Monsoon Monograph
(Volume 1)

Editors

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MINISTRY OF EARTH SCIENCES
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INDIA METEOROLOGICAL DEPARTMENT

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Monsoon Monograph Volume I

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एयर वाईस मार्शल (डॉ) अजित त्यागी
मौसम विज्ञान विभाग के महानिदेशक

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Director General of Meteorology &
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Preface

Monsoon is traditionally defined as a seasonally reversing wind system accompanied by seasonal changes in atmospheric circulation and precipitation. A semi-annual reversal in the wind direction, a typical characteristic of monsoon, is caused due to differential heating of continents and oceans and the Coriolis force. The term was first used in English in the then British India and neighbouring countries, to refer to the large-scale seasonal winds blowing from the Bay of Bengal and Arabian Sea from southwesterly direction bringing heavy rainfall to the area. The monsoon rainfall is considered to be that which occurs in any region that receives most of its annual rain during a particular season. The English word '*monsoon' has* its origin in the Arabic word *mausim* ("season"). The definition includes major wind systems that change direction seasonally and allows other regions of the world to qualify as monsoon regions

The unique physiographic features of south Asia, with the vast Asian continent spread over equatorial to polar latitudes in the Northern Hemisphere contrasted by the extensive water surface of the Indian Ocean, spread over the equatorial to Antarctic latitudes in the Southern Hemisphere, primarily supports the development of intense thermal centres of action due to differential heating of land and oceans. The resultant pressure patterns and the meridional circulations in summer and winter are further accentuated by the presence of high mountain massifs (Himalayan and Tibetan Plateau) of south Asia, leading to the establishment of the South-West (SW) and the North-East (NE) monsoon, respectively.

The importance of the monsoons for India is manifold. Out of these two, the SW monsoon is more important as it accounts for over 75% of the annual rainfall in most parts of India, outside Tamil Nadu and Jammu and Kashmir. The economy of India is substantially dependent on its agriculture, which, in spite of development of irrigation facilities, is primarily and largely, rain-fed. It is thus dependent on the quantum and distribution of rainfall during the SW monsoon season. As such, the failure of SW monsoon, adversely affects the agricultural production in India and, in turn, the Indian economy. The SW monsoon, which is the main source of rainfall for India, is characterised by a high variability, both on spatial and temporal scales. This variability is a major feature and reason of the dependency of Indian agricultural economy on the SW monsoon rainfall

Due to its great socio-economic importance and its challenge as a complex scientific problem impacting on the global scale, there had been extensive research work on Indian Summer Monsoon in India and abroad, for almost over four centuries. An exhaustive summary of the research work on Indian Summer Monsoon, particularly carried out in India, was documented by Dr. Y. P. Rao, in 1976, in the form of a Meteorological Monograph (No. 1/1976), entitled 'South West Monsoon', which served as a principal reference document for research workers and operational weather forecasters for more than past thirty years.

During past few decades, there have been new developments in the understanding of Indian summer monsoon, particularly in the light of availability of extensive data sets, new research methodologies including modelling and field campaigns. Also, the issue of global warming has raised

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several questions about monsoon circulation and thus, has added a new dimension to it. In view of these developments, India Meteorological Department (IMD), which is the nodal agency for national meteorological services for India, has decided to bring out a comprehensive publication in the form of **'Meteorological Monograph on Monsoons'**.

Considering its wide scope, this publication will be brought out in two volumes. This volume contains the chapters on scientific studies on monsoon, monsoons over other south Asian countries and elsewhere in the world, characteristics of Indian monsoon such as climatological aspects, onset; advance and withdrawal, components and semi-permanent systems of monsoon, operational procedures during monsoon as observed by IMD, observational aids used to monitor monsoon, orographic monsoon rainfall, extreme weather events related to monsoon, monsoon and agriculture, and northeast monsoon. The first volume also contains a chapter on Indian summer monsoon experiments.

The second volume contains chapters on synoptic systems during monsoon season, monsoon variability on different temporal scales, tele-connections of monsoon, monsoon oceanic aspects, monsoon simulation, predictability of monsoon using coupled general circulation models, forecasting of monsoon in short; medium; extended range and long range time scales, and aspects of monsoon in relation to climate change. The second volume also includes a chapter on modeling of forecast sensitivity of progress of monsoon northward from southern tip of India during the onset and advance process.

All the contributors of this publication are eminent and experienced meteorologists from leading organizations in India and abroad of national and international repute. With its wide ranging scope and contents, this publication is intended to serve as a comprehensive reference publication for both, operational meteorologists and meteorological research scientists.

I take this opportunity to thank all the authors from India as well as from outside India including the authors from south Asian countries for their valuable contributions in making the monograph a comprehensive reference publication on monsoon over south Asian countries. I also thank other members of the editorial board viz., Prof. G. C. Asnani, Dr. U. S. De, Dr. H. R. Hatwar and Dr. A. B. Mazumadar for judiciously editing the publication. I also thank Dr. Medha Khole, DDGM (WF) and Dr. D. R. Pattanaik, IMD New Delhi in coordinating the work and bringing out the publications in time. Also I would like to thank Shri S. B. Gaonkar, and staff members of section of I&D of O/o DDGM (WF); Mr. Yogesh Visale and the staff members of the DTP unit of ADGM (R) Office, Pune for page setting and printing of the document.

New Delhi

Dated

(Ajit Tyagi)

Director General of Meteorology
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CHAPTER 1

INTRODUCTION
SCIENTIFIC STUDIES ON THE SOUTH WEST MONSOON – PAST, PRESENT AND PROSPECTS

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1.1. Background and Past Pioneering Studies (1875-1947)

South West Monsoon Season, the harbinger of the life giving summer rains over South Asian countries of Bangladesh, Bhutan, India, Pakistan, Maldives, Nepal and Sri Lanka, is the presiding deity for the economic wellbeing, based on agricultural productivity, of the vast population over this region. Monsoon rains, since time immemorial have been the bedrock of India's food security and of the very survival of land and its people. As such India celebrates the life giving monsoon rains year after year. Knowledge about the reversals of winds over the Arabian Sea had become available at the turn of the Christian era and was utilized by ancient mariners for trade between west Asia and India for many centuries. Indian people, being land-locked, studied the rainy season of India without its connections with the wind system till about the 17 century. However, the links between the seasonal reversals of winds over the north Indian Ocean, their cross-equatorial sweep and relationship with the beginning of the summer rains over South Asia began to be investigated with the classical work of Halley (1686), based on the analysis of data observed during transits of ships over the vast Indian Ocean. Scientific studies about the weather and climate of India, in somewhat unorganized manner, were initiated since early 19th Century, by the enlightened employees of the East India Company when they began observing surface meteorological elements by using instruments at places of their postings and kept records. IMD (1975) and Sikka (2009 a) have provided some accounts of those early observations. Two major

discoveries came out of these investigations. Firstly, the existence of diurnal variability of pressure over massive scale of south Asia and secondly, the structure of the tropical cyclones in the north Indian Ocean. The latter was though the work of Piddngton in mid 19th Century, published in the Journal of The Royal Asiatic Society, Calcutta and in his book "Hornbook of Tropical Cyclones". Piddignton is credited with the coining of word 'Cyclone' from the Greek Word Cyclos (Coils of a Snake).

With the establishment of the India Meteorological Department (IMD) in 1875, scientific studies on the South West (Summer) and North East (Winter) monsoons of India were systematized and for the next 50 years or so they were based on surface meteorological data, recorded under a uniform system of instrumental observations and meticulously analysed by the pioneers of South Asian Meteorology – H.F. Blanford, John Eliot, W.L. Dallas, G.T. Walker, G.C. Simpson, J.H. Field and C.W.B. Normand and others. As upper air winds, temperature and moisture data began to be collected from 1905 to 1945, vertical structure of the troposphere over India was brought to light by the efforts of J.H. Field, C.W.B. Normand and early Indian officers who had joined IMD between 1920 to 1935 – prominent among them being S.K. Banerjee, S.C. Roy, V.V. Sohni, K.R. Ramanathan, S. Basu, B.N. Desai, S.N. Sur, S. Mull, A.K. Roy, G. Chatterjee and several others who are recognized as the stalwarts of SW Monsoon studies from India. Based on the surface data of nearly 20 years, Blanford (1870, 1886) authored two classical works – one the Indian Meteorologist's Vade Macum and the other on the Rainfall of India. Blanford is regarded by many as the most prominent tropical meteorologist of that era. In his studies (Blanford 1886), emphasized the prominence of seasonal surface heat low over now Pakistan region and the monsoon trough (MT), stretching from undivided NW India to North Bay of Bengal, in modulating processes in the SW Monsoon and regulating rainfall over South Asia. He, for the first time, showed connections of spatio-temporal variability in the position and strength of MT with episodes of excess (active monsoon) and lulls (break monsoon) in rains – the so called 'active-break' cycle of the rains on the intra-seasonal scale, which is responsible for the variability of monsoon both on intra-seasonal (IS) and inter-annual (IA) scales. During active monsoon conditions MT lies south of its normal position and cyclonic whirls form all across the trough (though mostly over North Bay of Bengal). They move across its length and breadth, spreading rains along their tracks. The near-surface winds are also strong in active monsoon conditions from central Arabian Sea to central Bay of

Bengal across the Peninsular India. During the break monsoon the MT hugs the sub-Himalayan region or even disappears north of it and cyclonic whirls more or less remain absent over central parts of India and across the Indo-Gangetic Plain and rains also remain suppressed over most of India except the foot-hills of Himalayas. Blanford (1884) had established the links between the IA performance of summer monsoon rains with the winter and spring season accumulation of snowfall over the Himalayas. According to his hypothesis years of excess (deficit) snowfall come associated with drought (excess rains) during the ensuing summer monsoon season. Several other facets of Indian weather and climate were also studied by Blanford. Between 1889 to 1903, John Eliot, who succeeded Blanford, and was designated as Director-General of observatories, was able to demonstrate that the onset of monsoons rains across mainland India was an important singularity on the annual cycle and was accompanied with the sudden rush of strong winds across near-equatorial North Indian Ocean (establishment of cross-equatorial gyre) and their abrupt strengthening over SW to SE Arabian Sea. He also showed that the strengthening of winds over the region was often accompanied with the formation of a cyclonic vortex on the leading edge of the rushing SW winds. Eliot (1884) had also found that during the peak monsoon months of July and August cyclonic vortices with intensity not as strong as true cyclonic storms (he called them minor cyclones of the SW monsoon season – now known as monsoon depressions or monsoon low pressure areas), formed over the north Bay of Bengal and entered Orissa-West Bengal coast with interval of 3 to 10 days. These low pressure systems (LPSs) distribute rains along their path within the MT on synoptic scale. Most of them disappear over land before they reach the western end of MT or merge with the seasonal low on reaching it. Under Eliot IMD, in 1897, organised a major effort of collecting surface meteorological data on near-real time basis from tropical south Indian Ocean across the equator into the north Indian Ocean. This could be called the very first Monsoon Experiment. This effort launched the operational preparation of Indian Monsoon Area synoptic charts in the IMD. One can say that the surface phenomenology of SW Monsoon and the role played by synoptic scale LPSs had been well explored by the pioneering work of Blanford and Eliot. Eliot had also prepared and supervised the publication of the first Climatological Atlas of India which was published in London in 1907. G.T. Walker, who took charge of the IMD in 1904, after a brilliant scholastic career in Mathematics at Cambridge University,

took upon himself the challenge of long-range forecasting (LRF) of the SW Monsoon rains on the scales of seasonal and large aerial portions of India. He believed that the seasonal rainfall over vast Indian area must have some precursors from other regions of the globe and adopted statistical correlation as the methodology for diagnosing these relationships (tele-connections) of the global weather systems. He spent nearly 20 years in this pursuit in India and for another 20 years on return to England (Walker and Bliss 1932) with rare passion. His efforts were of course helped by the global sweep of the British Empire for access to meteorological and allied data. He discovered the famous three-pressure oscillation, the Southern Oscillations (S.O), the North Atlantic Oscillation and the North Pacific Oscillation. He wrote many papers, using the labour of human computers in IMD, and finally established the system of LRF of monsoon in India based on linear multiple statistical correlation methodology. He also encouraged work on the development of upper air sonding systems, and introduction of instrumented kites and pilot balloon observations under the supervision of his young colleague Mr. J.H. Field. He also helped in initiating scientific investigations on the atmospheric electricity in India under his other colleague Mr. G.C. Simpson. He was handicapped by lack of scientific manpower during the First World War Years (1914-1918) and foreseeing the problem of getting meteorologists from England (as the U.K. Meteorological office was also growing in those years), he succeeded in convincing the Government of India for inducting capable Indians into IMD to hold positions as meteorologists. Only two Indians had worked as officer's between1880-1918, the first being Lalla Ruchi Ram Sahni for a short period of 2 years under Blanford and the second being Lalla HemRaj under Eliot and Walker. Walker, between1918-1924, inducted several young Indians into the service of IMD. The prominent among them were Prof. S.K. Banerjee, Dr. S.C. Roy and Prof. K.R. Ramanathan who came from academic background of college / university teaching. There was a time when the heads of the IMD (Walker himself), the U.K. Meteorological Office (G.C. Simpson) and the Canadian Meteorological Office (J. Patterssen) all had the work experience in India. Walker also contributed immensely for the growth of Indian science and served as President of an Indian Science Congress Session and as Member on the first Governing Council of the Indian Institute of Science (IISC), Bangalore. During 1875 to 1924, the Headquarters of the IMD functioned from Calcutta (now Kolkatta) and Simla (now Shimla), but Walker had proposed that it should be established in Poona

(now Pune). The grand building of the Pune Meteorological office was inaugurated in 1926 under the regime of Walker's successor, J.H. Field. The Headquarters was finally shifted to its present campus at Lodi Road, New Delhi in 1945 to facilitate the Second World War efforts for better coordination.

Field, who took over from Walker in 1924, was responsible for the introduction of upper air sondings by meteosonde ascents and pilot balloon ascents in India and had also done pioneering studies on the upper air structure of SW monsoon winds, temperature and moisture fields. He lead the efforts on the introduction of Aviation Meteorology and Agricultural Meteorology in India. C.W.B. Normand, who succeeded J.H. Field in 1928, was at the helm of affairs of IMD till 1945. He did remarkable work on the thermodynamics of SW Monsoon and importance of latent instability and convective instability to pre-monsoon thunderstorms over India. He had emphasized the Convective Instability of the First Kind in local convective processes but had missed the vital link between local scale convection and large scale convection on the synoptic scale which were later established by Charney and Eliassen (1964), under the concept of the Conditional Instability of Second Kind (CISK). In between Simpson, in his classical paper published in the Quarterly Journal of the Royal Meteorological Society in 1920, had even questioned the grand-scale sea-breeze hypothesis of SW Monsoon by mentioning that there was no monsoon rains in India when the land-sea contrast was at peak in summer (month of May) and also the rains disappeared in prolonged break monsoon spells when the landmass of South Asia became quite warm on the IS scale. Simpson propounded the box-like trapping of the moisture-laden winds within the geographical topographical barriers of India and the role moist processes (release of latent heat) in maintaining and modulating monsoon rains. Later, S.K. Banerjee, in 1930, produced the first mathematical model of the regional SW Monsoon and theoretically simulated the structure of surface pressure based on topographical features of South Asia. This was about a decade earlier than the discovery of conservation of absolute vorticity by Rossby and the importance of rotation (vorticity) in atmospheric processes. K.R. Ramanathan, who had joined IMD in 1928, did pioneering work on the structure of the tropical tropopause and the upper air over India, monsoon depressions and tropical cyclones etc. S. Mull did important work on formation of thunderstorms and cloud physics. Similarly, other researchers like B.N. Desai,

G. Chatterjee, S.N. Sur, A. K. Roy advanced monsoon research in the years between 1928 and 1945. During the Second World War (1940-1945) there was massive expansion of IMD. Many young officers joined IMD during 1930 – 1950, the prominent among them being P.R. Krishna Rao, R. Ananthakrishnan, L.S. Mothur, K.N. Rao, S.P. Venkateshwaran, P.R. Pisharoty, P. Koteswaram, Y.P. Rao, P.K. Das, Anna Mani and several others who contributed immensely to the development of meteorology in India in the post-Independent era and are also included among the stalwarts of Indian Meteorology during 1947-1980.

Scientific literature on monsoon meteorology began to grow from 1957 onward, primarily under the stimulus of International field campaigns of IGY (1957-58), IIOE (1963-66), ISMEX-1973, Monsoon-77, Summer MONEX - 1979, TOGA (1985-95) and the national field campaigns since 1988 under the MONTBLEX (1989-90), LASPEX (1995-96), BOBMEX (1998-1999), ARMEX (2003-2005). International atmosphere-ocean science community also began to take active interest in investigations on the SW Monsoon, since 1960s, and has advanced the subject vastly. Several excellent books, monographs and review articles have appeared on the monsoon studies detailing investigations on the different facets of the SW Monsoon, the prominent among them being by Ramage (1971), Rao (1976), Fein and Stephens (1987), Chang and Krishnamurti (1987), Keshvamurty and Shankar Rao (1992), Das (1986), Lighthill and Pearce (1981), Webster et al (1995), Pant and Rupakumar (1997), Asnani (1993, 2005), Gadgil (2003), Goswami et al (2006), Ding and Sikka (2006), Goswami (2005) and others. Y.P. Rao's monograph, which was written some 35 years ago, had produced comprehensive account of observational knowledge on the SW Monsoon accumulated between 1875-1974, based largely on vast amount of synoptic studies undertaken in the IMD during 1960s under their project Forecasting Manuals. This was an important monograph highlighting the knowledge on monsoon studies till 1970. However, in the last 3-4 decades investigations on the SW Monsoon have advanced remarkably, covering synoptic aspects but also monsoon diagnostics, monsoon prediction on short-medium-and-extended-range (including long-range) scales, monsoon simulations and dynamical long-range monsoon prediction and monsoon climate change etc. It is heartening that IMD has initiated the present studies in which articles are being solicited from specialists in different aspects of monsoon. The present article is aimed to provide

the salient aspects of progress made in the understanding and prediction of the SW Monsoon in the post-Independent era (1947-2008) of India and assess the contemporary scene as well as future prospects. Section 1.2 summarises prominent discovery-oriented research during 1947 – 2008. Section 3 gives the contemporary status and possible future projections based on international developments and the Indian national efforts.

1.2. Salient Aspects of Major Investigations on SW Monsoon during 1947 – 2008

1.2.1. Increasing Research Infrastructures and Facilities in India

From the establishment of IMD in 1875 to the dawn of free India in 1947, the following aspects of the SW Monsoon were either well known or on the horizon as a result of investigations primarily done by researchers in IMD by the pioneers and stalwarts of Indian Meteorology with the use of surface meteorological and emerging upper air data from meteosonde balloons and pilot balloons:

- MT and its importance,
- Strong winds in the lower troposphere over Peninsular India and possibility of reversal of monsoon westerlies in the mid-troposphere to broad easterly flow in the upper troposphere over India,
- Vertical extent of the monsoon over Peninsular India to about 600 to 500 hpa levels,
- Existence of high moisture content in the westerly monsoon flow upto mid-troposphere under active monsoon conditions and relative dryness of the mid-troposphere in ‘break’ monsoon conditions,
- Synoptic scale monsoon LPSs (depressions and low pressure areas), their broad westerly to northwesterly tracks along the MT and their role in distribution and modulation of monsoon rains,
- Active – break cycle of the monsoon on IS scale and the precursors for onset of ‘break monsoon’ and its revival on synoptic scale,
- Role of orographic features of India in trapping moisture within the country,
- Role of latent heat release in maintenance of the monsoon system rather than the land-ocean contrast,

- Possible role of land surface features, such as distance from the sea, rivers, valleys, lakes, hills and vegetative cover in distribution of rainfall across the country on climatological basis,
- Quasi-periodicity in the occurrence of droughts over different part of India varying between 3-4 years over semi-arid to arid regions and 8-10 years over hyper-moist region of NE India,
- Connection of Southern Oscillation negative phase with the monsoon droughts over India,
- Role of excessive snowfall over Himalayas in winter and spring seasons to ensuing monsoon season droughts and
- Statistical LRF of monsoon seasonal rains over India and empirical-based (sympotic-climatological) approach to short-range (1-3 days) monsoon forecasting.

With India's freedom, the process of all round development of the country began under the Five Year Plans, launched by the Government of India since 1950. Indian Meteorology has also gained from this thrust and India started playing a much more vital role in international meteorology under the aegis of World Meteorological Organisation (WMO). From IGY (1957-58) onward India became a partner and even a host to several international monsoon-related programs. Observational infrastructures such as radiosonde, rawins, radars, satellite coverage and Ocean Observing System were also progressively introduced in the Indian observational networks between 1950s between 2005. Tele-connection links within India were improved and India (New Delhi) became a hub on the Global Tele-communication System by 1970 for global meteorological data exchange under the WMO's arrangements. Electronic computer was first introduced for data processing in 1964 and Indian meteorologists began to use computers for scientific research and even Numerical Weather Prediction (NWP) since 1970s. Indian meteorologists got several opportunities to be trained in centers of excellence in advanced countries of Europe, USA, Canada, Australia and Japan and also collaborated in their research programmes as exchange or invited visitors. Formal training courses at advanced level were introduced in the Meteorological Training School at IMD, Pune by 1950s. Till 1950, IMD was the only organisation in India where meteorological training and

research was practiced. Indian Air Force established their separate Meteorology Branch in 1956. From 1950 onward universities began to introduce Master Degree level courses and Ph.d research programs. Several new centers like the Physical Research Laboratory, Ahmedabad in 1950, the Indian Institute of Tropical Meteorology, Pune in 1962, National Institute of Oceanography, Goa in mid-1960s, Space Application Centre and Space Physics Laboratory, Trivandrum under the Indian Space Research Organisation (ISRO) in 1970s, Mesosphere-stratosphere-Tropsphere Radar Centre (now known as National Atmospheric Research Laboratory) at Gadanki in 1980s, Centre for Atmospheric Research at IIT, Delhi and centre for Atmospheric-Ocean studies at Indian Institute of Science, Bangalore in 1980s, National Centre for Medium-Range Weather Forecasting (NCMRWF) at New Delhi (now in NOIDA) in 1988, Centre for Mathematical Modeling and Advanced Computing (CMMAC) under CSIR in Bangalore in 1990s, Indian National Centre for Ocean Information Services, Hyderabad in 1999 and Centre for Land-Ocean-Rivers at IIT Kharagpur, in 2006 etc. Other Government departments and ministries like DST, DOD, ISRO, CSIR, and MoES also began to support research in atmospheric sciences by special funding arrangement since 1980s onward. All these organizational expansions for atmospheric-ocean science have lead to overall tremendous growth in research in India on observational, diagnostic, modeling and theoretical aspects of the South West Monsoon in particular. Scientific Journals like the Indian Journal Meteorology and Geophysics (now Mausam) was introduced in 1950. Other journals followed, like the Proceedings of the Indian Academy of Science (Earth and Planetary Sciences, (now Journal of Earth System Science), Indian Journal of Space Radio and Physics, Journal of the Indian Geophysical Union, Vayu Mandal (the Bulletin of the Indian Meteorological Society). Mahasagar, Current Science, Vatavaran (the journal of the Indian Air Force Meteorological Branch), Rainbow (Journal of Indian Navy's Naval Oceanography and Meteorology Branch) Proc. Ind. Acad. Sci., Indian Science Academy, Annals of Arid Zones Research, Journal of Agrometeorology and some others began publishing research papers in monsoon-related subjects and other disciplines of atmospheric-ocean science in India. Indian researchers have begun to publish their researches in increasing numbers, since 1980s, in prestigious international journals like Nature, Science, Journal of Atmospheric Sciences, Monthly Weather Review, Tellus, Bulletin of the American Meteorological Society, Quarterly Journal of Royal Meteorological

Society, Pure and Applied Geophysics, Journal of Climate, International Journal of Climatology, Climate Change, Geophysical Research Letters, Journal of Geophysical Research and others. This has given on international visibility to research outputs of the Indian scientists. Besides research on the SW Monsoon became a hot topic among international monsoon research community and many researchers of repute in advanced countries have devoted their studies to unravel several phenomenological aspects on monsoon, its prediction on different temporal scales, simulation of monsoon with dynamical models, process studies on monsoon, monsoon climate change and other topics. Indian Meteorological Society was formed in 1956. Several Indian atmospheric-ocean scientists have been elected as Fellows of the Indian Academy of Science, Indian National Science Academy, National Academy of Sciences, Indian Meteorological Society etc.

Scientific conferences / symposia were introduced in IMD only in 1940s. They began to be organized more frequently since 1950s and new several such events are organized at national level every year in India. IMD at New Delhi organized the first major international symposium on Monsoons of the World in 1957. IITM organized its first international symposium on Tropical Monsoons at Pune in 1976. At present India increasingly plays host to international events too. The combined research output on the SW Monsoon has grown phenomenally since 1960s and it would be hard to review the present knowledge in detail. Therefore, we attempt only the salient or the most important aspects in this article. Although a large number of references to work done by Indian and foreign atmosphere-ocean scientists are quoted, many more could not be included but are available in references quoted in this article. Scientific literature on the study of the South West Monsoon is so vast that it is a stupendous task for anyone to provide an exhaustive bibliography since scientific studies began in the 17th Century. Other specialized papers quoted in this monograph deal with details of the developments in their respective areas.

1.2.2. Important Investigations on the Southwest Monsoon

1.2.2.1. Phenomenological Studies

a) Large-scale aspects of the SW Monsoon

Several workers in India as well as outside India have added new large-scale components of the SW Monsoon in different years since 1950. The prominent amongst them are

- Tibetan High in the upper troposphere (Flohn 1960, Koteswaram 1958b, Yanai and Song 1992, Wu et al 2004)
- Upper Tropospheric Tropical Easterly Jet Stream (Koteswaram 1958a)
- Global Scale East-West divergent circulation in the upper troposphere (Krishnamurti 1971) with wave number two structure with two divergent centres over Tibetan and Mexican highs and two convergent centres over the mid-Atlantic and mid Pacific troughs
- Low Level Jet (LLJ) off Somalia Coast (Findlatter 1969) and its extension over Peninsular India (Joseph & Raman 1966) which is the artery for moisture supply for monsoon in the lower troposphere. LLJ is also found to fluctuate on intra-seasonal and IA scales (Joseph and Sijikumar 2004)
- Maintenance of mean monsoon trough over India (Keshvamurthy 1968, Keshvamurthy and Awade 1970), heat and moisture budget of the monsoon trough (Anjaneylu 1969) and dynamics of tropospheric circulation during contrasting monsoon seasons (Kanamistu and Krishnamurti 1978, Awade et al 1984 and others)
- Marcarene High (Krishnamurti and Bhalme 1976)
- Presence of temperature inversions in the lower troposphere over the Arabian Sea, which were, for the first time, discovered from the aircraft reconnaissance during the IIOE period (Colon 1964). They were also studied with SMONEX aircraft data. Recently, during ARMEX – 2002, their role for widespread subsidence over the Arabian Sea and even inside northern and central India have been discussed by Bhat (2006) and Rao & Sikka (2005)

- Role of double equatorial trough, the so called northern hemisphere near equatorial trough and the southern hemisphere near equatorial trough over the Indian Ocean (Raman and Dixit 1964)
- Fluctuations on IS and IA scales of each one of the components of monsoon over western end of MT have been investigated in many studies and shown to be related to monsoon variability.
- Energetics of the northern hemisphere and summer circulation over the tropics and regional South West Monsoon regions were diagnosed in several studies during 1970s to 1990s (Murakami 1978, Kanamitsu and Krishnamurti 1978, Mohanty et al 1983, 2005, Awade et al, Ramesh et al 1984, and Rao et al 1999). These studies brought out the interactions between the planetary-scale waves, medium-scale waves and short-scale waves in the maintenance of circulations and differences in good and deficit monsoon years.
- Heat sources and moisture sinks fluctuations (Bhide et al 1998)

b) Synoptic and Meso-Scale Disturbances

Rahmatullah (1952) investigated the synoptic aspects of monsoon circulation over Indo-Pakistan region, which have remained under further investigations since then. Besides LPSs, already described, a major new synoptic system, known as Miller and Keshvamurty (1968) discovered the Mid-Tropospheric Cyclone (MTC) as a result of IIOE aircraft and conventional data analyses. Mak (1975, 1987) and Borde and Mak (1978) applied linear and non-linear theory to explain the instability of MTC. Carr (1977) and others have studied MTC and their scope to understand MTC processes from observational studies and the present day operational medium-range dynamical models. Rajeevan et al (2000), using satellite data, examined the asymmetry in the monsoon depression. Ding and Sikka (2005) have discussed in some details different aspects of synoptic scale systems. Meso-scale variability in the monsoon rainfall within the synoptic scale, although broadly known earlier, was much emphasized by different workers (Venkataraman et al 1974) has been recently reviewed by Sikka (2008a). Moolely & Shukla (1987), Sikka (2005), Jadhav (2004), Dash et al (2004) and others have shown the climatological aspects of LPSs, their quasi-periodicity, impact of their frequency distribution and the relationship between

frequency of LPS days and droughts, which was first discussed by Sikka (1980). Monsoon depression, traditionally being an important synoptic scale system of the SW Monsoon circulation has been extensively studied. Beginning with the work of Pisharoty and Asnani (1957), which emphasized the organized moist ascent in the SW sector of the depression, several attempts have been made since then to study dynamical structure of depressions. Study of monsoon depressions was an important component of the SMONEX. Different workers prior to and subsequent to SMONEX (Krishnamurti et al 1975, 1976, 1977, Daggupathy and Sikka 1975, Godbole 1977, Keshvamurty et al 1978 undertook several studies. Nitta and Masuda 1987, Saha et al 1981, Sander 1984, Warner 1984, Saha and Chang 1983, Ding et al 1984, Koteswaram et al 1987, Sarkar and Choudhary 1988, Rajamani and Sikdar 1989 and others) to diagnose structure and vertial motion fields associated with depressions. Some of these researchers used aircraft data (including air-borne radar) to understand organized convection of a depression, which had formed in the North Bay of Bengal during SMONEX in the first week of July 1979. Also researchers applied dynamical theories to understand the dynamical instability of the monsoon flow for formation of monsoon depression (Shukla 1978, Mishra and Salvekar 1980 and others).

1.2.2.2. Monson onset and advance

Onset of monsoon over the mainland of India, (Kerala Coast) known as MOK has been much studied since the first work of Ananthakrishnan et al (1967). Regular and false onset of monsoon has been investiaged (Fazallo & Webster 2003). Monsoon onset is also connected with the 30-50 day oscillation (Joseph et al 1994, 2006). Formation of monsoon onset vorex (Krishnamurti et al 1981) in the Arabian Sea was investigated with SMONEX-1979 aircraft data. Models have been used to understand its dynamics (Krishnamurti and Ramanathan 1982) and dynamical instability associated with it was investigated (Mishra et al 1985). Pearce and Mohanty (1984) showed that the major precursor for monsoon onset is the build-up of moisture up to mid-troposphere about 2 weeks in advance of MOK. Because of the importance of MOK in the annual calendar, many observational investigations have been undertaken in India as well as by the international scientific community to unravel features and variability of MOK. Atmospheric general circulation models

have been also used to diagnose MOK and its IA variability (Ratnam et al 2005 and Sikka et al 2010).

Most of these indicators or precursors related to rainfall, strengthening of winds and seasonal reversal in temperature distribution between southern India and Tibetan region are dynamically related to each other. These changes occur in a slow manner within upper troposphere, mid-troposphere and lower troposphere at different rates beginning from mid-March to get fully set or organized by mid-May when the big event of Monsoon Onset Over Kerala (MOK) is about to take place (Ananthakrishnan 1977). Several investigators have suggested objective criteria for MOK. Recently Goswami and Xavier (2005) have emphasized the change in mid / upper tropospheric meridional temperature gradient as an important parameter for objectively defining MOK. There are other studies on this subject too (Ramesh et al 1996, Joseph et al 2006 and others). Pasch (1983) introduced global onset of summer monsoon and Zhang and Lu (2004) have advocated globally unified monsoon onset and retreat indices. Even energetic of monsoon onset were examined by George and Mishra (1993). However, from the point of MOK, formation of a synoptic scale disturbance (low pressure area, onset vortex, mid-tropospheric vortex or trough along 8-10^0 N between 70-90^0 E) brings about the burst of rains over Kerala for which the operational meteorologists in India wait to declare that the MOK has indeed taken place. This is quite needed because the public perception of MOK, traditionally, is based on the jump in rainfall and onset of strong winds. As such IMD has recently developed multi-parameters criteria which include jump in rainfall over Kerala Coast, satellite observed cloud band, strengthening of winds at 925 hpa within 5-10^0 N and 70-75^0 E, to define the MOK. There are several other indices proposed by different authors to objectively define the MOK from different parameters. However, most of the years objective indices-based MOK and the IMD's operational MOK date agree within 2-3 days of each other. The mean date of MOK is 1 June with a standard deviation of one week.

Understanding of processes (such as burst of rainfall over Kerala, enhanced moistening of lower and middle troposphere over Peninsular India / West Coast, building up of the meridional temperature gradients in the middle and upper troposphere, appearance of the tropical easterly jet over southern India, shift in the

large scale trough patterns in mid-tropspherical global scale and disappearance (Yin 1949) of sub-tropical (westerly jet stream from India etc. and several others are interlinked dynamically and take place within about a week of each other and coinciding with the operational declaration of MOK.

Slingo et al (1988) investigated the role of physical parametrization schemes in predicting monsoon onset by dynamical models. The high-resolution model being used currently in NCMRWF has shown the capability to predict monsoon onset within about 2-3 days of the observed one. Satellite photographs and OLR data have also been used for monsoon onset in several studies (Lau and Chan 1986, Simon et al 2006 and others).

SW Monsoon rains begin to advance northward after the MOK has taken place. The advance process is not smooth but is punctuated by fast propagation for 3-5 days and then stops for a week or so to progress in steps along the West Coast, Peninsular India, Central India, NE India, Indo-Gangetic plain and finally reaching the borders of West Rajasthan by 15 July. On the average it takes nearly 45 days for the advance process to be completed with extremes lying between 30 days to 70 days (Ananthakrishnan and Soman 1980, Soman & Krishnakumar 1989, Li an Yanai 1996 and several others). Biswas et al (1998) have found that the hiastus or temporary cessation in the monsoon advance occurs due to increased activity of the mid-tropospheric westerly trough along 65-80^0 E and temporary cessation of the formation of synoptic scale disturbances over the Bay of Bengal. Flatau et al (2002) studied the delayed monsoon onset in 2002 season and Prasad and Hayashi (2005) have also examined the onset and withdrawal phases of monsoon over India. There is need to further investigate the causes for cessation in the advance process and determine the credibility of dynamical models to predict this cessation and its revival on the medium-range scale.

The performance of the seasonal rains over India, however, do not depend on MOK (Dhar et al 1980) but on the period taken for the monsoon from MOK to its full advance reaching upto west Rajasthan. Thus researchers may also focus their efforts to find whether the advance period can be forecast on long-range scale. All said and done the key role of formation and propagation of synoptic scale

disturbances in the onset and advance processes of monsoon is dominant and hence its diagnosis through dynamical models would provide guidance on medium-range scale. On the long-range scale, it is also worthwhile to examine the inter-annual variability in the period of monsoon advance based on multi-decadal runs of AGCMs or CGCMs available with different modeling centers in India and in other countries, using ensemble runs, to determine their utility to forecast the duration of monsoon advance just like these models are used to forecast monsoon seasonal rainfall. An attempt to understand this aspect of variability in the period of monsoon advance is made by Sikka et al (2010)

1.2.2.3. Monson Variability: Intra-Seasonal and Inter-Annual

The topic of IS and IA scale variability, earlier known as vagaries of monsoon or viccitudes of monsoon, has been much pursued by several researchers since the early period of monsoon meteorology (Alexandar et al 1978 and others). The prominent new facet since 1980s has been the northward movement of the monsoon cloud zone from near-equatorial Indian Ocean belt upto Himalayan foot hills at about 1^0 lat/day on 30-50 day scale. This is linked with the 'active-break cycle' of monsoon (Sikka and Gadgil 1980, Yasunari 1980). Several workers have done extensive work since then on this topic. Another IS oscillation of monsoon is brought about by biweekly oscillation first introduced by Krishnamurti and Ardanay (1980) and since then discussed by others (Chatterjee and Goswami 2004). The processes responsible for such oscillation on low frequency IS scale are yet to be fully known and modeled, though hydrological feedbacks and air-sea interactions may play crucial roles according to some authors already referred to. Dominant modes of monsoon variability amongst the IS & IA scales have been shown to have much resemblance to each other (Goswami and Ajay Mohan 2000, Webster et al 1995 and others) Goswami et al (2006) have extensively reviewed this aspect. Study of IS & IA variability of the monsoon from observational viewpoint has attracted many investigations. (Sikka et al 1987, Goswami et al 2003, Goswami 2005, Gadgil 2003 and others). AGCM-based studies to diagnose ARE & IA variability of the monsoon have been also performed by several investigators. (Ferranti et al 1997, Sperber et al 2000, Waliser et al 2000, 2003a, b, 2006 a, b, Goswami et al 2004, 2006, Bellenger et al 2007, Agudelo et al 2006). Thus huge information-base is available

on monsoon variability but the subject will continue to be pursued because of its importance. IA variability of monsoon controls the incidence of drought and excess monsoon seasons (Shukla 1987, Krishnamurthy and Shukla 2000, 2007, 2008 and Bhat 2006).

1.2.2.4. Off-shore Trough along West Coast of India and Embedded Meso-scale Vortices

Offshore north-south oriented trough in the lower troposphere along the West coast of India, like the east-west oriented MT along the Gangetic plain, is another important topic of the SW Monsoon studies which have gained importance since the work of George (1956). He had, for the first time, advocated that mesoscale cyclonic vortices form between mid-June to mid-August (peak monsoon phase) which propagate northward with a life duration of 1-2 days. The presence of such a vortex, within the active phase in the offshore trough, results in very high intensity rainfall on mesoscale in the vicinity of the vortex. Grossman and Durran (1981) Mukherjee et al (1984) and others since then have suggested that strong up-stream convergence near the West Coast is responsible for intense rainfall and later latent heat released in the rainfall tends to sustain mesoscale convection. Ogura and Yoshizaki (1988) further stressed the role of air-sea fluxes and orography in producing high intensity rainfall. Francis and Gadgil (2005) had found that most of the occasions of high intensity rainfall along the West Coast are organized due to the presence of MTC or east-west oriented mid-troposphere trough which results in the strengthening of large scale lower tropospheric monsoon flow with intense cyclonic shear within 2-3^0 north of the maximum wind in the vicinity of the west coast. ARMEX 2002 and 2003 field campaigns were focused on the study of offshore trough and embedded mesoscale vortices. Routray et al (2005), Bhaskar Rao and Prasad (2008) could simulate the formation of such vortices by using mesoscale models with ARMEX data. The subject needs further attention, as it is important for the incidence of high impact weather events along the West Coast during the monsoon season. Such events are accompanied with disastrous consequences and major losses to socio-economic sector and hence their skillful predictions 1-3 days in advance using mesoscale models would be of great societal benefit. Perhaps organisation of special

observational networks off and along the west coast, are need to initialize the models and improve prediction skill.

1.2.2.5. Active-Break Cycle of Monsoon

This has been an important theme discussed in Y.P. Rao's monograph. Even though traditionally 'break' monsoon was defined by Blanford (1886), taking into considerations factors like the MT hugging the foothills, rise of sea level pressure over central India, enhanced rainfall along the Himalayan foothills and cessation of rainfall over central India. The same criteria were used by Ramamoorthy (1969) in documenting past cases of monsoon break and by Raghvan (1973) and Keshvamurti et al (1980). Ramaswamy (1962) explained the incidence of break monsoon as a consequence of interaction between sub-tropical westerly circulation and tropical easterly jet stream. Keshvamurty et al (1980) examined the shift in quasi-stationary large-scale features of tropospheric circulation in active-break monsoon. Pant (1983) provided a physical basis for the break monsoon. Rao et al (2004) have found drying of the upper troposphere during the 'break' period and examined the conditions, which lead to 'break' monsoon. Rao (1988) also diagnosed dynamical causes for the active-break cycle. In the recent past there has been revival of interest in the study of break monsoon both in India and abroad. (In India, Krishnan et al 2000, De & Mukhopadya 2002, Gadgil and Joseph 2003, Prasad and Hayashi 2007 and Rajeevan et al 2008) and abroad (Rodwell 1987, Ferranti et al 1997, Annamalai and Slingo 2001 and others), several of these authors have defined break monsoon periods in varying manner than those by Ramamoorthy (1969) and hence the beginning of the break monsoon spells and their duration for similar spells differ amongst the recent studies. However, the fact remains that prolonged break of greater than 7-10 days are rather rare as majority of the breaks last for only 3-5 days. Hence short or moderate break monsoon episodes can be said to be part of oscillations of the MT but longer breaks lead to the disappearance of the MT or shifting its quasi-stationary position north of the sub-Himalayan belt. The latter leads to a sustained shift in the SW Monsoon rainfall regime for over a weak and it is in such epochs that above normal surface temperatures, lower-tropospheric inversions, strong northwesterly dusty winds and lower tropospheric anti-cyclonic vorticity persist over the Gangetic plain. The month of July 2002 was a characteristic example of an

extraordinary prolonged break monsoon which led to perhaps the worst drought in July (All India rainfall in July 2002 being in deficit 50% of the normal). During July 2002, lower tropospheric inversions due to subsidence were present for several days over MT region (Rao and Sikka 2005, Bhat 2006). During prolonged break monsoon spells, the monsoon airmass over MT region is replaced by the desert airmass from NW of India. During such dusty episodes which are accompanied with long period breaks, the aerosol load may even suppress the rain process due to indirect aerosol radiative feedbacks as lesser available moisture is being shared by many aerosol particulates, thereby possibly causing further suppression in rainfall. Aerosol layer present above the shallow clouds, further adds to the warming of the tropospheric layer and consequently subsidence is enhanced. Wang et al (2005) have used satellite data to study 'active' 'break' cycle and have suggested self inductive mechanism as a factor. Kusma Rao et al (2006) have suggested drying of the upper troposphere in transition from 'active' to 'break' monsoon conditions. Active-break cycle is an essential part of the monsoon intra-seasonal variability and is also observed in the model simulations even with climatological SST forced simulations. It will and should continue to attract the attention of investigators.

1.2.2.6. Western Disturbances and Troughs in mid-Troposphere mid-Latitude Westerlies

Pisharoty and Desai (1956) and Mooley (1957) had examined the role of western disturbances associated with passage of mid-latitude westerly troughs in mid-troposphere, on Indian weather. The topic was also well discussed in Y.P. Rao's monograph of 1976. Ramaswamy (1962) had emphasized that breaks in the SW Monsoon is a phenomenon of interaction between the TEJ and STJ, as the STJ shifts southward (28-32^0 N) and TEJ northward (15-20^0 N) during the southward penetration of westerly troughs in the peak monsoon season, Kalsi (1980) and others have re-looked at the role of western disturbances and monsoon circulation. Recently, Krishnan et al (2009) have investigated the model dependent internal feedbacks among monsoon – mid latitude interactions.. Rao (1976) have provided the upper level divergence ahead of the westerly troughs, which makes a pre-existing LPS to recurve, enhances examples when the monsoon rains. From the varied influence of western disturbances and mid-latitude troughs (suppression rain,

enhanced rains and normal rains) one tends to conclude that perhaps the stage of monsoon activity (active or near-break) dictates the response of the SW Monsoon to the influence of a western disturbance. Researchers may focus their efforts to use dynamical models-based diagnose to unravel the influence of western disturbances / troughs in westerlies on the monsoon processes in relation to the existing phase of the 'active' 'break' cycle of monsoon.

1.2.2.7. Meso-Scale Convective Systems (MCSs) and Cloud Scale Processes

Meso-scale, organized convection within the synoptic scale, has been recognized as the producer of high impact (very heavy rainfall to phenomenal rainfall of over 12 cm / day), since the work of Venkataraman et al (1974). These systems predominantly occur in high rainfall regimes like the West Coast of India, Sub-Himalayan range stretching from Pakistan eastward along Indian Himalayan belt, Nepal, Bhutan and NE India. Besides they also occur along the tracks of LPS over Central India, Gujarat, Bihar, Uttar Pradesh, Jharkhand, Punjab, and Haryana and even over Rajasthan. Rainfall in their vicinity on many occasions exceeds 25 cm / day and there are instances when the rainfall has even exceeded 50 cm / day. The recent episode of over 94 cm / day rainfall over Santa Cruz (Mumbai) on 27 July 2007 has been a devastating experience (Jenamani 2006, Vaidya & Kulkarni 2006, Shymala and Bardhan 2006). Bohra et al (2005), Sikka and Rao (2008), Deb et al (2007, 2008) and Litta et al (2007) have investigated this event with meso-scale models. There have been several studies to validate meso-scale models (MM5, WRF, ARPS etc.) for prediction of meso-scale weather (Routray et al 2005, Madan et al 2005, Dass et al 2007, Bhaskar Rao and Prasad 2005, Rama Rao et al 2007, DasGupta et al 2007, Bhaskar Rao and Ratna 2009 and others). Das et al (2006, 2008) found that the models do not give consistent skill in prediction of such events and the parameterization have to be turned as well as assimilation of data improved upon to get better prediction. Sharan et al (2007) have emphasized upon the recommendations on meso-scale modeling research which came out of the symposium on meso-scale processes in 2006 at IIT, Delhi.

