow northwestward over North Bay and then turn cyclonically around the oceanic trough southwestward towards the Indian peninsula. Thus, it has a long sea travel and is fully saturated with moisture by the time it arrives at the Tamil Nadu coast where it converges into the circulation around the high pressure cell over Central India. The resulting moisture convergence along the Andhra Pradesh-Tamil Nadu coast (indicated by a thick continuous line) along with the prevailing upper-air divergence over the area is responsible for producing the rainfall over Tamil Nadu. Heaviest rainfall occurs at the mountains of Tamil Nadu facing the moist winds.

4.7.3 Disturbances of the Withdrawal Phase

4.7.3.1 Western Disturbances

Closely following the southward movement of the equatorial trough, the belt of the subtropical westerlies shifts southward and W'ly waves (WDs) follow a more southerly track. Their influence on weather over the northern part of the Indian Subcontinent, especially western Himalayas, increases. They also influence the track of cyclonic disturbances which move up from the Bay of Bengal and the Arabian Sea towards the mountain.

4.7.3.2 Depressions and Cyclonic Storms

During the monsoon withdrawal phase, there is a spurt in cyclonic activity over the Bay of Bengal and the Arabian Sea. A larger percentage of the depressions develop into cyclonic storms and their places of origin shift continually equatorward from September to November, as shown by statistics over a 70-year period, 1891–1970 (Rao, 1981). It is estimated that in November, 90% of the depressions and cyclones formed over a wide area bounded by latitudes 7.5 and 15°N and longitudes 77.5 and 100°E. This continued southward movement and the widening of the area of cyclonic activity would be evident from Fig. 4.16 which shows the tracks of these disturbances, when it is compared with Fig. 4.11 for July. These changes appear to be characteristic features of the monsoon withdrawal phase. During the same 70-year period, the number of depressions and cyclonic storms that formed in the Arabian Sea was 21 and 15 respectively. About 50% of them formed over the area bounded by latitudes, 10–12.5°N and longitudes, 62.5–75°E.

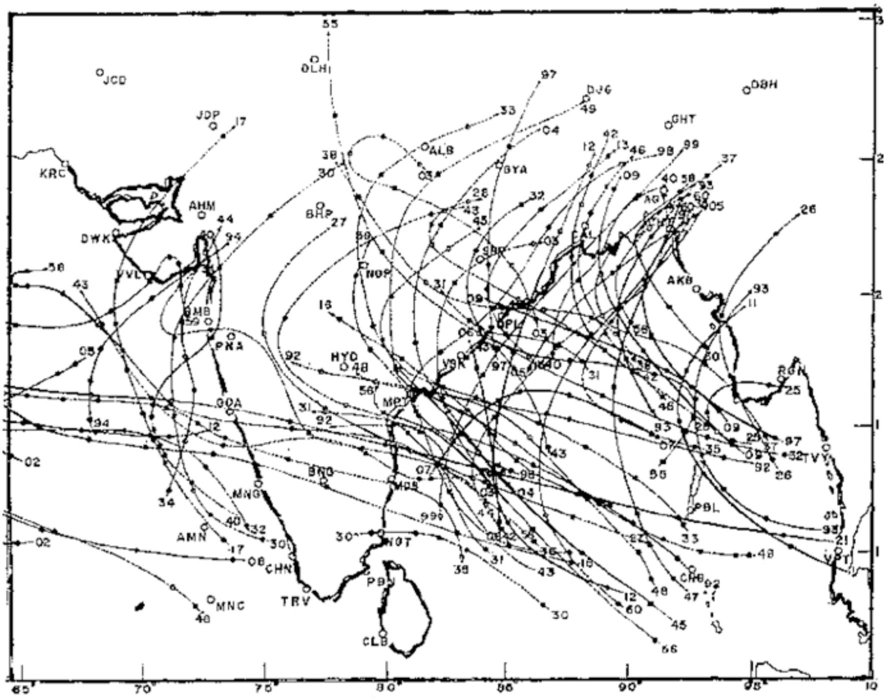

Fig. 4.16 Tracks of cyclonic storms that formed over the Bay of Bengal and the Arabian Sea in October during a 70-year period, 1891–1960 (after Rao, 1981)

In the Bay of Bengal, most of the depressions and cyclones after initial travel over the sea in a WNW direction recurved to a N/NE-ly direction after reaching the latitude belt, 15–20°N. Those amongst these which turned into severe cyclonic storms and entered land devastated many coastal belts. Low-lying river deltas which are particularly vulnerable in this regard suffered enormous losses due to high winds, storm surges and torrential precipitation.