Better understanding of the MCSs in the tropics resulted from GATE Studies (Houze and Betts 1981). Subsequent studies in other tropical regions showed that

convective organisations are essential component of synoptic scale but sometime occur independent of synoptic scale organisation such as in the form of severe local thunderstorms and cloud bursts. MCSs have their own hierarchy of spatial and temporal scales and have been extensively examined broadly under two aspects viz squall line and non-squall line MCSs. Over the SW Monsoon region, land surface processes (minor orography, vegetation, water bodies and state of soil) also impinge on organisation of mesoscale convection. Mountain valley winds, low level jets and even cloud scale microphysical processes can also influence meso-scale convection in the SW Monsoon season. Ramanathan's (1981) attempt was among the first to understand the role of moisture sinks and heat sources on cloud cluster scale convection over the Indian region using Soviet research ships data collected over Arabian Sea during Monsoon-77 field campaign in June 1977. This study showed that vertical eddy transport of heat and moisture to be higher in Arabian Sea cloud clusters compared to those in the West Pacific. Entrainment was found to be large (15 cm sec^{-1}) in the lower troposphere but deteriment was moderate (8 cm sec^{-1}) in the middle and upper troposphere. Johnson and Honze (1987, 1989) and Krishnamurti and Kishtawal (2000) and others have reviewed important aspects of MCSs in the Asian Monsoon System. They have emphasized upon the inter play of synoptic and diurnal scales, land-sea breeze system, differential radiative forcing between cloudy and nearly non-cloudy regimes in the organisation of MCSs.

Using research aircraft data collected during SMONEX-1979, Warner (1984) and Warner and Grumm (1984) have extensively investigated meso-scale cloud organisation within a monsoon depression system. These studies showed discrete NE ward propagating convective lines, similar to squall lines in the GATE region, with maximum reflectivity of about 40dBZ. Even ice particles were observed which pointed to the importance of cloud microphysics in deep convection. Sikka (2008 a), using Kolkatta Dopplar weather radar, has shown that convection near Kolkatta is very deep in monsoon depressions at the time of monsoon onset or after a long 'break' phase but is only moderate during active monsoon phase. This is because CAPE is high over the regions at the time of monsoon onset or at the time of revival of monsoon compared to that under active monsoon conditions. As Kolkatta radar data are now available for several monsoon seasons and similar radars are likely to

be functional along the MT, it would be good if the researchers use them to investigate meso-scale convection and its life cycle using Dopplar radar images.

There have been satellite-based observational studies too as the organisation of convection on cloud cluster scale (Laiang and Fritsch 1993, 1997, 2000, Rocca and Ramanathan 1997, Gambhir and Bhat 2000, Zuideman 2003, Islam et al 2004, Yang and Smith 2005, Ramesh 2007). According to these studies formation of cloud systems are highly favoured over landmass of South Asia during afternoon hours in the SW Monsoon season and morning hours are relatively free of cloud systems. Cloud systems in north Bay of Bengal are in continuous change and those lasting more than 6 hours form at night hours and decay during afternoon unlike those forming over the land mass of India. These features are less favoured over the central Bay of Bengal where cooler SST is being brought by the monsoon current (Rao et al 2006). Bhat (1998) and Gambhir and Bhat 2000, 2001) have shown good correlation between growth of cloud clusters and their size as well as depth and have also shown that majority of cloud clusters have life duration of 3-hours and decay within 100 km from where they had formed. Sorooshian et al (2002) and Yang and Smith (2006) studied diurnal cycle of convection over Indian monsoon area using TRMM data and confirmed late evening – night activity over land and late night – early morning activity over coastal regions. The interaction between the diurnal-scale and synoptic-scale in monsoon process needs to be investigated more extensively through observational and modeling studies. The expansion of observational networks, the installation of meso-scale networks along the MT region and the expansion of Dopplar radar network would provide the data for such detailed studies. Modernization of the observational systems by IMD and the Indian Air Force Meteorology Branch are likely to provide more information about the life cycle of MCSs and their diurnal variability within the synoptic scale. There is also a need for combining radar and satellite-based information about the development of meso-scale cloud system organisations over India in the SW Monsoon season. The importance of meso-scale cloud systems to monsoon rainfall could be derived from recently prepared 1^0 X 1^0 daily rainfall (Rajeevan et al (2006)) and from the TRMM / TMI 0.1^0 X 0.1^0 rain rates data which show that most intense rainfall is received for only 6 hours (Goswami and Ramesh 2007, Sikka 2008 a).

Many outstanding problems in monsoon convection still await further research, which includes its growth, propagation and decay and interaction between large-scale environment and the environmental flow around MCSs. The Science Plan of the Continental Trough Convergence Zone (CTCZ) multi-year (2009-2012) field campaign (DST 2008) has a special sub-program on cloud systems and MCSs and we hope that this area of research will get further attention as a result of stimulus provided by CTCZ.

1.2.3. Monsoon Process and Monsoon Tele-Connections

1.2.3.1. Atmospheric Boundary Layer and the South West Monsoon

Study of the interactions between the Southwest Monsoon and the atmospheric boundary layer (ABL) gained importance from the IIOE period onward. During the IIOE monsoon ABL was studied over the Arabian Sea by placing an instrumented buoy. IN SMONEX-1979 campaign a study of the ABL over the Arabian Sea was conducted by using a special technique called ‘gust probe’ on a low flying NCAR aircraft (Electra) and results were reported by a few investigators. It is over the Arabian Sea where the near-surface winds are strong and the surface is very rough that both the sea surface and the lower troposphere monsoon are exchanging fluxes of momentum, latent heat and sensible heat in more energetic manners. During SMONEX the U.S scientific team, lead by Prof. Seturman, and the Indian scientific team made attempts, leads by Prof. Narasimha, to examine the land ABL along Orissa coast. Rather strong low-level air-stream impinges on this coast after traveling over Peninsular and Central India and emerges over the North Bay of Bengal, which is the seat for active convection, warm SST and gensis of LPSs. These studies showed that the investigation of ABL over the entire east-west extent of the MT would be an interesting and challenging area of research. Accordingly, the IITM Pune and IISC, Bangalore planned a field program entitled Monsoon Trough Boundary Layer Experiment (MONTBLEX) during 1989 and 1990 monsoon seasons, under the DST sponsorship, to probe the complex interactions along the MT from its eastern end (moist end) across its western end (dry end). For this purpose several research organisations were involved and surface layer towers with slow and fast response sensors were installed at Khargpur (eastern end), Varanasi (central part), New Delhi (western part) and Jodhpur (western end). A Doppler Sodar and a

Kytone System were also used at Kharagpur. A research vessel was also placed in North Bay of Bengal during part of the experiment and interactions between air and sea were examined during formation of LPSs (Seetharamiah et al 2001). The entire quality controlled data sets were made available to research community in India and several research papers (Narasimha et al 1995) resulted. The data are still being used in scientific investigations. Some important results, which came out of, the MONTBLEX data referred to: -

(i) Increased instability in the ABL during weak / break monsoon under suppressed condition and decreased instability (increased stability) under active monsoon and organized convective conditions.

(ii) Momentum, sensible heat and latent heat exchanges are much influenced under the changing wind conditions as the monsoon oscillates between active and suppressed convective episodes. Such interactions are more marked along the western margin of the MT Air sea fluxes are also enhanced at the time of formation of monsoon depressions in Bay of Bengal (Sivaramakrishna et al 1997).

(iii) Special ABL formulation / parameterization are necessary along the MT where the wind is generally low in strength.

Since the western portion of MT oscillates more between active and suppressed convective episodes, it could be important for monsoon ABL processes. To investigate the role of land surface processes in monsoonal oscillations over western India, Indian atmospheric science community planned yet a more comprehensive field campaign entitled Land Surface Process Experiment (LASPEX) in Gujarat in the Sabarmati river basin during 1995 and 1996. For this experiment 4 surface layer towers, in quadrilateral formation around the central station at Agricultural University Anand, were deployed. At Anand a Doppler Sodar was also used and pilot balloon ascents were taken more frequently at near-by stations. Data collected have been used in several studies and results summarized by Sikka (2000) and Vernekar (2000, 2001) and others. During the BOBMEX – 1999 and the ARMEX – 2002, 2003 field campaigns also, researchers investigated monsoon ABL using research ships data over the Bay of Bengal, observations along the east coast and by using a surface layer instrumented tower in Goa on the West Coast, where the impact of convective outbursts under strong monsoonal episodes are important

(Tyagi et al 1994, Swati Basu et al 1998, 1999, 2000, 2001, 2002 and others). Mohanty et al 2003 and Sam et all 2007 have used ABL models and validated them against observations during BOBMEX and ARMEX. Kusma Rao (1996, 2004), Kusma Rao and Narasmiha (2006) have proposed heat flux scaling for weakly forced turbulent conditions. The CTCZ field campaign (under planning), to be conducted in 2009-2012, offers a unique opportunity to address several scientific issues related to monsoons interactions with ABL (CTCZ Science Plan 2008). Important aspects to be examined are the changes in fluxes and their role in active and weak phases of organized convection including changes in CAPE and CINE as convection episodes change. Other scientific issues to be examined include (i) surface heat balance over the ocean and its role in evolution of SST (ii) local feedbacks between ABL and monsoon and diurnal variation (iii) role of ABL for supply of moisture for formation of cloud systems within CTCZ (iv)cloud scale processes and role of ABL etc. For the above purposes data are planned to be collected on board two research vassels to be deployed at specific locations in the equatorial and north Bay of Bengal, 3 or 4 surface layer towers at suitable places over land (Kharagpur, Ranchi and Anand). A vast amount of slow and fast response surface layer data, radiosonde ascents on boardships and land stations are planned. Doppler sodars and weather radars are likely to be available during the 4-years CTCZ period (possibly aircraft probes too) and the ABL research community in India can look forward to an exciting opportunity for ABL research. The most important challenge would be to use the data for designing a specific ABL parameterization scheme for SW Monsoon region over which changes in lower tropospheric circulation, exchanges of fluxes are significant and the SST also changes on diurnal scale (about 0.5 ^{0} C), synoptic scale (1 to 1.5^0 C) and IS Scale (about 2^0 C) which may have impact on destabilization / stabilization of ABL.

1.2.3.2. SW Monsoon and Ocean Processes on Regional Scale

(a) Processes over the Arabian Sea and the Bay of Bengal

Traditionally, SW monsoon is considered to result from the land-ocean thermal contrast on large scale during the N.H. summer on a rotating earth. Indian Ocean provides energy to the monsoon atmosphere and the atmosphere in turn drives the vigorous southwest monsoon current in the Arabian Sea and the Bay of

Bengal. The importance of Indian and Western Pacific warm pools to monsoon processes has been much emphasized in the last 3-4 decades. North Arabian Sea and North Bay of Bengal begin to warm from March onward and reach peak warming toward mid-May. Air-Sea interaction over the Arabian Sea became a subject of monsoon studies beginning with IIOE (Colon 1964, and other studies) as the monsoon flow passes over cooler western Arabian Sea with suppressed clouds to warmer eastern Arabian Sea, which favours deep convection. The impact of the monsoon onset over the west and central Arabian Sea is large as the SST abruptly falls at the time of monsoon onset (Rao 1986 a, b). During July and August too, the Arabian Sea continues to cool but at a much slower rate with episodes of slight warming during weak monsoon and slight cooling during active monsoon as the monsoon oscillates on IS scales (Rao 1987, Vinaychandran 2004). The more vigorous circulation in the Arabian Sea, forced by stronger mean surface winds, cools the surface layer by exporting heat through southern boundary and into deeper ocean. Also north-south SST gradients over the Arabian Sea and the Bay of Bengal and their role in monsoon processes have been studied by Rao & Sikka (2005) and Shankar et al (2007). Gadgil (2000) has reviewed the role of ocean on monsoon processes. Durrand et al (2004) have investigated the role of ocean mixed layer in convective perturbations over the Indian Ocean.

Even during the peak monsoon season the SST over north Bay of Bengal remain quite high (near 29^0 C) with minor diurnal oscillations (0.5^0 C), moderate synoptic oscillations (~$1.0\ ^0$ C) and somewhat larger IS oscillations ($1.5 – 2^0$ C) (Sengupta et al 2001). The reasons for the Bay of Bengal to remain quite warm even though it is an area of organized moist convection with good frequency of genesis of LPSs, lies in the high salinity and relative stability of the near-surface ocean layer with prevalence of a barrier layer (Shenoi et al 2002 and several other). Bay of Bengal warm pool has been studied with BOBMEX data by Bhat et al (2001) and Bhat (2002) as latent heat fluxes are nearly 30-40% smaller than over warm pool of the Western Pacific at a given windspeed. Calculations with BOBMEX data showed that the combination of shallow mixed layer, presence of barrier layer in the salinity distribution and large net heat flux into the ocean, result in rapid recovery of SSTs after a convective event has given place to conditions under suppressed convection. BOBMEX data also provided glimpses of processes in the North Bay

(Vinaychandran et al 2002, Shankar et al 2007). However, little information on upper ocean processes is available from the southern Bay.

Fu and Fletcher (1985) emphasized the role of thermal contrast between Tibet and tropical Indian – West Pacific Ocean belt on the IA variability of SW monsoon rainfall. Recent studies have also shown presence of a tongue of relatively cool SST in the Southern Bay (Joseph et al 2005, Rao et al 2005) along the central part of the monsoon current, which is ascribed to the presence of Arabian Sea waters brought by the monsoon current.

Various hypothesis are available for the relative warm status of the north Arabian Sea and its role in the IS migration of cloud-band from near-equatorial Bay of Bengal northward (Sikka & Gadgil 1980). These hypotheses have been mentioned in CTCZ Science Plan (DST 2008) in which local in-situ rainfall, freshwater discharge from Ganga – Brahmaputra – Irrwadi and Mahanadi basins and tidal circulation in air-sea coupling have been invoked. Modeling experiments by Vinaychandran et al (1996), Kurian (2007) and Vinaychandran and Kurian (2007) also offer interesting insights into the problem. Besides fresh water discharge into the Bay, coastal upwelling also occurs along Andhra-Orissa coast. The IS meridional migration of monsoon cloud band has been linked to ocean-atmosphere coupling in the model experiments carried out by Krishnamurti et al 2007). Thus the Bay of Bengal and the Arabian Sea basins have a rich variety of ocean-atmosphere processes from diurnal to IS Scales, affecting the monsoon and offer potential opportunity to ocean-atmosphere researchers for further investigations with data to be collected under CTCZ through observational and modeling studies. Understanding of air-sea interactions on different spatio-temporal scales (diurnal to IS) is key to better understanding and more skillful predictions of monsoon on these scales.

(b) Mini-warm Pool in the SE Arabian Sea

Studies by Rao and Shivkumar (1999), Shivkumar and Shenoi et al (2005) and others have emphasized on the build-up of a mini-warm pool over the SE Arabian Sea from January to April, which was invoked to monsoon onset over Kerala

through formation of even monsoon onset vortex. This interesting scientific problem was investigated in several studies using special observations during ARMEX – 2002 & 2003 which besides a dense network of met-ocean buyos, special XBT lines on fortnightly scale were organized between Cochin and Mincoy region. These data revealed the presence of temperature inversions in the first 100-200 meters of the ocean surface. Such inversions in th sub-surface layers of the SE Arabian Sea in winter were first discussed by Thadathil and Ghosh (1992). Researches done by several investigators (Shankar et al 2004, Shenoi et al 2005, Haresh Kumar et al 2009) have shown that either the inversions are caused by the presence of low salinity Bay of Bengal waters over SE Arabian Sea during winter or due to changes in atmospheric fluxes during the winter season off Kerala coast and in Sri Lankan region. Some model results (Durrand et al 2007, Harish Kumar et al 2009) have shown that the break-up of the inversion could result in creation of the mini-warm pool over SE Arabian Sea in March. Other investigators (Rao and Sikka 2005) doubted the role of inversions in generating mini-warm pool. However, there is enough evidence that SE Arabian Sea in winter and pre-monsoon season is an interesting arena for important air-sea interactions which affect the ocean as well as the atmosphere with possible consequences on the build-up of the summer monsoon Data from XBT lines between Cochin and Minicoy and the ARGO floats offer potential opportunities to investigate the details of ocean-atmosphere processes over SE Arabian Sea.

1.2.3.3. Southwest Monsoon and Ocean Process on Seasonal Scale ENSO – Monsoon and IOD – EQUINOO – Monsoon Connections

(a) ENSO – Monsoon connections

As mentioned in Section 1, role of Southern Oscillation (SO) in modulating seasonal monsoon rains over South Asia was first discovered by Sir Gilbert Walker, Director General of Observatories in India in the early years of the 20th Century. Walker could only hint on the possible role of ocean in triggering opposite phases of SO. It was nearly 50 years later, when Prof. J. Bjerkenes of USA while investigating the abrupt warming of the eastern equatorial Pacific Ocean in the EL-Nino year of 1957-58, emphasized the physical connection between the coupled ocean and

atmosphere phenomenon as ENSO and named the associated east-west vertical circulation in the troposphere as “Walker Circulation” in honour of Sir G.T. Walker. Since the convective heat source along the east-west direction shifts on the equatorial area, its importance is driving tropical convection and mid-latitude circulation through Rossby wave propagation has become an important branch of global climate variability research. Since that time study of EL Nino / La Nina (cooling / warming of equatorial eastern Pacific Ocean) has become an important branch of the tropical ocean variability and links with the global atmospheric processes on inter-annual and decadal scales have been established, including those with the SW Monsoon El Nino-Southern Oscillation (ENSO) has been recognized as the biggest signal on inter-annual variability of the global climate system and its regional impacts on tropical /global circulations have been under active investigations. Under the World Climate Research Program (WCRP) of WMO & ICSU, a decade long international field program (1985-95) was carried out under the name of Tropical Ocean-Global Atmosphere (TOGA) with observations and modeling as tools for scientific study of ENSO and its climatic impacts with emphasis on prediction, using coupled ocean-atmosphere models. This research has brought a major shift in global and regional climate studies as a result of oscillations between ENSO warm phase (ElNino) and ENSO cold phase (La Nina).

Sikka (1980), and, subsequently Pant and Parthasarathy (1981) and Ramusson and Carpenter (1982), for the first time associated monsoon droughts over India with warm phase of ENSO. Subsequently several investigators worldwide have worked on the ENSO-Monsoon connections. Mooley and Paolino (1987) emphasized that it is in the warming phase that ElNino has impact on monsoon. Webster and Yang (1992) emphasized the selective interactions of Monsoon and ENSO. Some investigators have even emphasized the secular changes on decadal scale about this relationship (Torrence and Webster, 1999). Kriplani and Kulkarni 1997, Krishna Kumar et al (2006) and others have suggested that ENSO-Monsoon connections have weakened in the last two decades. There could be a climate shift from 1970 onward in the ENSO-Monsoon relationship. The connections, however, are not very intimate as the correlation between SST in El-Nino 3.5 region and Indian monsoon rainfall is -0.54. However, the fact remains that there is no other parameter except warm El-Nino event, which holds the relationship with deficit

monsoon on decadal scales so consistently, and the parameter has not lost significant correlation with monsoon seasonal rainfall over India. Besides some of the worst monsoon droughts over India (1918, 1951, 1965, 1972, 1982, 1987, 2002) have happened when warm El-Nino events was evolving (Mooley and Paolino 1987, Sikka 1999).

ENSO – Monsoon relationship is a 'tantalizing' one and many modeling experiments have shown it to be largely valid though again not on one to one basis. Beginning from the work of Keshvamurty (1982), who used GFDL GCM to investigate the role of east Pacific SSTs on monsoon, several investigators have studied monsoon-ENSO relationship using models (Ju and Slingo 1995, Soman and Slingo 1997, Yu et al 2002, Krishnamurthy and Kirtman 2003, Kug et al 2006, Kug and Kang 2006, Bellanger et al 2007, Kitoh 2007), have investigated the relationship using interactive coupled models and found that the ENSO-Monsoon connection to vary on decadal scale. Krishnamurthy and Kirtman (2003), Wu and Kirtman (2004), Duval and Roca (2004) and Bellenger et al (2007) have investigated the connection between Indian Ocean with the growth and termination of ENSO. Krishnamurthy and Goswami (2000) investigated monsoon-ENSO connections on decadal scale and Prasad and Singh (1996) have examined the seasonal variability of ENSO parameters. Goswami and Xavier (2005) have advocated that ENSO controls the length of the summer monsoon rainy season of India. There is potential predictability in monsoon long-range forecast provided coupled ocean-atmosphere models could predict the evolution of warm event. This aspect would be further discussed in seasonal monsoon prediction. ENSO-Monsoon relationship is likely to remain a topic of active research in the future too. There are suggestions that the relationship has secular variations on decadal and even centennial scales and could be modulated by global warming. Many Indian researchers have engaged their attention on observational modeling aspects of the relationship and these investigations would continue to be pursued.

(b) IOD-EQUINOO – Monsoon Connections

Indian Ocean Dipole (IOD) is an ocean-atmosphere coupled phenomenon in the equatorial Indian Ocean, which has come to light in recent years. (Webster et al

1997, Saji et al 1999, Murtugudde et al 2000 and others) Its connection with the SW monsoon has been investigated in several studies and also linked with the amplification of ENSO phenomenon (Ashok et al 2004, Yamagata et al 2003, Annamalai et al 2003, Behra et al 2003, Oh et al 2005 and others). IOD occurs in a rather narrow equatorial channel between the eastern equatorial Indian Ocean and equatorial west Arabian Sea in which upwelling off the eastern, Malaysian coast is enhanced by the strongest SE winds in some years such that SST is cooler in eastern equatorial Indian Ocean compared to the western end. The phenomenon begins to evolve in March-April and peaks in October-November and hence lasts for only 6 months unlike ENSO, which is of much longer duration. The reverse occurs in some other years too when the eastern end is warmer than the western end. In some respect it is similar to El-Nino-LaNina variability in the equatorial Pacific Ocean but the amplitude of the SST variations in IOD is much less than in the ENSO. Thus IOD has now been recognized as an important component of the near-equatorial Indian Ocean Climate Variability. Changes in the SSTs in equatorial Indian Ocean, as a result of IOD, impact on the organized convection in the region such that in the positive phase (eastern end cooler and western end warmer) of the IOD convection along the eastern end is reduced. In the negative phase (eastern end warmer SST) convection over the eastern end is enhanced. If the eastern end is warmer (a negative IOD phase is operating), enhanced convection over 80-100^{0}E in the equatorial region would influence the Monsoon Hadley Cell such that organized convection is reduced over the continental India, thereby weakening the performance of seasonal SW Monsoon rains over mainland India. Although most of the positive phases of IOD occur when warm ENSO event is progressing, there are IOD events, which occur, independent of ENSO. Hence some researchers feel that IOD is an independent mode of ocean-atmosphere variability of the equatorial Indian Ocean. Gadgil et al (2004, 2007) have emphasized the atmospheric part of the IOD and used the term EQUINOO to describe it in which sea level pressure gradients along the equatorial Indian Ocean are affected thereby leading to modulation of winds over the region and also the intensity of organized convection along equatorial Indian Ocean. Gadgil et al (2004) have further shown that monsoon droughts have much better coherence if the incidence of warm ENSO and EQUINOO occur in phase such that the reduced convection over India due to both phenomena may enhance the effect of large-scale subsidence over main land India. This would

explain as to why all warm ENSO events are not connected with droughts as in some events EQUINOO favours good performance of monsoon over India and hence annuls the effect of warm ENSO event. Essentially ENSO and IOD / EQUINOO events redistribute near-equatorial convective heat sources, whether in the equatorial Pacific Ocean (ENSO) or in the equatorial Indian Ocean. IOD / EQUINOO in isolation or in combination with ENSO are responsible for affecting monsoon convection over land by modulating equatorial (Walker Cell – ENSO) or Monsoon Hadley Cell (IOD / EQUINOO). Since the discovery of IOD and EQUINOO there have been several observational and modeling studies in Japan and USA on their evolutions and their relationship in the Asian monsoon system. Thus IOD – EQUINOO has become another topic of hot pursuit and more and more researchers are getting interested in it. For example, Francis et al (2007) have suggested that the positive phase of IOD could be triggered by severe-cyclones over the Bay of Bengal in April-May. Some studies have shown that IOD is not an independent phenomenon but is triggered by ENSO. The connections between the ENSO, IOD and Asian Monsoon are very intriguing and Indian researchers may devote more attention to this branch of monsoon research. The potential role of the inter-annual, decadal and secular variability in ENSO-IOD-EQUINOO Monsoon connections through observational and modeling studies should be exploited for further insight into the relationships. The recent efforts by international community to put near-equatorial moorings along 100-50^0 E would help in observational research. For use in monsoon seasonal forecasting evolution of IOD / EQUINOO has to be predicted just like ENSO by the beginning of May. Therefore, ocean-atmosphere coupled models are to be used for finding potential predictability of IOD / EQUINOO. The signal in SST oscillations is rather weak in equatorial Indian Ocean and hence definition of the incidence of IOD / EQUINOO, based on SST fluctuations, has to be quite precise.

1.2.3.4. South West Monsoon and Himalayan – Eurasean Snow Connections

In the early years of IMD Blanford (1884) had suggested that snow cover over the Himalayas in winter and spring seasons could influence the rains in the ensuing SW Monsoon season over the Indian sub-continent. This was also supported by Sir Gilbert Walker (1910 to 1924). Walker (1928) and Walker & Bliss (1932) through

their statistical correlation approach. It was adopted as a parameter in Walker's statistical long-range forecasting of the SW Monsoon rains over the subcontinent. The relationship between Himalayan snow cover and the SW Monsoon rains has a physical base in the thermally driven monsoon circulation and consequently the high albedo of snow would delay the sensible heating over the region as considerable incoming radiation would be used to melt the snow before the region is heated. Thus over snow covered region, not only the extent of snow cover but its depth, and the amount of snow melt and its evaporation all become important factors in altering sensible heating during spring and summer over South Asia.

Data on snow cover extent over the Himalayan region began to be collected by visual observations over different parts of Himalayas under arrangement of the IMD and then empirically quantified to be used in Walker's regression equation for LRF of monsoon rains. These data are available in IMD Pune for many past years. The estimation was no doubt faulty but this is all what could be done in past years over difficult terrain. Thus a comprehensive climatology of snow cover over Himalayas could not be established over the region till space-based (Satellite) sensing became possible since 1970s. Hahn and Shukla (1976) revived the work of Blanford on snow-monsoon relationship by affirming that satellite derived snow cover over extensive Eurasia rather than limited Himalayan belt is a better precursor for LRF of monsoon. This was further validated by Dickson (1984) and Parthasarathy and Yang (1995), Yang (1996). Kriplani et al (1996). Kriplani & Kulkarni (1999) even connected Nimbus 7 observed snow mass and the Soviet snow mass data respectively with performance of monsoon over India on IA scale.

Grosiman et al (1994) have established a negative partial correlation of -0.50 between snow cover over the northern hemisphere and the Southern Oscillation Index (SOI) which is significant at 99% level and found that SOI explains about 65% of snow cover variability over the Himalayas and Tibetan region. Thus higher (lower) snow extent over the Himalayan Tibetan region during an warm ENSO (Cold La Nina) event is implied which would result in deficit (excess) monsoon rains from the normal. Barnett et al (1988) were the first to test snow cover-monsoon relationship by using AGCM experiments and found it to be valid. Since their work many researchers have validated the hypothesis by using different AGCMs, which

establishes the robustness of the results, (Yasunari et al 1991, Kitoh 1994, Shankar Rao et al 1996, Douville and Royer 1996, Vernekar et al 1995 and others). In India, Prof. S.K. Dash and his group in IIT, Delhi has written several papers validating the snow-monsoon hypothesis by their sensitivity experiments using AGCMs. IIT Delhi is now preparing dynamical seasonal monsoon rains forecast using spring-snow cover as boundary forcing. However, the results remain mixed as other factors rather than snow boundary forcing alone also modulate the seasonal rains.

Bamzai and Shukla (1998) have extended the climatology of Satellite snow cover and snow depth over different regions of the Northern Hemisphere and studied their connections with the performance of the Indian Monsoon. Sikka (1999) has reviewed the vast literature and found that the snow-monsoon relationship has shown some weakness in recent years and uniqueness of monsoon droughts or excess seasons with excess or deficit snow cover could not keep pace with extension of satellite data over the last 3 decades. Evidence has been shown, in other investigations too, that there is secular negative trend in the N.H winter / spring snow cover – possibly due to global warming. Some workers have even linked ENSO to fluctuations in Eurasean snow cover and through ENSO to monsoon. As in the case with the relationships of other parameters with monsoon performance, there could be complex decadal fluctuations in monsoon rains which may introduce changing patterns on the decadal scale. National Remote Sensing Agency (NRSA) has introduced Himalayan Snow cover Information System (HIMSIS), based on retrievals from the Indian Remote Sensing Satellites, and this data when extended in future for several decades may help in diagnosing decadal scale variability of snow cover and also the influence of climate change on extent of snow cover. Himalayan cryosphere is important to India as the snow melt augments the water resources in the Indus, Ganga and Brahmputra river basins in the pre-monsoon season. India's climate change studies give importance to Himalayan cryosphere and ecology. This area of research is to remain important for the next 2-3 decades and there is scope for Indian meteorologists to exploit studies of the Himalayan cryosphere and its relationship with monsoon.

1.2.3.5. Monsoon and Aerosols (Atmospheric Brown Cloud) Connections

In 1970s Prof. Bryson of USA had propounded a hypothesis that the Rajasthan desert (Thar Desert) had resulted due to human interaction (over grazing) as the region was not as dry in the distant past some centuries ago as at present. The desert aerosols, suspended in lower-mid tropospheric layer, trap the incoming solar radiation, stablise the lower atmosphere and thereby cause subsidence. The role of atmospheric aerosols, particularly the anthropogenic component of it, is to make the sky hazy by scattering. This is contributed by most of the accumulation mode fraction of aerosols and impinges on radiative forcing at surface and in the lower mid troposphere. Absorbing aerosols (desert dust and elemental / black carbon (EC / BC) as by product of combustion) would lower the single scattering albedo. Coarse desert dust particles, raised by dust raising winds and dust storms, even obscure the sun for days in May-June over the Indo-Gangetic Plain (IGP). These particles under the present day industrialized environment get coated with EC / BC, making the anthropogenic mineral dust coated aerosol layer more absorbing. The influence of aerosols on atmospheric radiative forcing may be enhanced if the elevated aerosol layer is situated above the low level shallow clouds as is the case in weak monsoon conditions. Since 1999, as a result of INDOEX field phase, this anthropogenic aerosol layer is being referred to as Atmospheric Brown Cloud (ABC) and has become a hot research topic; particularly with regard to its influence on regional climate forcing (Ramanathan et al 2007, 2008). ABCs are stated to mask the green house warming by surface dimming while at the same time enhance green house warming of the atmosphere by absorbing solar radiation particularly if the ABC layer is above the shallow clouds. Their indirect radiative effects, by sharing moisture content of the air among many particulates, can also suppress vertical growth of clouds. INDOEX made a paradigm shift with regard to our understanding of the aerosols over South Asia. They can be transported thousands of kilometers away from their source regions before they are removed from air. IGP has now been recognized as a hot spot for ABC studies.

Several modeling studies have been performed with regard to the influence of aerosols (ABCs) on the Asian summer monsoon and its regional components like the East Asian Monsoon and the SW Monsoon. Ramanathan et al (2005), Chung

and Ramanathan (2006), Menon et al (2002), Adhikari et al (2007), Meehl et al (2008) and others have chracterised the seasonal cycle of aerosols over South Asia and used AGCMs / CGCMs to simulate their effect on monsoon. In India, Chakraborty et al (2003) used AGCM to simulate the effect of absorbing aerosols on the SW Monsoon and found that physical para-meterizations play important roles in modulating their impacts / sensitivity. Recently Sikka (2008 b and 2009) has shown that even in pre-monsoon (hot season) and SW Monsoon season there are episodes when the aerosols load over the region remains high (AOD > 0.7) followed by spells of low load (AOD < 0.3). According to him there is a fluctuating regime of AOD and each spell lasts for 3-7 days. Thus there is no semi-persistent aerosol load but fluctuating one and the AGCM/CGCM experiments in future could be done under fluctuating aerosol radiative forcing regime so as to estimate their influence on monsoon processes. He hypothesized that the high aerosol load could support the long break monsoon episodes in some years and a stronger monsoon disturbance would be needed to offset this suppression of monsoon activity. He also suggested that since dust particles have ice-nucleating properties, their presence in western India may lead to enhancement of rainfall through microphysical effects when a weather disturbance reaches that area. Aerosols indirect effect on rainfall over India has been examined by Tripathi et al (2007). Also Moorthy et al (2007) have used in-situ and satellite data to examine the role of dust from the Indian desert in reducing incoming solar radiation by their absorption effect.

There are two hypothesis with regard to the effect of ABCs on the SW Monsoon:

(a) ABCs can weaken the monsoon circulation through hydrological feedbacks and decrease the SW Monsoon rains or even increase the frequency of droughts as it has happened since 1960s (Ramanathan et al 2005). This has to be viewed with caution as even in the period 1910-1925, when the ABCs were not present, Indian Monsoon rainfall passed through higher frequency of droughts or the multi-year spread of droughts (mega-droughts). Aerosols through hydrological feedbacks of ABCs (direct and indirect) effects can also influence monsoon by changing meridional trpospheric temperature gradients between aerosol deficient equatorial oceanic area and land dominated parts of South Asia with heavy loads of aerosols (Surface cooling effect of aerosols) can also influence monsoon processes.

(b) The second hypothesis is by Lau et al (2006a) in which the elevated Heat Pump over IGP during May-June is invoked to enhance the upper tropospheric warming and divergent circulation over South Asia / Tibet region. Presence of elevated layer of the aerosols within middle of upper tropospheric layers could enhance heating of the troposphere regionally over IG Plain and the Himalayan-Tibetan regions. This would enhance early monsoon season rainfall (June) or earlier advance of the monsoon over IGP (Lau et al 2006 a, 2006 b). Lau et al (2008) have proposed a field program to test this hypothesis.

Indian aerosol research community has recognized the potential of anthropogenic aerosols to monsoon processes on different scales in fact they have launched an instrumental road campaign in winter. A multi-disciplinary campaign was organized in March-May 2006 - ICARB Campaign (Moorthy et al 2008) to monitor different aspects of aerosol distribution over India and over the adjoining Bay of Bengal. ISRO has a good network of in-situ land stations over India and the IMD has plans to install more stations under their Environmental Monitoring Program. IITM, Pune has an ambitious plan to launch Cloud Aerosol Interaction and Precipitation Enhancement Experiment (CAIPEX) to be carried out in 2009-2012 to unravel the role of aerosols to monsoon processes in different phases of the monsoon. Aerosol research in India has identified location of potential source regions (NW India, IG Plain, Central India, Western India, Indian Seas) within India and in neighbouring countries of West Asia and Africa. The contributions of different sources differ with seasons and from episode to episode within a season (even in SW Monsoon Season). Organic matter from forests, biomass burning, industrial and transport sectors combustion and advection of massive aerosol loads from sea spray of the Arabian Sea are major contributors to local and regional sources. Three scales of aerosol-monsoon interactions have been also identified viz., (i) micro-physical cloud scale (first and second indirect effect), (ii) meso-scale perturbations to heat budget in terms of sensible heat and latent heat (direct effect) and (iii) large scale pattern (persistent break monsoon). To study aerosol-monsoon interactions we need to understand tropical-sub-tropical meteorological interactions in active-break monsoon cycle, biomass burning in IGP and impacts of aerosols on seasonal monsoon. Bridging the aerosol-monsoon knowledge would involve studies planned during CTCZ and CAIPEX, and the international Joint Aerosol-Monsoon Experiment

(JAMEX), involving insitu (Lau et al 2008) aircraft and satellite train (like CALIPSO – Cloud Sat and Parasol satellites). Aerosol may be a factor in driving the monsoon rainfall variability only after meteorology brings them in good measure to persist for several days over the MT (IGP) region. We need to identify and understand possible modulations of monsoon by aerosols using observations and modeling tools. The Indian CTCZ Program (2009-2012) has included study of aerosols as a sub-program under Cloud system and Aerosol component. Thus both under the CAIPEX and CTCZ programs intensive and extensive campaigns with in-situ observations, satellite monitoring, aircraft flights and micro-pulsed lidar studies are planned. Similarly under the Asian Monsoon Years (AMY 2008-2012) campaigns, China, Korea, Japan and Mangolia have plans to study Monsoon-Aerosol connections. Aerosol studies planned under CTCZ, CAIPEX, JAMEX and AMY could add another dimension to monsoon research in the next decade. Huge amount of new data are likely to be available which could be used by Indian atmospheric scientists to take forward monsoon – aerosols (ABCs) connections through observations and modeling studies during the next decade. The subject is important in other seasons too particularly with regard to the impact of aerosols on the South Asian Cryosphere. Studies by Hasnain (2007) and others have shown that the glaciers over South Asian region are melting at faster rate. This could be due to enhancement of global warming due to increasing deposition of black carbon over the high mountains. The Indian Himalayas in western and eastern regions are covered with forests. Forest fires have been also increasing in frequency in these regions. Forest fires are source of ash / black carbon which is deposited on the glaciers. This black ash could reduce the reflectivity (albedo) of white ice and may accelerate glacier melt. Therefore, there is a need to include aerosol and snowmelt chemistry studies over the higher reaches of Himalayas at least over a few selected stations / glaciers. Anthropogenic aerosol dumps could affect glaciers being past repository of climate and data are needed as to whether glacier retreat could be influenced by aerosol dumps. Himalayan glaciers are buried under thick rubble of surface moraine and glacier ice underneath has a dusty appearance due to accumulated mineral dust which may percolate into the crystal lattice of glacier ice and may have a layer of desert aerosol deposit which is becoming coated with black carbon of anthropogenic origin. This aspects need to be studied by collecting dust samples over the glaciers. Some experiments have been conducted on Indian Himalayan glaciers about the

role of aerosol dumps on glacier region (Raina and Sangewar 2008) but the results are not conclusive.

1.2.3.6. Modeling of the SW monsoon on different scales

Atmospheric modeling using computers began in 1950s. It took roots from short-range NWP scale (1-3 days) to climate simulation since 1956, medium-range scale prediction (3-10 days) since 1980s and seasonal scale dynamical prediction since 1980s. Work on short-range NWP scale in an organized manner began in India in 1960s (IMD 2009) and on medium-range scale since 1988 (NCMRWF 2009, Sikka and Shukla 2009), and meso-scale since 1990s. Advances made on each scale of prediction are significant. Operational predictions are now made within IMD, NCMRWF and the Indian Air Force. Research centres like IITM, Pune, Andhra University, CMMACS Bangalore, IIT Delhi, IIT Kharagpur and others carry out research studies on dynamical models of monsoon on different scales.

Systematic research on application of NWP in India both at IITM and IMD began in early 1970s. IMD developed an operational system based on quasi-geostrophic model by the end of 1970s. Research on PE model was conducted at both organisations during 1980s, (Das 1978, Singh et al 1980, Singh and Sugi, 1985). IMD began using regional PE models for operational NWP at the end of 1980s and since then rapid progress has been made (Sarkar and Bedi 1987, Basu 1999, Roy Bhowmik et al 2001, Rama Rao 2005, Basu 2005, 2007 and others). Krishnamurti et al (1983, 1984) were the first to use AGCMS for medium-range forecasting of monsoon weather. Basu (1990, 1991) used NCEP model and analysed sensitivity of the model to vertical resolution. Since 1992 medium-range monsoon prediction model is in operational use at NCMRWF and several papers have resulted from NCMRWF since then. In the IMD (2009) several specialists have contributed articles on different scales of monsoon modeling and hence here we mention only the major advances or the contemporary stress being given on the modeling of the SW Monsoon. Even though most of the models, on all scales, being used for operations and research have been adopted from USA or European versions, it is to the credit of Indian scientists that they have used them for different purpose with great dedication. It would not be wrong to say that the future work on

the SW Monsoon, for operations and research, which was till 2000 based on synoptic approach, would shift more and more to use of dynamical models. Therefore, it is desirable that young scientists are trained in modeling, though not totally being unaware of synoptic and climatic understandings. As computational and data facilities and other infrastructures for research are improving at a rapid pace in India, younger scientists would benefit a lot if they devote their efforts on model developments, model validations and model diagnostics to test the monsoon processes as brought out in earlier synoptic and climatological approaches, which had formed the basis for Y.P. Rao's monograph on the South West Monsoon.

(a) Present Status in Short-Medium-and Meso-scale Dynamical Prediction of the SW Monsoon

Starting from 1950s use of dynamical models for predicting weather on the above scales has made tremendous advances in advanced countries. These advances came as a result of developments in theory, computational aspects, data availability upto stratosphere levels, data initialization and data assimilation systems, introduction of higher resolution in the models and parametrization of a variety of physical processes for the ABL, convection, radiation etc. Many researchers, mostly in advanced countries of Europe, North America, Japan and Australia, have contributed to original research in modeling. Availability of enhanced computational power, no doubt, has contributed to the introduction and application of more and more sophisticated models. Weather prediction on all scales has now become a technology. As a variety of models became available they are found to give results which differ in detail about weather incidence, its intensity etc. Uncertainty in NWP approach has been now well realized. Even though the dynamical core of most of the models is similar, uncertainty in weather prediction is produced due to several factors; (i) variable data availability, (ii) differences in initialization and data assimilation procedures (iii) differences in model resolutions and (iv) differences in parametrization of different sub-grid scale processes etc. There are also special problems in the tropical weather prediction because sub-grid as well as large-scale convective processes play very important roles in tropical weather dynamics. Sikka and Rao (2008) had hinted about the uncertainty in high impact weather events in the SW Monsoon season. Recently, during the SW monsoon season of 2008, IMD

had made special arrangements to obtain daily model products from advanced global weather prediction centres like ECMWF, NCMRWF, U. K. Met. Office, Japan Met. Agency and NCEP as well as the meso-scale models being used in India (IMD, NCMRWF and Indian Air Force). The differences in 1-5 days forecasts among the models with respect to large scale evolution of the phases of the monsoon, predicted tracks of monsoon transient disturbances (LPSs) and distribution of rainfall became obvious in day to day discussions of the model results as well as in their detailed analysis. Significant differences in model forecast make the decision-making by the operational forecasters a big problem. The solution for removing such differences has been proposed by the researchers in terms of forecast ensembles either using the same model but perturbing the initial conditions slightly by using different dynamical approaches (ensemble of a model) or using multi-model ensembles (Krishnamurti et al 2000, 2001, 2004, 2007) in which each model is trained and its weakness and strength are statistically assessed and based on their performance weightage is provided to average the products (mostly rainfall or tracks of transient disturbances) of different models. Under the joint efforts of NCMRWF, IMD and IITM, a project was implemented in monsoon season of 2008 to assess the 1-5 days rainfall forecasts of NCMRWF, U.K. Met Office, NCEP, JMA and ECMWF models as well as multi-model ensemble forecasts. The results of the study (Mitra et al 2009) are under preparation. Tentatively it is found that the models differ significantly in terms of 3 cm or higher rainfall in 24 hours even on one-day forecast basis and these differences became quite large after 3 days. Also the multi-model ensemble forecasts show significant differences for higher than 3cm/day rainfall and for forecast period beyond 3-days. Therefore, weather forecasters have to keep in mind such significant uncertainty in monsoon rainfall forecasting for period beyond 3 days and for even moderate rainfall intensity on local and sub-regional scale. The users of the forecast must also be made aware about the uncertainty in quantitative rainfall forecasts. However, continuous research in the next few years is likely to reduce this uncertainty. Roy Bhowmik and Durai (2008) have also applied a modified MME Scheme for forecasting monsoon rainfall over India.

Meso-scale high resolution forecast models began to be developed in academic research centres in USA and a variety of models are now available (ETA, MM5, RAMS, WRF, HWRF etc.). These models are well configured to be used with

different resolutions and physical packages. Embedding them in different AGCMs for using lateral boundary conditions can use the models. All these models are being deployed in various operational and research centres in India. However, since meso-scale observational networks are not yet available in India, the data assimilation system for meso-scale models is not adequate. Besides the distribution of details of surface in homogeities (vegetation, water bodies, detailed orography, soil moisture etc.) is also not yet quantified. The prediction products from these models differ significantly among themselves. Various schemes for convective parametrization have been used by several researchers in India (Krishnamurti et al 1988, Vaidya and Singh 2000, Das et al 2001, 2002 and others). No consensus is available in the Indian research community as to which of the scheme should be preferred under daily evolution of the monsoon for meso-scale and synoptic-scale and even for medium range prediction models. Similarly meso-scale high resolution models like MM5, WRF, ARPS etc. have different choices for convective parameterization and have been tested by several researchers in India (Roy Bhowmik, 2003, Das et al 2006, DasGupta 2007, Rama Rao et al 2007, Abhilash et al 2007, 2008). The researchers have found that there was no consistent skill of the high-resolution meso-scale models in predicting the location as well as intensity of the high impact weather events during the monsoon season. False alarms also occur on several occasions. IMD has introduced district scale daily weather prediction since 2007 by combining model outputs with synoptic experience (Lal et al 2006) as well as by multi-model ensemble (Roy Bhowmik and Durai 2008). However, the skill for district level forecasting needs improvement as false alarm rates on meso-scale rainfall remains large. This area needs more concerted research by inclusion of data from meso-networks. Cloud resolving models with non-hydrostatic approach is an alternative approach, which would need high computing power for operational use. Work in the Frontier Research Institute in Japan (Saito et al, 2008), under the leadership of Prof. Matsuno, has shown that cloud resolving super parameterization produces may offer better results for producing cloud-cluster scale convection within the overall envelop of large scale convection in the tropical monsoon regime. This is another break through on the horizon for more skillful monsoon prediction. Similarly assimilation of non-synoptic data in short-medium-and meso-scale models have been used including Dopplar

Weather data (Bohra et al 1998, Roy Bhomik et al 2005, Abhilash et al 2007, Errrico et al 2007, Xavier et al 2008)

(b) Monsoon Climate Modeling and Present Status

Simulation of the SW Monsoon circulation and associated rainfall is one of the most challenging scientific tasks and in the last 3 to 4 decades many attempts have been made with varying degree of success. Godbole (1973) had made the first attempt in India to simulate SW Monsoon in a 2-dimensional Y-Z plane model along 80^0 E and had emphasized the role of Himalayan orography in the north and ocean to south of India. Since the pioneering work of Manabe et al (1974) and Hahn and Manabe (1975), on simulating Indian monsoon climate with atmosphere general circulation models (AGCM), the monsoon modeling community world-over has made tremendous progress both in terms of improving the physical parameterizations and model resolution. During the two decades of 1980-2000 several AGCMs and even coupled GCMs in different countries have been used to simulate the SW Monsoon circulation and rainfall with different resolutions and model physics (Fennessy et al 1994, Sabre et al 2000, Lau et al 2004, Wang et al 2005 a, Sajani et al 2007, Rajendran et al 2008). The simulated monsoon show differences in detail (Gadgil and Sajani 1998), though almost all models capture the large scale circulation aspects in lower and upper troposphere but differ in the simulation of MT south of the Himalayas (Sperber and Palmer 1994, Fennesey et al, 1994, Kar et al 1996, Krishnan et al 2001, Nanjudiah et al 2005, Ratnam et al 2005, Sajani et al 2007, Saji Mohandas et al 2008). Influence of land surface and biospheric processes on monsoon simulation has been investigated by Sud and Smith (1985) and Sud et al (1990) and others as well. Hence fundamental challenge in simulation of monsoon is still present (Wang et al 2005 a). Also simulation of monsoon seasonal rainfall over India and its inter-annual variability differ from model to model, as evidenced from studies based on AMIP project and other research on the topic (Sperber and Palmer 1994, Gadgil and Sajani 1998, Kang et al 2002 and others). The simulations are sensitive to model resolution, model physical packages, initial conditions, prescription of SSTs, representation of orography, and the time of beginning the simulation process etc. In India too some research groups viz. IITM, Pune, CDAC, Pune, IITD, IISC, Bangalore, CMMACS. Bangalore, NCMRWF, NOIDA began using

different AGCMs for monsoon simulation since 2001. Even ensemble runs have been used for the purpose (Kar et al 1996, 2000, Ramakrishna et al 1998, Basu 2001, Patnaik and Satyan 2000, Krishnan et al 2001, Ratnam and Krishna Kumar 2005, Ratnam et al 2005). Sensitivity studies to orography, (Saji Mohandas et al 2008), SSTs, (Deb et al 2006) and model physic have been investigated. Recently, Kitoh (2004) has systematically examined the role of Tibetan-Himalayan orography on monsoon simulation with the high resolution model of MRI and found that a critical threshold of 0.6 of the present orography is needed to get somewhat proper simulation of the Indian monsoon. Again the ensemble simulation experiments of the Indian monsoon rainfall have been reported by Sajani et al (2007), forced with observed SST to study its IA variability. They found that despite that simple ensemble mean, capturing the features of climatological Indian Monsoon pattern and rainfall anomalies, the model shows certain systematic bias over the Asia-Pacific region. Hence the IA variability of rainfall over India is not adequately represented unless bias correction is applied for each Julian day separately at each grid point. The low-frequency variability of the monsoon on 30-40 day scale is also not well captured in the model simulations. In the 20 years model runs of NCEP T-170 AGCM and NCEP coupled model (CFS) too there are differences in IA variability of monsoon onset, monthly and seasonal rainfall over India. Evolution of rainfall on pentad scale and lack of realistic northward propagation mode of the monsoon have been found. Although tremendous progress has been also made since 1970 in dynamical monsoon simulation, AGCMs and CGCMs still show significant differences in terms of mean rainfall on seasonal scale (June to September) and monthly scale, IA variability on seasonal and monthly scales, timings of major episodes in monsoon cycle (onset, active, break and withdrawal). The revival of the monsoon through observed northward progressing convective organisations is mostly missing in all AGCMs and even CGCMs. Higher resolution models are capable of better simulation of the orographic rainfall and the position of MT as well as other characteristic circulation features of monsoon in the troposphere. As such high resolution AGCMs with 3-D Var and 4-D Var data assimilation are now operationally deployed in several countries including India. We are still to make progress before climate models would be trusted in all respects for monsoon simulation close to observations (particularly in rainfall) and used for prediction of monsoon climate. Indian monsoon community is aware of the limitations and would

benefit from the ongoing national and international efforts for more realistic monsoon simulation by using climate models.

(c) Dynamical Monsoon Seasonal Prediction and its Present Status

Long Range Forecasting (LRF) of monsoon has a long history in India beginning with the work of H.F. Blandford (1884) and Sir G.T. Walker (1910-1924, 1928) in early years of the 20th Century. The status of statistical LRF has been discussed at different times by different researchers (Banerjee 1950, Normand 1953, Jagnnathan 1960, Das 1987, Rajeevan 2001 and others). For over 100 years now operational LRF in India is being done on statistical regression equations basis, first introduced by Walker, though periodically the parameters used have been changing by rejection of those parameters which have lost significance with seasonal rainfall and introduction of new parameters those which have gained significance (Hasternath 1988, Thapliyal 1981, 1982, Gowarikar et al 1991, Rajeevan, 2001, Guhathkurta et al 1999 and several others). As extreme rainfall seasons are difficult to predict through statistical methods, international and national efforts in India have been devoted increasingly in the last decade to use dynamical climate models forced by boundary condition of SST and snow cover as propounded in the pathmaking discovery by Charney and Shukla (1980) and further reviewed by Shukla (1998) and others. Climate prediction models evolved from numerical models used for weather prediction. Over the years climate prediction models diverged from weather prediction models. As such instead of assimilation of data in NWP models, climate models have laid emphasis on importance of representation of Earth-System complexities in coupled modeling approach. Using the SST boundary forcing approach enormous work has been done to predict seasonal monsoon rains over India and special projects like AMIP, CMIP, PROVOST, DEMETER, SHIVA and climate of the 20th Century have been launched in Europe and USA. Krishnakumar et al (2005, 2006) have also assessed seasonal monsoon forecast from different AGCMs and multiple coupled models respectively but consistent success could not be found. Even a special project in India under Seasonal Prediction of Indian Monsoon (SPIM) in which 5 different models available in different research organisations (IITM, Pune, IISC, Bangalore, NCMRWF, NOIDA, IIT, Delhi and CDAC, Pune) participated with 20 years of simulations with observed SST boundary

forcing. The results of SPIM did not show any significant consistent skill in seasonal prediction over India. It is now recognized that the skill through SST forcing exists over the tropical oceanic regions but it becomes even insignificant over land-locked India. There could be two possibilities for the loss of skill over India. Firstly, seasonal surface processes could be very crucial over the Indian monsoon and secondly internal dynamics of the monsoon, which operates on IS and synoptic scales, could vitate the outcome (Palmer and Anderson 1994, Shukla 1998, Goswami 1998, Webster et al 1995, Kang et al 2004, Krishnamurti et al 2005, 2006, Kumar et al 2007 and several others).

It has been now well recognized that prediction of summer monsoon rains over India remains as one of the most challenging problems in meteorology and progress is likely to proceed on slow trajectory till some revolutionary new idea is presented. Computational power is not doubt important for this effort but more than that is a totally new idea. Better than the present boundary forcing approach may, perhaps, be needed to achieve spectacular progress in this discipline. However, since the problem is so very important for India as the Indian GDP fluctuations, inspite of present Indian economy being more dependent on the service and industrial sectors, are vitally dependent on monsoon. In the drought of 2002, the agricultural GDP fell as much as 12%. Indian researchers are engaged in climate modeling for hardly over a decade and still have limited computational power. They have to pursue dynamical seasonal prediction inspite of several limitations. However any promise about the likely success is to be given in a restrained manner as extensive work in the last 20 years has not shown high prediction skill. Though the potential predictability of seasonal summer rains through boundary forcings have been well realized over Tropical Ocean, it has yet to be skillfully demonstrated over land-ocean regions like India. Perhaps internal dynamics and land surface-atmosphere ocean processes play significant role in the evolution of monsoon which are responsible for deterioration is the skills of model predictions. The systematic error of ensemble mean prediction is particularly large over the Indian monsoon region and sub-tropical Indo-Pacific region where the air-sea interactions are active. Individual ensemble members differ considerably among themselves and the confidence is even lower on monthly scale (Sikka et al 2010). Therefore, there is an opinion that the present prediction experiments may not be appropriate for correct

predictions in the Indian region (Kang et al 2004 and others). Statistical post-processing and bias correction approach for each grid point or super-ensemble technique may be attempted to enhance the predictability of summer rains over India. Indian monsoon researchers may explore model runs from centres of advanced research in Europe, USA and Japan and work on the bias removal technique for each model to enhance their skills in seasonal prediction over India. This hard and tedious work is to be done by Indian monsoon science community to explore enhancing the skill of dynamical seasonal monsoon prediction. New insights and constraints of the methodology of NWP may be brought to bear on the problem of quantifying and reducing uncertainty in SW Monsoon climate prediction. The issue being debated in scientific circles is whether the notion of seamless prediction, a core project of the WCRP strategic development in the next two decades, would be suitable for predicting climate over South Asia – a region where coupled land-atmosphere-ocean processes are crucial. Even if India adopts seamless prediction approach, enormous infrastructure support in computing and well-trained manpower would be needed, which may take us one decade to develop, even if we start today. Perhaps it would be wise to become partner in the global initiative for seamless prediction being considered under the WCRP as a World Climate Research Facility. If such a world facility is established, India could participate in it with its own agenda focused on climate prediction of the SW Monsoon.

1.2.3.7. Monsoon Climate Variability and Climate Change Research

(a) Monsoon Climate Variability

Study of climatic under the discipline of climatology for meteorological elements, temperature, pressure, rainfall etc) at surface began in India soon after a few decades of instrumental data were available. As the observations extended to higher layers of the troposphere upper air climatological elements were also prepared. Climatology refers usually to the averages of various elements based on at least 30-year data (decadal climatology) and long-term climatology may be based on multi-decadal averages of say 100 years. In India IMD has prepared decadal climatology and long-term climatology of surface elements. For SW Monsoon rainfall is the most dominant element, which has been studied since the establishment of IMD in 1875. At present there are two well recognized series of monsoon rainfall of

India – one based on the mean rainfall of 310 fixed stations, excluding the hilly stations (the IITM rainfall series after Parthasarathy et al (1994) and the other based on all IMD observatories (including hilly regions) - the IMD series (from 1900 to the present). IMD Rajeevan et al (2006) have recently prepared 1^0 X 1^0 rainfall climatology based on data of 1950-2000 and have now extended it backward to cover period up to 1900. The long-term seasonal monsoon rainfall for India as a whole is about 85 cm in the IITM series and about 89 cm in the IMD series. Year to year (inter-annual) and decade to decade (inter-decadal) averages of rainfall for India as a whole in both the series fluctuate somewhat about 3% of the long-term mean. The statistics of fluctuations is worked out as standard deviation (SD) or coefficient of variability (SD X 100 **/** Mean). For both the Indian monsoon rainfall series, the coefficient of variability (C.V) on IA scale is about 10% for India as a whole. Spatial distribution of IA C.V. over India is quite large as it is about 8% for hyper-moist NE India and even 30-40% over the semi-arid and arid parts of India.

Variations of rainfall climatology (departure from climatic mean or normal) over decades to a century are known as climate variability on IA an multi-decadal scales. IA variability (C.V.) of All India rainfall for the SW monsoon is 10%. Multi-decadal (30 years) climatic variability of monsoon rainfall for India as a whole is about 3% and records show that the rainfall in the period 1900-1930 and 1960-1990 was lower and in the period 1930-1960 was somewhat higher than the long-term normal rainfall. This had lead to the concept of epochal behavior of the monsoon rainfall of India on multi-decadal scale (Joseph 1976 and others since then). A monsoon drought occurs on the scale of India when monsoon seasonal rainfall is 10% below the long-term normal. In the multi-decades of less than normal rainfall (1900-1930 and 1960-1990) there is higher frequency of monsoon droughts. As such these decades could be called as those of incidence of mega droughts or decades of bursts of monsoon droughts more persistent monsoon droughts In the decades 1930-1960, the drought frequency was less as only 2 drought seasons (1941 and 1951) occurred in these 30 years. Again the period 1961-1990 witnessed higher frequency of drought-incidence over India. Recently, Meehl and Hu (2006) have investigated the incidence of periods of mega droughts using long control runs of global coupled models and connected them to multi-decadal SST anomalies in Pacific Ocean and tropical ocean mid-latitude connections. Sikka (1999) has

extensively studied monsoon droughts and discussed their possible causes and societal impacts. Sikka (2003 C) has also studied monsoon floods (monsoon excess years when the season rainfall for India exceeds 10% of the long-term normal). The recent severe drought of 2002 has been studied by Sikka (2003 a), Majumdar et al (2005), Bhat (2006) and others. Study of Indian monsoon climate variability has become an intensive subject of scientific pursuit among atmosphere – ocean science community in India and abroad and the dominant modes of IA monsoon climate variability have been found. EOF analysis of daily rainfall for the monsoon season for drought and excess monsoon years has been examined by Sikka (2008) and are found to follow typical patterns. Kailas and Narasimha (2000) have applied wavelet analysis to bring out quasi-cycles in monsoon rainfall. IA monsoon rainfall variability has been linked to the large-scale circulation anomalies on the regional and global scales by several workers such as Snow-Monsoon, ENSO-Monsoon, IOD-Monsoon relationships as well as with changes in circulation regimes of northern hemisphere such as North Atlantic Oscillation or blocking highs in the troposphere and in the neighbouring warm pool of the Western Pacific (frequency of formation of tropical cyclones and their tracks). Rehman et al (2007) used TRMM / TMI satellite data to study evolution of contrasting monsoon years. On the regional scale monsoon drought over India is usually associated with long ‘monsoon break’ spell or more frequent occurrence of mid-season breaks in the monsoon, less number of days of monsoon LPSs etc (Sikka 1980). Decadal changes in SST forcing over eastern Pacific may also impact on decadal monsoon variability (Kucharski et al 2006). As yet studies on multi-decadal monsoon variability are rather few but it is feasible that this could be linked to 11-22 years solar cycle, lunar tidal cycle, ocean-atmosphere interactions comprising of ENSO and IOD occurrences, natural variability in ocean circulation and variations in solar flux on decadal scales or some other causes yet to be determined. It is hoped that by a better understanding of cycles of variability of sub-systems of the coupled global climate system it may be possible to understand the causes of multi-decadal monsoon variability. Thus understanding of the mechanisms for decadal scale monsoon rainfall over India is a topic which would need attention of researchers of India. The internal dynamics of the coupled climate system and externally forced climate variability interact on decadal scale. Decadal prediction combines both the need for accurate initial conditions and the need for climate change scenarios. Then the expertise of seasonal prediction and centennial

scale climate change scenarios is to be pooled to develop methodology for decadal prediction of the SW Monsoon.

(b) Monsoon Climate Change

Although instrumental records of monsoon rainfall of India do not exist prior to mid-19th Century, there have been several palaeo-climatic studies, based on proxy data (pollens, tree rings, corals etc) which suggest that the monsoon rainfall in the past had changes on multi-century to multi-millienia scale. Some of these studies were reviewed in the Special Issue of Mausam (2001). As scientific data are not available to determine causes of climatic change due to natural causes, researchers have used AGCMs to study climate changes under ice ages due to changes in orbital parameters of the sun or Holocene warm period by prescribing ice-age boundary conditions or the orbital parameters of sun in the mid-Holocene warm period. Results of ice-age boundary conditions experiments with AGCMs indicated that under ice-age boundary conditions the SW Monsoon circulation did not extend beyond near-equatorial region. On the contrary in mid-Holocene warm period, (Kutzback 1982) monsoon rains were atlealst 50% more than in the present climate and the climate of the Thar desert was definitely more moist.

The present emphasis on the anthropogenic (human-induced) climate change (not due to natural climate change) has emerged from 1980s onward as the Physical Basis of the Climate was presented under the World Climate Research Program (WCRP) which linked climate to interactions among different components of the climate system (Atmosphere, Ocean, Cryosphere, Biosphere). These interactions are very complex and occur on all time scales. There has been a major shift from earlier study of climatology to the study of present climate science or climate dynamics. As a result of this new international programs under the WCRP (TOGA, CLIVAR, WOCE, Cryosphere and Climate etc.) and under the International Geosphere-Biosphere are being implemented since 1980s. The emphasis of these programs is to understand climate variability on IA to inter-decadal scales with a view to their predictions. Thus climate prediction is the new horizon to be achieved in the next two decades. The issue of uncertainty in the climate prediction, just like it is for weather prediction, has also gained importance in recent years. Multi-model or

super-ensemble climate prediction is offered as a possibility to reduce uncertainties in climate prediction. The decadal scale prediction of the SW Monsoon is only in the future for assessment of its success. The job of climate prediction is to say what the climate would be decades or centuries from now. The answer to this question has become very relevant to the societies under the threat of anthropogenic climate change. The answer depends on physical factors and impinges on important economic decisions which the societies have to take to overcome the threat of climate change. Physical factors like GHGs emissions depend on human activities which may undergo changes in the next 100 years or so in respect of industrial and agricultural production, transport and energy sources and so on. On the basis of such factors and even demographic and geopolitical forecasts scenarios are constructed. Climate models are then run under such scenarios to calculate the associate climate. This is the essence of climate prediction.

Application of AGCMs and CGCMs to understand anthropogenic climate change has progressively increased to prepare global and regional climate change scenarios under increasing load of different greenhouse gases (GHGs) and even with the inclusion of scattering and absorbing aerosols. This research was steered by the IPCC at the international level. Anthropogenic global climate change, which was considered as only probable in 1980 has now in the latest IPCC Report (AR-4, IPCC 2007), is considered as certain in terms of global surface temperature change of about 2 to 3^0 C by the end of the 21st Century, unless action is taken to reduce the GHGs emissions.

In India, concern was expressed about climate change in 1886 by Blanford and in 1910 by Walker. Blanford (1886) had examined the incidence of monsoon droughts over central India as possibly resulting from deforestation which had occurred during 1860-1880 period for using wood (deforestation) for the building of rail coaches and railway sleepers for the Great Peninsular Railways. Blanford conducted experiments around two protected forests for several years and based on the analyses of data discounted possibility of large scale decrease in rainfall by deforestation. Walker (1910) showed that for climate change to be detected rainfall series for several decades would be needed to remove the natural fluctuations in monsoon rainfall. Parthasarathy and Dhar (1976), Sarkar and Thapliyal (1988) and

Thapliyal and Kulshretha (1991) examined the multi-decadal rainfall series of Indian monsoon rainfall available by that time (unlike with Walker) and they did not find any trend in annual and monsoon rainfall of India but presence of only a fluctuating rainfall regime on multi-decadal scale. Other observational studies since then have also discounted presence of any signature for anthropogenic climate change in monsoon rainfall series. Recently Guthakurta and Rajeevan (2008) have examined the monsoon rainfall and have found that only 3 sub-divisions viz. Chattisgarh, Kerala and Jharkhand, showed decreasing trends in last 100 years. On the monthly scale, they have found increasing trend in June and August rainfall and decrease in July rainfall in several sub-divisions but statistically not yet significant. However, more careful analyses of multi-decadal trends in monthly rainfall series over different parts of India are needed before any firm conclusion can be drawn. Multi-decadal scale variability in All India monsoon series is now well recognized as there was an increasing trend in rainfall between 1901-1960 and a decreasing trend between 1961-2009. Zonal winds at 850 hPa level over Peninsular India have been also found to have decreased between 1950-2000 period and the number of monsoon depressions have also shown decreasing trend since 1950s. Number of good monsoon years have been also higher during 1921 to 1961 period, and number of bad monsoon years declined in these four decades. No. of bad (good) monsoon years have increased (decreased) between1962-2002. Since 1988 there has been only one excess monsoon year and for most of the years rainfall over India has been on the negative side of the normal. The present continuation of negative trend of the last 40 years, even in the presence of global warming, is somewhat puzzling.

With regard to surface temperature change since the beginning of IMD's observational network, some research workers in IITM and IMD have shown that the surface temperature change (warming) for India as a whole over the last century has been about $+0.6^0$ C per century which is close to what is the global average. Similarly some researchers in India have examined trends in the monsoon depressions and tropical cyclones but except for decadal scale fluctuations, which may have been due to multi-decadal climate variability, no long-term signal could be firmly established. Recently, Ramesh and Goswami (2007), based on re-analysis data, have suggested reduction in temporal and spatial extent of the Indian summer rainfall. This may be due to multi-decadal scale variability of rainfall in the last 40

years or so rather than a signature of anthropogenic climate change. There have been other studies on the impact of climate change / global warming on Himalayan cryosphere as data in the last 100 years have shown a decreasing trend in Himalayan glacier melt over India, Nepal, Pakistan and China. (Kulkarni 2007, Hasnain 2007 and others). Also the snow extent over Himalayan in winter / spring seasons has shown decrease both in observations and modeling results. Changes in Himalayan cryosphere have implications on the water resources of major Asian rivers like Indus, Ganges, Bramahputra, Mekong and Yangzee.

By mid-1980s AGCM experiment runs for climate simulation under doubling CO_2 were published by some centres in USA, Sikka and Pant (1991), for the first time, used these runs to determine climate change signatures for rainfall and temperature over India. Later other workers in India (Lal et al 2001) used AGCMS outputs from USA, Germany and Australia to show that the Indian monsoon seasonal rainfall in double CO_2 scenario may increase by about 5% without sulphate aerosol and many even decrease by 5% by including anticipated increasing load of sulphate aerosols. Ramanathan et al (2005) showed that with increasing black carbon aerosol loadings, monsoon drought frequency would increase by 2040 and rainfall may decrease but after 2040, the increase in CO_2 would overcome it and by 2080 Indian monsoon season rainfall may even show some increase. Kriplani et al (2007) have used the outputs from IR-4 coupled atmosphere-ocean models and found that only about 4 models show reasonable agreement in the model rainfall climatology and its IA variability over India to the observed climatology. They to study anthropogenic climate change scenario for India under double CO2 climate scenario then used these 4 models. Rajendran and Kitoh (2008) have used a super-high resolution (20 km) global model of MRI Japan to study Indian monsoon in future climate projection. They reported a projected overall intensification of all India monsoon rainfall with strong regional modulation in response to the anticipated increase in GHGs contributions. According to this study spatially varying rainfall patterns may occur with widespread increase over interior India and significant reduction in orgraphic rainfall over west coast of India and NE India. The model also projected substantial but spatially heterogeneous increase in extreme (heavy) rainfall events over parts of India by the end of the 21 Century. Annamalai et al (2007) have also examined the impact of global warming in AR-4 GCM experiments in respect of

ENSO-Monsoon relationship. Studies using a Regional Climate Model (RCM), nested with an AGCM, have shown increase of rainfall with more abundant water vapour to the southern portion of India. IITM scientists. Rupakumar et al (2006), used a RCM nested with U.K Met. Office AGCM and found that in double CO_2 there would be 5 to 10% increase in Indian monsoon rainfall but decrease in number of rainy days which meant that the heavy rainfall spells may increase. Increase in non-rainy days may have serious consequences for agriculture and increase in heavy rainfall days may lead to more rivers in floods under this climate change scenario. Several other papers on the subject have been published (Stephenson et al 2001, Pal et al 2001, May 2002, Meehl and Arblaster 2003, Knopp et al 2008 etc). Goswami et al (2007) while analyzing the gridded rainfall series of India found that increased frequency of heavy rainfall over Central India showed increase in recent decades between1950 to 2000. This has occurred despite land use changes over several parts of India which impact on climate variability and change. However, when Sikka (2009) examined the data extended till 1900, he again found multi-decadal fluctuations in heavy rainfall events over central India. According to his analysis in earlier decades too frequency of heavy rainfall was close to that found by Goswami et al (2007). Recently, Tewari (2008) has examined the impact of land use changes in Himalayan Lake region. Similar impacts need to be studied over other Indian region too like coastal areas and the Indo-Gangetic Plain.

Anthropogenic climate change is likely to impact on several sectors like agriculture, water resources, ecology, medicinal plants, Himalayan glaciers, desertification, coastal zones etc. Several Indian authors have written sector-specific papers on the impact of climate change.

Sinha (1996), Gadgil et al (1999), Lal et al 1998, Aggarwal (2003) and Sikka (2003), Krishna Kumar et al (2004), WWF (2003) and others have assessed the impact of climate change on agriculture. Attri (2006) has provided an overview of climate change and agriculture over India. Rise in surface temperature under global warming may add to the problems in agricultural productivity by increase in evapo-transpiration and hence increase in crop water demand.

Attri and Rathore (2008) and others have suggested decrease in wheat production over India as a consequence of global warming under climate change. Several papers have been also written on the mitigation of climate change in India. There are even suggestions that increase in anthropogenic aerosols in recent decades have made negative impact on rice production (Auffhammer et al 2006). Increasing concentration of surface ozone due to anthropogenic causes may cause produces high toxicity to reduce crop production. Climate and Health (Sikka and Kulshrestha 2005) is another emerging area of research in India. Effect of climate change on human health is an area where Indian meteorologists and medical specialists can work together.

Human-induced Climate Change of monsoon is of strategic importance to several sectors of the Indian economy. It has policy implications too. The Government of India is ceased of the possible emerging impacts of climate change upon India's development and fight for poverty reduction. It has established several missions to address the problem. One of the missions is ceased of the strategic knowledge on climate change. Sustainability of India's economic development in the face of climate change threat is the challenge of next two to three decades. Combined efforts of research organisations in several government departments and collaboration among climate scientists and social scientists as well as economists will help policy makers to face climate change threat and even mitigate it. Atmospheric scientists need to work closely with social scientists and economists to help in framing long-term policies for reducing the impact of climate change. At the IITM, Pune a separate Centre for Climate Change is being established in 2009-2010. Similarly a Centre for Glaciology has been established at Wadia Institute of Himalayan Geology, Dehra Dun and the establishment of an Institute for Himalayan Glaciology is under consideration of the Government. Indian meteorological community has a role for contributing to the growth of climate change science in the next decade and providing inputs for policy initiatives to combat climate change.

1.3. Assessment of Progress – Contemporary Scene, Future Prospects and Concluding Remarks

Indo-Aryans, who were the early inhabitants of India some 5000 years ago, used to look up at the sky, observe the clouds, look at the fluttering banner on the mast to help an eye on the changing wind and feel the rains. They had no measuring instruments but with their intuition and logic for theory they even tried to forecast weather by building relationship with the phase of the moon and other heavenly bodies as their societal functions depended on weather (rains). Mythology and meteorology remained mixed since the 5th Century B.C. as specific gods were believed to control specific weather phenomenon. There are records to suggest that in India some kind of a rain-gauge was being used by the 4th Century B.C. to measure rain and its annual / seasonal distribution on All-India basis. From those early observations to the early 19th Century knowledge on the SW Monsoon was based on experience passed on from generation to generation of learned people and the weather forecasts on fortnightly to seasonal basis were built on the basis of astronomical considerations. Instrumental observations did not exist. Even though instruments like thermometer, hydrometer, barometer and anemometer were invented in Europe by the 17th Century, they were not introduced in mainland India till the end of the 18th Century.

Research on the SW Monsoon had begun in the 17th century, based on instrumental observations of the data collected by trading ships. Theoretical meteorologists of the 18th and 19th centuries contemplated upon observations to understand the underlying causes of the monsoon and assigned them to the thermal contrast in summer between the land and ocean and rotation of the earth. In the later years of the 19th century with the establishment of an organized system of collection and keeping of meteorological observations over India, and their scientific analyses, under the central authority of the IMD, the knowledge-base began to be quantified and has increased substantially over the years. It was further extended into the tropospheric wind and thermodynamic regimes in the first half of the 20th Century. From the beginning of the 20th Century Meteorology has taken giant steps in India. Long Range (Seasonal) monsoon rainfall forecasting began in 1880s before even the short-range weather forecast system was organized in India which acurred

towards the end of the 19th Century. Long-range weather forecasting was put on firm statistically based by 1920s by the work of Sir Gilbert Walker. Operational short-range weather forecasting remained anchored to synoptic-dimatological (empirical) techniques till the end of the 20th Century (for over 100 years) though developmental research on NWP began in India in organized manner by 1970. NWP has made great strides in India since then and particularly after the establishment of NCMRWF in 1988, global model began to be used for operational purpose for forecast upto 5 day since 1994. At present a fairly high resolution model T-254 / L64 is being used with data assimilation of different kinds into the model and a lower version model (T-80 / L-16) is used in ensemble mode. AGCMs began to be used for research in different research centers in India on monsoon simulation and dynamical prediction for seasonal monsoon in a vigorous manner by year 2000. Meso-scale models with data at the boundaries coming from AGCM's, were introduced for research by 1995 and are now in operational use for prediction of intense weather events and district-level weather forecasting.

Improvements in observational and infrastructure, particularly after India's Independence, support helped in improving knowledge base on the SW Monsoon and in better weather forecasting. New and extensive observations helped in determining the large scale components of the tropospheric circulation (global and regional) which are considered important for maintenance and modulation of the monsoon weather. Phenomenological studies on various facets of the monsoon, as discussed in some detail under section 2, and including these not discussed have added a new dimension to the complex multi-scalar interactions within the regional system (diurnal to IA scales), responsible for the monsoon variability on the respective scales. It has been shown that the SW Monsoon interacts strongly with neighbouring regional systems (mid-latitude circulation in N.H and S.H, near-equatorial region of the Indian Ocean, tropical western Pacific circulations emanating on the warm pool). Even the global aspects of the monsoon have come to light and the role of ENSO-Monsoon, IOD-Monsoon, Cryospere-Monsoon, Aerosol-Monsoon, Land-Surface-Monsoon, Biosphere-Monsoon etc.have been investigated. All the above aspects have improved our knowledge and understanding of monsoonal process enormously since the establishment of IMD in 1875 through national and international efforts. However, a thought among well meaning critics lingers that the

operational weather and climate prediction on the large-scale, synoptic scale and local scales on different time domains has not shown corresponding improvement in skill. Prediction on the extended range scale (a few weeks ahead) are yet in infancy though effort is underway in India (Goswami et al 2007 to use an analogue-based technique with OLR data and abroad Waliser et al and others (2005) based on the development and passage of MJO. The forecast are still not specific in the eyes of the users and need improvement. Partly this impression is built because in India quantitative aspects of weather predictions are not fully adopted to bring the performance of model prediction on disturbed weather days, which really matter, although a beginning has been made with respect to prediction of the position of the tropical cyclones and to some extent with regard to long-range monsoon forecasts. Also the weather dynamics in convectively driven tropical region like India is very complex. However, it is necessary and even possible to substantially improve the accuracy of weather forecasts over India. The science is well known, the observing systems in India are under modernization in IMD and other agencies, the high resolution AGCMs and meso-scales models are available and can be further utilized under collaboration with other global operational and research centers. However, lack of sufficient well-trained scientific capacity in the county is one limitation, which could be removed by a concerted effort of national agencies under governmental stimulus and support. There are too few well-trained scientists in the country pursuing many monsoon research areas and as such the research efforts remain sub-critical. Most of the data being used for monsoon research is based on weather satellites and other analysis products from USA, Europe and Japan as there are bottlenecks in accessing the data from India, though efforts are on to remove these bottlenecks. It is hoped that the current plans in India have the potential to provide Indian society with the benefits of advances that have taken place, since 1980s, in weather and climate prediction in advanced countries. One of the biggest problems, recently recognized in weather and climate predictions is uncertainty in prediction. Different models provide quite different forecasts to monsoon transients and seasonal prediction coming out of resolution, physics, data assimilation etc. in models. Palmer and Williams (2008) have suggested stochastic physics in climate modeling to overcome uncertainty. Park et al 2008 has suggested new ways for combining ensembles, which are different from super ensemble approach of Krishnamurti et al (2006). The next decade research would confront the uncertainty

in predictions and reduce it so that forecast users can depend on the forecasts more profitably for societal use to project India's image of a developed country by 2020. This could help in several sectors of the economic development as well as reducing loss of life and property when disastrous weather events or extreme climate events strike the country. The threat of climate change in the SW Monsoon System looms large and opinions at the expert levels differ. The gaps in different climate change scenarios need to be reduced so that effective policies are introduced for mitigation and safeguarding the Indian economy from climatic shocks.

Our review has shown that Indian researchers have to move forward leaving their restrictive regional and synoptic approaches to adopting inter-regional global approaches and model-based studies. Reliable weather prediction during the SW Monsoon season continues to be critically important for protection of life and property, particularly under high impact weather conditions. Recently the need for monsoon climate prediction has increased dramatically as the issue of human induced climate change has gained importance. Monsoon climate predictions, on skillful basis with high reliability, are needed to guide policy and prepare for extreme monsoon seasons. Investment decisions to install infrastructures to adapt to climate change have also become critical. Coupled processes within the atmosphere, ocean, cryosphere, land and biospheric continuum of the earth system influence monsoon weather and climate and they have to be understood and modeled in the coupled context. Observations on the total earth system are needed for understanding and predictions. Synoptic understanding is very useful to determine the effectiveness of models. More coordinated programs like the ones recently undertaken for multi-model ensemble forecasting and district-base weather forecasting have to be undertaken. Model diagnostics should be adopted to carefully determine the differences evolving in development of synoptic and large scale systems and compared to the knowledge available about them through observations. If dynamical structures of systems do not evolve realistically (close to observations) in the models with growing forecast times, and model energetic differ from the observed ones, the reason for these vital differences, which vitate the forecast quality, have to be understood. Physical parameterization for convection is crucial and differences amongst them provide significantly different forecast. Proper season-base convective schemes and even synoptic system-base schemes must be determined

for use in operational NWP models. Data assimilation is very crucial for improving the quality of weather forecasts on all spatio-temporal scales. Weather and Climate Satellites such as the A-Train constellations (Parasole, Calipso, Aqua, Cloudsat etc.) with their global sweep are providing vital data for monitoring the earth's planetary ecosystem. Earth System models of the near future would be able to assimilate these data sets. Space-based satellite observations are now indispensable for a global / regional monitoring of weather and climate. Models need data on all scales to test their validity in forecasting. Climate science is a young multi-disciplinary science. Indian climate scientists have to march together with international climate science community to advance climate prediction from inter-annual to centennial scales. A well-coordinated effort is needed between operational and research centers in India to work on the important area of the understanding and prediction of weather and climate of the SW Monsoon system. Increasing use of satellite data and precipitation data in assimilation systems, as proposed by Errico et al (2007), offer tremendous possibility for further research. Diurnal variability of rainfall occurs on continental scale as suggested in several observational studies (Krishnmurti and Kishtawal 2000 and others). Model rainfall forecasts must be diagnosed to compare the diurnal variability of rainfall in the models vis-a-vis observations. Indian weather satellite data have not been effectively used so far for monsoon studies and only a handful of studies have used satellite products. Data access for this purpose at the hand of young scholars in the universities needs considerable improvement. A revolution in weather and climate prediction is on the horizon (Shukla et al 2009) under international effort and new important programs are underway under World Weather Research Programs (WWRP), WCRP and IGBP as well as under regional collaboration such as the Asian Monsoon Years (AMY) and Indian CTCZ and CAIPEX. World is marching toward as seamless prediction system under the Earth Sciences System of Systems Programs (ESSP) which will confront the challenges of resolution, complexity, length and number of ensembles and data assimilation system. The Japanese experience of simulating global climate and weather system under non-hydrostatic dynamics and cloud-resolving physical framework is a revolutionary approach to captilise on the seamlessness of the weather-climate framework. Indian research community should be geared to take advantage of new emerging approaches on software and hardware. Computing industry is marching forward to supply the requisite hardware to the atmosphere-ocean science

community and using multi-model ensembles under operational conditions. Under the WMO's TIGGE umbrella the entire model outputs from global centers, including those from ensemble forecasting systems, is now available for researchers and even an ensemble of ensembles is on the horizon. Improved Re-Analysis Programes are in the offing. Monsoon researchers in India are encouraged to reach these new datasets for in-depth research on different facets of monsoon. The survey of literature in this article and the science plans of campaigns on CTCZ, AMY, and Monsoon-Aerosols etc. offer new insights into monsoon research. Monitoring of the equatorial Indian Ocean by a chain of moored buoys, other Ocean atmospheric, land surface, cryospheric observing systems are likely to place new data sets at the hands of researchers in the next decade. The scope of monsoon research is changing in a revolutionary manner and all these efforts are expected to improve the skill of weather and climate prediction of monsoon for societal purposes. While the skill in the monsoon weather forecasting has improved over the past two decades, uncertainty in predicting climate variability and climate change looms large. Multi model ensembles of monsoon climate prediction are to be diagnosed and skills improved by removing biases.

Understanding monsoon climate change is crucial to help address problems facing rapidly expanding Indian economy which is increasingly using energy and hence GHGs emissions are increasing from India even though India's per capita emissions are still very low. Therefore, science and technology has to play a dominant role in defining solutions to monsoon climate change aspects also. India needs a policy framework and a comprehensive approach to climate change so that actions and policy converge. In this effort climate scientists have important role as in better and better climate predictions economists and social scientists would provide credible solutions to policy makers for combating any possible monsoon climate change. Decision-making would require more precise information so as to narrow down the uncertainty limit.

The new challenge of seamless weather and climate prediction can be more effectively met under a global program in which developing nations should have stakes depending on their priorities to reduce uncertainty in regional weather and climate predictions. Better access to weather and climate information systems

nationally and globally would help in addressing to societal needs. High quality sustained observations on weather and reliable climate monitoring systems would facilitate research and operations. Additional components in the observational strategies like aerosols; chemistry cryophilic status and dynamic vegetations ought to be added to provide a comprehensive picture of monsoon weather and climate regimes. Scientific capacity must be strengthened through national effort and international collaboration to take full advantage of the possibility emerging in weather and climate research and operations. Awareness is required to be created among professional's users of weather and climate forecasts and the lay public about the present status, complexities and future prospects. Looking ahead exchange of data, scientific information and knowledge as well as national and international cooperation would help in further advancement of research on the SW Monsoon. The SW Monsoon has been recognized as a dominant circulation in the northern hemisphere summer being the seat of organized convective heating of the atmosphere. There has been some success in modeling it as a coupled atmosphere-ocean-land system, understanding its variability on different spatio-time scales and also in its prediction on synoptic and meso-scales. However, much is yet to be achieved with respect to its skillful prediction on extended, seasonal and Inter-annual to decadal scales and the global atmosphere-ocean science community has accepted as a challenge to improve its predictability in the coming decade. Assessing the impact of global warming on its multi-faceted performance is another big challenge. We are hopeful that by 2020 when the achievements of the next decade are documented by someone, he would conclude with great optimism the achievements of the coming (2010-2020) decade just like we have provided the progress on SW Monsoon Research since the famous monograph on SW Monsoon by Y.P. Rao written 3 decades ago.

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CHAPTER 2

MONSOONS ELSEWHERE IN THE WORLD

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2.1. Introduction

The phenomenon which is commonly known as the monsoon can be looked upon, from the meteorological standpoint, as the response of the state of the atmosphere to the annual migration of the sun relative to the earth, between the Tropic of Cancer in the northern hemispheric summer and the Tropic of Capricorn in the southern hemispheric summer. The monsoon is actually an extremely complex and intricate combination of physical processes that operate not only in the atmosphere, but involve land and ocean as well. Monsoons should therefore be found over the entire tropical belt of the earth and possibly even extend into the sub-tropics. That we do not find a monsoon everywhere in this region, is more because of a restrictive definition of the monsoon than otherwise.

The first scientific explanation of the monsoon was postulated by Sir Edmund Halley, a British mathematician and astronomer in whose honors a comet has been named. Drawing from the experiences of various mariners and navigators who were well-acquainted with conditions in the tropics, Halley made an extensive analysis of the global patterns of trade winds that blew over the Atlantic, Pacific and Indian oceans and the seasonal change of wind direction associated with the monsoons. In 1686, he presented to the Royal Society in London his own hypothesis (Halley 1686) that the monsoon was caused by the differential heating between the Asian landmass and the Indian Ocean. In other words, the monsoon has the character of a giant land-sea breeze that reverses its direction twice during a year. In April, when the sun starts heating the land, the southwest monsoon begins and blows until October; then the land cools and the northeast monsoon blows in the winter until April.

The beauty of Halley's empirical proposition lies in its simplicity. That is why it has survived for more than three centuries and is still talked about. It is difficult to

discard it altogether and it is even more difficult to offer an equally elegant alternative. In fact, land temperatures over the Eurasian continent and sea surface temperatures over the Indian Ocean are the two factors that have continued to dominate all efforts to understand and predict the monsoon, but of course in an increasingly complex manner. However, Halley's definition of the monsoon is based only upon the reversal of wind and it does not involve rainfall (Fig.2.1). From the practical point of view, rainfall is a very important byproduct of the wind reversal process, perhaps more important than this process itself. In this respect it would be unfair to blame Halley as in his times; a global observation system hardly existed. It is only now that we know that rainfall over the entire tropical belt and the adjoining subtropics shows an annual oscillation (Fig. 2.2 and 2.3).

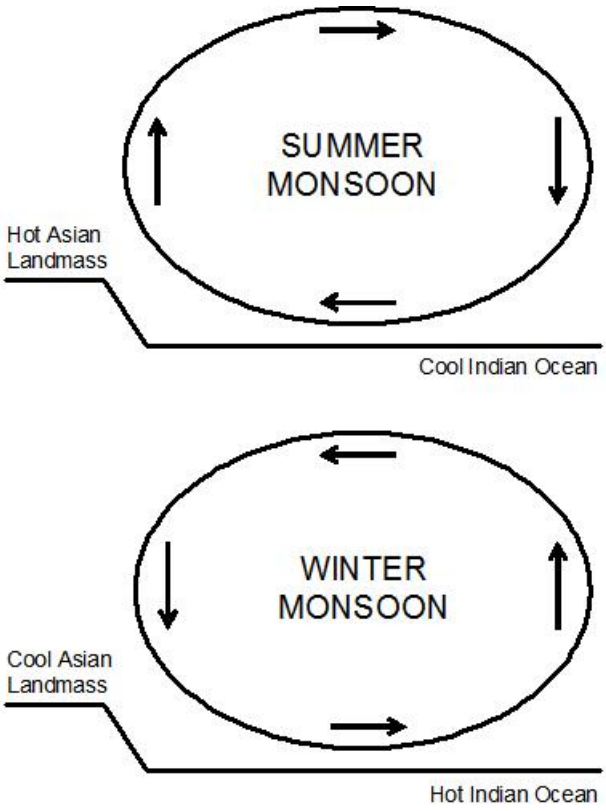


Fig. 2.1: A conceptualisation of Halley's hypothesis that the Indian monsoon is a giant land-sea breeze between the Asian landmass and the Indian Ocean (from Kelkar 2008)

The oceans had been a generally data-sparse region for long, but particularly so in the case of rainfall. It was only after the launch of geostationary meteorological satellites like INSAT, that it became possible to indirectly estimate the large scale precipitation over the oceans (Arkin, Rao and Kelkar 1989) and the real breakthrough came with the Tropical Rainfall Measurement Mission (TRMM) satellite that carried 3.onboard precipitation radar. Under the Global Precipitation Climatology

Project (GPCP), rain gauge measurements and satellite-based rainfall retrievals have been skillfully blended and extensive rainfall data sets and statistics have been compiled. With the help of the GPCP global analyses it is now possible to quantify the seasonal alternation of the precipitation patterns in the monsoon domains of the world, something that Halley could only have conceptually visualized in 1686, in the absence of any data to support his argument. The GPCP maps which are based upon data from 1979 onwards, clearly depict how in January (Fig. 2.2), the Indian subcontinent is almost devoid of any significant precipitation, the rain belt associated with the ITCZ having shifted to the south of the equator over the Indian Ocean. In July (Fig. 2.3), on the other hand, some of the rainiest areas of the world are over the Bay of Bengal, northeast India, and the west coast of India in association with the southwest monsoon.