Table 4.2 lists a few of the deadliest cyclones on record which took heavy toll of lives through storm surge drowning during the monsoon withdrawal phase.

Table 4.2 Statistics of some *killer tropical cyclones* with record storm surge

Cyclone	Date	Surge (m)	Death-toll
Bangladesh (Buckergunge)	November 1, 1876	9.0–12.0	100,000
Calcutta (Kolkata)	October 5, 1864	12.0	50,000
Bangladesh	November 13, 1971	7.0	300,000
Andhra Pradesh	November 17, 1977	5.0	9000

4.8 Variability of the Indian Summer Monsoon Rainfall

Observations show that the summer monsoon rainfall over India for the period, June–September, is not steady but varies on time scales ranging from a few days to a year or several years.

Since large-scale variability often leads to disastrous floods and droughts which cause miseries to people and affect the economy of the country, the problem of variability has been studied extensively ever since the time of Blanford (1884, 1886) who first initiated the study. He was followed by Walker (1910a, b, 1914) and several scientists in India and abroad (e.g., Mooley, 1975, 1976; Hahn and Shukla, 1976; Kanamitsu and Krishnamurti, 1978; Bhalme and Mooley, 1980; Angell, 1981; Shukla and Paolino, 1983; Mooley and Parthasarathy, 1983, 1984; Rasmusseen and Carpenter, 1983; Parthasarathy, 1984). To date, there has been a large body of literature on the subject of variability of Indian summer monsoon rainfall. An excellent review of some of the recent studies has been provided by Mooley and Shukla (1987) as well as by Krishnamurti and Surgi (1987) and readers interested in details of these various studies may refer to the original papers mentioned in the references.

4.8.1 Interannual Variability

For studying the rainfall variability, different workers used different sets of data. Mooley and Parthasarathy (loc. cit.) used data from a network of rain gauge stations which were fixed and evenly distributed over the country (one raingauge station per district) covering a period of 108 years from 1871 to 1978 (this was later extended to 114 years from 1871 to 1984) but by excluding the hilly stations (where rainfall depended upon height and was of a different pattern from rest of the stations) from the existing network. The season of rainfall considered in these studies is from June to September. On statistical tests, they found the series of rainfall data used by them to be homogenous. Mooley and Parthasarathy (loc. cit.) used the following statistical criteria for describing the various parameters of the variability.

4.8.1.1 Statistical Criteria

$$\text{Mean or average } (\underline{x}) = \sum_{i=1}^{i=n} x_i/n \tag{4.8.1}$$

$$\text{Standard deviation (SD) } \sigma_x = \left\{ \sum_{i=1}^{i=n} (x_i - \underline{x})^2/(n-1) \right\}^{1/2} \tag{4.8.2}$$

$$\text{Co-efficient of variation (CV)} = \text{Standard deviation} \times 100/\text{Mean} = (\sigma_x/\underline{x}) \times 100 \quad (4.8.3)$$

$$\text{Inerannual variability} = \sum_{i=1}^{i=n-1} [x_{i+1} - x_i]/(n-1) \quad (4.8.4)$$

where x_i denotes the monsoon season rainfall for the ith year, $\underline{x}$ (x underlined) is the average rainfall of the total number of years n, and σ_x is the standard deviation of the monsoon rainfall.

Mooley and Parthasarathy describe the monsoon rainfall in terms of a standard unit, which is equal to deviation from mean divided by the standard deviation. Figure 4.17 shows the all-India summer monsoon rainfall in standard units for each year of the period, 1871–1984.

According to Mooley and Parthasarathy, a small deviation up to 5% on either side of the long-term average value can be considered as normal or average rainfall.

Some of the statistical properties of the all-India summer monsoon rainfall and its long-term variability as given by Mooley and Parthasarathy (loc. cit.) are given in Table 4.3.