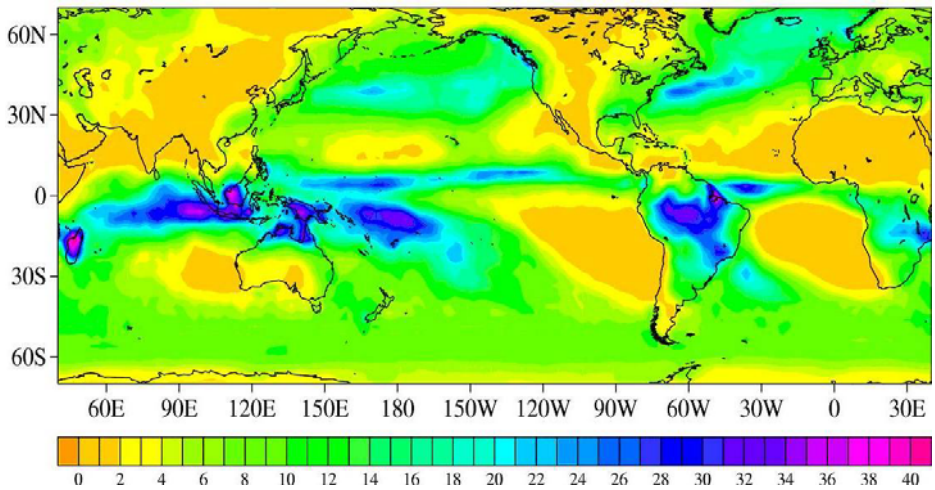


Fig. 2.2: Average January precipitation in cm (from Global Precipitation Climatology Project GPCP 2008)

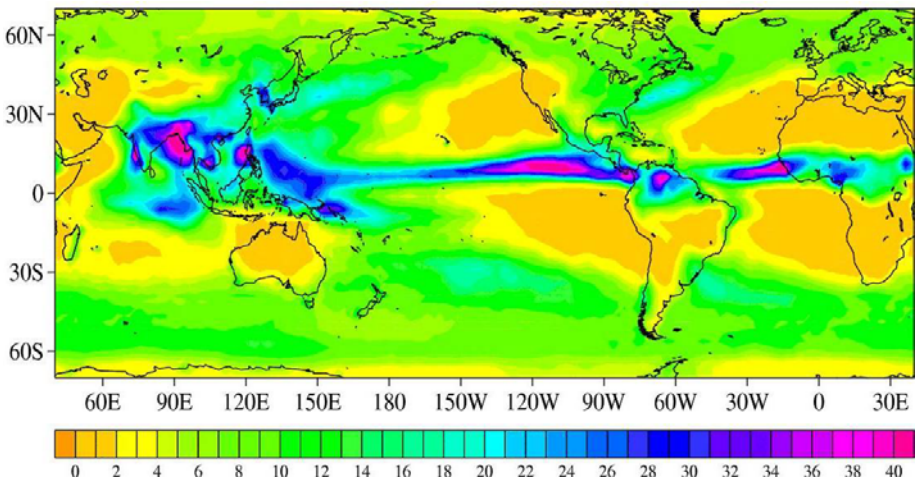


Fig. 2.3: Average July precipitation in cm (from Global Precipitation Climatology Project GPCP 2008)

2.2 Delineation of Monsoon Regions

Although by the twentieth century it was known that monsoons prevailed in many parts of the world, it was Ramage (1971) who made the first attempt to delineate the monsoon regions on the basis of objectively prescribed criteria. He stipulated that for any given geographical region to qualify being called a monsoon region, the wind patterns over it in January and July must be distinctly different and should satisfy certain basic considerations. Ramage's criteria were that the prevailing winds in these two months should blow in a preferred direction at least 40% of the time, they should have a minimum strength of 3 m/sec and there should be a change of at least 120º in the prevailing wind direction between the two months. On this basis, he delineated the monsoon region as the geographical area bounded by the latitudes 35º N and 25º S, and the longitudes 30º W and 170º E (Fig. 2.4). Although Ramage's criteria were objective, they were still arbitrary to a certain extent, and they restricted the monsoons to south and Southeast Asia, northern Australia and tropical Africa. These came to be regarded as the traditional or classical monsoon domains.

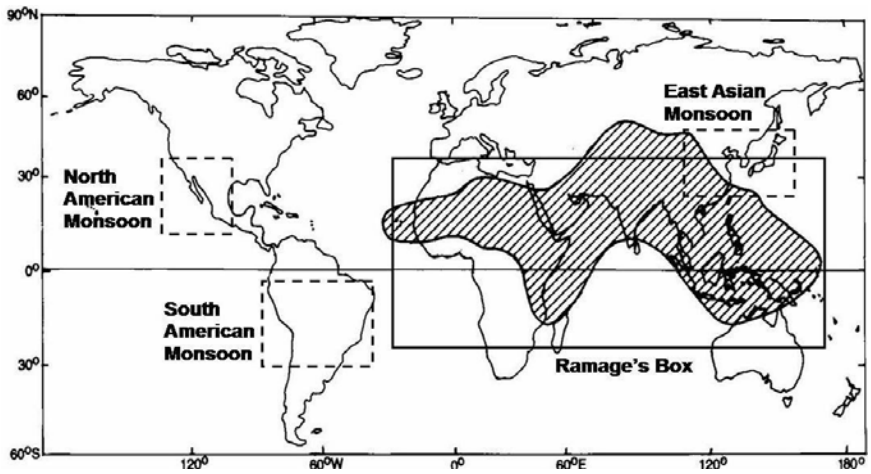


Fig. 2.4: Classical African, Asian and Australian monsoons (hatched area), Ramage's delineation, and the new monsoon zones shown by dashed lines (from Kelkar 2008)

2.3. Monsoons and the ITCZ

Perhaps the most inclusive definition of the monsoon was that given by Asnani (1993) in which the role of the Inter-Tropical Convergence Zone (ITCZ) was brought into consideration. The ITCZ that circles the globe (Fig. 2.5) is a region of lower tropospheric wind discontinuity with horizontal velocity convergence and net upward motion. While the ITCZ band has a large latitudinal width, the band as a

whole exhibits a north-south movement in association with the march of the sun, and the belt of heavy tropical rainfall also shifts along with it. The seasonal wind reversals and changes in precipitation patterns are not just confined to the traditional monsoon domains, but they also occur elsewhere in the ITCZ region.

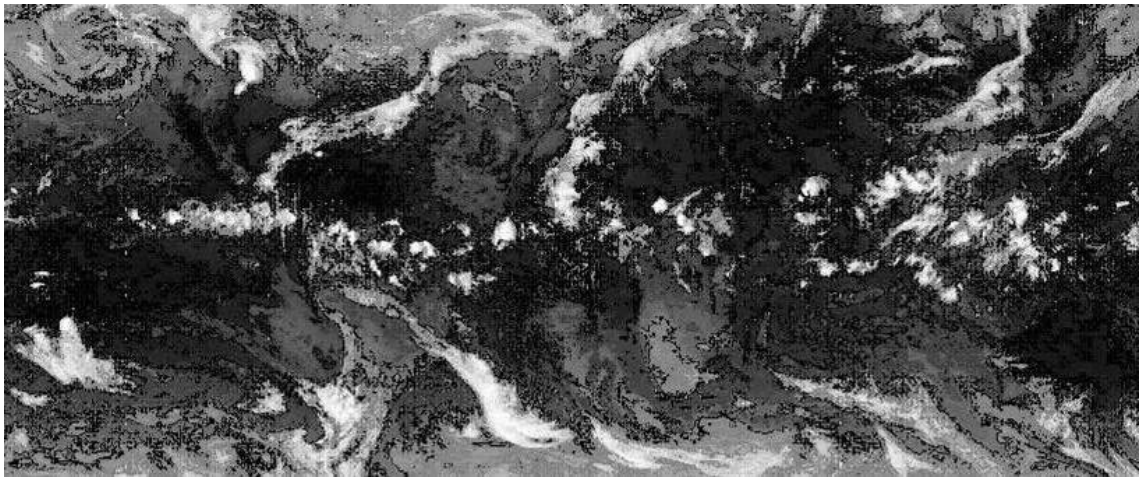


Fig. 2.5: A typical montage of satellite images showing the ITCZ as a band of cloud clusters encircling the globe (from Kelkar 2008)

An important feature of the ITCZ is the large regional variations in its seasonal alignment (Fig. 2.6). Over the eastern Pacific and Atlantic oceans, it remains to the north of the equator throughout the year. In other regions, it moves from the north of the equator in northern summer to the south of the equator in southern summer. Over land the ITCZ is located over the warmest regions, while over the sea it is located over the highest sea surface temperature (SST) regions. From west Africa to southeast Asia, there is a discontinuity between the westerlies in the near-equatorial region and the easterly trade winds on either side of the ITCZ. The westerlies are largely the southeast trade winds which have changed direction after crossing the equator. Over the Atlantic and Pacific oceans, the discontinuity is between the northeast and southeast trades of the two hemispheres.

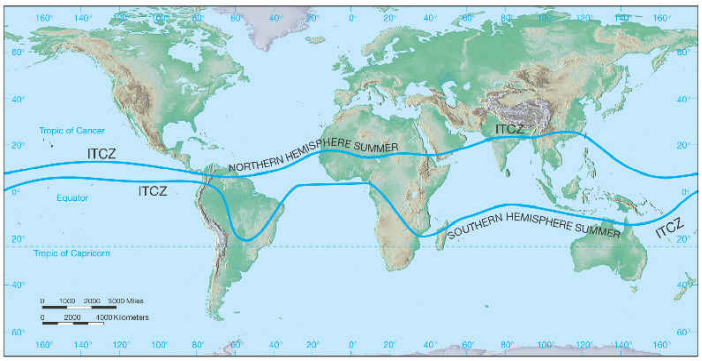


Fig. 2.6: The alignment of the ITCZ in January and July

Asnani's definition of the monsoon was simple: 'the monsoon is where the ITCZ is', but it paved the way for what could be called the globalization of the monsoon. Going by this inclusive definition, the monsoons could now be said to prevail over the entire area covering the global tropics and adjoining subtropics (Asnani 2005a, 2005b). Therefore, our contemporary view of the monsoons is not just in terms of the seasonal reversal of winds, which of course remains an important consideration, but more in terms of the amplitude of the seasonal change, which is larger over the monsoon regions than elsewhere in the tropics. The cause of the larger amplitude is that over the continental regions the land surface heating is determined by the net radiation, whereas the ocean heating is influenced additionally by the winds that drive the ocean circulation. In the non-monsoonal oceanic regions, the location of the maximum SST does not vary with the seasons (Gadgil 2007).

As far as the south Asian monsoon is concerned, the monsoon oscillation is stronger in the northern hemisphere than in the southern hemisphere, and it is stronger over south and Southeast Asia than elsewhere in the northern hemisphere. This can be attributed to the Himalayan mountains and the elevated Tibetan plateau producing diabatic heating over a large area of the middle troposphere, the Indian Ocean to the south providing abundant moisture supply and the strong meridional gradients of temperature (Asnani 2005c).

Despite the fact that over large areas of the Pacific and Atlantic oceans the ITCZ does not oscillate between the southern and northern hemispheres, the monsoon is currently being viewed as a persistent global scale overturning of the atmosphere which occurs over the entire tropics and subtropics and has an annual cycle (Trenberth et al 2000). Embedded within this global monsoon circulation are the more significant and well-known regional monsoons of Asia, Australia and Africa. Against the concept of a global monsoon, Hadley's classical notion of a large scale thermally driven land-sea breeze appears to be greatly idealized and region-specific. In fact, over some of the newly identified monsoon regions such as north and south America and east Asia, the wind reversal factor is not significant.

2.4. Asian Monsoon

The Asian monsoon in its totality is by far the largest monsoon system of the world. Over land, the Asian monsoon region includes the Indian subcontinent and the Indo-China peninsula, and extends northeastwards across mainland China further into Korea and Japan. Over the ocean, it covers the South China Sea and the

northwest Pacific Ocean. However, the Asian summer monsoon system could be said to consist of two large and distinct regional subsystems: the Indian southwest monsoon or the south Asian monsoon and the east Asian monsoon. These two subsystems are indeed interrelated, but they are also capable of operating independently of each other at times. The phenomenon of the Asian monsoon has been treated exhaustively by Asnani (2005d) and a concise description of the east Asian summer monsoon has been given by Kripalani et al (2007).

The east Asian summer monsoon is largely a subtropical phenomenon and its most significant is the quasi-stationary front that extends from south China to Japan. Different parts of this long-winding front bear different local identities: like Mei-yu in China, Chang-ma in Korea and Bai-u in Japan. The front can be traced along the northwestern periphery of the north Pacific subtropical high whose alignment and strength have a direct influence on the monsoon (Fig. 2.7). When the high extends more to the west, the low level jet to its northwest gets strengthened, bringing more moisture into the Yangtze River basin and leading to increased monsoon precipitation. The west Pacific warm pool which is situated at the southern edge of the high also influences the East Asian monsoon.

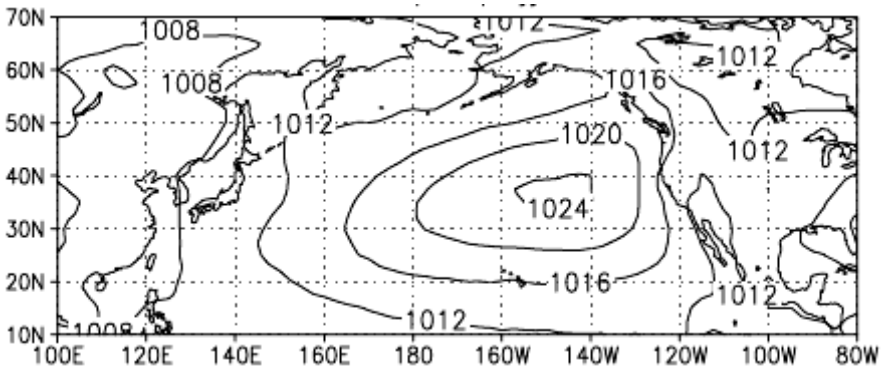


Fig. 2.7: Average June-August sea level pressure in hPa over the East Asian monsoon Region showing the north Pacific subtropical high (from Kripalani et al 2007)

The climatological mean date of onset of the monsoon over the South China Sea is around the middle of May. This results as a combination of the tropical deep convective rainfall over the equatorial region and the rainfall associated with the subtropical front over south China. After this onset has taken place, the monsoon rain belt runs from the Arabian Sea across the Bay of Bengal, through the South China Sea, to the northwest Pacific Ocean. It is associated with the characteristic southwesterly winds that connect the south Asian and east Asian monsoons. The monsoon then advances to the north and northeast across the Yangtze river basin

into south Japan by middle of June. It subsequently enters north China, Korea and north Japan. The advance of the east Asian monsoon does not proceed at a steady pace and it could be very rapid or extremely sluggish in phases. The southeastern plains of China receive a mean annual precipitation of over 150 cm. The northwestern regions are much drier, the Gobi desert getting only 25 cm of rain, and in some parts even less than 10 cm, annually. The average June-August precipitation pattern over the east Asian monsoon region is shown in Fig. 2.8.

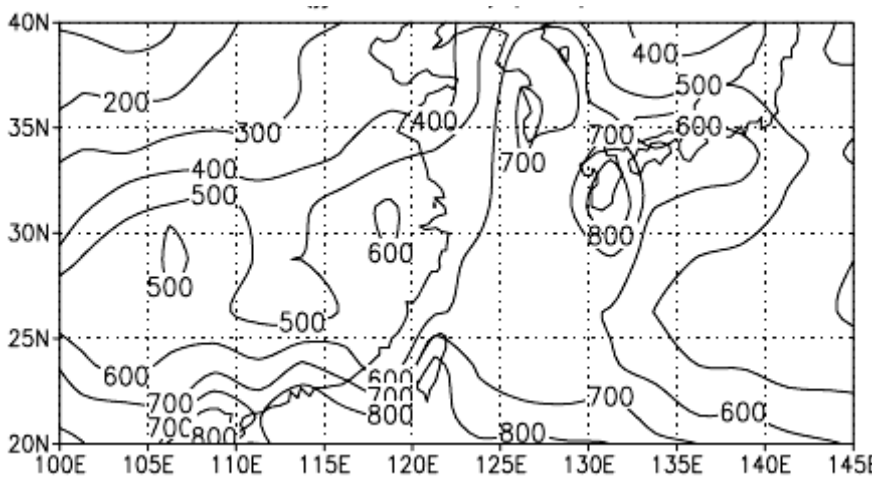


Fig. 2.8: Average June-August precipitation in mm over the East Asian monsoon region (From Kripalani et al 2007)

The Asian monsoon was the subject of intense investigation under the GEWEX Asian Monsoon Experiment (GAME) which was carried out under the overall umbrella of the Global Energy and Water Cycle Experiment (GEWEX) between 1996 and 2005. The objectives of GAME were: to understand the role of the Asian monsoon in the global energy and water cycle, to improve the simulation and seasonal prediction of the Asian monsoon, and to assess the impact of monsoon variability on the regional hydrological cycle. Japan, China and Korea were the main players in GAME. Special observation programmes were conducted over four selected regions: Siberian Taiga forests and Tundra area, the Tibetan plateau, the Huai-he river basin in China, and the Chao Phraya river basin in Thailand. The last phase of GAME was devoted to a detailed analysis of the observations and modeling studies.

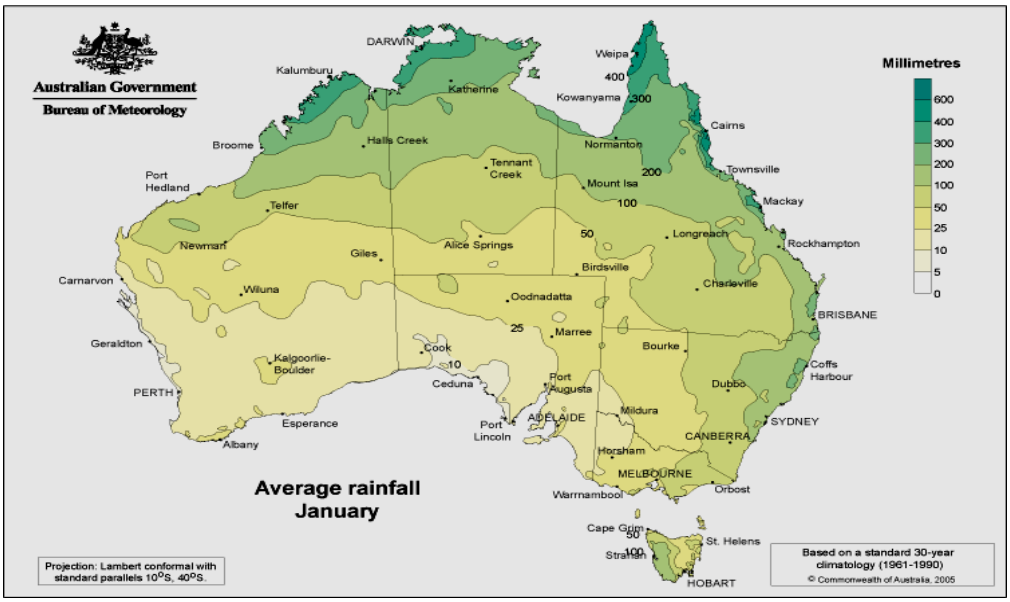
GAME has been followed by another programme called Monsoon Asian Hydro-Atmosphere Scientific Research and Prediction Initiative (MAHASRI) whose objective is to develop a hydro-meteorological prediction system for the region. The planning for MAHASRI began in 2005 and it has the support of Japan, China,

Thailand, Vietnam, Bangladesh and other countries. While MAHASRI is essentially a continuation of the hydro meteorological component of GAME, it has some wider objectives. An expansion of the target scientific field from air-land interactions in GAME to air-land-sea interactions is envisaged. The area is also to be expanded beyond the GAME area and the investigations are to cover the winter monsoon as well. MAHASRI is expected to establish a prototype system that would demonstrate the recording of hydro meteorological observations and data transmission in real time, which will help in flood management in Asian monsoon countries. It will also generate weekly to monthly scale dynamic probabilistic monsoon predictions for water resources and agricultural management in these countries. For this purpose, intensive observational programmes will be carried out for limited periods.

Another new international experiment called the Asian Monsoon Year (AMY) has been mooted recently and it has since been accorded a formal status by the international community with Japan and China playing a major role in its planning and execution. The AMY08 experiment of 2008-2009 is a crosscutting observation and modeling effort aimed at understanding the radiation-monsoon-water cycle interaction and the ocean-land-atmosphere interaction of the Asian monsoon system, and on improving monsoon prediction. The impact of elevated aerosols on the radiation-monsoon-water cycle interaction is one of the new areas of investigation under the AMY. A wide range of scientific questions are being addressed by AMY (AMY 2009).

2.5. Australian Monsoon

The monsoonal character of the climate over north Australia has been known since long. Almost 90% of its annual rainfall occurs during the southern hemispheric summer months from November to April. Fig. 2.9 shows the contrasting rainfall patterns that prevail over Australia in January and April. There is a marked seasonal shift from a dry easterly lower tropospheric wind regime in winter to a moist westerly one in summer (Fig. 2.10). The heated Australian mainland and the cooler waters of the Pacific Ocean to the north provide the thermal contrast required for the monsoon to develop. It has now become the practice to regard the north Australian monsoon as part of a larger phenomenon called the Australian-Indonesian monsoon, covering the area south of the equator to north Australia and between 100 and 170° E longitudes.



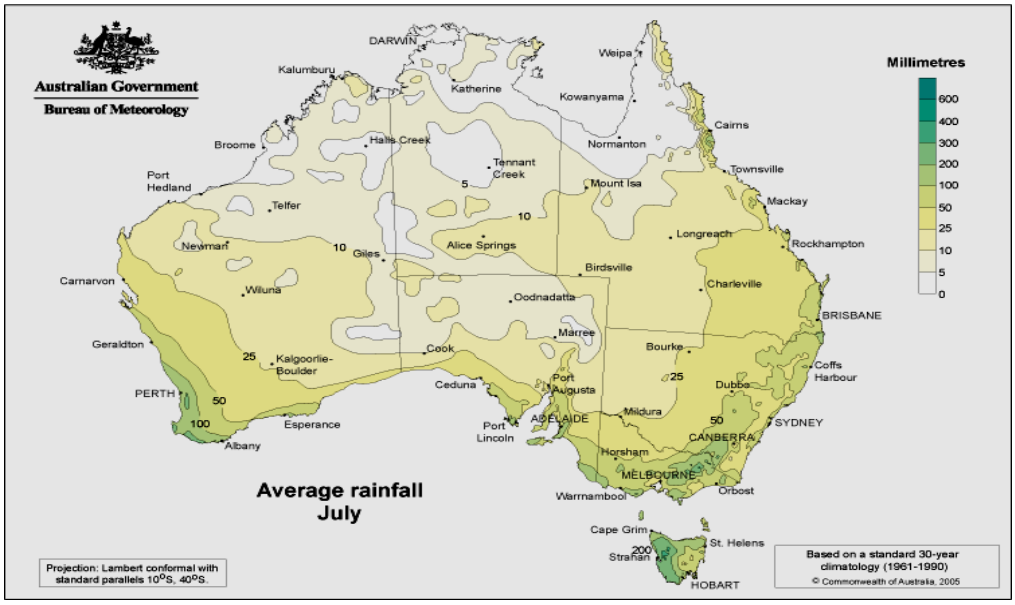


Fig. 2.9: Average January and July rainfall in mm over Australia (from BOM 209)

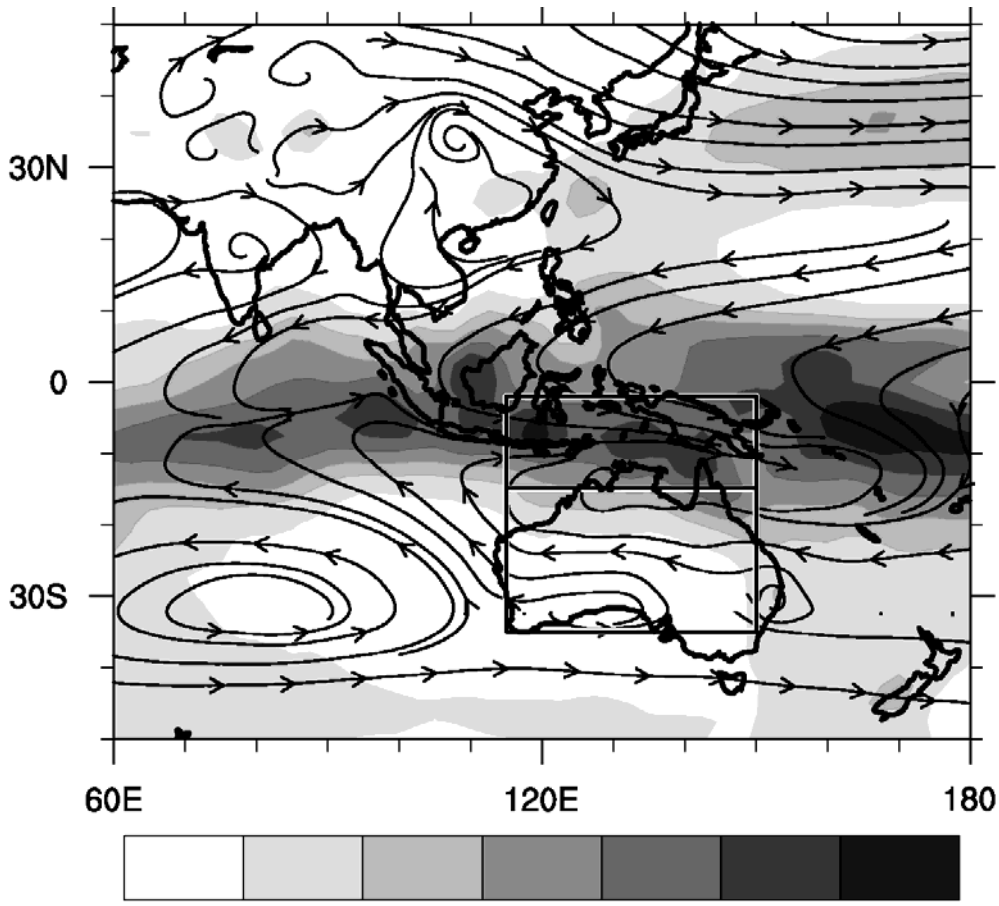


Fig. 2.10: Average Dec-Feb 850 hPa streamline pattern and precipitation in mm/day ranging from white 0-2 to black 12-14 (from Hung et al 2004)

A heat low develops over northwest Australia and the ITCZ runs through it. The maximum monsoon rainfall occurs over the region of confluence of the ITCZ and the South Pacific Convergence Zone (SPCZ). The Australian monsoon is triggered by cold surges from the South China Sea and surges in the low level southerlies along the west Australian coast. The deep tropical convection over north Australia and Indonesia provides the upward branch of the Walker circulation over the Pacific Ocean which governs the southern oscillation.

The onset date over Darwin is around 25 December but it has a standard deviation of about a fortnight (Hendon et al 1990). The onset of the Australian monsoon has been investigated in depth by Hung et al (2004) who found that the land-sea thermal contrast and barotropic instability act as a seasonal preconditioning for the onset. The onset of the Australian summer monsoon generally coincides with the arrival of the eastward-propagating MJO. When the MJO is weak or absent, the onset gets triggered by some other disturbance like the midlatitude trough.

There are many commonalities between the Indian southwest monsoon and the north Australian monsoon (Asnani 2005e). The north Australian monsoon has a well-marked onset and withdrawal. There are intra-seasonal and inter-annual variations in the intensity of the Australian monsoon (Holland 1986). There is a monsoon trough, equator ward of which there are lower level westerlies and upper level easterlies, and monsoon depressions form over the adjoining oceans. Although the characteristics of the Australian monsoon quite resemble those of the Indian southwest monsoon, the absence of tall mountains like the Himalayas in Australia makes a lot of difference and the diabatic heat supply from the land to the atmosphere remains confined to the lower levels. The net result is that the Australian monsoon is not as intense as the Indian southwest monsoon. Many specific features of the Australian monsoon were investigated during the BMRC Australian Monsoon Experiment (AMEX) conducted during 1986-1988, details of which have been described by Holland et al (1986).

2.6. North American Monsoon

In both north and South America, there are regions which receive more than half of their annual rainfall during the summer months, with a relatively dry winter season, like in other summer monsoon regions of the world (Asnani 2005f and 2005g). They also exhibit many of the basic features that characterize other major global monsoon regimes, such as the land-sea temperature contrast, a thermally direct circulation with a continental rising branch and an oceanic sinking branch, land-atmosphere interactions, intense low level inflow of moisture into the continent, and associated seasonal changes in regional precipitation.

In the monsoon regions identified by Ramage (1971), north and south America had not found a place. In fact, until the late 1970s, the very existence of monsoons over north and South America had been doubted. However, the Southwest Arizona Monsoon Project (SWAMP) which was implemented in 1990-1993, the ongoing North American Monsoon Experiment (NAME) and its 2004 field campaign, and the Monsoon Experiment South America (MESA) have all resulted in a significant improvement in the scientific understanding of the north and south American monsoon systems. It has by now been widely accepted that the monsoon does indeed prevail over parts of north and South America and it is characterized by large-scale seasonal changes in the wind and rainfall patterns.

The North American monsoon is also known as the southwest U. S. monsoon, Mexican monsoon or Arizona monsoon, or just as the summer thunderstorm season. This is because the core of the North American monsoon is actually situated over northwestern Mexico, but it influences much larger areas of the southwestern U. S. (Adams et al 1997, Douglas et al 1993).

The moisture supply that sustains the North American monsoon originates from two main sources. Middle and upper level moisture comes from the Gulf of Mexico, usually when it has warm SSTs. Lower level moisture comes from the Gulf of California, particularly when it is warm and SSTs off the Baja coast are relatively cooler, setting up a thermal gradient. Another phenomenon that comes into play is the gulf surge, which occurs when the low level jet brings moisture directly into Arizona from the Gulf of California.

The wet season begins in the middle of June and continues up to the end of September (Fig. 2.11). Over Tucson, Arizona, the total June-September rainfall is 15 cm on the average, with the months of July and August each receiving about 5 cm. However, the total monsoon rainfall can vary widely from year to year. For Tucson, the driest monsoon year was 1924 when the rainfall was less than 25% of the normal, and the wettest monsoon was in 1964 when it was 230% of the normal (NOAA/NWS 2009).

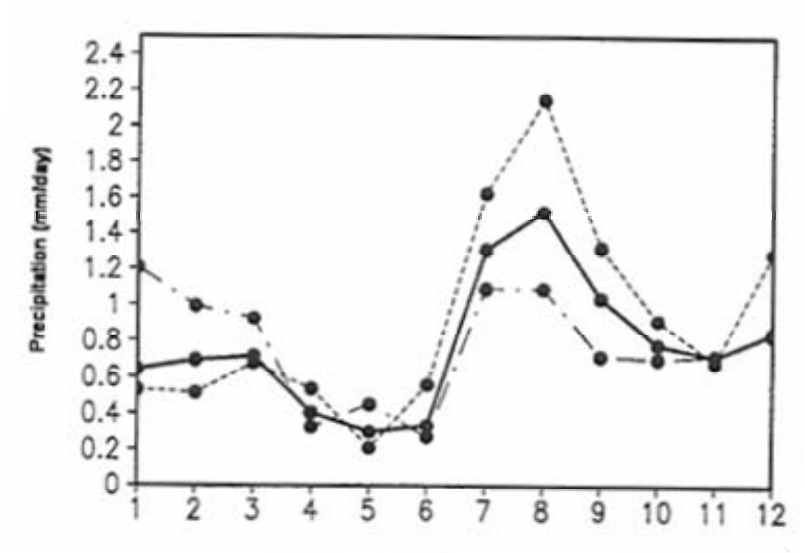


Fig. 2.11: Monthly average precipitation in mm/day over Arizona and New Mexico, dash-dot line representing dry monsoons, dashed line wet monsoons and the solid line a composite of all monsoons (from Higgins et al 1999)

The North American monsoon circulation pattern begins to evolve around late May or early June over southwest Mexico. By mid-June or later the monsoon comes to northwest Mexico, and by early July to the southwestern U. S. Large scale synoptic conditions over North America can help or block the moisture advection from the Gulf of Mexico into the southwestern U. S. In June, the 850 hPa monsoon ridges is situated over Mexico (Fig.2.12) and it actually prevents middle and upper tropospheric moisture from moving north into Arizona. However, by July, this ridge shifts northwards into the southern Plains and southern Rockies (Fig. 2.13) facilitating middle and upper level moisture incursion from the Gulf and low level moisture surges from Mexico into Arizona.

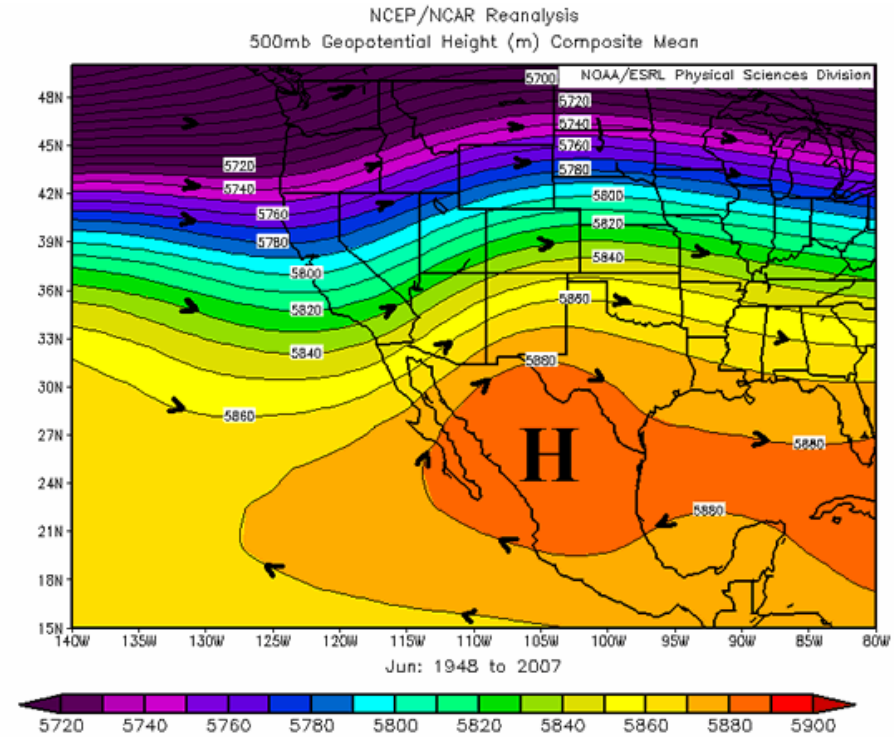


Fig. 2.12: Average June 500 hPa geopotential height in m showing the ridge over northwest Mexico which blocks the moisture in the Gulf of Mexico from entering southwest U. S. (from NOAA/NWS 2009)

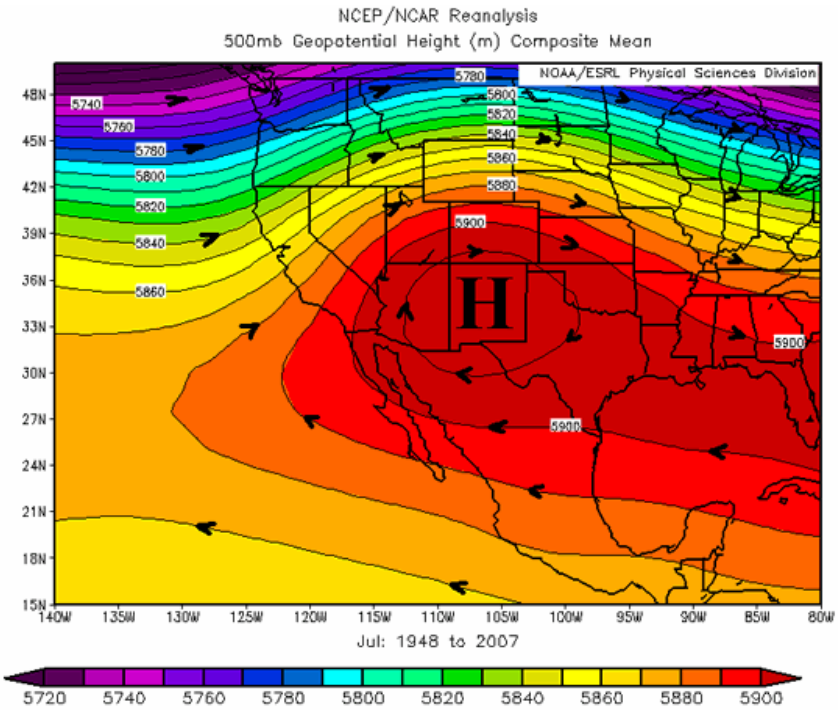


Fig. 2.13: Average July 500 hPa geopotential height in m showing strong ridge located over the Great Plains which allows moisture flow from the Gulf of Mexico into Southwest U. S. (from NOAA/NWS 2009)

As the low level moisture flow is not always very strong or continuous, it is the higher level moisture influx, which is influenced by the location and strength of the monsoon ridge that largely modulates the monsoon rainfall. This is one of the causes of the high variability of the rainfall on different temporal and spatial scales.

July and August are the months of peak monsoon activity, barring occasional breaks. There are frequent moisture incursions from the Gulf of California, commonly called Gulf surges in this period. The actual mechanism of the Gulf surges is quite complex and has been the subject of continuing research (Zehnder 2004). However, the surges are known to be produced because of an increased thermal gradient and pressure imbalance between the two ends of the Gulf and evaporation over the warm Gulf waters. By the end of August, the monsoon ridge begins to weaken and return southeastwards. The retreat is completed by the end of September, by which time the jet stream strengthens and moves deeper into the U. S.

Under the North American Monsoon Experiment (NAME), an international team of scientists from the United States, Mexico and Central America carried out a major field campaign during the summer of 2004. The aim was to develop an improved understanding of the North American monsoon system that would lead to an improvement in the precipitation forecasts. This field campaign provided an

unprecedented observing network for studying the structure and evolution of the north American monsoon. Higgins et al (2007) have given an overview of the results and addressed the motivating science issues as well as some unresolved questions. Johnson et al (2007) have studied the multi-scale characteristics of the flow from the large scale to the mesoscale using atmospheric sounding data from the enhanced observing network. Becker et al (2008) have examined the structure of the diurnal cycle of precipitation associated with the North American monsoon using a diverse set of observations, analyses and forecasts from the NAME field campaign of 2004.

2.7. South American Monsoon

Just as in the case of the North American monsoon, the existence of a monsoon over South America had not been accepted for a long time. Easterly winds prevail over eastern South America throughout the year and the seasonal reversal of surface winds is not immediately apparent. However, Zhou et al (1998) showed that when the annual mean is removed from summer and winter composites of surface winds, the characteristic reversal in anomalous low-level circulation does show up. Further supporting evidence has become available more recently, as a result of coordinated research and observational efforts, like the South American Low Level Jet Experiment (SALLJEX) organised in 2002-2003.

Most of the South American landmass lies within the tropics. However, surface conditions vary between the Amazonian rain forest and the dry Altiplano plateau. Moisture supply from the Pacific Ocean is blocked by the Andes mountain range, but it is brought in by the easterly winds from the Atlantic Ocean and by mid-latitude systems. It is now known that the South American monsoon regime does show a distinct life cycle from onset to decay, the onset usually being characterized by a sharp change from hot, dry conditions to cool, wet ones. The South American monsoon makes its onset in the middle of August over the equatorial Amazon. From there it spreads rapidly both eastward and southeastward across the Amazon basin and covers Brazil typically within just four to six weeks (Fig.2.14). November-February is the months of heaviest monsoon precipitation. Thereafter convection weakens and gradually retreats towards the equator, bringing the rainy season to an end in March.

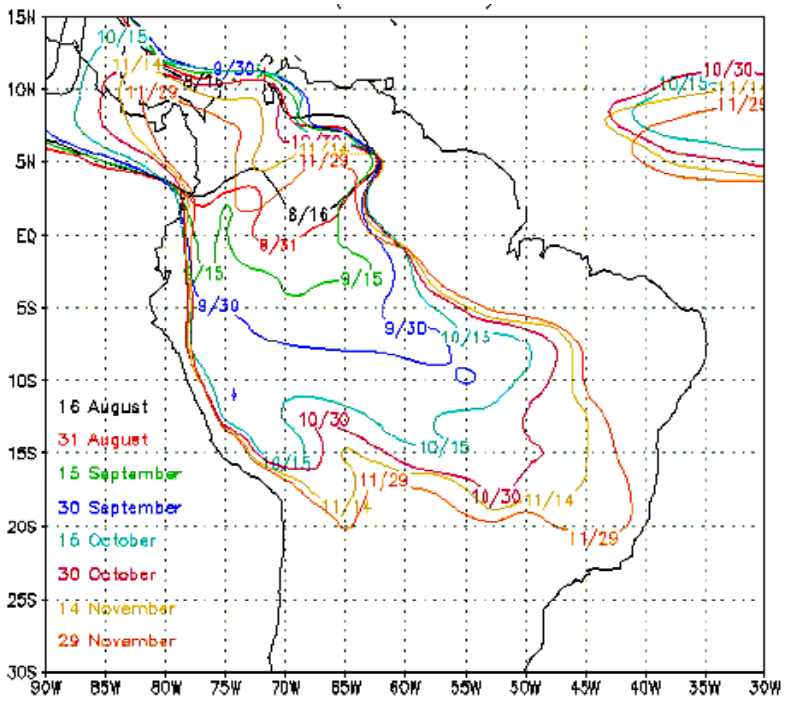


Fig. 2.14: Average dates of onset of the South American monsoon (from VAMOS 2009)

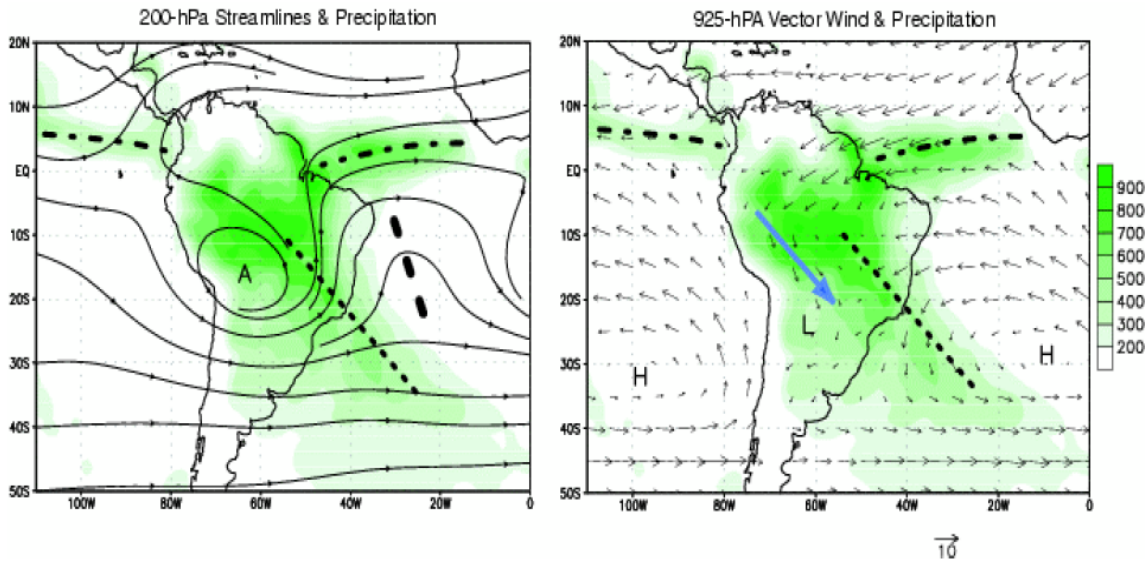


Fig. 2.15: Average Dec-Feb 200 hPa streamline pattern showing the Bolivian high (A), and 850 hPa wind pattern showing the subtropical surface highs (H), the Chaco low (L), the low level jet (thick vector), SACZ (dotted line), ITCZ (dash-dot line), the Nordeste trough (heavy dashed line), and precipitation amounts in mm as shaded green areas (from Nogués-Paegle et al 2002)

Over South America, the upper levels of the atmosphere are characterized by a high pressure area centered near 15° S, 65° W over the Altiplano, known as the Bolivian high, and a low pressure area located over northeast Brazil, called the Nordeste trough. These features are observed throughout the year but they are most pronounced during summer and at the 200 hPa level. At lower levels, there is an easterly flow from the Atlantic Ocean which is deflected southward by the Andes

mountains. During the period of peak monsoon activity, a band of deep convection known as the South Atlantic Convergence Zone (SACZ) is seen to extend from central to southeast Brazil on towards the Atlantic Ocean. SACZ is the Atlantic counterpart of the South Pacific Convergence Zone (SPCZ) which is important to the Australian monsoon. A schematic representation of these synoptic features is given in Fig. 2.15 with the precipitation pattern superimposed. The thick arrow in Fig. 2.15 represents one of the strongest low level jets in the atmosphere, called the South American LLJ or SALLJ. It plays an important role in the transport of moisture from the Amazon to the La Plata basins and has a strong influence on the precipitation.

The South American monsoon exhibits variability on the intra-seasonal scale, primarily in association with the movement of the SACZ, the location and strength of the SALLJ and the migratory mid-latitude systems. The inter-annual and longer term variability is attributable to oscillations like the MJO, NAO and ENSO.

Currently a major experiment called the Variability of the American Monsoon System (VAMOS) is in progress under the umbrella of the Climate Variability (CLIVAR) programme which in turn is a component of the World Climate Research Programme (WCRP). VAMOS is a concerted attempt towards achieving a better understanding of the north and South American monsoon systems in their totality (Higgins et al 2003, Nogues-Paegle et al 2002 and VAMOS 2009). VAMOS has launched two complementary international programmes called the North American Monsoon Experiment (NAME) and Monsoon Experiment South America (MESA). The common objectives of NAME and MESA are to understand the key components of the two American monsoon systems, their variability and their role in the global water cycle, to build observational data sets and to improve the simulation and monthly to seasonal scale prediction of the monsoons. It also has sub-experiments and field campaigns such as the South American Low Level Jet Experiment (SALLJEX) designed to study specific phenomena like the low level jet (Vera 2006).

The results of NAME and MESA have already started providing new insights into various aspects of the American monsoon systems such as moisture transport processes, structure and variability of the South American low level jet, and the diurnal cycle of precipitation in the core monsoon regions. They are also helpful in model development and hydrological applications. As research on the American monsoon systems proceeds further, a unified view of the climatic processes modulating continental warm season precipitation is expected to emerge (Vera et al 2006).