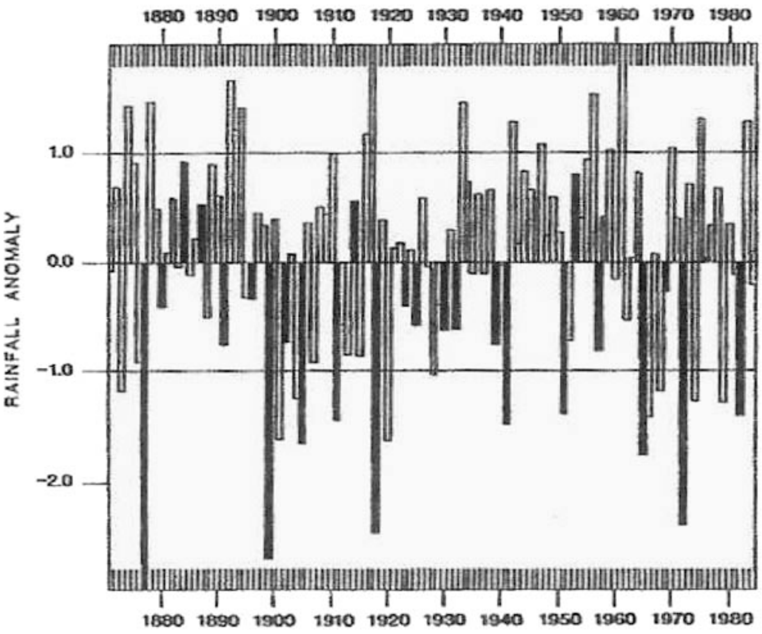

Fig. 4.17 All-India summer monsoon rainfall in standard units (deviation from normal divided by standard deviation) for each year during the period, 1871–1984 (after Mooley and Parthasarathy, 1984)

Table 4.3 Statistical properties of the all-India summer monsoon rainfall, 1871–1984

Property	Value	Property	Value
Mean	852 mm	Highest Rainfall	1017 mm
Mean/annual	78.1	Deviation from mean	19%
Median	864 mm	Range	413 mm
Lower quartile	800 mm	Coefficient of variation	9.7%
Upper quartile	908 mm	Mean interannual variation as % of mean	11.9
Standard deviation	83 mm		
Lowest rainfall (1877)			604 mm
Mean interannual variation in terms of standard deviation			1.22
Deviation from mean			–29%

4.8.1.2 Floods and Droughts

As should be expected, large excess (deficit) of rainfall in a year leads to large-scale floods (droughts) in India. Mooley and Parthasarathy (1983) from the results of their study lay down the following criteria for occurrence of these events:

$$\begin{array}{ll} \text{Flood if} & [(x_i - \underline{x})/\sigma_x] > 1.28, \\ \text{Drought if} & [(x_i - \underline{x})/\sigma_x] < 1.28 \end{array}$$

Using the above criteria, they worked out the years of large-scale droughts and floods in India during the period, 1871–1984, and identified the following years as those of worst flood and drought years in India:

Floods	1892, 1917, 1956, and 1961
Droughts	1877, 1899, 1918, and 1972

4.8.2 Factors Likely Responsible for Interannual Variability

Since the time of Blanford (1886), there has been frantic search on a global scale for parameters which likely influence the interannual variability of the Indian summer monsoon rainfall. Studies have pointed at several factors around the globe and even at extra-terrestrial ones, with varying degrees of correlation, but the relationship with most cases have turned out to be fragile and undependable. Only a few have passed the rigorous tests of confidence. To date, the following factors appear to be promising in this regard:

(a) Dates of onset and withdrawal of summer monsoon;
(b) Eurasian snow cover;
(c) Sea surface temperature, El Nino and Southern oscillation;
(d) Soil moisture, vegetation and albedo of the earth's surface

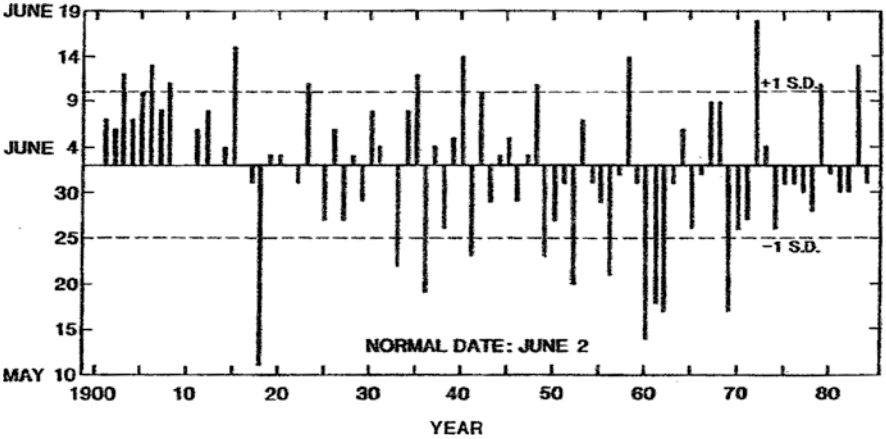

Fig. 4.18 Dates of onset of summer monsoon over southern Kerala, 1901–1984, showing the mean date and the limits of one standard deviation on either side of the mean (Mooley and Shukla, loc. cit.)