2.8. African Monsoon

An exhaustive description of the monsoon systems that prevail over different parts of Africa has been given by Asnani (2005g). The region of central Africa between 15º N and 15º S receives the highest rainfall over the continent while Sahara in the north and Kalahari and Nahib in the south are desert areas. The coastal strip of North Africa receives its rainfall in association with extratropical westerly waves and low pressure systems which move across the Mediterranean Sea in winter. Over West Africa, the ITCZ has a north-south oscillation, but it remains north of the equator throughout the year (Fig. 2.6). However, the air on both sides of the ITCZ being dry, there is no rain in the vicinity of the ITCZ. In northern hemispheric summer, the ITCZ reaches its most northerly location giving rains over Sahel and areas to its south up to the Gulf of Guinea. When the ITCZ moves southwards, Kenya and Uganda receive what are called 'short rains' in the months of October to December and as it goes northwards once again; they get 'long rains' from March to June. Tanzania gets its rains between November and March. The onset and withdrawal of the summer monsoon do not have a clear association with the migration of the ITCZ because of the influence of major orographic features like Lake Victoria and the Great Rift Valley and local topography

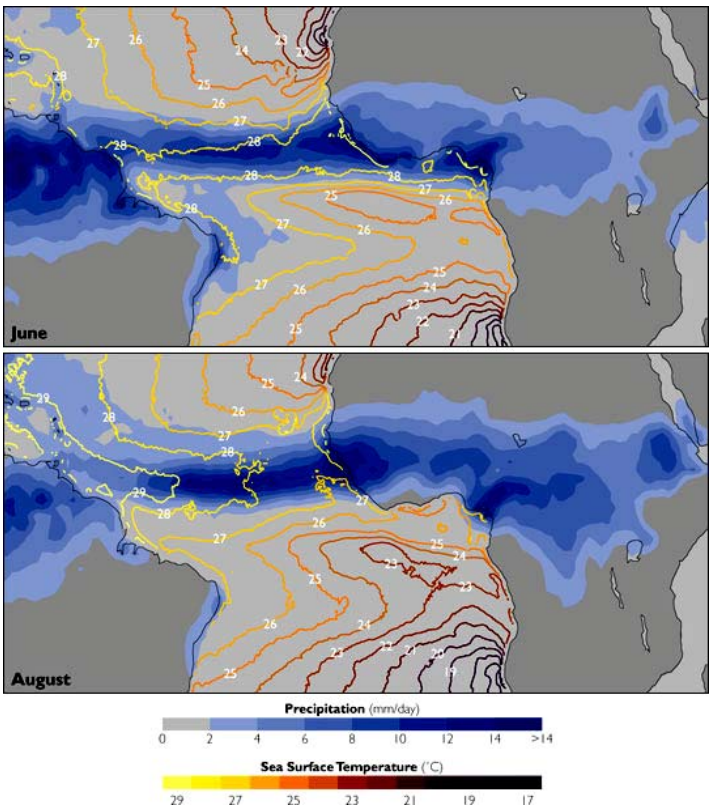


Fig. 2.16: Average June and August precipitation in mm/day and SST in °C over the West African monsoon region (from NASA 2009).

In a recent investigation based on TRMM precipitation data (Gu et al 2004), the African monsoon season has been found to consist of two distinct sub-seasons (Fig. 2.16). The first is in late spring and early summer, when the West African Coast near the Gulf of Guinea, at 5° N latitude gets the heaviest rainfall, under the influence of strong sea surface temperature gradients off the coast. The second is in late summer when the heavy rainfall region shifts to 10° N in association with atmospheric easterly waves. Rainfall in the two periods is anti-correlated.

Sanjeeva Rao and Sikka (2007) have discussed the commonalities and differences, as well as possible interactions between the African and Indian monsoon systems. Both monsoons develop in close proximity and have similar large scale atmospheric features, but the Indian monsoon is much stronger than the African monsoon and extends further north in the peak phase in July. The Indian monsoon is dominated by a 30-50 day oscillation and there is a possibility of its being modulated by an eastward propagating Madden-Julian Oscillation (MJO), when travelling over the near-equatorial African region. This interaction can be detected through eastward moving low outgoing longwave radiation (OLR) pulses and corresponding fluctuations in the 250 hPa velocity potential. A comparison of the occurrence of drought over India and the Sahelian and sub-Sahelian regions of Africa has shown that although the drought years may not exactly match, there is a signal of a common multi-decadal mode in the monsoon variability over the two regions.

An international project known as the African Monsoon Multidisciplinary Analysis (AMMA) is currently under implementation. This major initiative is aimed at improving the knowledge and understanding of the West African monsoon and its variability on time scales ranging from daily to inter-annual. Among the monsoons of the world, the West African monsoon is marked by an extremely high inter-annual and decadal-scale variability (Le Barbe et al 2002). This region which had wet conditions during 1950-70, suffered from a dramatic change to drought conditions thereafter, resulting in devastating environmental and socio-economic impacts on food and water security.

The present observational and monitoring system over the West African region is inadequate for the purposes of numerical prediction and also for studying the diurnal, seasonal and annual cycles of monsoon rainfall. From a wider perspective, the latent heat release through deep convection in the ITCZ over Africa is a significant heat source that affects the global circulation, including Atlantic

Ocean hurricanes, many of which have their origins in West African systems (Landsea et al 1992). AMMA aims at investigating the West African monsoon in a holistic manner, including the role played by the Saharan region which is the world's largest source of atmospheric dust and aerosols. One of the objectives of AMMA is to study the phenomenon of the West African monsoon from the global and regional scales down to the mesoscale and submesoscale and how the scale interaction results in the observed features of the monsoon and its variability. The AMMA project is motivated by an interest in fundamental scientific issues and also by the need for an improved prediction of the West African monsoon. Since 2006, over 400 scientists from more than 25 countries, representing more than 140 institutions have been involved in AMMA. The latest information on this project can be obtained from the AMMA web site on the internet (AMMA 2009).

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CHAPTER 3

Monsoon over other South Asian Countries

I. ***Pakistan Summer Monsoon***
II. ***Monsoon over Nepal***
III. ***Climate and Monsoon of Bhutan***
IV. ***Monsoon over Bangladesh***
V. ***Monsoon Climate of Myanmar***
VI. ***Monsoon Climatology of Sri Lanka***
VII. ***Seasonal Rainfall Characteristics and its distribution in the Maldives***

CHAPTER 3(i)

PAKISTAN SUMMER MONSOON

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3(i).1. Introduction

Pakistan is located along the western border of India and bound on the north by lofty Himalayas and on the northwest by the Hindukush and Suleman mountain ranges. Monsoon being the low level airmass does not cross these ranges, while travelling east to west from the Bay of Bengal. In this way Pakistan is at the tail-end of the Western Edge of the South Asian Monsoon. It is the major rain producing system over the South Asian region from Bangladesh in the east and Pakistan on its western extent. It is commonly known as “Southwest Monsoon”. The name “Southwest Monsoon” is used both for the phenomena of rains associated with southwesterly surface winds and the rainy season from June to September during which they bring huge amount of precipitation. The Monsoon, in meteorological paralance, is referred to the seasonal reversal of winds. This shifting of wind is gradual keeping pace with the rise in land temperature over the Indian peninsula. Monsoon march starts from India around end May or early June and progresses steadily toward west arriving Kashmir and Pakistan by the end of June or early July.

For centuries, people have tried to understand why the mighty Indian monsoon brings rain in each June and dries up each September. Prior to the invenion of weather satellites and advanced computers, climate scientists had only a basic understanding of how great atmospheric forces interact to make a monsoon. Now that is changing, and the age-old dream of predicting a monsoon's exact schedule and rainfall seems tantalizingly close, although there are still inter-annual

variabilities of monsoon. Scientists say that the two key ingredients needed to make a monsoon are a hot land mass and a cooler ocean which may further be attributed to the specific heat capacity of land and water. This heats up the air masses over the land, that results in its expansion and rise. As the air rises, it is replaced with cooler, moist, and heavier air from over the ocean. During the dry season, the winds blow offshore, from land to sea. As the monsoon begins, the wind starts blowing onshore, from sea to land. Since the Indian Ocean is bounded to the north by the largest land mass on the planet, the effects of differential heating" are especially intense over the Indian Peninsula (Y.S. Gao, 1959; Krishnamurti and Ramanathan, 1982). In addition, there exists the annual variation of summer monsoon which could be associated with the seasonal or annual oscillations of atmospheric circulation (Yasunari, 1985) and the interaction between ocean and atmosphere (Li, 1990).

3(i).2. Monsoon Season and its significance for Pakistan

Southwest monsoon prevails over Pakistan generally from July to September and brings lot of rain associated with meso-scale systems such as tropical monsoon depressions, and meso-scale monsoon lows. The monsoon-depressions form over the Bay of Bengal and move westward over the land and reach Pakistan. Their intensity, life span and distance travelled depend upon the depth of Monsoon system at their center. Meso-scale monsoon lows develop as a result of strong convection under moist and unstable atmospheric conditions which commonly prevail during the monsoon season. Sometimes, they are so devastating that they inflict huge losses to the lives, property and infrastructure. Due to associated heavy downpour, the streams and rivers are flooded beyond their capacity. Fertile soils from the top layer are eroded from one location and laid down over another place; also resulting in piling up loads of sediments in water reservoirs reducing their storage capacity.

Summer monsoon is widely celebrated for the life it brings to the forests and the farm lands ~~as~~ and also it terminates the hot and dry spell of May and June. But the rains are also feared for the death and destruction that floods can bring to the communities in their paths. The fate of the people is linked to the course and behavior of the monsoon. If the rains come in a delayed manner, farmers will sow on

less area or no planting, due to the fear of drought. If there is a lack of rain or breaks in the rain or no rains at the critical stage of development such as reproductive phase, plant seedlings may not be able to survive. If the rains are too heavy, young plants and seedlings can be washed away. All these factors can greatly increase the price or decrease the availability of food in the region. But the worst disaster may occur when the rains do not come at all in the whole season. The failure of the rains is the first link of the chain that leads to starvation and famine. Today, however, the improved storage facilities and irrigation systems have reduced the chance of famine. The four years long worst drought of 1999-2002 triggered by the strongest El Nino of 1997-98 in the history of Pakistan will always be remembered due to its severity and persistence. It engripped 2/3 of Pakistan but Balochistan and Sindh provinces suffered the most where crops and orchards dried up due to poor rainfall. Drought swallowed 240 human lives and over 10000 animals resulting into massive migrations and thus increasing the pressure on already limiting resources in neighboring areas.

Residual moisture retained by the cultivated soil from the summer monsoon rainy season is of great importance for the plantation of winter crops. As monsoon retreats generally in September, the summer is over and cooling starts causing reduction in evaporative demand of the atmosphere defending the loss of moisture from the soil. October and November are dry months in Pakistan when summer winds shift to set the winter pattern. This is the time for sowing of winter crop especially the staple food crop wheat. In rainfed areas conserved soil moisture of late monsoon season is a blessing for farmers for timely sowing of their crops as the delay in sowing cause significant reduction in yield. Therefore, the farmers pray for a good and extended monsoon season to harvest sufficient moisture to carry forward for optimum crop sowing and early establishment. However, recent observations show that monsoon has been following an early retrieval from Pakistan which leaves the farmers at the mercy of early winter rains for Rabi (winter) plantation.

3(i).3. Summer Climatic Features of Pakistan

Having peculiar location, Pakistan receives both summer and winter rain but the monsoon contribution to the total annual rain is about 60% whereas winter share is around 20% only. The rainfall activity occurs over Pakistan in two distinct periods namely winter (January to March) and summer monsoon (July to September). In the transition season, i.e., April to June and October to December, southern parts of the country particularly Sindh hardly gets any rain, while relatively good enough in other parts of the country such as hills of the Khyber Pakhtoonkhaw, Gilgit-Baltistan, Kashmir and sub-mountainous regions of the Punjab. In Northern parts of Pakistan, the rainfall activity continues throughout the year, but the amount is relatively small in the above mentioned transition seasons. Sometimes, an early onset of summer or winter rains strikes out June and December from the list of transition months. The case of early monsoon onset i.e. in June instead of July had been common in first decade of 21st century when frequency of onset in June remained 4 out of 10.

In April, low pressures begin establishing about tropic of cancer in the South Asian region due to increase in heat. Peninsular India, south of 20°N has a tapering shape and becomes narrow south of 15°N. The latter portion comes under considerable maritime influence as it is surrounded by ocean from three sides. This makes the heating over land more prominent to the north of 20°N and hence the axis of the low remains generally at more northerly latitude over India and Pakistan. Another low pressure, generally known as ‘Heat Low’, develops due to intense heating over Indo-Pakistan and becomes well marked in June with its main center over central parts of Pakistan. The low pressure picks up its intensity as the summer season progresses towards its climax. The day temperature shoots beyond 50°C in areas around the center of heat low encompassing central parts of Pakistan. The horizontal extent of the heat low ranges from few hundred to several hundred kilometers and it experiences different changes in its location and intensity with approaching weather systems. The vertical profiles over the area reveal that heat low becomes weaker and weaker with height. It generally exists in the first 1.5 km and is over- laid by the Sub-Tropical High (STH) pressure belt. There is a frictional convergence in the heat low with a weak ascent and subsidence is likely above. A

more stable layer occurs in the lower part of the column of subsidence due to vertical shrinking. The core of the heat low lies over 27°N - 30°N and 62°E - 67°E over southern half of Pakistan which is out of reach of the maritime air mass. The heat low takes different shapes with the advance of rainy season and with the changing circulation pattern during significant weather events. An extensive investigation of the heat low is required to determine its role in providing the driving force to bringing the monsoon system to Pakistan. The pressure pattern over Pakistan during the month of June is characterized by a relatively high pressure over Northern and Southern parts of Pakistan and low pressure over Central parts. A considerable pressure gradient exists along the coastal belt.

A trough commonly known as “monsoon trough” lies over northeast Pakistan and northern India. It extends up to the head Bay of Bengal with axis almost parallel to the Himalayan range on its north. Summer rainfall activity over Pakistan and India is driven by the axis of the monsoon trough as it acts as highway for low pressure systems to travel from east to west transporting huge amount of moisture. Western and northern mountains draw boundary for the monsoon regime in Pakistan as monsoon airmass remains at low level and has a limited vertical extent. Normally the pressure gradient in the south of the monsoon trough is stronger as compared to its north. The different alterations in its geometrical shapes in east-west orientation are associated with various precipitation characteristics over temporal and spatial scales. Similarly, the north-south march of the monsoon trough in connection with the Heat Low brings heavy downpours as well as the monsoon breaks halting the monsoon activity temporarily over Pakistan. During the active period of the monsoon, the trough is generally located in the central and northern Indian Peninsula. The break monsoon phenomenon is a reverse of the active monsoon spell over central and northern India. During the monsoon break the monsoon trough hugs the Himalayan rim, the low-level easterlies disappear entirely along the Indo-Pak plains and are replaced by west-northwest flow along the periphery of the Himalayas, resulting in decrease of rainfall over most of the area. Merger of the monsoon trough with the Inter-Tropical Convergence Zone (ITCZ) is also responsible for the dramatic changes in the intensity, areal extent and concentration of monsoon precipitation.

Hence, the monsoon trough holds the major control of Pakistan summer monsoon precipitation.

3(i).4. Normal Precipitation Pattern

Due to the peculiar orographic features, such as the Himalayas, Karakoram and Hindukush ranges in the north and northwest respectively along with the Tibetan Plateau on the northeast, play an important role in modifying the weather systems. The high mountains act as a “blocking high” to the movement of low pressure systems from the south toward north. When there is sufficient moisture feeding from the Bay of Bengal or the Arabian Sea in association with some active westerly systems, the heavy rainfall is produced sometimes in the northern parts of Pakistan. This is primarily the situation in which devastating floods occur in the rivers and cause heavy losses downstream to agriculture and infrastructure.

In the summer season (June to September), the monsoon depressions, originating from the Bay of Bengal are primarily responsible for the rainfall activity over Pakistan. Monsoon onset takes place over the sub-continent in the last week of May whereas in Pakistan it occurs around a month later i.e. the first week of July. From the geographical distribution of normal summer monsoon (July-September) rainfall over Pakistan based upon climatic normals of 1961-90, it can be inferred that in a broader sense, northern half of Pakistan receives substantial amount of precipitation in summer from monsoon whereas southern half is considerably drier. The monsoon precipitation does not reach up to the Southwestern parts of Pakistan and they remain practically dry during the summer monsoon season. These regions are beyond the reach of winter precipitation also; therefore the climate is desert like.

Onset of monsoon in Pakistan generally takes place during the first week of July from northeastern parts of Punjab adjoining Kashmir. It advances first toward west until blocked by western mountain barrier and simultaneously spreads over the Indus plains of Punjab and Khyber Pakhtoonkhaw till mid July. A southwesterly current from the Arabian Sea establishes at the same time relatively weaker than the northeasterly flow along foothills of the Himalayas from the Bay of Bengal. It

produces rain over the southeastern parts of the Sindh province in the presence of meso-scale convective activity. The advance of the monsoon over the entire subcontinent is a rather slow process taking on average nearly 40-45 days from its start off Kerala around 31st May to its culmination by mid-July over central Pakistan. Pakistan receives summer monsoon associated with southeasterly current in the north which set up in late June or early July and southwesterly current in south established in middle of July (Rasul, et at 2004). Onset over Kashmir and northeast Pakistan takes simultaneously on 1st July with a deviation error of 15 days. Similarly on southern half of the country, rains start around 15 July with 16 days standard deviation (Rasul, et at 2005). During the last decade, a tendency of early onset has been observed with a common occurrence in the last week of June. Although there were some evidences of early onset as early as around mid June in the past but presently persistent early arrivals and early withdrawals make their connections with global warming and climate change which have resulted into enhanced variability both in time and space.

3(i).5. Mechanism of Heavy Rainfall

During the last ten years, it has been observed that the frequency of heavy downpour events has overwhelmingly increased resulting into historic flash, reverine and urban flooding in an unusual manner. Cloud burst in twin cities of Islamabad-Rawalpindi on 23rd July, 2001 poured 620mm rainfall in only 10 hours which resulted in history's worst urban flooding. More than 200mm per day rain, in any part of the country, is enough to chock the drainage system and paralyze the civic life. Such events have been regularly occurring in different parts of the country every year since 2001. Persistent extremely heavy rainfall in Khyber Pakhtoonkhaw and Northern Areas in the last week of July and early August 2010 generated super flood in the River Indus. Another extremely unusual great meteorological anomaly in monsoon has been registered recently in 2011, when the hyper arid areas of Sindh province received more than 1000mm rainfall in a period of four weeks, which these areas normally receive in five years. These unusual rains caused history's worst rainfall flooding in Sindh. The stagnant water destroyed the standing crops and wet soil did not allow the plating of winter crops especially wheat crop over a large tract

of agricultural land. These meteorological extreme precipitation events are attributed to the global warming and climate change (Chaudhry, 2011).

In the past, heavy rains in Pakistan were usually associated with the tropical depressions formed over the Bay of Bengal reaching Pakistan crossing the Indian peninsula. But now their frequency and intensity has considerably decreased. Either they seldom gain depression intensity or they die immediately after landfall. However, if some tropical depression or monsoon low reaches Pakistan, it merges with already exiting meso-scale lows in southeastern parts of Sindh or western Rajasthan and reinforce their energies in the presence of additional moisture supply from the Arabian Sea. In case of the presence of a western disturbance passing across the northern latitudes of Pakistan, cold advection in the mid tropospheric level takes place which tends to suppress the warm and moist rising airmass. Such coincidence develops a strong vertical circulation and results in a persistent heavy downpour in areas under its influence. In the north,—strong convection with continuous supply of moisture from the Bay of Bengal as well as the Arabian Sea and presence of western disturbance are the commonalities but here inverted “V” shaped topography at the junction of Himalayas and Hindukush also play an important role in enhancing the lifting process (Rasul et al., 2004; Rasul et al., 2005).

3(i).6. Recent Research on Pakistan Monsoon

Rasul, et al., 2008 discussed the extremities of Indian Summer Monsoon on its western extent over Pakistan in relation to the change in weather pattern, persistence and energized physical processes. They stated that summer monsoon is lifeline for billions of people and brings many folds larger mass of water than winter monsoon due to its unique thermodynamic features. Indian monsoon generally prevails over the region from June to September; however, Pakistan being located at the western edge of the South Asian monsoon zone receives it almost a month later and early retrieval. Since 1990s, it has been observed that the onset of monsoon over Pakistan has moved toward middle of June about two weeks earlier than normal. Similarly, the peak has been shifting to the end of July instead of mid August and likewise an earlier wrap-up. Main source of summer rains in Pakistan was

attributed to the monsoon depressions formed over the Bay of Bengal and enter Pakistan after moving west/northwestward over India. This depression formation activity is considerably decreased in recent years possibly due to slightly cooling trend of the Bay water surface.

In Pakistan, the frequency of occurrence of long dry spells (drought) as well as floods has increased. The most alarming feature is the increased frequency of extreme precipitation events during the monsoon season. Diagnostic analyses on a number of such cases revealed that the direct links are established with the global warming. Increase in the Sea surface temperature over the north Arabian Sea appeared as a fuel for activation of weakening depressions entering from Indian peninsula. Frequent interactions of westerlies across the eastern periphery of Sub-Tropical High (STH) with monsoon currents along foothills of Himalayas resulting in localized heavy downpour have been observed. Rise in extreme precipitation events during pre-monsoon and monsoon season has clear links with Indian Ocean Dipole, Madden Julian Oscillations, Tibetan High and surrounding climatic conditions. Westerly waves are, although, shifted slightly northward but extension of their troughs has elongated toward the southern latitudes and frequency to reach 30°N extent is increased. Low level easterly jet has gained strength and thermal regime represented by equivalent potential temperature is heated up resulting into enhanced meso-scale convective activity. Heavy downpour of July 2001 in Islamabad-Rawalpindi (Chaudhry, et al., 2005), in Karachi July 2003, 2007 and 2009 are the examples of summer monsoon getting vigorous behavior.

Syed et al., 2010 investigated the intraseasonal summer rainfall events and monsoon onsets over the western edge of the South-Asian monsoon. A thermodynamic structure leading the active phase (AP) of the Western Edge of the South-Asian Monsoon (WESAM) is investigated. The APs seems to have significant contribution in the mean seasonal rainfall in the region. A few days before APs, the upper level warm anomaly appears over the north Hindu Kush-Himalaya (HKH) region and it is reinforced by surface heating yielding the column average warming. The height anomalies are baroclinic with the low-level anticyclone being located at the east of warming. The low-level anticyclone causes the moisture convergence at

the core WESAM region. As the region keeps warming, the height anomalies and associated low-level convergence become stronger. The AP starts when the low level moisture convergence is strong enough to overcome the preexisting stable atmospheric condition due to the upper level warming. The proposed mechanism of APs has some resemblance with large scale south Asian monsoon onset, whereas conventional south Asian monsoon intraseasonal oscillations (ISOs) do not show clear relationship with APs of WESAM.

In an effort to find relationship between extratropical influences on interannual variability of monsoon Syed, et al., 2011 conducted a detailed study using different data sets. Based on NCEP/NCAR reanalysis and CRU precipitation data, a conditional maximum covariance analysis is performed on sea level pressure, 200 hPa geopotential heights and the SAM rainfall by removing the linear effects of El-Nino Southern Oscillation from the fields. It is found that two modes provide a strong connection between the upper-level circulation in the Atlantic/European region and SAM rainfall: the Circumglobal Teleconnection (CGT) and the Summer North Atlantic Oscillation (SNAO). The structures in the 200 hPa heights of both modes in the Atlantic region are similar in the Atlantic region, and their southeastward extension to South Asia (SA) also corresponds to upper-level ridges (in their positive phases) in slightly different positions. Nevertheless, the influence of both modes on SAM rainfall is distinct. Whereas a positive CGT is related to a widespread increase of rainfall in SAM, a positive SNAO is related to a precipitation dipole with its positive phase over Pakistan and the negative phase over northern India. The physical mechanisms for the influence of CGT and SNAO on SAM are related to the upper-level geopotential anomaly which affects the amplitude and position of the low-level convergence. The small displacements of the centers of these responses and the low level cold advection from the north east of SA in case of SNAO explain the differences in the corresponding SAM rainfall distributions. These findings are confirmed with the relatively high-resolution coupled climate model EC-Earth, which gives confidence in the physical basis and robustness of these extratropical variability modes and their influence on the South-Asian monsoon rainfall.

Sajjad, et al., 2011 examined the role of mid-latitude circumglobal wave train in the 2010 flood over Pakistan that caused severe damage to agricultural crops, live stock and infrastructure. They pointed out that the co-occurrence of mid-latitude wave train and upper level blocking high over western Russia was associated with the July 2010 heavy rainfall over northern Pakistan. The eastward propagating mid-latitudinal wave train in association with the north-south oriented blocking high favored extra-tropical cold air down in to the northwestern Indo-Pak and adjacent areas. The confluence of the wave train and blocking high is further associated with enhanced low surface pressure anomalies over Pakistan and adjacent areas of Afghanistan, Iran and northwestern Arabian Peninsula. Consequently, an intensification of the surface low pressure over northern Pakistan and adjacent areas favoured the moist southerly flow to intrude deep over northern Pakistan where orographic lifting in association with upper level trough fostered convection and hence precipitation. In an additional analysis covering the period from 1951-2007, they found a similar wavelike pattern in the mid-latitudes associated with past heavy rainfall events over northern Pakistan. However, compared to past events, the mid-latitude circulation associated with July 2010 rainfall was much more intense that made it the worst flood event of the recorded history.

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CHAPTER 3(ii)

Summer Monsoon over Nepal

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3(ii).1. Introduction

Nepal is a land locked and predominantly mountainous country situated at the foothills of the Himalayas between 26° 22' N and 30° 27' N latitude and 80° 04' E and 88° 12' E longitude. With an area of about 147,181 sq km, the country extends about 885 km in the east –west direction and on an average width of approximately 190 km in the south- north direction. The altitudinal extent ranges from 60 m in the south to more than 8000 m above mean sea level in north. Physiography of Nepal is presented in Fig.3(ii).1 along with the development regions. The rapid changes in the altitude and aspect have given rise to a wide range of climatic conditions in Nepal (Nayava, 1974).

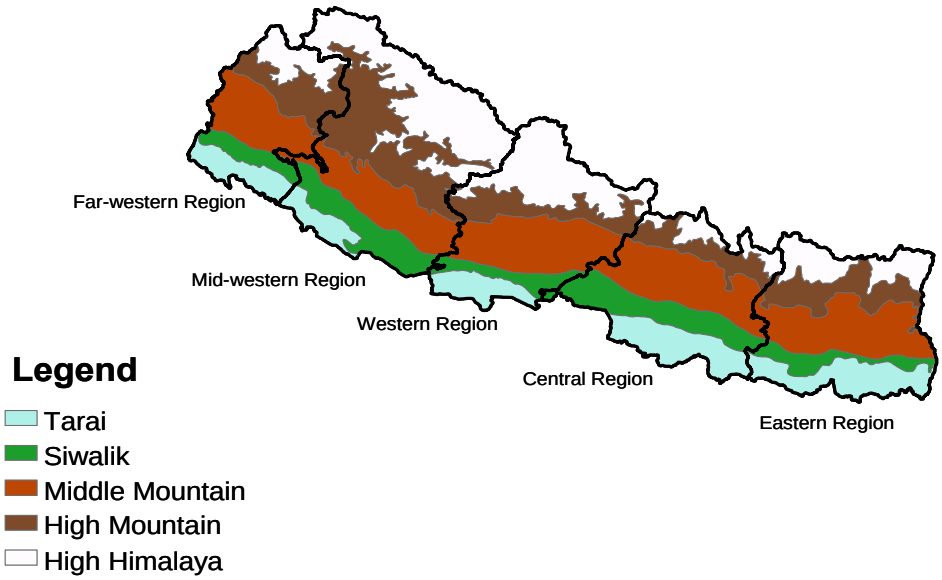

Fig. 3(ii).1: Physiographic division of Nepal

Temperature in Nepal closely follows the geographical configuration of the country. The hottest part of the country is in the southern Terai belt where maximum temperature reaches more than 40°C. The highest maximum temperature so far recorded is 46.4°C in Dhangadhi, in the western Tarai region of the country. The valleys have especially pleasant climate with temperature range mostly between 0 and 30°C. The coldest part of the country is high Himalaya in the north.

Eighty percent of the total annual precipitation in Nepal is attributed to summer monsoon rain (Nayava 1974; Shrestha 2000). Fig.3(ii).2 depicts All Nepal Mean Monthly Rainfall, which shows July as the rainiest month. Fig.3(ii).3 shows the percentage of rainfall in different seasons. In general, premonsoon and postmonsoon periods are dominated by thunder activities producing more precipitation in the east than in the west Nepal contributing about 12% and 5% to the total annual rainfall respectively. However, during winter months, the western disturbances produce more precipitation in the western than in the eastern parts of the country. Winter is the driest season for Nepal which accounts for about 3% of the total annual rainfall.

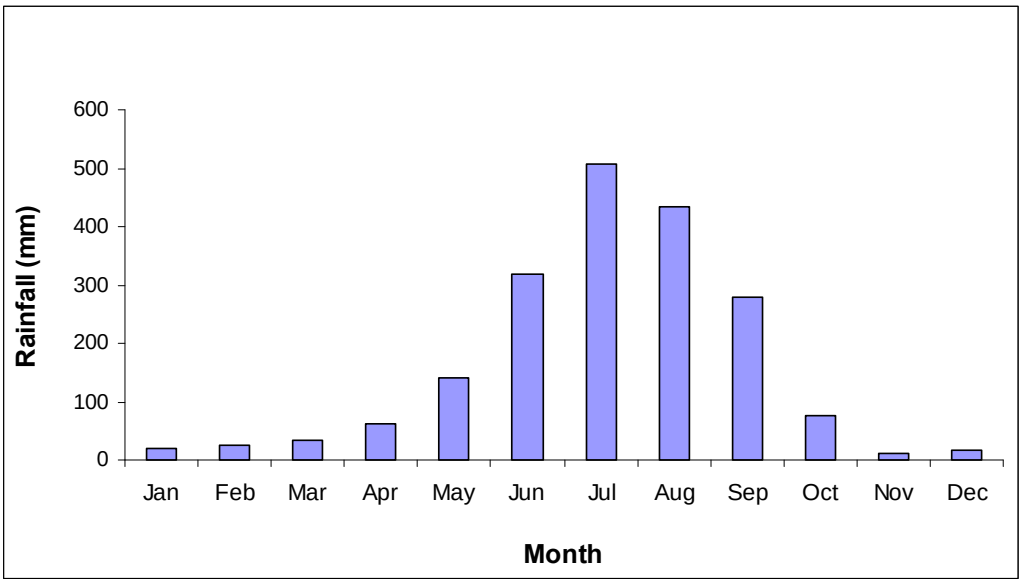

Fig. 3(ii).2: All Nepal Mean Monthly Rainfall distribution (1971-2000).

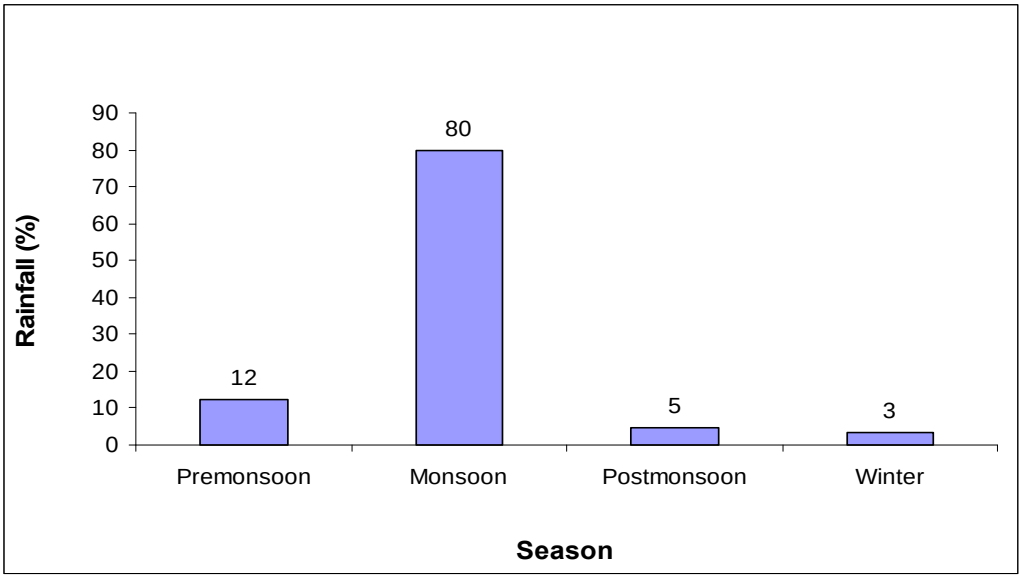

Fig. 3(ii).3: Seasonal contribution of All Nepal Rainfall (1971-2000).

3(ii).2. Climatic features of Nepal

According to the De Martonne Aridity Index, most of the eastern, central and western parts of the country are humid to very humid (Fig.3(ii).4). Mid and far-western regions have mainly sub-humid climate. The northward of western and mid-western regions is largely semiarid to dry which lie mainly in the leeward side of the Himalayan range (to be particular Annapurna - Dhaulagiri range). Some portion of the Far-western region of the country bordering India have humid climate due to the influence of winter rain in these regions. The southern plains (Tarai) are mainly sub humid. In the leeward side of the Annapurna – Dhaulagiri range, potential evapotranspiration is always greater than precipitation. Therefore, these regions have a continual water deficit. However, in the windward side with very humid condition, precipitation is more than the potential evapotranspiration throughout the year. These regions therefore have water excess throughout the year. The Thornthwaite's climatic classification also reveals the similar features.

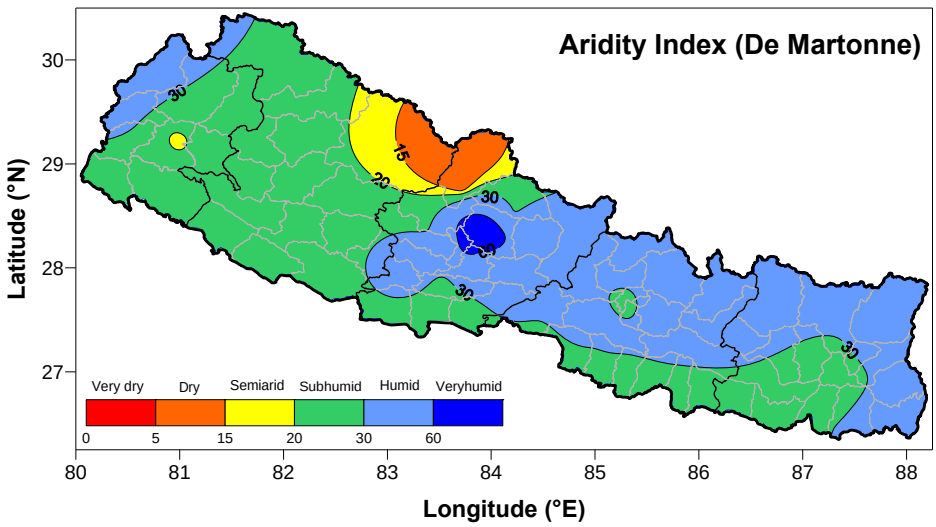

Fig. 3(ii).4: Aridity Index map of Nepal (De Martonne)

3(ii).2.1. Floods and droughts

Nepal is prone to weather related disasters due to its rugged, steep topography, and fragile young geological conditions which is aggravated in the recent time with deforestation that are occurring in the country. Water induced disasters are most common ones, especially during monsoon season. Floods and landslides are the most common weather related disasters caused by the heavy rainfall. Every year they cause immense damage to lives and properties. Because of the altitude and orientation of the topography, considerable variation in precipitation is observed from place to place resulting in some places susceptible to floods and some places vulnerable to drought.

Twenty four hour extreme rainfall map shows that the southern lower mountains (Siwalik) and southern plain area (Tarai belt) are especially high intensity rainfall area and so susceptible to flooding (Fig.3(ii).5). Flood generally originates in the upstream of a river basin and as it emerges out of the Siwalik and Mahabharat ranges, it extends and covers the vast plain area of the Tarai. Since flood occurs along a river and its periphery, Karnali, Babai, Tinau, Narayani, Bagmati, Kamal, Sapta Koshi, Kankai and Mechi are the more flood prone river basins. The precipitation show increasing trend in total and heavy precipitation events at most of the stations (Baidya et. al, 2008). Most of the stations below 1500 m show increasing trend, whereas the trends are not clearly defined above 1500 m (Baidya et. al, 2008).

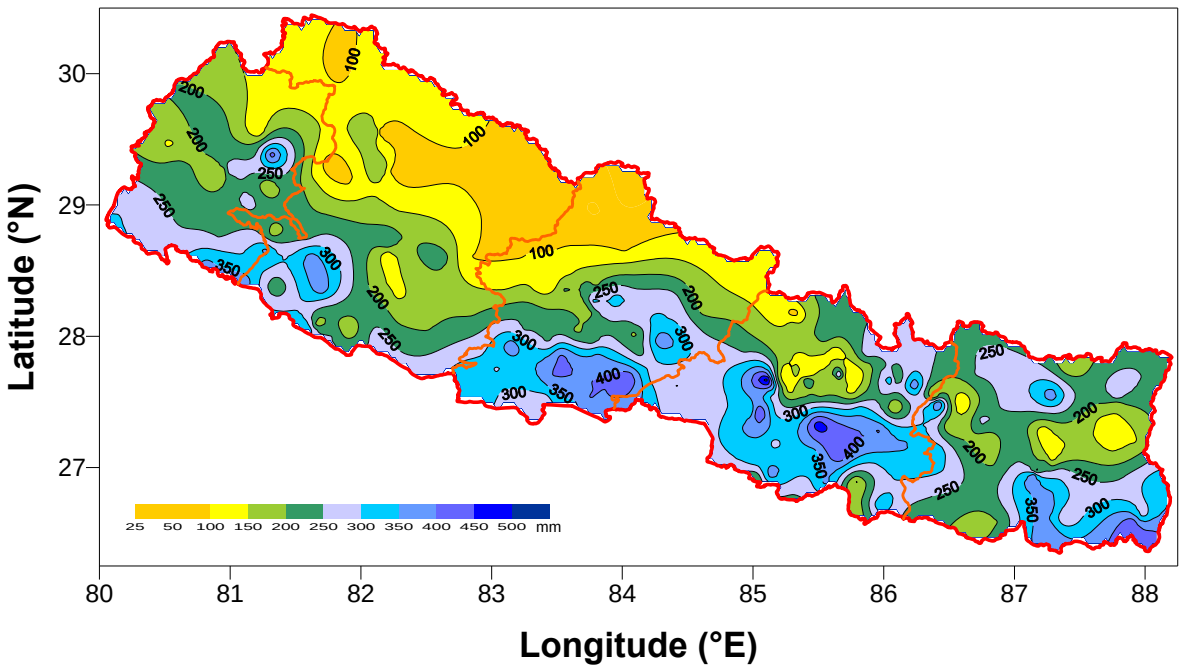

Fig. 3(ii).5: 24 hour extreme rainfall distribution (1971-2000)

Standardized Precipitation Index (SPI) and Soil Moisture Index (SMI) analysis show that drought of all categories occur in Nepal in all seasons (Nepal hazard risk assessment, 2010). Moderate drought frequently occurs in all seasons. But severe and extreme droughts in comparison are less common. Fig. 3(ii).6 depicts the number of stations in percent recording drought during monsoon season. Altogether 40 stations for the period 1977 - 2007 have been used for the analysis. Western parts of the country are more droughts prone especially during winter and pre monsoon season in comparison to other parts (Nepal hazard risk assessment, 2010).

Analysis of soil moisture index shows that highest moisture surplus occurs during monsoon season whereas there is highest moisture deficit during post monsoon season. In winter, there is moisture surplus in western half of the country but deficit in the eastern half (Nepal hazard risk assessment, 2010). It shows the influence of the weather systems namely, the western disturbances in the winter season.

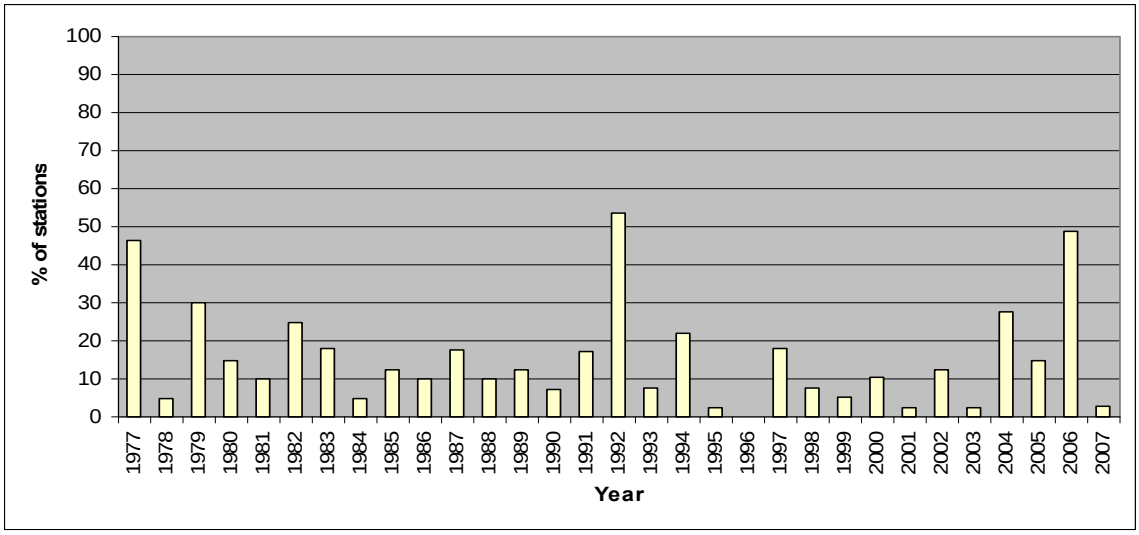

Fig. 3(ii).6: Drought occurrence during monsoon season

3(ii).2.2. Weather situation associated with monsoon in Nepal:

The weather systems associated with monsoon season are mainly driven by the semi-permanent features which remain more or less in the same location for number of days. However, there is another category of weather system as well, known as monsoon disturbances which are migratory in nature.

Monsoon disturbances such as depressions are the major migratory systems which transfer moisture from Bay of Bengal to higher altitude towards Himalaya and into Tibet. The effectiveness of this transport is controlled by the depression path, and the frequency of depression that originate from Bay of Bengal.

The onset of monsoon in Indian subcontinent takes place after a strong low level moisture flux associated with southwesterly cross equatorial flows off the coast of Somalia is well developed (Krishnamurti, 1981). Such a situation is related with development of heat lows over northwest India and adjoining areas of Pakistan. An elongated area of low pressure system established at this time, running roughly parallel to south of the Himalayan range plays an important role in bringing monsoon activities in Nepal.

During monsoon, the thermal condition of land drives moist air flow vertically from low lying Tarai plain in the south of Nepal and adjoining areas to the hills and

mountains including the Tibetan Plateau which greatly strengthen land sea thermal contrast regionally and serve to enhance the intensity of the monsoon. The vertical structure of horizontal wind, relative humidity distribution, and upward motion fields are observed with the moisture laden monsoon flow over mountain extending to 500 hPa level bringing cloudy weather condition in Nepal and adjoining areas. The elongated trough axis exhibits considerable day to day variation in the region which has a major role in controlling the monsoon activity there by changing the weather situation in the country.

During the monsoon onset, dramatic changes occur in large scale atmospheric structure which includes rapid increase in vertical moisture and the duration of daily rainfall. The onset of monsoon is characterized by the abrupt initiation of rainfall activity in the country.

3(ii).3. Monsoon onset and withdrawal in Nepal

Monsoon depressions are very important synoptic features during the monsoon season in the region. Formation of strong mesoscale depression over the head Bay of Bengal and its north westward propagation generally in the second half of June is the major cause of monsoon onset in Nepal. Lang and Barros (2002), conducted intense precipitation observation in Nepal Himalayas and confirmed that monsoon onset were generally associated with development of monsoon depression over north Bay of Bengal and its progress towards northwest.

Onset and withdrawal of monsoon have been defined by Xavier et al (2008) as the day when the tropospheric heat source shifts from south to north or north to south respectively. A characteristic of large scale onset of Indian summer monsoon is an abrupt increase in kinetic energy of the low level monsoon flow (Krishnamurti, 1985).

In a mountainous country like Nepal, declaration of monsoon onset is a complicated task, as it is still based on the synoptic feature. The monsoon enters into Nepal from East. Once it enters Nepal, topography of the country plays an important role on distribution of rainfall. The monsoon winds ascend up the hills and mountains

resulting in a remarkable variation in rainfall from south to north and at the same time, from east to west (Shrestha 2000). In general, there is a decreasing tendency of precipitation from east to west under the influence of the monsoon flow that advances from Bay of Bengal towards the northwest direction, facilitated by the topographical configurations of the region. The windward side of the mountain barrier receives a lot of rainfall while the leeward side receives comparatively less rainfall. As a result, there are some high rainfall pockets, which are favored by the topography and its orientation (Fig.3(ii).7). The total monsoon precipitation map shows the highest precipitation area of more than 450 cm centered over southern flank of Annapurna range and the driest part with less than 50 cm on the lee side of the same range.

During monsoon season, most of the Tarai and Siwalik get more than 80% and the central mountainous regions receives between 75% - 80% of the annual total rainfall (Fig.3(ii).8). However, the low precipitation area of western high altitude regions receives below 65% and rest of the country gets about 65% to 75% of the annual precipitation during the monsoon season (Devkota et.al. 2006). The mean monsoon rainfall for Nepal amounts to be 1422.8 mm (Shrestha, 2000).

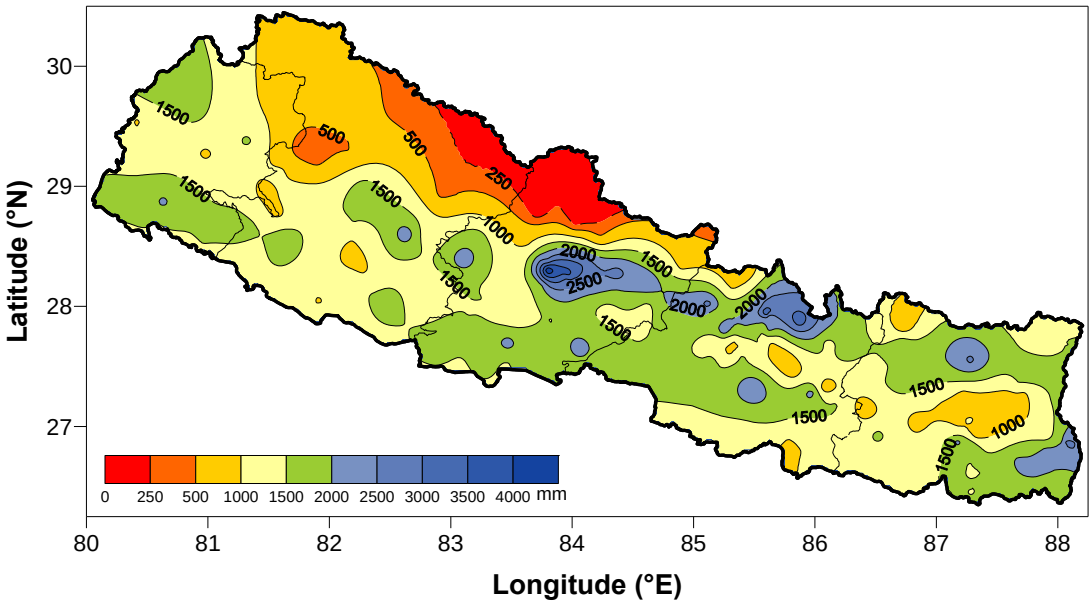

Fig. 3(ii).7: Spatial distribution of mean monsoon rainfall over Nepal (1971-2000).

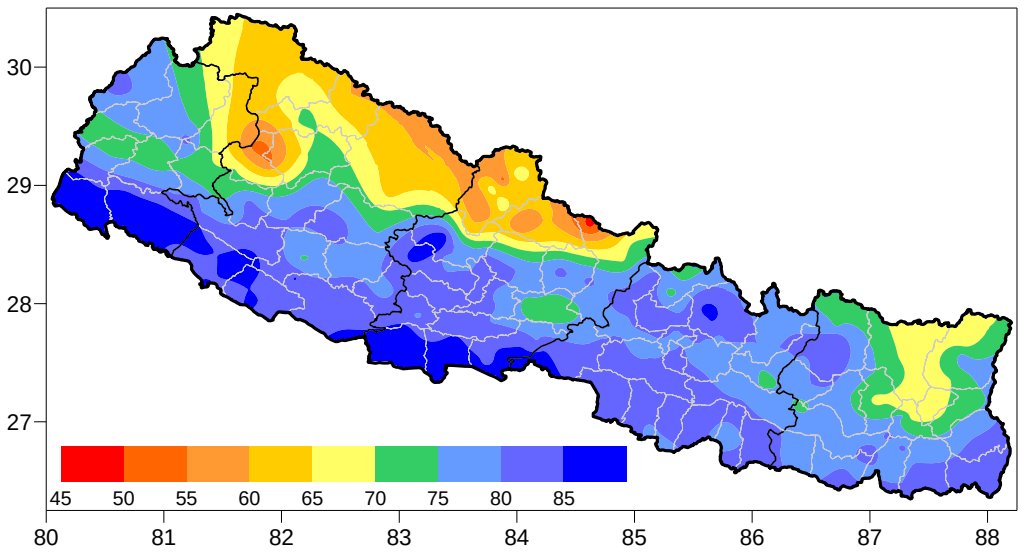

Fig. 3(ii).8: Contribution of monsoon total rainfall to annual total (%).

Onset of monsoon generally takes place in Nepal in the south eastern part of Nepal resulting rainfall associated with thunderstorm activities and extending the rainfall in the northern hills. Examples of onset of monsoon in Nepal in the satellite images where depressions play important roles are shown in Fig. 3(ii).9. Once the monsoon sets in, it gradually advances towards westward and covers whole Nepal within 5 to 7 days.

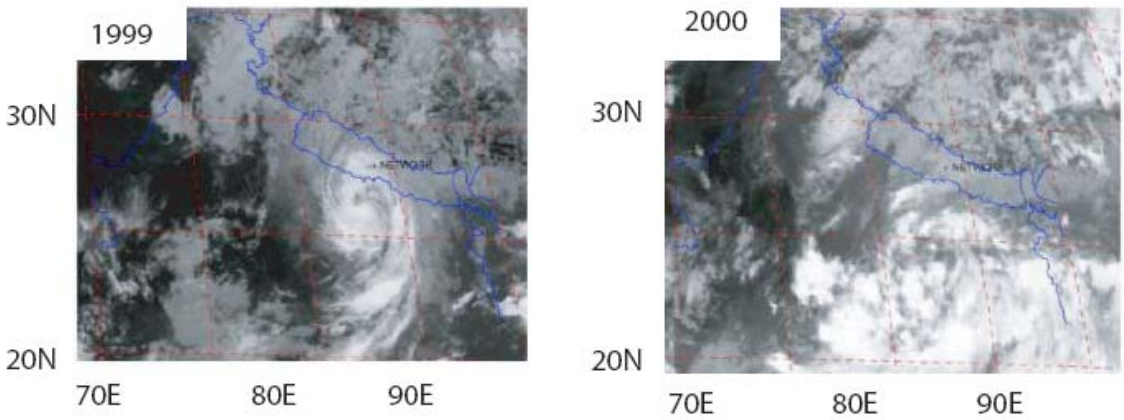

Fig.3(ii). 9: Satellite images of Onset of monsoon in Nepal in the year 1999 and 2000

depending upon intensity of monsoon flow. Climatologically, normal date of onset of monsoon in Nepal is 10th June (DHM, 1977). Study made for the period 1989-2010 on monsoon onset shows that the date of onset of monsoon in south eastern part in Nepal varies between 31st May to 22nd June. The earliest date of onset and the

delayed one differ by 22 days with a standard deviation of 6 days. Fig.3(ii).10 shows a trend of monsoon onset and it is observed to be delaying by 0.23 days per year.

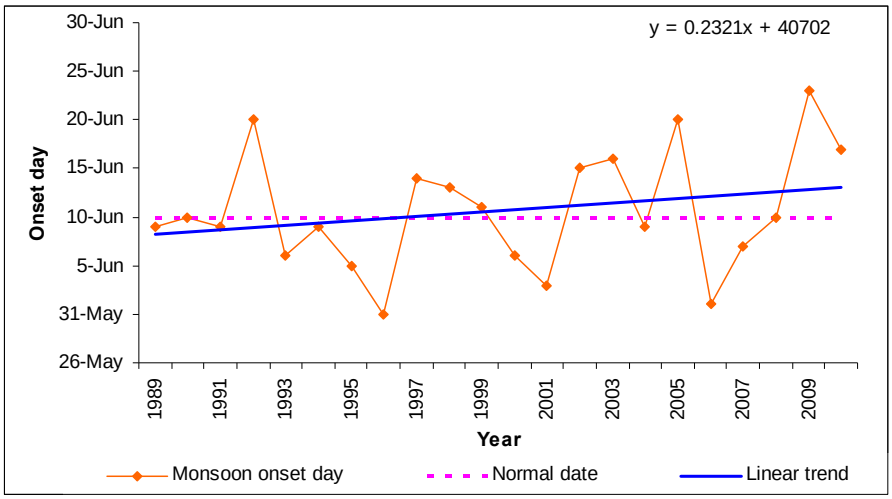

Fig. 3(ii).10: Onset date of Monsoon in Nepal

The withdrawal of monsoon and its gradual equator ward movement and deceleration of low level westerly flow is heralded by the seasonal cooling of the Asian continent (Ramage, 1971). In such a situation, the anticyclone over upper level moves from North to South. The anticyclone over Tibetan plateau at this time, shifts from north of the Himalaya to south towards the northwest India and adjoining areas which plays an important role in monsoon withdrawal from northwest India and leading to withdrawal from western part of Nepal. Then the monsoon gradually withdraws from central Nepal and finally from eastern Nepal. Fig.3(ii).11 depicts the time evolution of withdrawal of monsoon from Nepal.

The long time average normal date of withdrawal of monsoon from eastern Nepal is 23rd September (DHM, 1977). Study made for the period 1989-2010 on monsoon withdrawal date shows a variation between 19th September to 17th October. The earliest and most delayed withdrawal varies by 28 days with a standard deviation of 8 days and an increasing trend of monsoon withdrawal is observed to be delaying in recent years (Fig.3(ii).11). And the average withdrawal date of monsoon is noted to be 29th September. The difference between the withdrawal dates may be due to the different period. Fig.3(ii).11 indicates a delayed withdrawal of monsoon by 0.75 days per year.

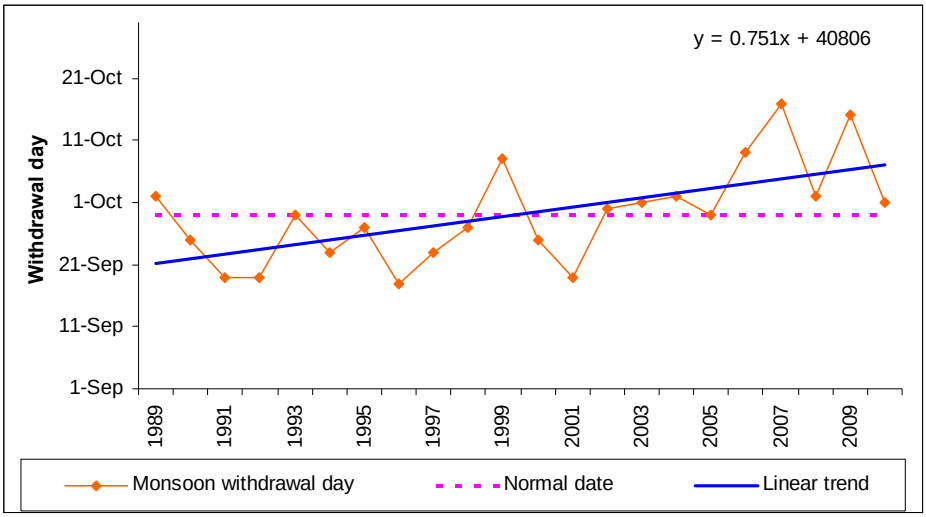

Fig. 3(ii).11: Withdrawal of monsoon in Nepal

Trend of monsoon rainfall and extreme events

In an agricultural based economy of the country like Nepal, rainfall during the monsoon months have significant role for food security and national planning. As the monsoon months usually starts from June and remains till September, the wettest month is noted to be July. A linear trend of all Nepal monsoon rainfall for the period from 1971 to 2008 does not show significant trend (Fig.3(ii).12).

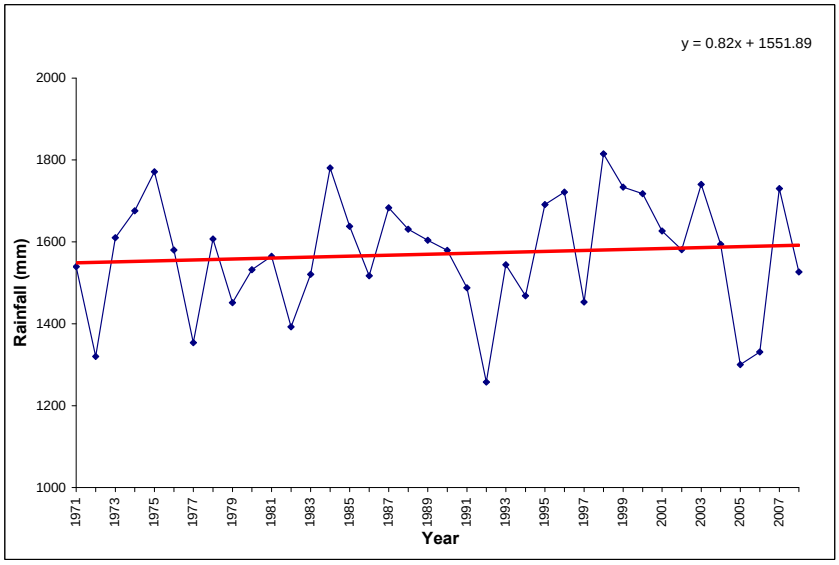

Fig. 3(ii).12: Annual variation of All Nepal Monsoon rainfall and solid heavy line is the trend line.

Fig.3(ii).13 shows the interannual variation of the number of rainy days (rainfall equal to greater than 1 mm/day is considered as rainy days). The trend of number of rainy days receiving rain 1mm/day shows a decreasing trend in Nepal. However, numbers of rainy days receiving 100 mm/day are observed to be increasing (Fig. 3(ii).14). This shows that heavy rainfall activity in the country is increasing.

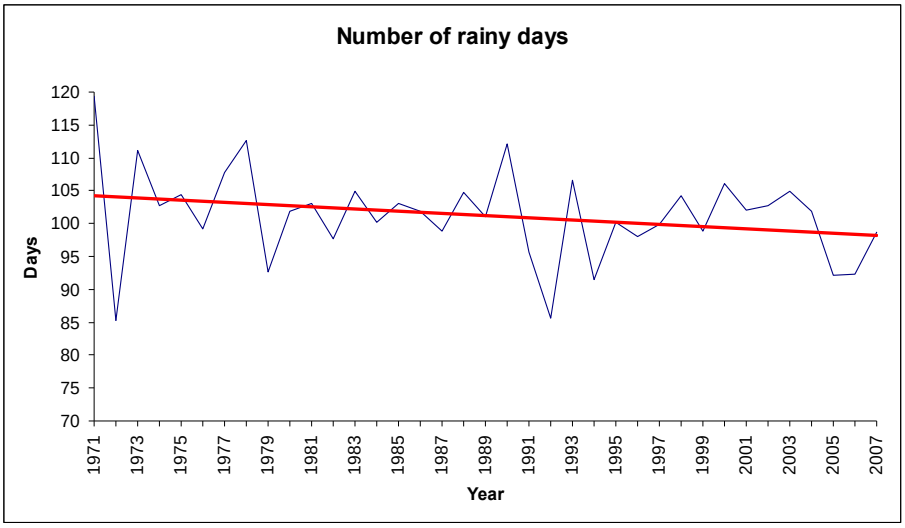

Fig. 3(ii).13: Trend of number of rainy days (≥ 1 mm/day)

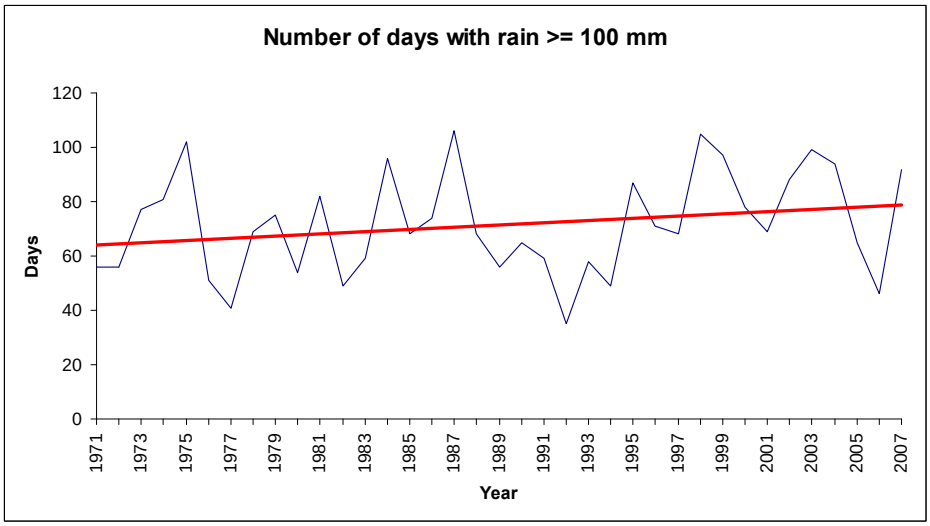

Fig. 3(ii).14: Trend of heavy rain (≥ 100 mm/day).

3(ii).4. Thunderstorm activities in Nepal

Thunderstorms are very common during the monsoon season in the country. A preliminary study on thunderstorm in Nepal based on 10 years data from 2000 to 2009 from different synoptic stations installed at various places in the country shows that thunderstorm occur in all months of the year in Nepal (Shakya, 2010).

The study is carried out using the day time observations from 00:00 to 12:00 UTC. The time period from 00:00 to 06:00 is termed as forenoon observation (FN) and from 06:00 to 12:00 UTC as afternoon observation (AN), and both the values when added is considered as the total (T) (Fig. 3(ii).15). It shows that the maximum thunderstorm days are noted to be in the month of May during pre monsoon season whereas the lowest is observed in the winter month December.

In the month of May, afternoon thunderstorms are frequent in comparison to forenoon thunderstorms. All Nepal average total number of thunderstorm days in the month of May is observed to be about 3 days, whereas in July, during the mature phase of monsoon season has relatively less number of thunderstorms in Nepal.

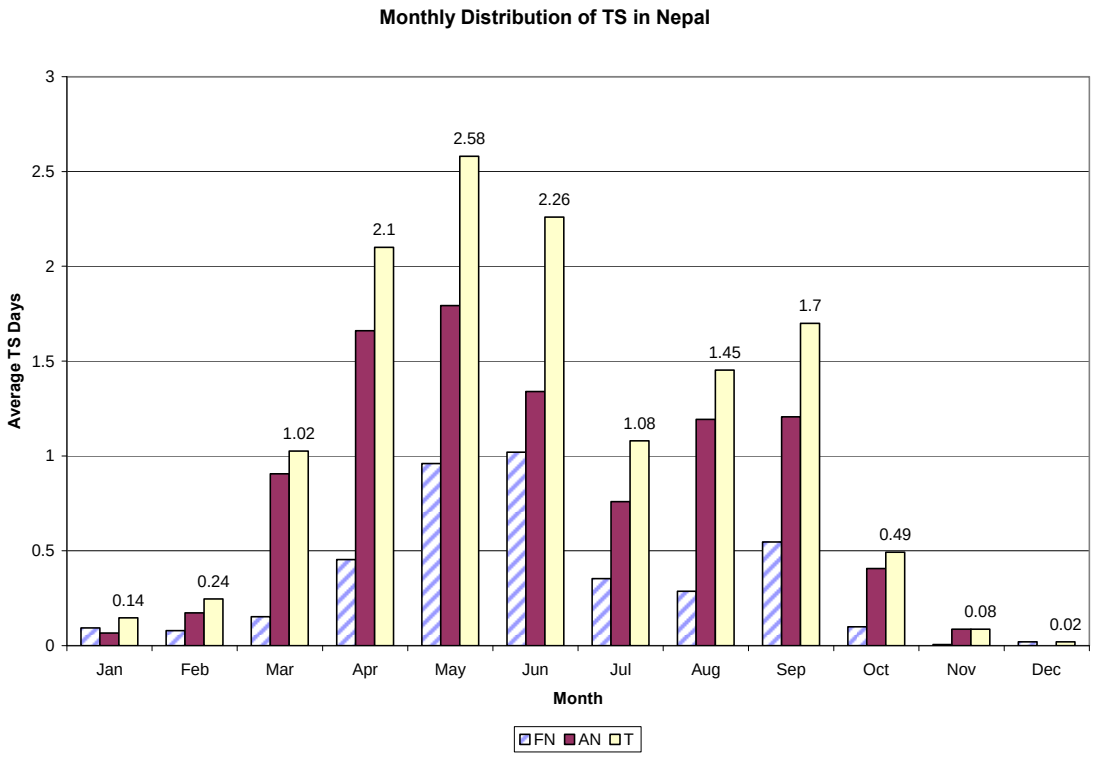

Fig. 3(ii).15: All Nepal monthly distribution of Thunderstorm days in Nepal

3(ii).5. Intraseasonal and interannual variability of monsoon rain

Active and break period of monsoon significantly affect rainfall at different parts of India and Nepal. The fluctuation of monsoon trough to north and south direction from the normal position results in break and active phase in monsoon. Southward shifting of monsoon trough causes active monsoon in central part of India with decreasing rainfall intensity in foothills of Nepal and adjoining northeastern parts of India. Likewise, northward shifting of monsoon trough towards foothills of the Himalayas leads to widespread and heavy rain at various places of Nepal.

Likewise, many studies have been made on Interannual variability and it has been identified that ENSO has been recognized as an important atmospheric feature of the global climate system.

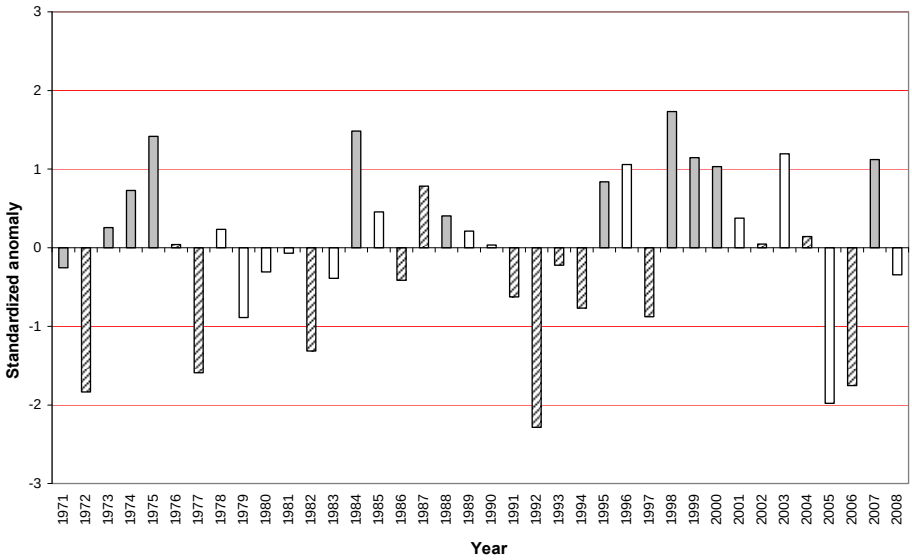

Fig. 3(ii).16: Interannual variation of all Nepal monsoon rainfall (stippled bars are El Nino years; shaded bars are La Nina years).

Shrestha (2000) showed that monsoon rainfall has a large interannual variation and there is tele-connection between Nepal monsoon and warm and cold phases of ENSO. The years with deficient (excess) monsoon rain are most of the time associated with the negative (positive) value of Southern Oscillation Index. Fig. 3(ii).16 shows interannual variation of all Nepal monsoon rainfall. In most of the El

Nino years, the rainfall observed to be less than normal. The year 1992 had most deficit rainfall in Nepal.

3(ii).6. Conclusion:

In a country like Nepal where there is no sufficient irrigation facility, monsoon rain plays an important role. The early or delayed onset of monsoon has profound influence in various sectors like agriculture, water resources etc. Timely onset of monsoon and the adequate amount of rainfall help better agricultural products which strengthen the national economy, whereas excess rain or delayed onset of monsoon create havoc causing floods and landslides or droughts respectively in the country.

The present study on monsoon indicates a decreasing trend in the number of rainy days in the country but, the heavy rainfall such as 100 mm per year is increasing which implies greater chances of floods and landslides in the country.

Likewise, thunderstorms are other weather events during the monsoon season which brings natural calamity. Thunderstorms are frequent in the month of May during pre monsoon season and as the monsoon progresses the frequency of thunderstorms becomes less.

Hence, it is essential to understand the vagaries of monsoon in order to utilize this knowledge effectively in the development process of the country.

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CHAPTER 3(iii)

CLIMATE AND MONSOON OF BHUTAN

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3(iii).1. Introduction

Bhutan is situated in the eastern Himalayas and is mostly mountainous and heavily forested. Bhutan shares it territory with China in the north, south and south east by Indian states of Assam. Sikkim to the west, West Bengal to the southwest, and Arunachal Pradesh to the east.

3(iii).2. Weather and Climate system

The weather in Bhutan is erratic due to dominance of mountainous topography. As such, its climate also varies along with the variation in elevation and like most of the places in Asisa, Bhutan is also affected by monsoons. The low lying regions of the country experience hot days and warm nights, the high lands experience cooler days and cold nights. As such Bhutan is broadly divided into four climate zones as given below.

Climate zones:

- Subtropical Zone
- Temperate Zone
- Sub-alpine Zone
- Alpine zone

Sub-Tropical Zone

- Ranges from 200m to 2000 meters above sea level.
- Winter mean temperature 15 Degree Celsius
- Summer mean temperature 28 Degree Celsius
- The total annual rain fall is usually above 2000mm.

Temperate Zone

- Ranges from 2000m to 3000 meters above sea level.
- Winter mean temperature varies from 5°C to 8°C.
- Summer mean temperature varies 10°C to 22°C.
- The total annual rain fall varies from 1500mm to 2000mm.

Sub-alpine Zone

- Ranges from 3000m to 4000 meters above sea level.
- Heavy Snow fall and frost in winter.
- The total annual rain fall varies from 1000mm to 1500mm.
- Annual Mean Temperature is below 8 degree Celsius.

Alpine Zone

- It refers to the vegetation or climatic condition found on the high lofty mountains
- Temperature is below freezing point
- The permanent snow line starts at between 4800meters to 5000meters.

3(iii).3. Seasons

Bhutan experiences four kinds of seasons in a year.

- Winter(December,January,Febuary)
- Summer(June,July,Agust,September)
- Autumn(October, November)
- Spring(March,April,May)

The above four seasons are grouped under three monsoon seasons:

3(iii).3.1. Pre-monsoon Season (March, April, May)

During the pre-monsoon season, most of the places in Bhutan experiences intermittent rainfalls with gradual increase in temperature. In the high lands it is usually dry dominated with strong gust of winds. During these periods the westerly disturbance is experienced sometimes accompanied by strong winds.

3(iii).3.2. Monsoon Season (June, July, August, September)

Monsoon season last longer than the other two seasons. During this season almost every places in the country receives plenty of rainfall. The weather in most of the places is wet and humid. It is also the hottest time of the year. The monsoon usually starts during early of June and ends during the mid of September. However, there occurs a variation 1-2 weeks in onset and withdrawal of the monsoon. Monsoon is also the hectic season for paddy plantation in Bhutan.

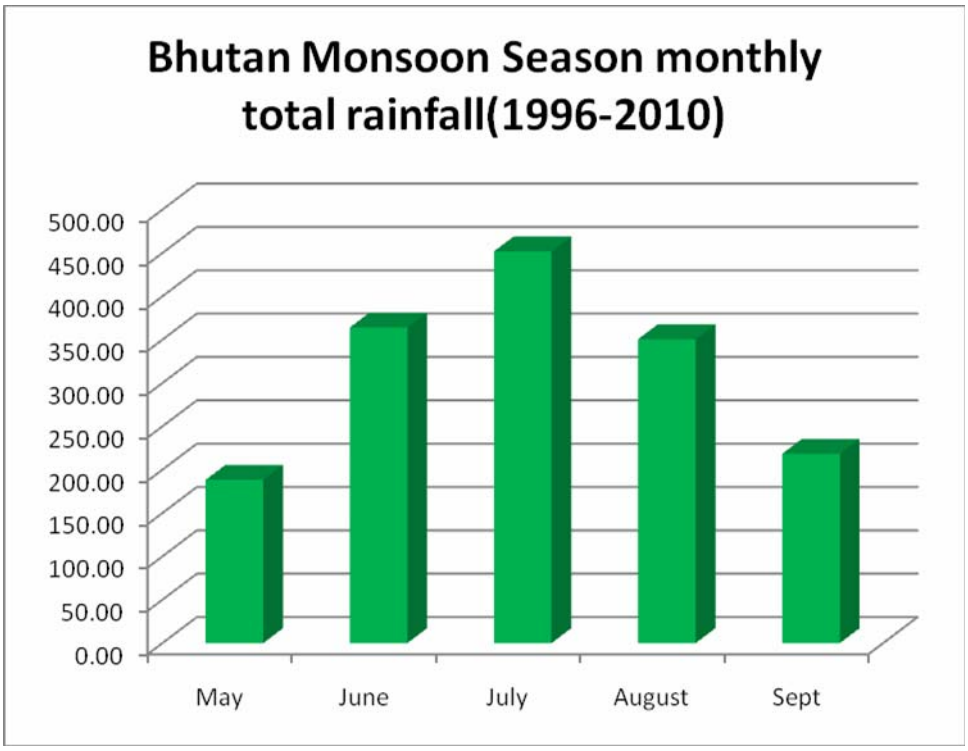

Fig. 3(iii).1: Monsoon season monthly total rainfall (1996-2010).

3(iii).3.3. Post Monsoon (Mid of September, October, November, December)

During this period, it is usually dry and warm. The post monsoon season is also gold season in Bhutan as it's the harvesting season. However, in the later part of this season, the temperature usually drops and in some higher places, it drops beyond freezing point. Higher altitude places experience snow with cold laden winds. The lower foothills experience light showers and low temperatures.

3(iii).4. Short and long term forecasting

At present Department of the Meteorology in Bhutan provides short term forecast only. The long term forecast is not done and the Department of Meteorology in Bhutan is working towards developing a long term forecast.

Nevertheless, Bhutan with assistance from RIMES has generated its own meteo-gram and WRF models for short term weather forecasting (Fig.3(iii).2). During 2010, Department of Meteorology in Bhutan has also developed METGiS for enhancing the weather forecasting system (Fig.3(iii).3).

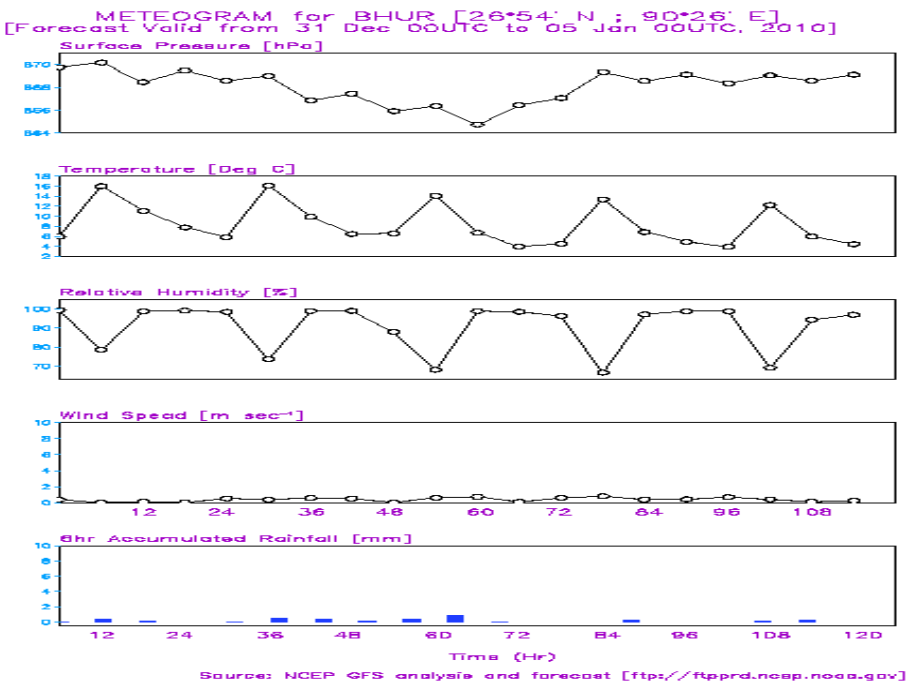

Fig. 3(iii).2: Meteogram generated for daily weather forecast.

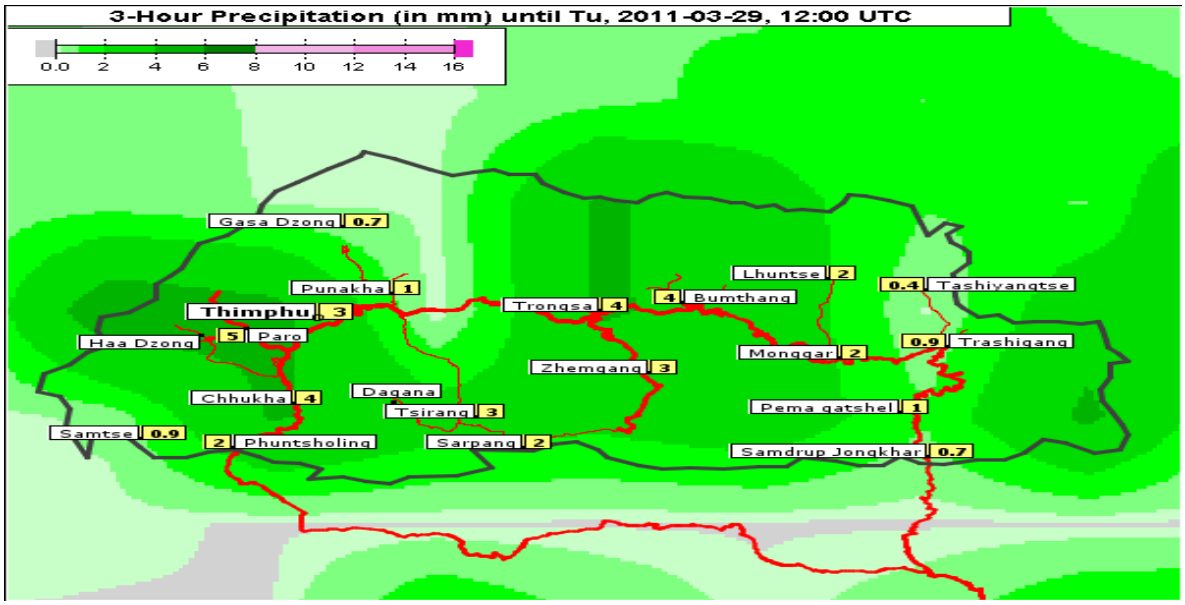

Fig. 3(iii).3: METG is generated for 3 hour precipitation.

3(iii).5. Meteorological Station Networks

The collection of meteorological data was started in early 1990s, however continuation of data collection and quality could not be maintained as there was no separate agency to collect meteorological data at that time. With institution of Meteorology Section,Hydro-met Services Division under Department of Energy, Ministry of Economic Affairs in late 1990s, the continuity of meteorological data collection was maintained.

With the rise in global issues on climate change and importance of hydro-met data for research studies and also to devise preventive and mitigation measures to tackle climate change issues, the Royal Government of Bhutan has upgraded the Hydro-met Services Division into Department of Hydro-met Services (DHMS) on 21 July, 2011 with the broader mandates to look after the issues of Hydro-meteorological aspects.

The Meteorological stations are looked after by the Meteorology Division, DHMS. At present Bhutan has 103 meteorological stations situated all over the country constituting 20 Agro-meteorological stations, 76 Climatological stations and

7 Automatic Weather Monitoring stations of 10m height and GSM system for data transmission in real time (Fig.3(iii).4)

The meteorological parameters such as rainfall, temperature, humidity, wind, solar radiation and evaporation were being recorded. Meteorological data for above parameters are available from 1996 to till date with Meteorology Division.

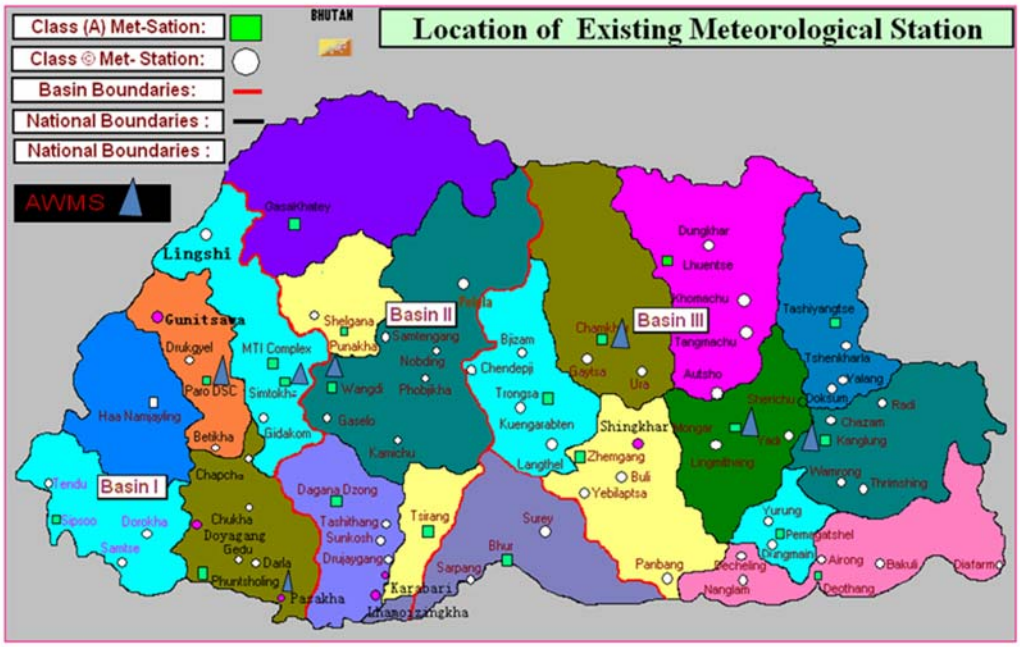

Fig. 3(iii).4: Map showing the meteorological station network.

CHAPTER 3(iv)

SUMMER MONSOON OVER BANGLADESH

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3(iv).1. Introduction

Bangladesh is located in the tropical region. The weather and climate of Bangladesh are strongly influenced by the monsoon from the Indian Ocean (Southwest monsoon). The agriculture and other industries in Bangladesh are favored by the tropical monsoon climate. Therefore the onset of southwest monsoon is very important for Bangladesh agriculture. The onset of the southwest monsoon is preceded by an increase in temperature over almost all the monsoon regions. Prior to the Southwest monsoon season, the monthly mean surface air temperatures exceed 33-35°C in the land areas of northwest India and adjoining areas (Joshi, 1990). The surface air temperatures, particularly the daytime temperatures, drop dramatically with the onset of the southwest monsoon, the monthly mean temperature falling to less than 30°C (Pant, 1997). The progressive development of heat lows before and during the monsoon season is considered to be one of the major causative factors of the monsoon circulation (Rao, 1976). Jay (1987) said the monsoon rain and winds are the end result of heating patterns produced by the sun and the distribution of land and ocean. The winds in the coastal areas are southwest or southerly with speeds of 5-10 m/s and it continues and persists throughout the active phase of the southwest monsoon. An important part of the anatomy of the monsoon circulation is the mean trough running from West Pakistan to the North Bay of Bengal. This trough extends from the sea level to 500 hPa, with southward slope (Keshavamurty, 1970).

Technically monsoon season in Bangladesh is the period lying between the date of onset and date of withdrawal of the monsoon air current over the area. But in the country, the span between the date of onset and date of withdrawal varies. However, over the whole country, June to September is considered as the period of southwest monsoon and the total rainfall of these four months as monsoon rainfall. In consequence of the rapid rise of temperature due to northerly position of the sun in May and early June over the Asian mainland, the air pressure decreases. At the end of May, the Asian high pressure region is replaced by a fairly deep low pressure area extending from Sudan in Africa to west Rajasthan and then to West Bengal. A trough of low pressure forms over Indo-Gangetic plain with its axis extending from west-northwest to east-southeast direction from north-west India along Ganges valley up to the head of the Bay of Bengal. This is the 'monsoon trough'. The axis of the monsoon trough is not stationary, but exhibits periodic movement to the north and south of the Indo-Gangetic plain covering Bangladesh. Its movements have an important bearing on short period variation of monsoon rain over the region.

Though the season is well known as the South-west monsoon season all over Bangladesh due to the prevailing wind direction over the country, but the condition is slightly different in southwestern parts of the country. The monsoon wind in the south Bay of Bengal is mainly directed towards the Myanmar coast; a part of this air stream also advances northwards and is then deflected by the Arakan hills westwards up to the Ganges Plains. The result is that, at the head of the Bay of Bengal the main direction of monsoon wind is more from south-west and south. But a few kilometers northwards from the coast the direction is southerly to easterly. Beyond the central part of Bangladesh, the direction is easterly all over the North Bengal. So, over northwestern part of Bangladesh, the surface wind during the monsoon is mainly southeasterly to easterly instead of southwesterly.

A feature of considerable interest is the existence of a belt of strong easterly wind along the southern periphery of the upper anticyclone. The narrow belt of strong easterlies is observed between 200 and 100 hPa levels. These easterly winds which often record the speed exceeding 100 knots are known to the meteorologists as the easterly jet stream of the tropics. The core of the easterly jet stream is located at about 150 hPa (13 km above mean sea level). A careful study of the easterly jet

stream suggest that its core is located at a higher altitude than the core of the westerly jet in extratropical latitude. Over India and adjoining area, the axis of the easterly jet extends from the southern tip of the peninsula to about 20°N (along Kolkata). In addition to the easterly jet stream over peninsular India, there is westerly jet stream to the north of the Himalayas. The northward movement of the sub-tropical jet provides one with the first indication of the onset of monsoon over Bangladesh and adjoining India. On the other hand, the location of the easterly jet moves north to south in phase with the northward and southward movement of the axis of the monsoon trough.

The South Asian summer monsoon is characterized by strong intra-seasonal variability. Rainy periods over South Asia, called ‘active phases’, alternate with dry spells, called ‘breaks’, during the monsoon season (Goswami, 2005, and references therein). Since the intensity of the rains does not exhibit a strong intra-seasonal variability, the seasonal mean precipitation over the continent depends strongly on the respective durations of these dry and wet phases. Consequently, the inter-annual variations of the monsoon are also linked to the intra-seasonal variability. In particular, the spatial patterns of the intra-seasonal variability are similar to those of the inter-annual variability. Understanding the processes at play in the intra-seasonal variation of the monsoon may be relevant to the understanding of the monsoon system at longer time scales. Indeed, a reliable prediction of the monsoon would allow people to optimize agricultural yields (Gadgil and Rao, 2000) and to prepare for extreme events.

During the monsoon, the seasonal mean precipitation field exhibits two longitudinal bands of maximum rain, one over the equator and the other over the monsoon trough around 20°N (Sikka and Gadgil, 1980). The intra-seasonal variability can be seen as a seesaw between these two tropical convergence zones (TCZs), as shown by the study of composites (Goswami and Mohan, 2001). Spectra of intra-seasonal variability in precipitation and wind exhibit two dominant modes: the 10–20-day and 30–60-day modes. The 10–20-day mode seems to be associated with disturbances propagating from the Pacific warm pool to South Asia that appear as Rossby waves deviated pole ward by the mean monsoon flow (Chatterjee and

Goswami, 2004; Wang and Xie, 1997). On the other hand, the 30–60-day mode is therefore crucial to the seasonal forecast of the onset.

Southwest monsoon is the principal ‘rainy season’ which contributes 71% of the average annual rainfall of Bangladesh (Table 3(iv).1) with 54 to 93 rainy days. The amount of rainfall of the season varies annually from 1180 to 1985 mm. But it varies from 1600 mm in the western part of Bangladesh to 4200 mm in the northeastern and southeastern parts of Bangladesh. This is because of monsoon season does not necessarily mean continuously uniform rain all over the country throughout the entire tenure of the season but the distribution of rainfall suffers from remarkable spatial and temporal variation over different years and from one region to the other in the same year. This spatial and temporal variation of rainfall depends on a number of complex factors some of these are constant or static while others vary from year to the next. Important among these are:

- Location of the place with respect to the moisture bearing monsoon air current,
- Position of land and water,
- Dates of onset, withdrawal and consequent span of monsoon,
- Break in the monsoon and its duration,
- General strength of monsoon,
- Position and movement of the axis of the monsoon trough
- Frequency and movement of depressions, and
- Formation of other low pressure systems.

The part of the country receives rainfall primarily from the Bay of Bengal branch of the monsoon which usually comes from the south, southeast and east. Therefore, as one move from the south to the north and east to west in the area will experience gradually less rainfall over western part of Bangladesh. But for rainfall in eastern and northeastern part of the country the hills located in the eastern side of Bangladesh act as the major powerful factor. As a result, the south facing slopes which almost directly stand on the way of moisture laden monsoon winds receive maximum rainfall and the amount decreases towards south with increasing distance from the hilly region.

The strength of monsoon which determines strength of all these parameters varies considerably from one year to the other causing the annual variation of monsoon rainfall in the country. Distribution of rainfall during the season depends to a great extent on the position and movement of the axis of the 'monsoon trough'. Monsoon trough is a tongue like extension of the low pressure from the seasonal semi-permanent low pressure area in the north western part of India to the head of the Bay of Bengal covering some parts of Bangladesh along the southern margin of the Ganges Plain. The trough is a zone of convergence of westerly winds of the Arabian Sea branch of the monsoon air current to the south and the easterly winds of the deflected Bay of Bengal current to its north. It is, therefore, an area of convergence of two branches of monsoon air current and hence is the area of well distributed moderate to heavy rainfall.