(a) Variability in Dates of Onset and Withdrawal of Monsoon: During a period of 84 years, 1901–1984, there was a large variability in the dates of onset of summer monsoon in Kerala, the southernmost state of the Indian peninsula, as may seen from Fig. 4.18 (Mooley and Shukla, loc. cit.).

According to Fig. 4.18, the long-term mean date of onset of summer monsoon over southern Kerala is 2 June with a standard deviation of 8 days. Most of the dates of onset lie within two standard deviations. The extreme dates are 11 May 1918 and 18 June 1972. However, as Mooley and Shukla remark, 'it is strange that both these extremes occurred in drought years. While a late onset in 1972 may be consistent with a drought in that year, the same cannot be said about an early arrival in 1918. This inconsistency only highlights the nature of the problem and the fact that an early or late arrival or departure of monsoon alone is not uniquely related to the overall behaviour of the monsoon during a season. Other factors may be involved in determining the observed variability of the rainfall.'

(b) Eurasian Snow Cover: Blanford (1884) was, perhaps, one of the first to point out that excessive snowfall in the Himalayas during the winter and spring was prejudicial to the subsequent monsoon rainfall over India. His observation was substantiated by Walker in 1910 and the inverse relationship was made use of in monsoon rainfall forecasts. However, on account of uncertainty in observations of snow cover, the use of this parameter was discontinued after 1950. The advent of earth-orbiting satellites which started observing the earth's snow cover changed the situation. Wiesnet and Matson (1976) on the basis of snow cover data furnished by satellites commented that 'the December snow cover for the northern hemisphere was a very good predictor of the following January–March snow cover'. Subsequent studies (e.g., Hahn and Shukla, 1976; Dickson, 1983, 1984) found an apparent inverse relationship of the Himalayan snow cover with the Indian summer monsoon rainfall.

In regard to the suspected inverse relationship between the two parameters, Mooley and Shukla (1987) observes: 'An inverse relationship between the Eurasian snow cover and summer monsoon rainfall is understandable, since large and persistent Eurasian snow cover would substantially reduce the rate of heating of the concerned land masses during spring and summer and thus delay the onset of the monsoon and prevent normal monsoon activity. It may be mentioned that the snow cover area as measured by the satellite is not quite a representative parameter to assess the amount of snow. The snow cover area in two cases may be the same but depths may be quite different, and the impact on monsoon would be quite different in the two cases'.

It may be hoped that with rapid development of satellite technology which would enable both horizontal and vertical extent of snow cover to be measured, a reliable correlation will be found between Eurasian snow cover and the Indian summer monsoon rainfall.

(c) Sea Surface Temperature, El Nino and Southern Oscillation: Need for an intensive study of the variation of sea surface temperature (SST) in the context of the Indian summer monsoon rainfall came to the fore after Bjerknes (1969) demonstrated the linkage between the ocean and the atmosphere and interpreted the Southern Oscillation and the Walker circulation in terms of air-sea interaction and El Nino and La Nina events in the equatorial Pacific Ocean. He found that in general a positive SST anomaly in the equatorial western Pacific was associated with heavy rainfall and a negative anomaly with deficient rainfall over India.

In a La Nina year, the equatorial eastern Pacific ocean is cold (SST anomaly negative), but the equatorial western Pacific including a part of the eastern Indian ocean, lying between about 70 and 160°E, is warm with a positive SST anomaly. With such a distribution of SST anomaly, the Southern oscillation index is positive and the Walker circulation has its ascending branch over the equatorial western Pacific and the descending branch over the equatorial eastern Pacific. The result is normal or heavy rainfall over the western Pacific including the India-Australia sector and drought condition over the eastern Pacific. The situation reverses in an El Nino year when the warm anomaly in the SST shifts to the equatorial eastern Pacific Ocean with a cold anomaly over the western Pacific. In such a year, the rainbelt shifts to the eastern side of the ocean and the India-Australia sector experiences drought conditions with abnormally deficient rainfall.