The position of the trough is not stationary and shows marked north-south oscillation. The zone of heavy rainfall also shifts to the north and south along with the movement of the trough. When the monsoon trough shifts to the north and lies very close to the Himalayas, the northern part of Bangladesh and adjoining areas receive heavy to very heavy rainfall causing flood but at the same time the amount of rainfall is remarkably reduced in the central to southern parts of Bangladesh causing practically a 'dry spell'. This is known as 'break chart' condition to the meteorologists. Alternate flood and drought conditions over the northern and southern Bangladesh are often resulted from prolonged persistence of the trough over any of the regions. So, monsoon season is by no means a period of continuous rain in any part of the country. There are alterations of bursts of general rain with partial or general breaks. This pulsatory character of monsoon rainfall is one of the most significant features of the period.

A considerable portion of the monsoon rainfall particularly over Bangladesh and adjoining areas is received in association with the frequency and movement of low pressure areas or depressions from the Bay of Bengal. Usually about eight such low pressure systems form in the Bay of Bengal in the season and travel north-westwards across Gangetic West Bengal. The occurrence of depressions in quick succession is responsible for continuous heavy rain and consequent flood over Bangladesh and the Gangetic West Bengal especially during July and August.

Sometimes, such systems after reaching Gangetic West Bengal following their usual track take right turn and move in a northeasterly direction towards northern Bangladesh across Southern part of the Sub- Himalayan West Bengal causing continuous heavy rain and flood in Bangladesh. Occasionally, feeble low pressure areas also form over Head Bay of Bengal and coastal areas of Bangladesh, remain stationary for two to three days and cause continuous widespread heavy to very heavy rain over Bangladesh.

Table 3(iv).1: Station wise monthly and seasonal normal monsoon rainfall in Bangladesh

Station name	Lat. (°N)	Long. (°E)	June	July	August	September	Seasonal Total
Dhaka	23.77	90.38	340.4	373.1	316.5	300.4	1330.4
Mymensingh	24.72	90.43	394.3	436.3	318.1	335.3	1484.0
Tangail	24.25	89.93	311.2	345.1	253.1	298.1	1207.5
Faridpur	23.60	89.85	345.1	339.8	308.5	264.2	1257.6
Madaripur	23.17	90.18	384.3	401.5	351.5	246.5	1383.8
Chittagong	22.27	91.82	607.3	727.0	530.6	259.3	2124.2
Sandwip	22.48	91.43	652.8	835.1	695.0	395.0	2577.9
Sitakunda	22.58	91.70	548.4	726.8	545.6	316.4	2137.2
Rangamati	22.53	92.20	504.8	572.6	435.2	259.6	1772.2
Comilla	23.43	91.18	329.8	415.5	316.0	226.6	1287.9
Chandpur	23.27	90.70	383.4	424.3	360.6	247.0	1415.3
Maijdi Court	22.87	91.10	532.7	790.2	637.1	359.1	2319.1
Feni	23.03	91.42	529.9	731.5	536.1	324.8	2122.3
Hatiya	22.43	91.10	572.0	698.7	566.2	385.0	2221.9
Cox's_Bazar	21.43	91.93	801.9	924.6	667.1	330.1	2723.7
Kutubdia	21.82	91.85	638.3	763.7	488.9	299.8	2190.7
Teknaf	20.87	92.30	968.1	1029.7	898.9	402.1	3298.8
Sylhet	24.90	91.88	818.4	819.2	612.6	535.9	2786.1
Srimangal	24.30	91.73	449.7	339.4	299.3	278.5	1366.9
Rajshahi	24.37	88.70	264.6	320.7	273.9	295.9	1155.1
Ishurdi	24.13	89.05	288.6	335.6	261.2	282.8	1168.2
Bogra	24.85	89.37	343.8	406.1	285.3	310.1	1345.3
Rangpur	25.73	89.23	417.4	464.8	376.1	383.0	1641.3
Dinajpur	25.65	88.68	335.3	433.6	387.7	383.8	1540.4
Sayedpur	25.75	88.92	435.2	350.0	350.0	456.3	1591.5
Khulna	22.78	89.53	335.6	329.8	323.5	254.7	1243.6

Mongla	22.60	89.50	323.8	342.7	344.4	313.0	1323.9
Satkhira	22.72	89.08	296.6	375.4	297.3	280.1	1249.4
Jessore	23.18	89.17	298.7	304.1	291.8	236.9	1131.5
Chuadanga	23.27	90.7	235.4	351.7	232.8	297.1	1117.0
Barisal	22.75	90.37	408.4	407.3	371.3	259.4	1446.4
Patuakhali	22.33	90.33	547.1	572.8	484.8	380.8	1985.5
Khepupara	21.98	90.23	510.0	650.1	479.9	357.7	1997.7
Bhola	22.68	90.65	465.2	444.5	395.5	264.2	1569.4
Country	-	-	459.4	523.0	420.4	318.2	1721.2

Table 3(iv).2: Station wise monthly and seasonal normal rainy days in Bangladesh

Station name	Lat (°N)	Long(°E)	Jun.	Jul.	Aug.	Sept.	Seasonal Total
Dhaka	23.77	90.38	14	17	16	13	60
Mymensingh	24.72	90.43	16	18	16	14	64
Tangail	24.25	89.93	14	16	14	13	57
Faridpur	23.60	89.85	14	17	16	13	60
Madaripur	23.17	90.18	15	18	17	13	63
Chittagong	22.27	91.82	16	19	17	13	65
Sandwip	22.48	91.43	18	21	19	15	73
Sitakunda	22.58	91.70	16	19	18	14	67
Rangamati	22.53	92.20	18	21	20	15	74
Comilla	23.43	91.18	15	17	16	12	60
Chandpur	23.27	90.70	14	18	17	12	61
Maijdi Court	22.87	91.10	17	21	21	14	73
Feni	23.03	91.42	17	22	19	15	93
Hatiya	22.43	91.10	18	20	21	15	74
Cox's Bazar	21.43	91.93	19	22	21	14	76
Kutubdia	21.82	91.85	17	20	17	13	67
Teknaf	20.87	92.30	21	25	24	15	85
Sylhet	24.90	91.88	22	25	22	18	87
Srimangal	24.30	91.73	18	17	18	14	67
Rajshahi	24.37	88.70	12	15	14	13	54
Ishurdi	24.13	89.05	12	20	14	13	59
Bogra	24.85	89.37	14	17	14	13	58
Rangpur	25.73	89.23	18	21	17	17	73
Dinajpur	25.65	88.68	12	16	14	13	55

Sayedpur	25.75	88.92	15	14	13	13	55
Khulna	22.78	89.53	14	17	16	13	60
Mongla	22.60	89.50	16	19	17	15	67
Satkhira	22.72	89.08	14	19	16	14	63
Jessore	23.18	89.17	17	21	21	16	75
Chuadanga	23.27	90.7	13	17	13	14	57
Barisal	22.75	90.37	16	20	19	13	68
Patuakhali	22.33	90.33	17	22	19	15	73
Khepupara	21.98	90.23	17	22	20	17	76
Bhola	22.68	90.65	20	24	23	17	84

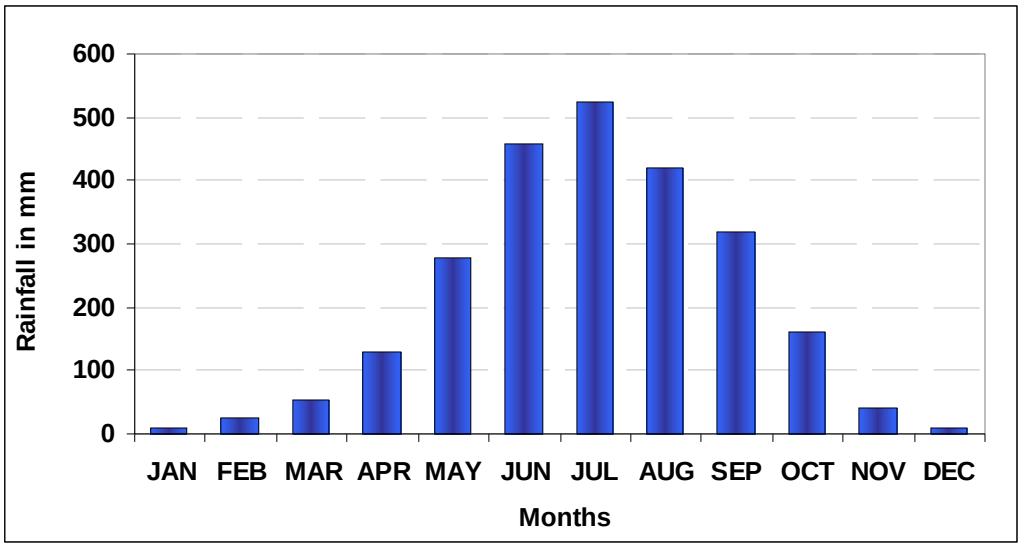


Fig. 3(iv).1: Monthly distribution of monsoon rainfall in Bangladesh

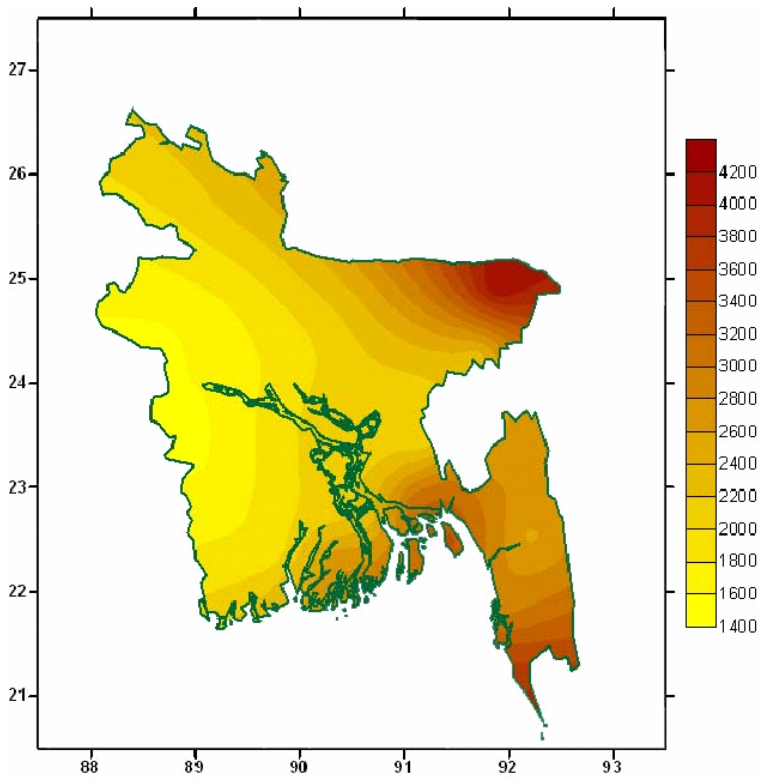


Fig. 3(iv).2: Spatial distribution of monsoon rainfall in Bangladesh

3(iv).2. Data used and methodology:

i. Monthly maximum temperature and minimum temperature and rainfall of 13 meteorological stations of BMD during the period of 1951 to 2010 have been used in the study (shown in Fig. 3(iv).3).

ii. In all cases, there were some missing data that has been filled in by Long Period Average (LPA). Missing data has been incorporated by Kriging method using Win Surfer software.

iii. Time series of monthly (June, July, August & September) and monsoon season mean temperatures (maximum and minimum) and rainfall and linear trends are calculated using MS-Excel.

iv. Anomalies of rainfall, temperature, rainy days are calculated using climatological mean values and displayed using Win Surfer software.

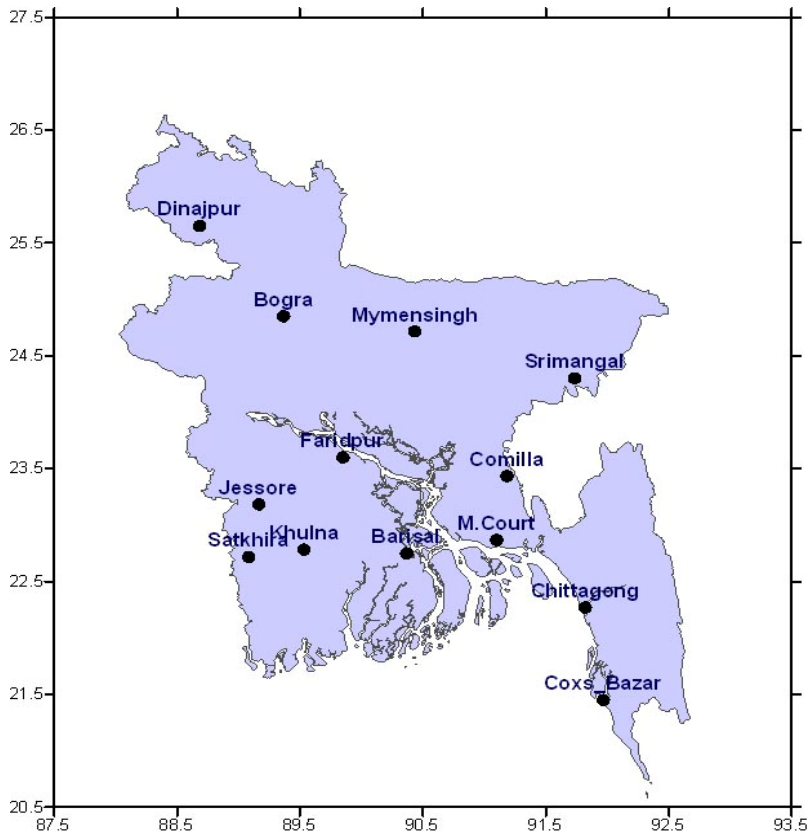


Fig.3(iv).3: Thirteen Key Meteorological Stations used in the study

3(iv).2. Onset and withdrawal of monsoon in Bangladesh

The summer winds originate in an area of high pressure in the southern Indian Ocean, and cross the equator before blowing onshore. The air thus acquires abundant moisture on its northward journey, which fuels convection and storm cloud development during the summer monsoon. The summer monsoon arrives in southern India in late May or early June, and gradually advances northwards and westwards, reaching Pakistan by early July. The monsoon begins to retreat from Pakistan by the beginning of September, and usually withdraws from southern India by early December. This pattern of advance and withdrawal gives the Indian subcontinent its characteristic seasonal rainfall pattern. Pakistan has a short summer rainy season, with generally light rainfall, whereas areas like the Ganges-Brahmaputra Delta have a longer, heavier monsoon.

The onset and withdrawal of the broad scale monsoon occur in many stages and represent significant transitions in the large-scale atmospheric and ocean circulations in the Indo–Pacific region (e.g., Rao, 1976; Murakami and Nakazawa, 1985; Lau et al., 1998; Hsu et al., 1999; Wu and Zhang, 1998). While there exist no widely accepted definitions of these monsoon transitions, at the surface the onset is recognized as a rapid, substantial, and sustained increase in rainfall over a large scale while the withdrawal marks the return to dry, quiescent conditions. Typically, rainfall amounts over South Asia increase from below 5 to over 15 mm day^{-1} during onset (Anathakrishnan and Soman, 1988, Soman and Kumar, 1993). It is also known that the onset experiences spatial coherency over a large scale. In general, the first rains of the monsoon occur over Burma and Thailand in mid May and extend subsequently to the northwest, so that by mid June, rains have advanced over most of India and Bangladesh. Near India, the onset occurs initially across the peninsula's southern tip in early June, progressing northwestward across most of the country in the following month.

The onset and withdrawal of the monsoon is characterized by rapid seasonal transitions in atmospheric circulation and precipitation over a large area of the Asian continent (Hsu et al.,1999). Objectively defining the timing of the onset and withdrawal of the monsoon, however, has been a difficult task, often being based on atmospheric variables exceeding some arbitrary threshold given with little (and sometimes no) explicit justification. Example thresholds for calculation of monsoon onset include- the day on which a 5 day running mean rainfall index exceeds 5 mmd^{-1} and persists continuously for 5 days while, over the next consecutive 20 days, the number of days with rainfall greater than 5 mmd^{-1} exceeds 10 days (Zhang et al., 2002); wind speeds in excess of 8 ms^{-1} for more than 7 consecutive days (Taniguchi and Koike, 2006); increases in precipitable water above the 'Golden Ratio' (Zeng and Lu, 2004); and normalized (+1 to -1) vertically integrated moisture transport rising above 0 (Fasullo and Webster, 2003). Even a few studies (Goswami and Xavier, 2005) have attempted to calculate monsoon season length and withdrawal. These remain valuable contributions and highlight the variety of methods and variables that can be used to diagnose monsoon season transitions. Still, a more objective determination of monsoon onset and withdrawal, without the requirement of a priori selected thresholds and critical values, would be valuable.

Southwest monsoon arrives first in the extreme south-eastern part of Bangladesh and initially moves towards the north. After diverted by the Meghalaya Plateau, the flow turns towards the west. The mean arrival dates of the monsoon in the extreme south-eastern coastal part and in the extreme north-western part are 2 June and 15 June respectively. For deriving the onset and withdrawal of the monsoon, BMD uses one of the most objective techniques using data readily available from meteorological stations. The criteria used in Bangladesh for onset of monsoon are as follows:

- There are three consecutive rainy days in a five day period.
- The consecutive rainy days must have not less than 5 mm each day.
- The accumulated rain of the five rainy days must not be less than 25 mm.
- The low level wind direction must change to southwest or southerly.
- The upper level wind (initially 300 hPa level) must change to easterly.

Probabilistic early dates of arrival of the summer monsoon in Cox's Bazar area in 1 year out of 4 years, 10 years, and 25 years are 28 May, 23 May, and 19 May, respectively. Probabilistic late dates of arrival in 1 year out of 4 years, 10 years, and 25 years are 8 June, 13 June, and 16 June, respectively. Withdrawal of summer monsoon proceeds in the opposite direction to the arrival. Mean withdrawal dates of the summer monsoon from the extreme north-western part and extreme south-eastern part of the country are 30 September and 17 October, respectively. Standard deviations of both arrival and withdrawal dates in different parts of the country vary from 7 to 10 days. Probabilistic early dates of complete withdrawal of the summer monsoon from Bangladesh in 1 year out of 4 years, 10 years, and 25 years are 10 October, 5 October, and 1 October, respectively. Probabilistic late dates of complete withdrawal in 1 year out of 4 years, 10 years, and 25 years are 23 October, 28 October, and November 1, respectively.

During the early part of May, a narrow zone of air mass discontinuity lies across the country from south-western corner to the north-eastern part, separating the hot dry air from the dry interior of northern India and the warm moist air from the

Bay of Bengal. As the season progresses, this discontinuity becomes weak and gradually it retreats towards northwest, finally disappearing by early June, making room for the onset of the summer monsoon. The withdrawal of the summer monsoon from Bangladesh begins in the late September or very early October from the north-western part of the country, and finally leaves the country by the first week/ middle of October through the south-eastern corner of the country. The retreat of the summer monsoon is accompanied by the northerly and north-westerly winds of the winter monsoon. The normal dates of onset and withdrawal of monsoon are given in Fig. 3(iv).4.

Analysis reveals that during 1991 to 2011 the onset dates of monsoon in Bangladesh as earliest as 26th May in 2009 and as latest as 16th June in 1996 as given in Fig.3(iv).5. Similarly, the withdrawal dates of monsoon in Bangladesh is as early as 05 October in 1994 and as late as 18th October in 2007 as depicted in Fig. 3(iv).6.

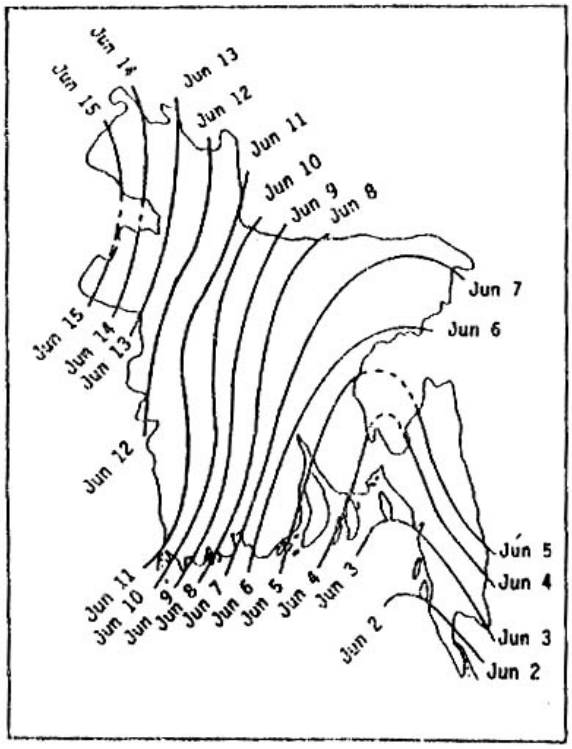


Fig. 3(iv).4(a): The normal onset dates of monsoon in Bangladesh

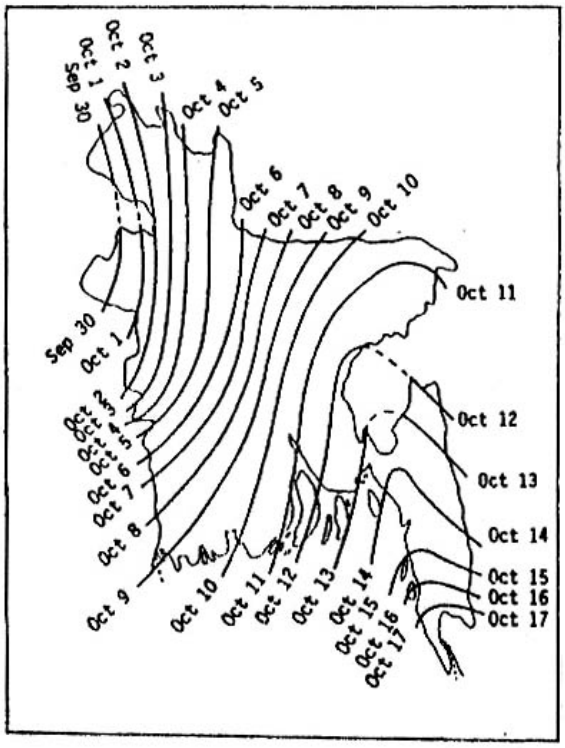


Fig. 3(iv).4(b): The normal withdrawal dates of monsoon in Bangladesh

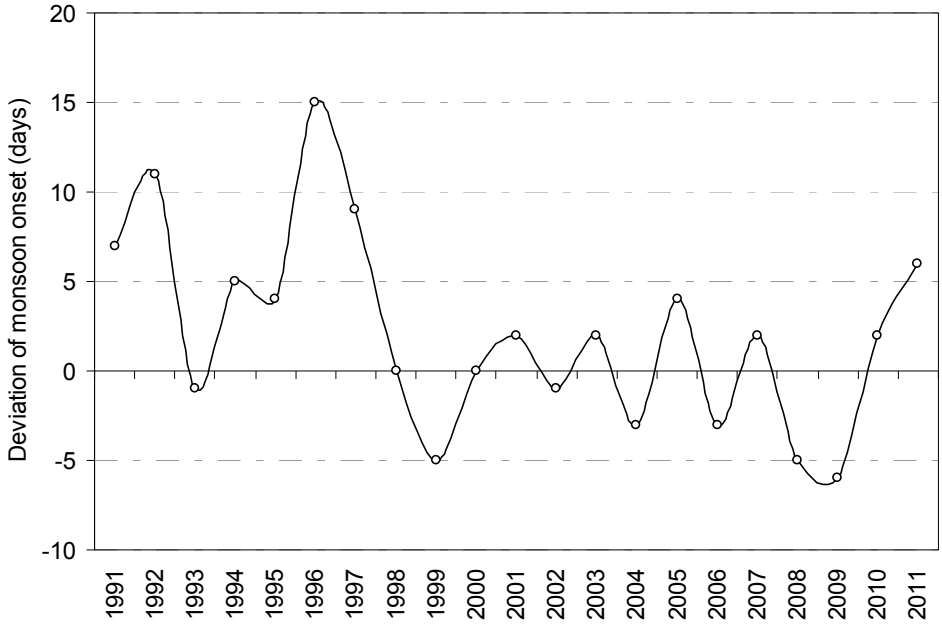


Fig. 3(iv).5: Variation of monsoon onset dates of Bangladesh during 1991-2011

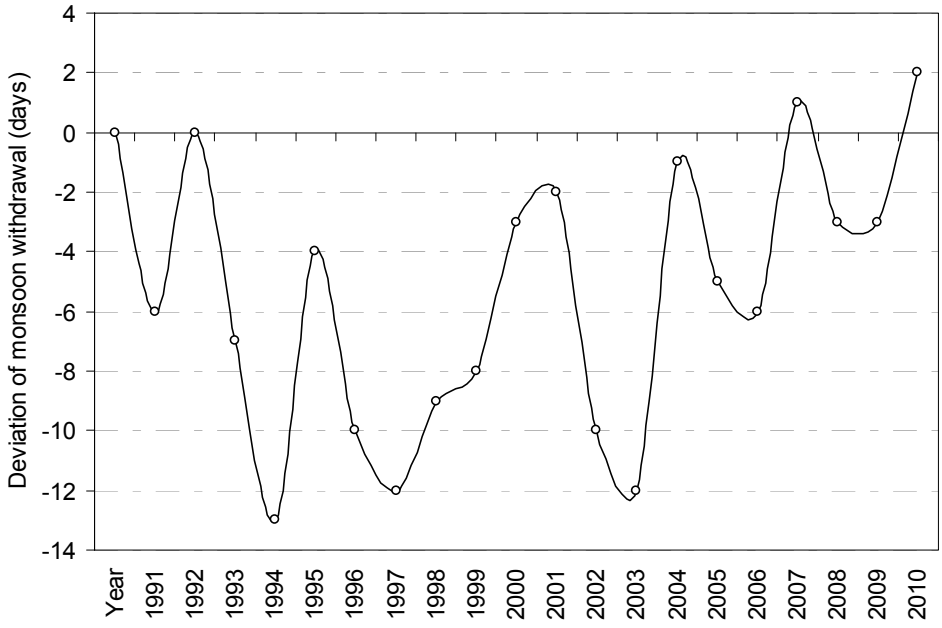


Fig. 3(iv).6: Variation of monsoon onset dates of Bangladesh during 1991-2010

3(iv).3. Temporal variation and spatial distribution of Temperature in Bangladesh

3(iv).3.1. Temporal variation of Maximum Temperature of monsoon season in Bangladesh during 1950-2010

The inter-annual variability of maximum temperature of monsoon season is low compared to pre-monsoon season in Bangladesh as revealed in Fig.3(iv).7. During this season maximum temperature was nearly steady (+ 0.002°C/year) during 1950-1982 but its trend was significant positive afterwards (+0.033°C/year) and hence the overall trend was positive (+0.0199°C/year) during the observed period. Country average highest maximum temperature of 32.8°C was recorded in 2009 followed by 32.6°C and 32.4°C in 2005 and 1998 respectively.

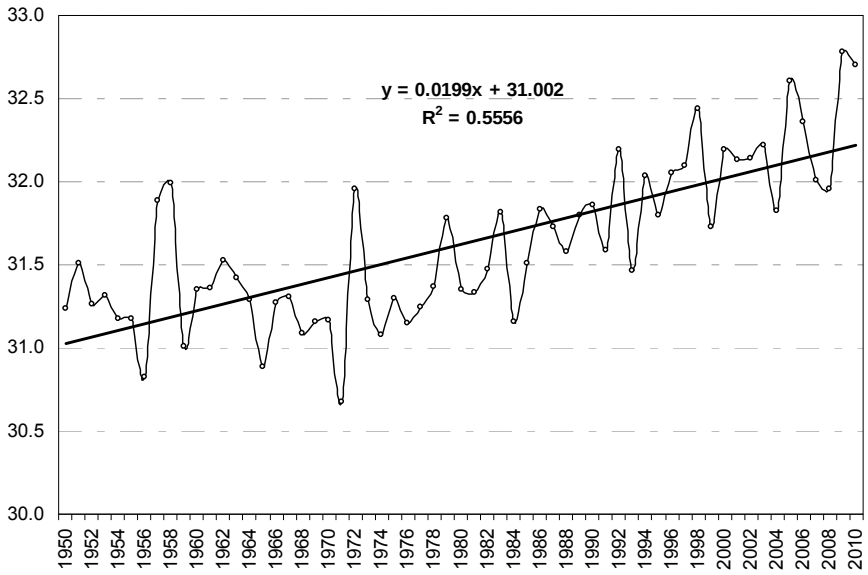


Fig: 3(iv).7: Temporal variation of maximum temperature of monsoon season in Bangladesh during 1950-2010

3(iv).3.2. Temporal variation of Annual Maximum Temperature of Bangladesh during 1950-2010

The inter-annual variability of annual maximum temperature is high in Bangladesh. During 1950-1984 the trend of it was little negative (-0.0065°C/year) but it was significant positive (+0.0177°C/year) afterwards and therefore the overall trend was slight positive (+0.0092°C/year) during the observed period as shown in Fig.

3(iv).8. The country average highest maximum temperature of 31.4°C was recorded in 2009 followed by 31.2°C 1999 and 2006.

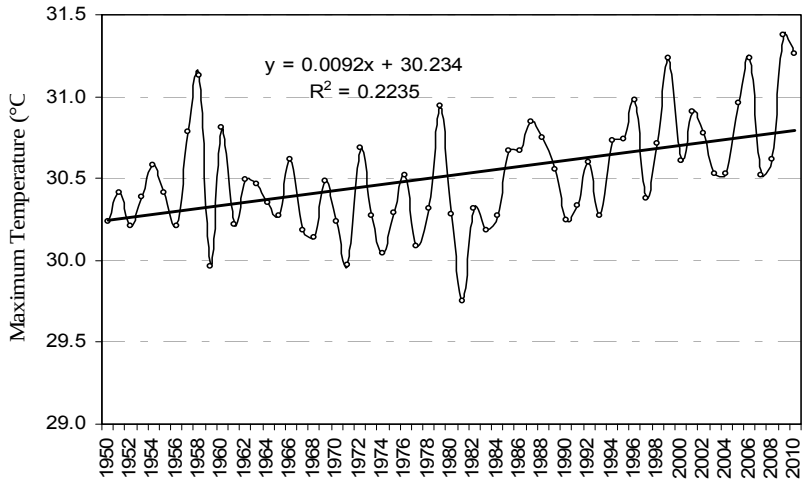


Fig: 3(iv).8: Temporal variation of annual maximum temperature of Bangladesh during 1950-2010.

3(iv).3.3. Temporal variation of Minimum Temperature of monsoon season in Bangladesh during 1950-2010

The inter-annual variability of monsoon minimum temperature is low in Bangladesh. The trend of country average minimum temperature was considerable negative (-0.0113°C/year) during 1950-1976, but its trend was stable (-0.0003°C/year) during 1976-1997 and significant positive (+0.0216°C/year) afterwards. Therefore, the overall trend was positive (+0.0081°C/year) during the observed period. Country average lowest minimum temperature of 25.0°C was recorded in 1976 followed by 25.0°C and in 1974 and 25.1°C in 1971 as given in Fig. 3(iv).9.

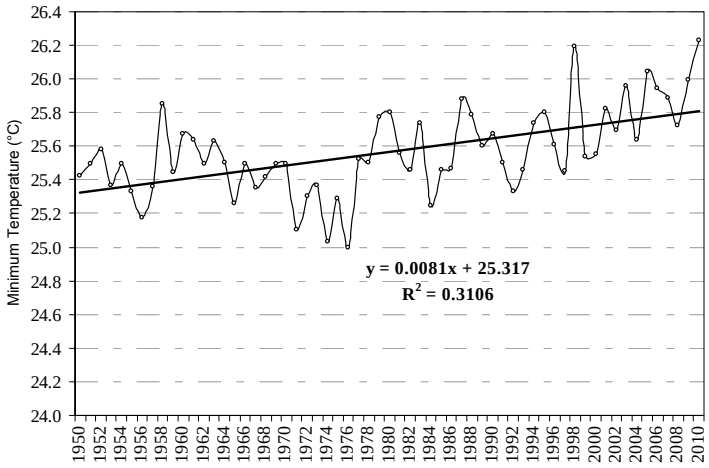


Fig.3(iv).9: Temporal variation of minimum temperature of monsoon season in Bangladesh during 1950-2010

3(iv).3.4. Temporal variation of annual Minimum Temperature of Bangladesh during 1950-2010

The inter-annual variability of annual minimum temperature is low in Bangladesh. The trend of country average minimum temperature was nearly steady (+0.0005°C/year) for the period of 1950-1984, but it was significant positive (+0.0225°C/year) afterwards. As a result, the overall trend was significant positive (+0.0107°C/year) during the observed period. Country average lowest minimum temperature of 20.5°C was recorded in 1962 and 1971 followed by 20.6°C in 1950 and 1957 as given in Fig. 3(iv).10.

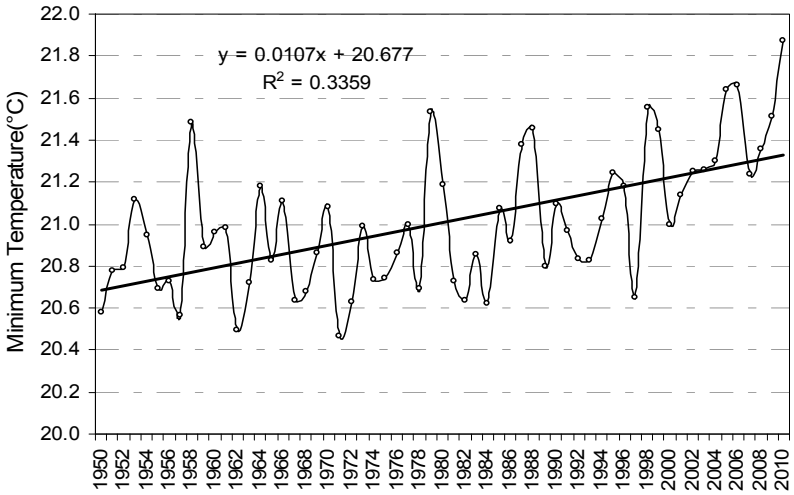


Fig. 3(iv).10: Temporal variation of annual minimum temperature of Bangladesh during 1950-2010

3(iv).3.5. Temporal variation of annual mean temperature of Bangladesh during 1950-2010

The inter-annual variability of mean annual temperature of Bangladesh during 1950-2010 is illustrated in Fig. 3(iv).11. Fig. 3(iv).11 shows that mean temperature of Bangladesh varies between 25.2°C to 26.6°C during the observed period having positive trend. The trend of country average annual mean temperature was nearly steady (+0.0008°C/year) for the period of 1950-1980 but it was considerable positive (+0.0261°C/year) afterwards. As a result, the overall trend was significant positive (+0.0099°C/year) during the observed period. The lowest mean temperature of 25.2°C was recorded in 1971 and 1981, but the highest mean temperature of 26.6°C was observed in 2010, followed 26.4°C in 2006 and 2009.

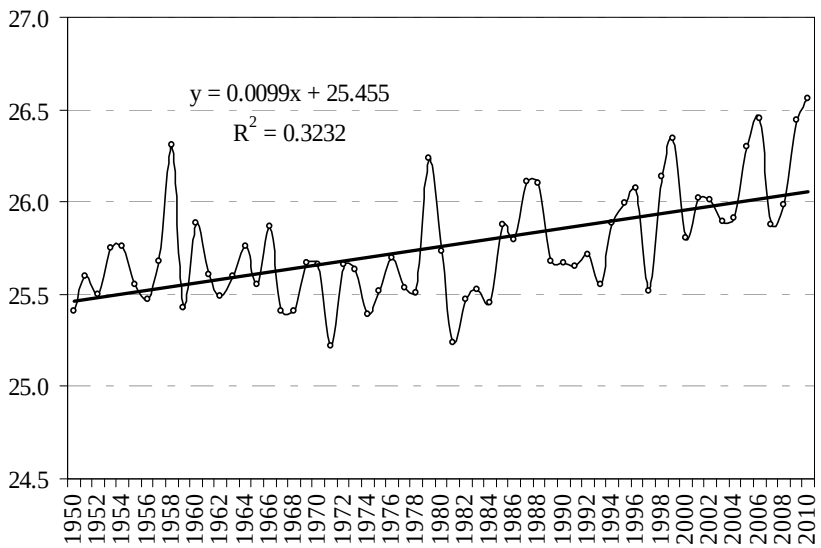


Fig. 3(iv).11: Temporal variation of annual mean temperature of Bangladesh during 1950-2010

3(iv).3.6. Daily variation of maximum, minimum and mean temperature in monsoon season during 1951-2010

Analysis reveals that due to abrupt change in wind, pressure system, cloud coverage there is an abrupt change in maximum temperature immediately after onset of monsoon which falls rapidly from 33.2°C to 31.3°C during the first half of June and then remains almost constant till July. After that, it raises by about 0.5°C in September as the strength of monsoon circulation weakens, intensity of rainfall and cloud coverage decreases. Further, raise of maximum temperature of about 0.5°C is

observed in September when the monsoon circulation weakens further over the country and accordingly southerly flow retreats, cloud coverage decreases, pressure distribution pattern becomes irregular as depicted in Fig. 3(iv).12(a). Similarly, with the progress of the season, moisture intrusion is taking place and continues in the lower troposphere, and hence minimum temperature becomes stable and lies within 25.4°C to 25.8°C during June and July. In August, as the monsoon flow becomes weak, minimum temperature raises a little, but in September as the monsoon flow becomes considerably weak, pressure distribution pattern becomes irregular, accordingly northerly flow introduces temporarily and Sun's position starts to shift southward, the minimum temperature starts to drop as low as 24.7°C as illustrated in Fig. 3(iv).12(b). Responding with the maximum and minimum temperature, the mean temperature of the country changes, but as minimum temperature of the season remains almost steady, the temporal variation of mean temperature pattern is very similar to maximum temperature with lower values as given in Fig. 3(iv).12(c).

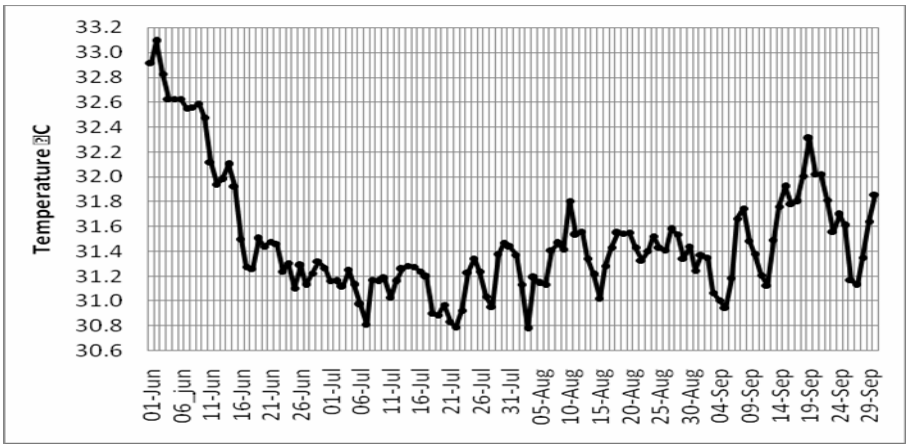


Fig. 3(iv).12 (a): Daily variation of normal maximum temperature of Bangladesh during monsoon season

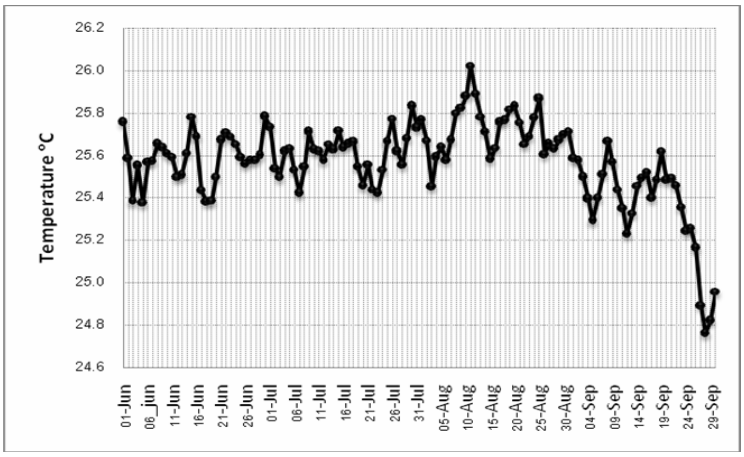


Fig. 3(iv).12(b): Daily variation of normal minimum temperature of Bangladesh during monsoon season

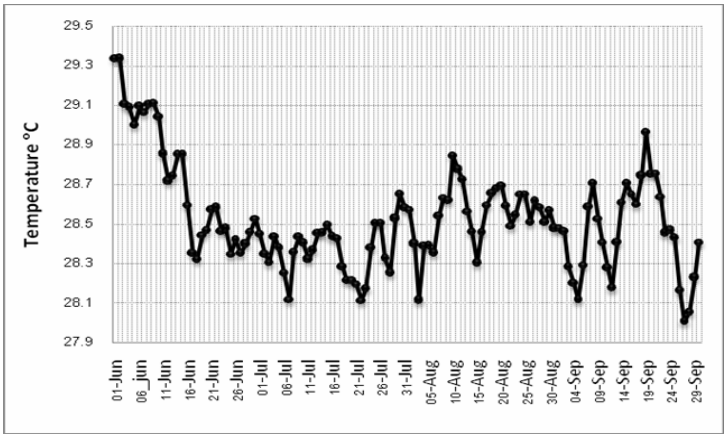


Fig. 3(iv).12(c): Daily variation of normal mean temperature of Bangladesh during monsoon season

3(iv).4. Temporal variation and spatial distribution of monthly and seasonal rainfall in Bangladesh

3(iv).4.1. Daily variation of rainfall during monsoon season in Bangladesh

Analysis depicts that with the progress of time from 01st June to last week of June, the average daily rainfall of Bangladesh increases with the amounts of 10-17mm and then remains steady till last week of July having slight variability in rainfall amounts. During first half of August, the amount of rainfall decreases to 08-10mm and then increases to12-14mm and remains steady till first few days of September and decreases afterwards. During the last week of September, amounts of daily rainfall

increases again and then decreases quickly with the withdrawal of monsoon from Bangladesh and the details are illustrated in Fig.3(iv).13.

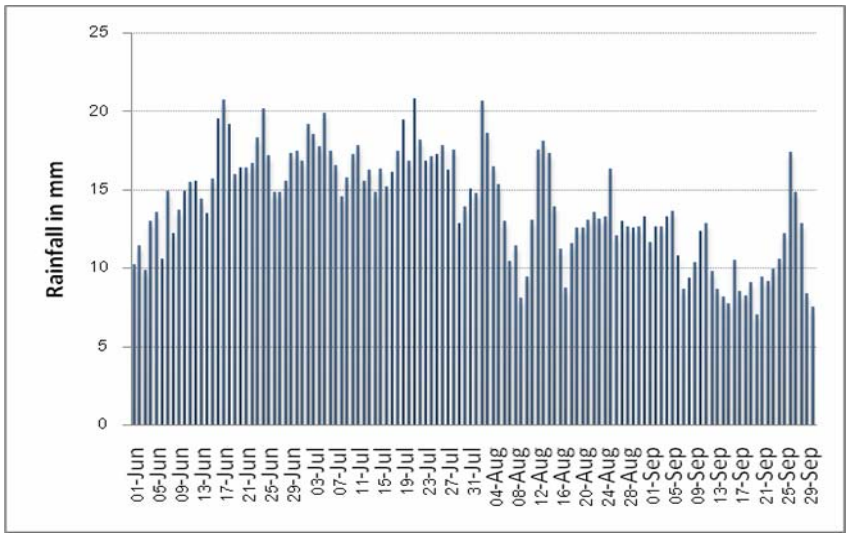


Fig. 3(iv).13: Daily variation of country average normal rainfall of Bangladesh during Monsoon season

3(iv).4.2. Temporal variation of monthly monsoon rainfall of Bangladesh during 1951-2010

The temporal variation of monsoon rainfall of June is illustrated in Fig. 3(iv).14(a). Fig. 3(iv).14(a) reveals that variability of June rainfall is high due to the causes of late onset of monsoon, weak phase of monsoon circulation, absence of low level circulation which some times initiating formation of monsoon depression over North Bay, position of monsoon axis further south to its normal position and strong westerly which extend longer duration of prevailing pre-monsoon characteristics over Bangladesh and some other causes. Analysis reveals that country averaged rainfall were 300 mm or below in 1958, 1967, 1975, 1980, 1987, 1998 and 2009; 400 mm or below in 1959, 1960, 1964, 1972, 1979, 1981, 1983, 1985, 1989, 1992, 1997, 2000 and 2005; 700 mm or above in 1954 and 1984; 600 mm or above in 1988, 1993, 2001 and 2003; 500 mm or above in 1953, 1956, 1961, 1963, 1965, 1966, 1968, 1976, 1977, 1978, 1982, 1991, 1993, 2001, 2002 and 2007. The lowest amounts of rainfall of 202.6 mm was recorded on 1998 followed by 258.0 mm and 266.7 mm in 2009 and 1958 respectively during the observed period but the highest amount of rainfall of 787.7 mm was recorded in 1954 followed by 726.3 mm in 1984. The tendency of amount of rainfall is decreasing (- 0.701 mm/year) during the observed period.