Following the investigations of Bjerknes (loc. cit.), there was a spurt in studies of Indian summer monsoon rainfall in relation to the zonal anomaly of SST in the Pacific Ocean. Several studies (e.g., Angell, 1981; Mooley and Parthasarathy, 1983, 1984; Sikka, 1980a; Rasmussen and Carpenter, 1983) were undertaken to find the degree of relationship of the Indian summer monsoon rainfall with the southern oscillation and the El Nino events by working out the correlation between them. Each of these studies used data for different years and often different parameters. Angell (1981) who used data for the period 1868–1977 obtained a correlation coefficient around –0.6 between all-India monsoon rainfall and SST anomaly over the equatorial eastern Pacific ocean (0–10°S, 90–180°W) one to two seasons later. The relationship is highly significant. Mooley and Parthasarathy (1983) who used a dataset for the years, 1871–1978, found a significant relationship between the monsoon rainfall over India and the El Nino events in the equatorial Pacific. They considered 22 moderate and severe El Nino events and found that in all the severe El Nino years, the all-India monsoon rainfall in standard units was less than –0.60, with the exception of 1884 when it was +0.92. However, on careful examination, they found that in that particular year, 20.4% of the country had, indeed, experienced drought conditions as would be expected in an El Nino year, but the deficiency was offset by heavy rainfall in the remaining parts of the country caused by other more influential factors.

In another investigation (Mooley and Parthasarathy, 1984) using the same dataset examined the relationship between the Indian monsoon rainfall and the SST anomaly in the Pacific in 3-monthly periods preceding or following the monsoon season and found inverse relationships that were significant at the 1% level for each of the JJA, SON and DJF seasons, and at the 5% level for the MAM season. They found the relationship to be consistent and stable. Sikka (1980a) used a different set of data. He used the Line Islands precipitation data as indicator of El Nino events in the Pacific and sought to relate them with the monsoon rainfall over India. He found a general association of El Nino events with deficient rainfall over India. Rasmusson and Carpenter (loc. cit.) found that in 25 El Nino years, the Indian monsoon rainfall was below the median rainfall in 21 years and below the mean in 19 years. They thought that the association had some predictive value.

Shukla and Paolino (1983) examined the relationship between the Indian monsoon rainfall and composites of normalized Darwin pressure anomalies (3-month running mean) for heavy monsoon rainfall years and deficient monsoon rainfall years during the period, 1901–1981, and found that the tendency of Darwin pressure anomaly before the monsoon season was a good indicator of the subsequent monsoon rainfall over India. According to them, a negative tendency between the DJF (December–February) season and the MAM season was found to be associated with good monsoon rainfall years and a positive tendency with poor monsoon rainfall years.

The results of their investigation are shown in Fig. 4.19. They expressed the view that whenever the tendency showed a large negative value, non-occurrence of drought in India could be predicted with a very high degree of confidence.

(d) Soil Moisture, Vegetation and Albedo of the Earth's Surface: There appears to be a symbiotic relationship between the albedo of the earth's surface on one hand and soil moisture, vegetation and rainfall on the other. The atmosphere over a region with high albedo tends to make up for the loss of solar radiation by large-scale subsidence which inhibits precipitation and perpetuates conditions which lead to high albedo. This is likely to happen particularly over the dry desert regions of the subtropics and the marginal lands where an increase of albedo caused by overgrazing or deforestation can lead to long-term reduction of precipitation, and continuation of dry desert conditions. On the other hand, a lowering of albedo by afforestation and vegetation can help to reduce subsidence and thereby promote more convection and precipitation to occur over the region. Afforestation and vegetation thus allowing long-term increased precipitation helps to increase soil moisture which in turn

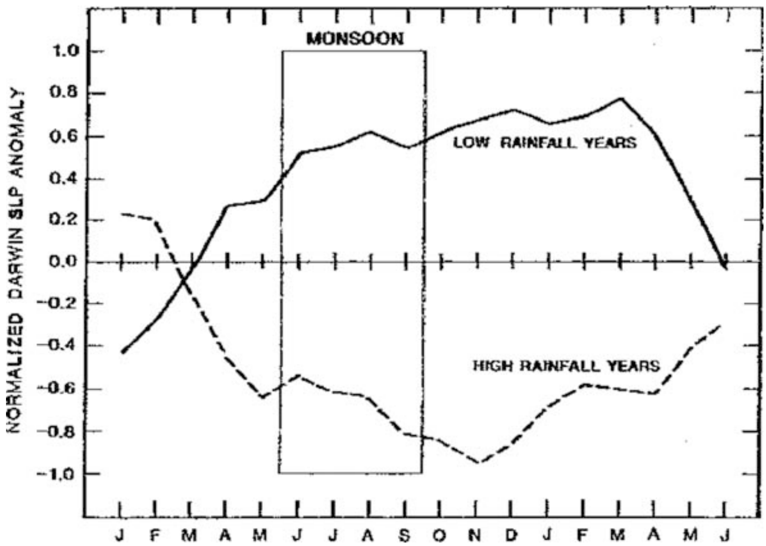

Fig. 4.19 Composite of the normalized Darwin pressure anomaly (3-month running mean) for years of heavy (good) monsoon rainfall and deficient (poor) monsoon rainfall (after Shukla and Paolino, 1983)

reduces albedo and increase rainfall. The validity of these cyclic processes has been well demonstrated by a series of numerical experiments using GCMs by Charney (1975); Charney et al., (1977); Shukla and Mintz (1982); Sud and Smith (1985) and several others.

Besides the global and regional factors mentioned above, attempt has been made to link the variability of the Indian summer monsoon rainfall with such extraterrestrial factors as solar activity as judged by the sunspot numbers. Some of the original studies in this direction were carried out by Gilbert Walker (1915a,b,c). However, the results of his studies showed no significant or consistent relationships with the mean annual rainfall over India.

4.8.3 Intraseasonal Variability

The variability of monsoon rainfall on time scales ranging from a few days to several weeks are caused by the familiar westerly and the easterly waves that move across the Indian longitudes and interact with the quasi-stationary monsoon wave, and also by low-frequency intraseasonal oscillations of the atmosphere during northern summer.

4.8.3.1 Variability on Scale of 3–7 Days – Active and Break Monsoons

IActive and break monsoon cycles occur frequently during the monsoon season. They have drawn the attention of meteorologists over a long time. Defining a

Table 4.4 Statistics of break in summer monsoons, 1888–1967

Month	No. of breaks	No. of break (days)	Average duration (days)	Longest break (days)	Most frequent duration (days)
July	53	306	5.8	17	4
August	55	356	6.5	20	3
1 July–31 August	5	47		21	

monsoon break as a disappearance of the monsoon trough over India from mean sea level and 850 hPa maps for at least a couple of days at a stretch, Ramamurty (1969) catalogued the breaks in July and August from 1888 to 1967. His statistics are presented in Table 4.4.

Ramamurty found from his long-period statistics that August is slightly more susceptible to 'break days' and longer breaks, particularly around the middle of the month. He also found that in the long record, there was no break in 12 years. Another important effect of the W'ly wave trough on the monsoon is a north/northeastward recurvature of the track of a monsoon depression or cyclone from its usual west/northwestward track if it happens to come under the influence of the W'ly trough.

4.8.3.2 Variability on 30–50 Day Time Scale

Atmospheric oscillations on this time-scale are known as Madden-Julian Oscillations (MJOs) after the name of their discoverers (Madden and Julian, 1971, 1972) and are believed to be excited by large-scale *tropical convection*. All subsequent studies (e.g., Murakami, 1976; Yasunari, 1979; Sikka and Gadgil, 1980; Murakami, 1984) have suggested a strong relationship of these oscillations with the slow meridional movement of a zonally-oriented low-level trough of low pressure associated with penetrative convection, cloudiness and heavy rainfall from the equatorial region to higher tropical latitudes in the Indian monsoon region Krishnamurti and Subrahmanyyam (1982) found that active and break monsoon cycles over the Indian longitudes were closely coupled to the meridional movement of these low-level troughs associated with rainfall. Oscillations on this time scale over the equatorial latitudes which moved slowly towards the pole have also been noted in the relative angular momentum of the earth's atmosphere from the datasets of the zonal wind by Rosen and Salstein (1981). Interestingly, they also found a strong signal on this time scale in the variations of the length of the day computed from lunar laser ranging observations.