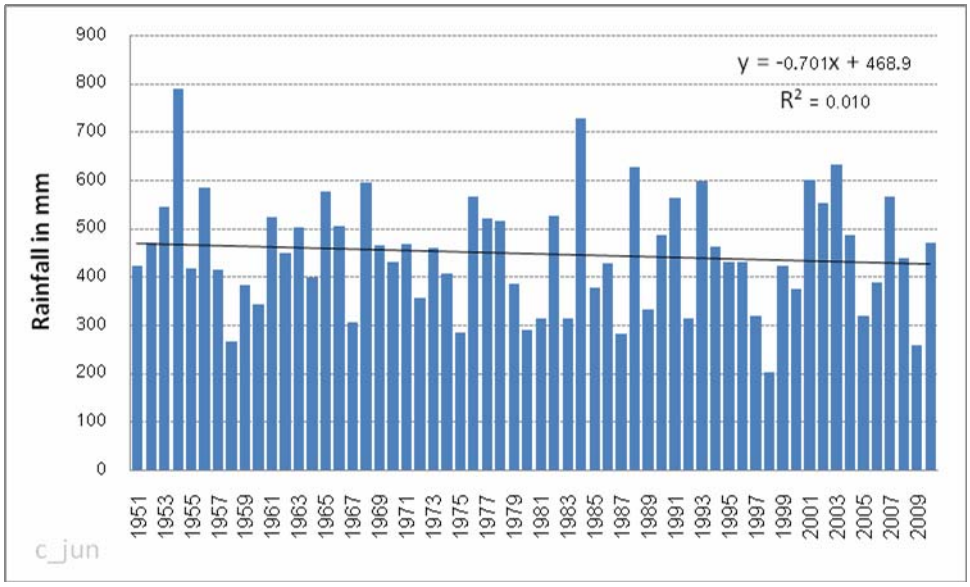


Fig. 3(iv).14(a): Inter-annual variation of June rainfall in Bangladesh during 1951-2010

Fig. 3(iv).14(b) illustrates the temporal variation of monsoon rainfall of July in Bangladesh. Fig. 3(iv).14(b) reveals that the rainfall of July is comparatively higher than that of June as a result of distinctive influences of monsoon circulation over Bangladesh. But the amount of rainfall varies due to the changing pattern of monsoon circulation, movement and duration of active and break phases of monsoon, presence and absence of low level circulation leading to the formation of monsoon depression over Bay of Bengal, position and shifting of low level jet and upper air anticyclonic circulation. Analysis reveals that country average rainfall of July varies mainly within the range of 350 to 500 mm it was lower than 350 mm in 1966, 1972, 1979, 1985, 1994, 2001, 2003 and 2010; 600 mm or above in 1964, 1970, 1974, 1987, 1997, 1998 and 2007. The lowest amount of rainfall of 268.0 mm was recorded in 2010 followed by 298.2 mm, 317.8 mm and 320.1 mm in 1994, 1972, 1966 respectively. The highest amount of rainfall of 762.3 mm was recorded in 1987 followed by 679.9 mm and 664.5 mm in 1998 and 1974 respectively. The tendency of the amount of rainfall is slight positive (+0.006 mm/year) during the observed period.

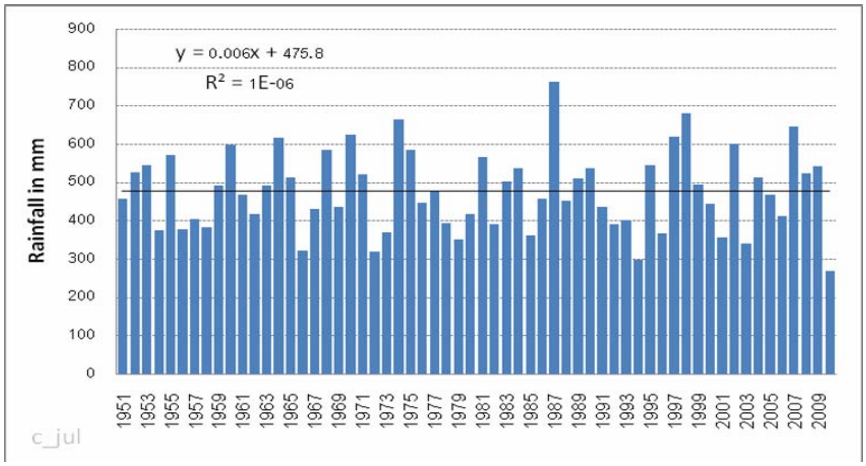


Fig. 3(iv).14(b): Inter-annual variation of rainfall in July over Bangladesh during 1951-2010

In August, the amounts of rainfall is comparatively lower than July but the variability is high as a result of active and weakening phases of monsoon circulation over North Bay and Bangladesh, movement and duration of active and break phases of monsoon with the shifting of monsoon axis, presence of low level circulation leading to the formation of monsoon depression over Bay of Bengal and crossed Bangladesh coast, position and shifting of low level jet and upper air anticyclonic circulation. Analysis reveals that country average rainfall of August varies mainly within the range of 300 to 400 mm, but it was lower than 300 mm in 1957, 1963, 1973, 1975, 1977, 1986, 1989, 1990, 1992, 2001 and 2003; 500 mm or above 1956, 1959, 1965, 1969, 1983, 1987, 1998 and 2009. The lowest amount of rainfall of 152.8 mm was recorded in 1989, followed by 245.9 mm in 1990. The highest amount of rainfall of 688.5 mm was recorded in 1983 followed by 601.5 mm and 566.6 mm on 1969 and 1987 respectively. The trend of the amount of rainfall is significant negative (-0.735 mm/year) during the observed period. The details are specified in Fig. 3(iv).14(c).

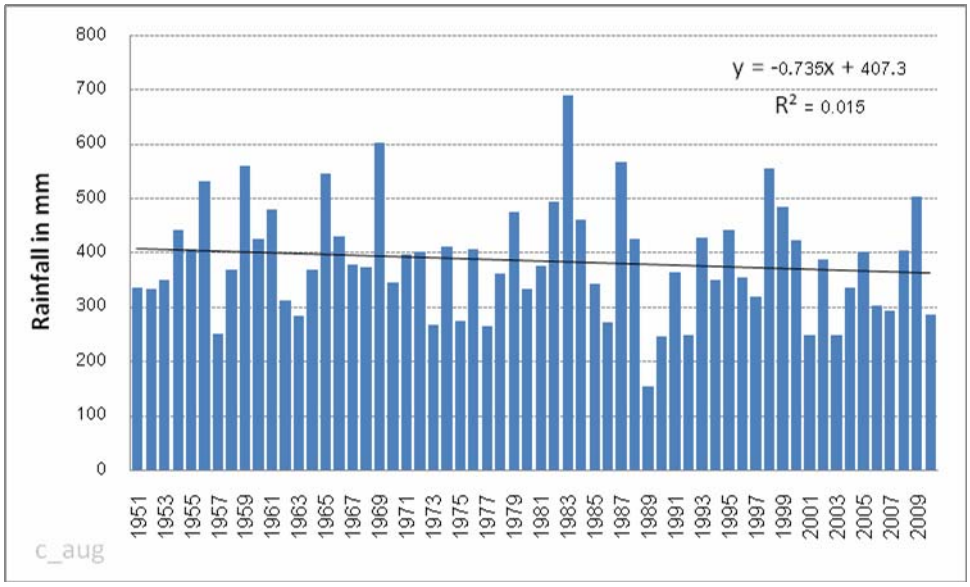


Fig. 3(iv).14(c): Inter-annual variation of rainfall in August over Bangladesh during 1951-2010

In September, the country average rainfall is comparatively lower than June, July and August, but the variability is high as a consequence of longer durable active and weakening phases of monsoon circulation over North Bay and Bangladesh, weakening stage of low level circulation over Bay of Bengal, southward shift of low level jet and weakening upper air anticyclonic circulation. Analysis reveals that country average rainfall of September varies mainly within the range of 200 to 350 mm but it was lower than 200 mm in 1954, 1972, 1977, 1994 and 2003; 400 mm or above in 1959, 1967, 1986, 2004 and 2006. The lowest amount of rainfall of 149.3 mm was recorded in 1994, followed by 159.0 mm, 183.2 mm and 189.5 mm in 1972, 1977 and 2003. The maximum amount of rainfall of 592.8 mm was recorded in 2004, followed by 527.0 mm and 439.0 mm on 1986 and 2006 respectively. The tendency of the amount of rainfall is significant positive (+0.370 mm/year) during the observed period as specified in Fig. 3(iv).14(d).

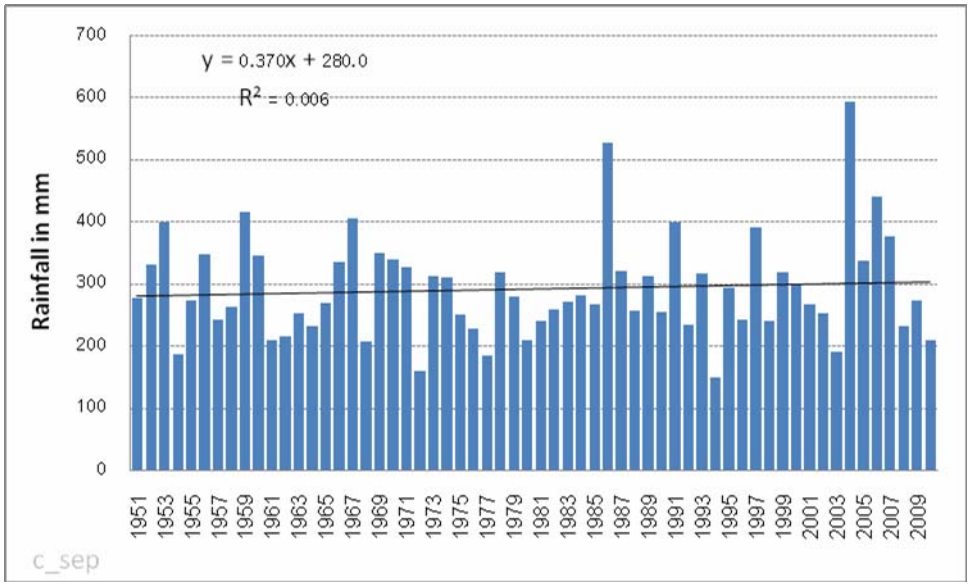


Fig. 3(iv).14(d): Inter-annual variation of rainfall in September over Bangladesh during 1951-2010

3(iv).4.3. Temporal variation of monsoon rainfall of Bangladesh during 1951-2010

Country average monsoon rainfall depends on the time of onset and spreading of monsoon over Bangladesh, duration of consecutive active and weakening phases of monsoon circulation over North Bay and Bangladesh, position and strength of upper air anticyclonic circulation, low level circulation associated with monsoon circulation over Bay of Bengal, position and shipment of low level jet over the subcontinent and withdrawal phase of monsoon. Analysis reveals that country average rainfall is nearly 1500 mm, but it varies mostly between 1400 to 1600 mm. It was lower than 1300 mm in 1957, 1958, 1962, 1972, 1980, 1985, 1989, 1992, 1994 and 2010; 1600 mm or above in 1953, 1954, 1956, 1959, 1965, 1969, 1974, 1985, 1987, 2002, 2004 and 2007. The lowest amount of rainfall of 1183.8 mm was recorded in 1992 followed by 1231.3 mm, 1234.6 mm and 1249.8 mm in 2010, 1972 and 1980. The maximum amount of rainfall of 2004.0 mm was recorded in 1984, followed by 1930.4mm and 1925.8 mm on 1987 and 2004 respectively. The tendency of the amount of rainfall is slight negative (-1.06 mm/year) during the observed period as specified in Fig. 3(iv).15.

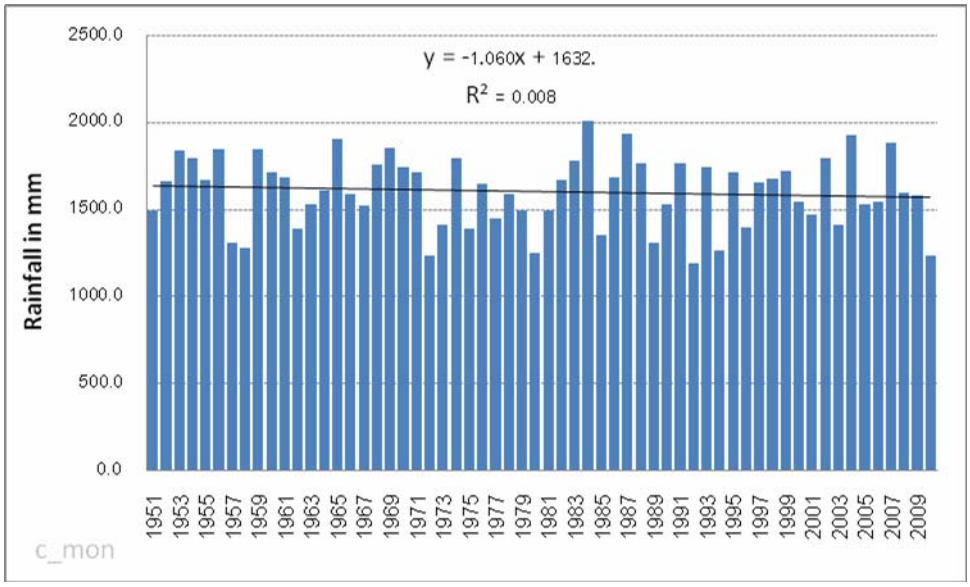


Fig. 3(iv).15: Inter-annual variation of monsoon rainfall in Bangladesh during 1951-2010

3(iv).5. Spatial distribution of monthly and seasonal rainfall and rainy days during monsoon season in Bangladesh

3(iv).5.1. Spatial distribution of monthly monsoon rainfall of Bangladesh during 1951-2010

Analysis reveals that due to onset of monsoon, meeting of monsoon flow with westerly flow and hill slopes, the maximum amount of June rainfall is recorded in the southeastern part of Bangladesh, but the secondary maximum is observed in the northeastern part of the country as monsoon reaches earlier and there are hill slopes in the eastern side. Rainfall is comparatively lower in the central and western part of county which is comparatively flat and monsoon reaches 5-7 days later. The maximum amount of rainfall of 968.1 mm is recorded at Teknaf, followed by 818.4 mm at Sylhet, but the minimum amount of rainfall of 235.4mm is recorded at Chuadanga as shown in Fig. 3(iv).16(a). The distribution pattern of July rainfall is similar to June, but the maximum amount of rainfall is recorded in the southeastern part of the country with the secondary maximum in the northeastern part as shown in Fig. 3(iv).16(b). The maximum amount of rainfall of 1029.7 mm is recorded at Teknaf, followed by 924.6 mm at Cox's Bazar, but the minimum amount of rainfall of 304.1 mm is recorded at Jessore.

In August the higher amounts of rainfall are recorded in the southeastern part of the country followed by northeastern part, but lower amounts of rainfall are recorded in the west-central part of the country as shown in Fig. 3(iv).16(c). The maximum amount of rainfall of 898.9 mm is recorded at Teknaf, followed by 695.0 mm at Sandwip, but the minimum amount of rainfall of 232.8 mm is recorded at Chuadanga. In September, the higher amounts of rainfall are recorded in the northeastern part of the country, followed by northern part, as there are westerly influences and local convection is taking place during the weakening stage and break monsoon period, but lower amounts of rainfall were recorded in the central part of the country as shown in Fig. 3(iv).16(d). The maximum amount of rainfall of 535.9 mm is recorded at Sylhet followed by 456.3 mm at Sayedpur, but the minimum amount of rainfall of 226.6 mm is recorded at Comilla.

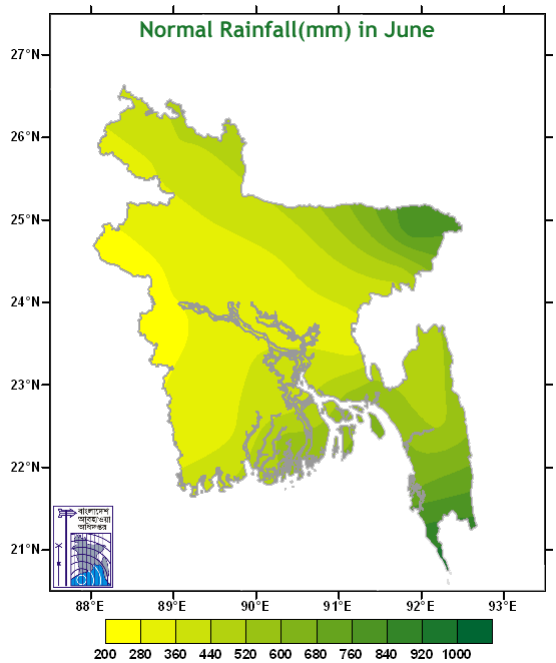


Fig. 3(iv).16(a): Spatial distribution of normal rainfall of Bangladesh in June

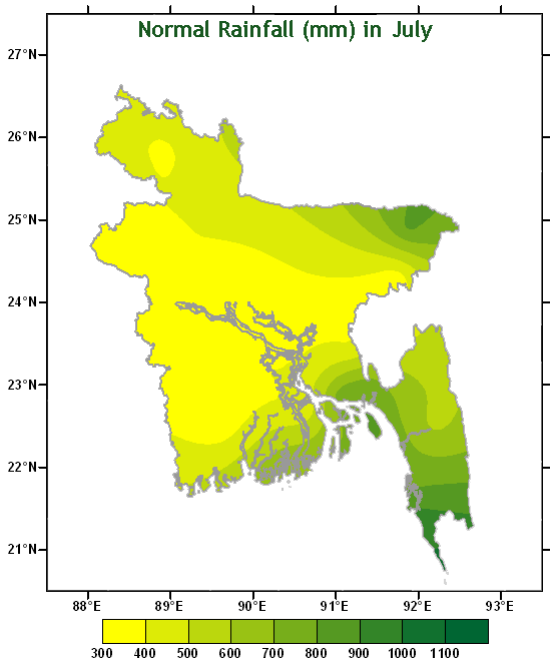


Fig. 3(iv).16(b): Spatial distribution of normal rainfall in July

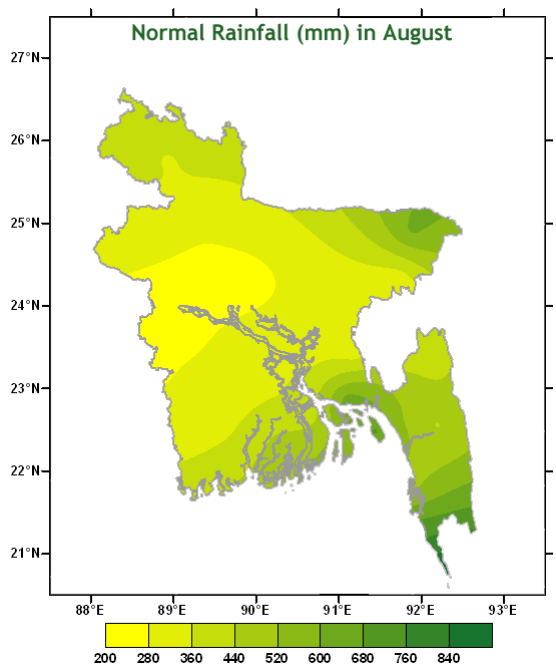


Fig. 3(iv).16(c): Spatial distribution of normal rainfall of Bangladesh in August

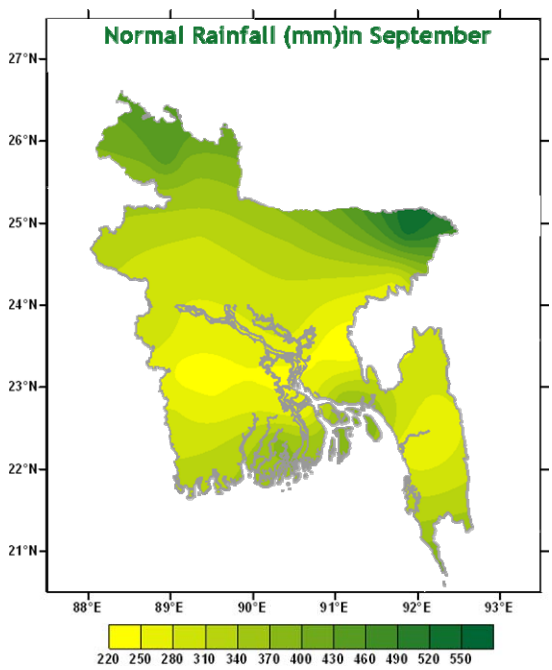


Fig. 3(iv).16(d): Spatial distribution of normal rainfall in September

3(iv).5.2. Spatial distribution of seasonal rainfall of monsoon season in Bangladesh during 1951-2010

Spatial distribution of seasonal monsoon rainfall reveals that the higher amounts of rainfall are recorded in the southeastern corner of Bangladesh, followed by northeastern parts due to seasonal flow pattern, distribution pattern of atmospheric variables and geophysical conditions, but the lower amounts of rainfall are recorded in the central and western parts of the country as revealed in Fig. 3(iv).17. The maximum amount of average rainfall of 3298.8 mm is recorded at Teknaf, followed by 2723.7 mm at Cox's Bazar, but the minimum amount of rainfall of 1117.0 mm is recorded at Chuadanga.

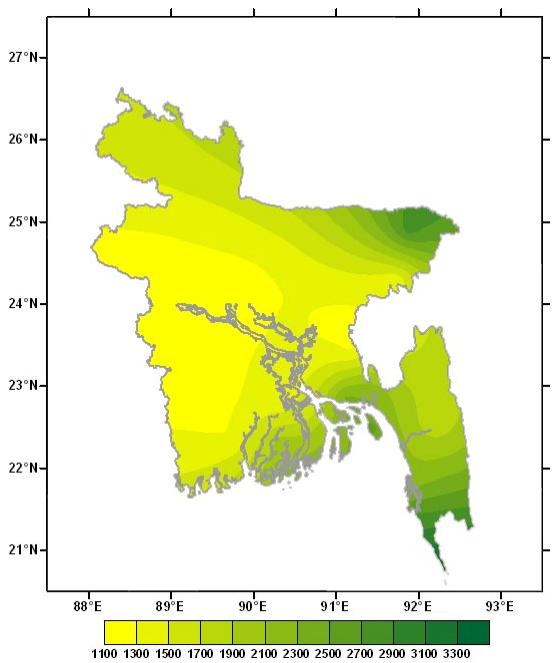


Fig. 3(iv).17: Spatial distribution of normal monsoon rainfall in Bangladesh

3(iv).6. Spatial distribution of monthly and seasonal rainy days in the monsoon season in Bangladesh during 1951-2010

3(iv).6.1. Spatial distribution of monthly rainy days in the monsoon season in Bangladesh during 1951-2010

Spatial distribution of the number of rainy days of June is given in Fig. 3(iv).18(a) which reveals that the higher numbers of rainy days are located in the north eastern part of the country followed by southeastern part, whereas lower number of rainy days are found in the central to western parts with the extreme lowest in the extreme northwestern part. The maximum 22 rainy days are observed at Sylhet, followed by 21 rainy days at Tenkaf, but the minimum 12 rainy days are observed at Dinajpur and Ishurdi.

In July, the number of rainy days are higher in the northeastern part of the country followed by southeastern part, whereas lower number of rainy days are found in the central and extreme northwestern parts, but the distribution pattern is little different from June as illustrated in Fig.3(iv).18(b). The maximum 25 rainy days are observed at Sylhet and Teknaf, followed by 24 rainy days at Bhola, but the minimum 14 rainy days are observed at Sayedpur.

In August, the number of rainy days are higher in the southeastern coastal districts of the country followed by northeastern part, whereas but lower number of rainy days are found in the north-central and extreme northwestern parts and the distribution pattern is deviated from June and July as shown in Fig.3(iv).18(c). The maximum 24 rainy days are observed at Teknaf, followed by 24 rainy days at Bhola, but the minimum 13 rainy days are observed at Sayedpur.

In September, the number of rainy days are higher in the northern districts of the country followed by southern coastal areas and southwestern part, whereas lower numbers of rainy days are found in the central and extreme northwestern parts and the distribution pattern is different from all other monsoon months as depicted in Fig. 3(iv).18(d). The maximum 18 rainy days are observed at Sylhet, followed by 17 at Bhola, Khepupara and Rangpur, but the minimum 12 rainy days are observed at Chandpur and comilla.

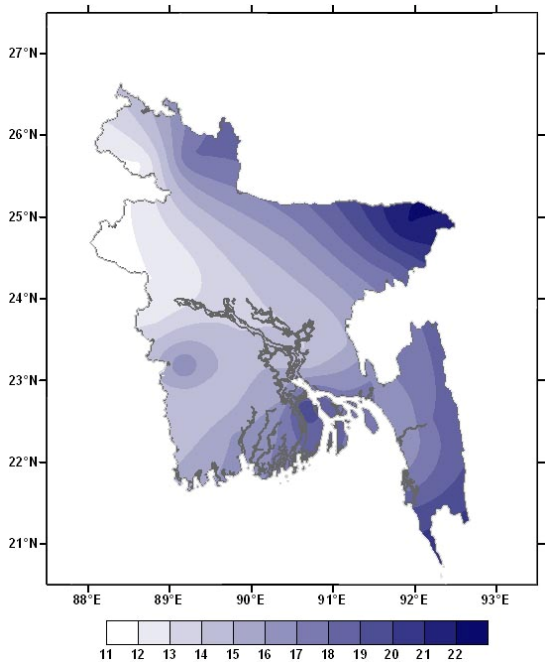


Fig.3(iv).18(a): Spatial distribution of normal rainy day in June

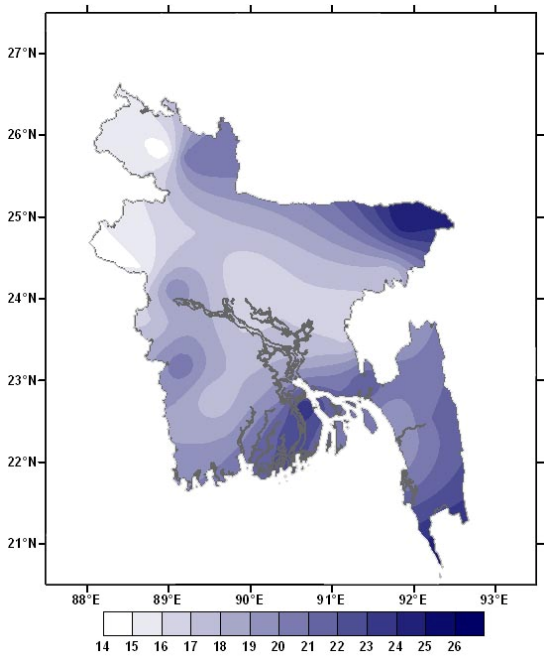


Fig. 3(iv).18(b): Spatial distribution of normal rainy day in July

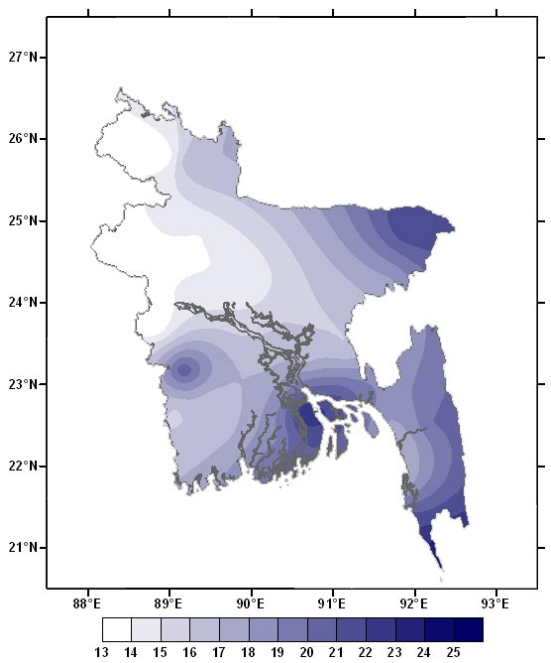


Fig. 3(iv).18(c): Spatial distribution of normal rainy day in August

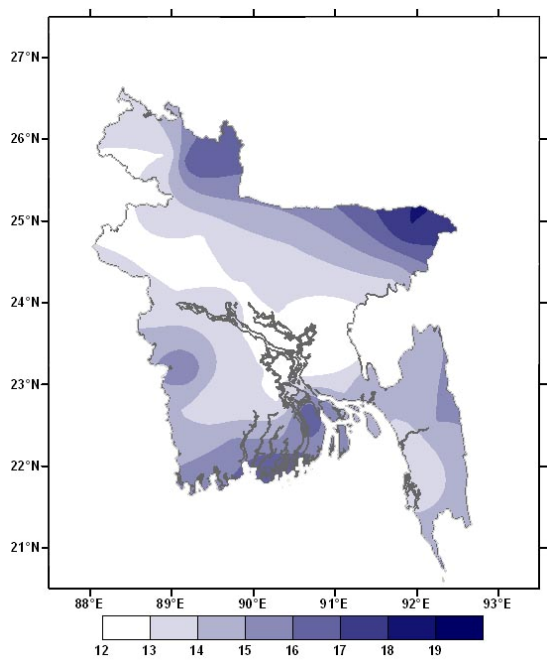


Fig. 3(iv).18(d): Spatial distribution of normal rainy day in September

3(iv).6.2. Spatial distribution of seasonal rainy days of monsoon in Bangladesh during 1951-2010

The number of rainy days are higher in the northeastern part of Bangladesh followed by southern coastal part, whereas lower number of rainy days are found in the central and extreme northwestern parts and the distribution pattern is different from monsoon months as depicted in Fig. 3(iv).19. The maximum 87 rainy days are observed at Sylhet, followed by 85 at Teknaf, but the minimum 54 rainy days are observed at Rajshahi.

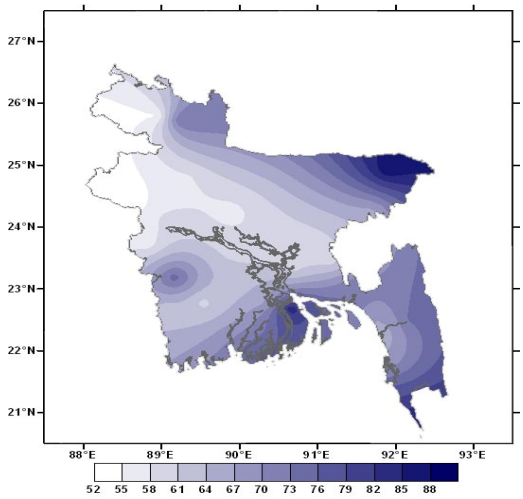


Fig. 3(iv).19: Spatial distribution of seasonal rainy day in Monsoon

3(iv).7. Spatial distribution of the decadal mean temperature anomalies of Bangladesh in monsoon season

3(iv).7.1. Spatial distribution of mean temperature anomalies during 1951-1960 in monsoon season

For better understanding the temperature climatology of Bangladesh during monsoon season spatial distribution of the anomalies of the recorded decadal mean temperature is analyzed and illustrated in Fig. 3(iv).20-Fig. 3(iv).25. During the decade 1951-1960, southern parts of Bangladesh and Rangpur region were under positive anomaly, but the other parts of the country were under negative anomaly as given in Fig. 3(iv).20. The peak positive anomaly of 0.5°C was observed at Barisal, but the lowest negative anomaly of -0.2°C was found at Faridpur.

3(iv).7.2. Spatial distribution of mean temperature anomalies during 1961-1970 in monsoon season

In the decade 1961-1970, positive anomalies were found in northwestern part and Barisal region and negative anomalies were observed in other parts of the country as shown in Fig. 3(iv).21. The peak positive anomaly of 0.3°C was observed at Khulna, but the lowest negative anomaly of -0.2°C was found at Cox's Bazar.

3(iv).7.3. Spatial distribution of mean temperature anomalies during 1971-1980 in monsoon season

In the decade 1971-1980, positive anomalies were found in northwestern part and northern side of Bangladesh, but negative anomalies were experienced in other parts of the country as shown in Fig. 3(iv).22. The peak positive anomaly of 0.3°C was observed at Dinajpur, but the lowest negative anomaly of -0.1°C was found at Faridpur.

3(iv).7.4. Spatial distribution of mean temperature anomalies during 1981-1990 in monsoon season

In the decade 1981-1990, negative anomalies were found in northwestern part, but the positive anomalies were experienced in other parts of the country with the significant positive anomalies in the central and southeastern parts as shown in Fig. 3(iv).23. The peak positive anomaly of 0.3°C was observed at Cox's Bazar, but the lowest negative anomaly of -0.4°C was found at Dinajpur.

3(iv).7.5. Spatial distribution of mean temperature anomalies during 1991-2000 in monsoon season

In the decade 1991-2000, negative anomalies were seen over Comilla, Dinajpur, Khulna, and Mymensingh regions, but positive anomalies were observed in other parts of the country with the significant positive anomalies in the southeastern part and some regions of western part as depicted in Fig.3(iv).24. The peak positive anomaly of 0.8°C was observed at Cox's Bazar, but the lowest negative anomaly of -0.1°C was found at Dinajpur.

3(iv).7.6. Spatial distribution of mean temperature anomalies during 2001-2010 in monsoon season

In the decade 2001-2010, mean temperature anomalies were uniquely positive over the country with the significant positive anomalies in the southern, central and west-central parts of the country as shown in Fig. 3(iv).25. The peak

positive anomaly of 1.2°C was observed at Cox's Bazarm but the lowest negative anomaly of -0.1°C was found at Dinajpur.

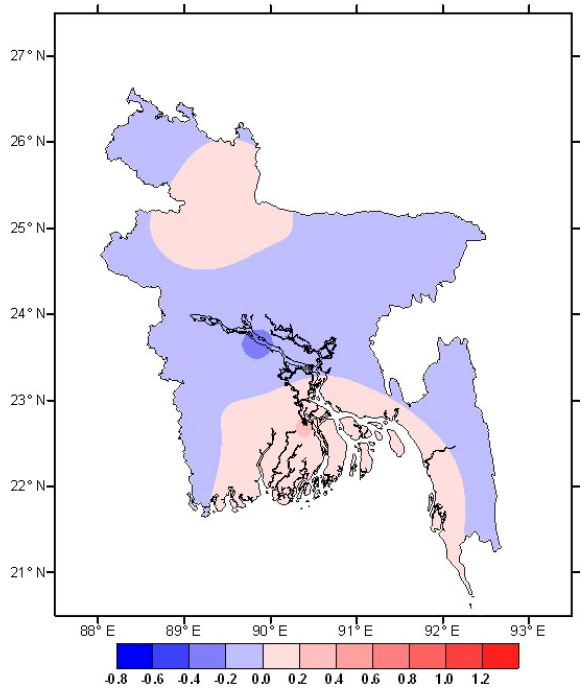


Fig. 3(iv).20: Distribution of mean temperature anomaliy during 1951-1960

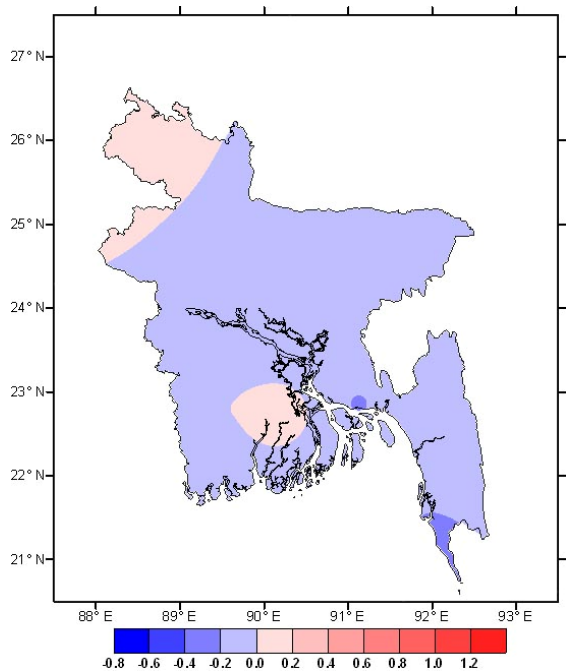


Fig. 3(iv).21: Distribution of mean temperature anomaly during 1961-1970

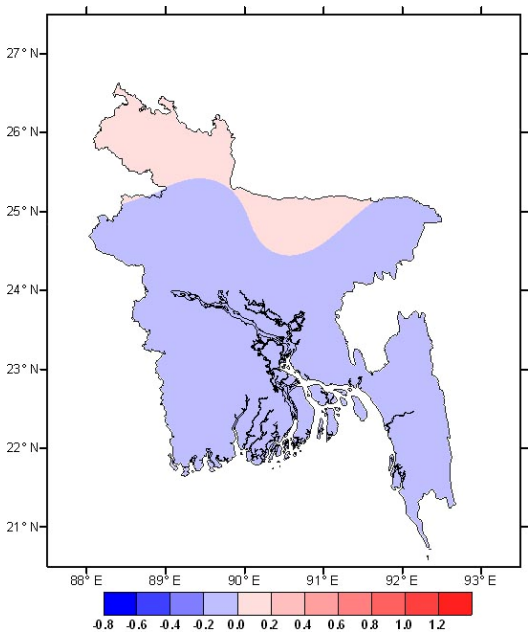


Fig. 3(iv).22: Distribution of mean temperature anomaly during 1971-1980

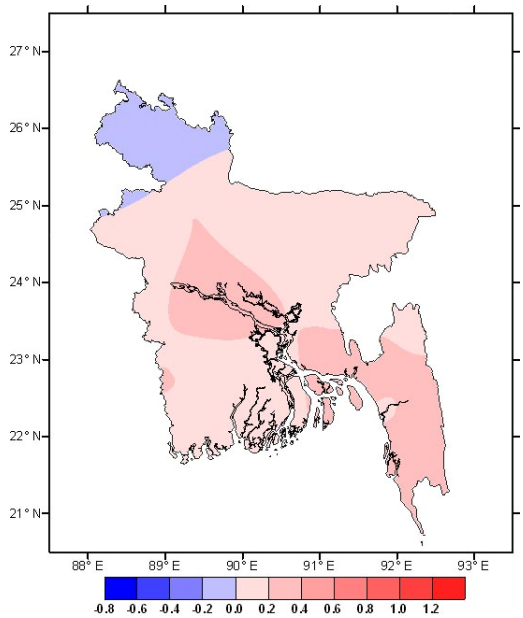


Fig.3(iv).23: Distribution of mean temperature anomaly during 1981-1990

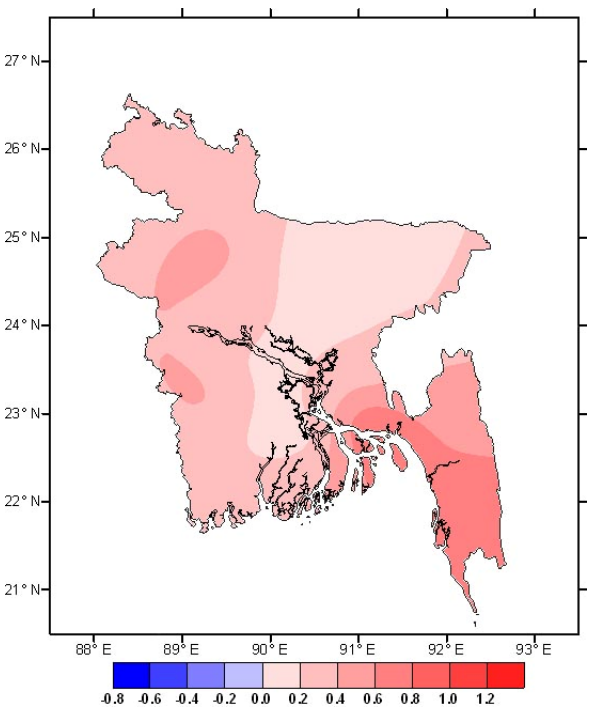


Fig. 3(iv).24: Spatial distribution of mean temperature anomaly during 1991-2000

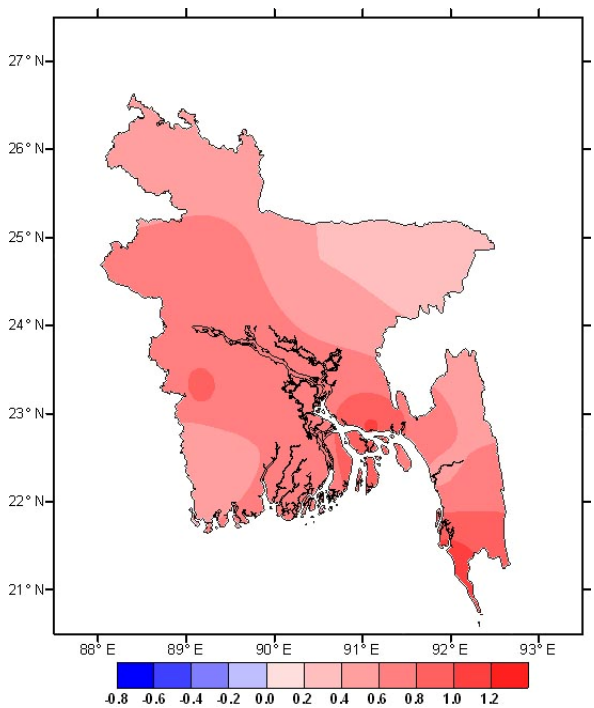


Fig. 3(iv).25: Spatial distribution of mean temperature anomaly during 2001-2010

3(iv).8. Heavy Rainfall situation of Bangladesh during monsoon season

Bangladesh is one of the most flood prone countries. The floods in Bangladesh mostly occur during monsoon season due to heavy rainfall. But most of the heavy rainfall events occur in monsoon months and the northeastern, eastern and southeastern regions of the country are most susceptible for this meteorological phenomenon. Heavy rainfall often disrupts our daily activities and hampers our works and sometimes it is very hazardous. The main socio-economic sectors affected by heavy rainfall are- agriculture, food security, urban/town planning and construction, energy, water resource management, fisheries, forestry, human health and social services, disaster management, transportation (air, land, water), tourism, sports and leisure etc. In BMD, intensity of rainfall has been categorized as in given Table 3(iv).3.

Table 3(iv).3: Classification of rainfall category in Bangladesh

Category of rainfall	Intensity (mm)
Trace	≤ 3.0 mm
Light rain	= 4.57- 9.64 mm
Moderate rain	= 9.65- 22.34 mm
Moderately heavy rain	= 22.35- 44.19 mm
Heavy rain	= 44.20- 88.90 mm
Very heavy rain	≥ 89 mm

Therefore, heavy rainfall has been considered in this study only when the total rainfall during 24 hours is higher than 22 mm. For better understanding, intensity of different categorized rainfall- moderately heavy, heavy and very heavy rainfall has been analyzed separately. Total heavy rainfall has therefore been considered and analyzed by the total amount of these three categories.

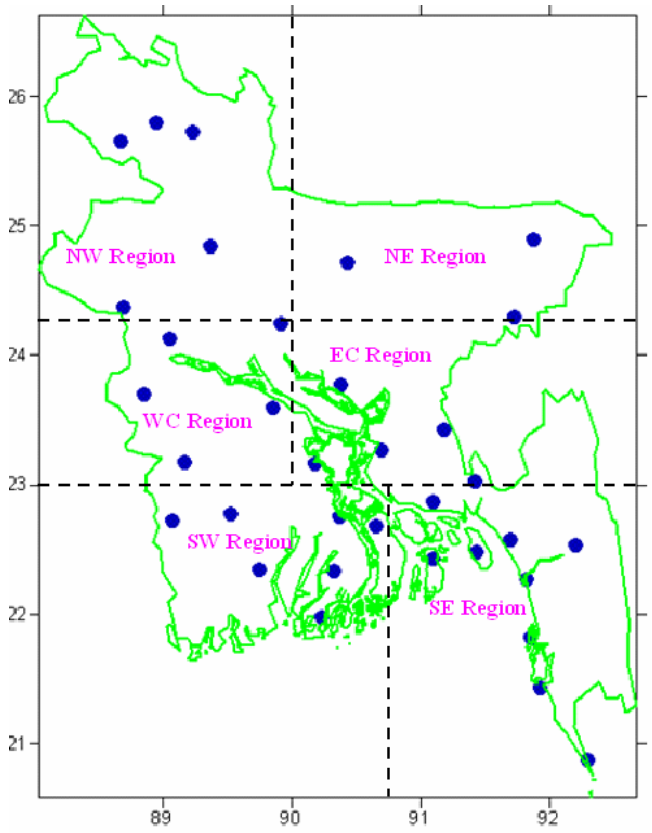


Fig. 3(iv).26: Location of different regions considered under this study

As heavy rainfall has the area specific variation it is essential to analyze it by separating the country in different parts. To do this job, the whole country has been divided into six regions as in Fig.3(iv).26. The regions and the included stations of respective regions are-

a. Northeast region (NE region): Mymensingh, Sylhet and Srimongal.
b. Northwest region (NW region): Rajshahi, Rangpur, Dinajpur, Syedpur and Bogra.
c. East-central region (EC region): Dhaka, comilla, Chandpur, Feni and Madaripur.
d. West central region (WC region): Tangail Faridpur, Ishurdi, Jessore and Chuadanga.
e. Southeast region (SE region): Chittagong, Sandwip, Sitakunda, Rangamati, Maijdicourt, Hatiya, Cox's Bazar, Kutubdia and Teknaf.
f. Southwest region (SW region): Khulna, Mongla, Satkhira, Barisal, Patuakhali and Bhola

The average values of the located stations of each region are considered as the regional representative value.

3(iv).8.1. Analysis of heavy rainfall frequency of June

The frequency distribution of moderately heavy, heavy, very heavy and total heavy rainfall of June and their respective trend are given in Table 3(iv).4. The higher total heavy rainfall frequencies are observed during 1951-1966 and 1977 to onward with inter-annual variation. The highest annual average frequency is 10.46 in 1954, followed by 9.38 in 2001 and the lowest frequency is 2.73 in 1975, but the country-averaged frequency of it is 6.17. The linear trend of it is +0.0092/year during the observed period as shown in Fig. 3(iv).27(a) Regional analysis reveals that the highest total heavy rainfall frequency is 8.37 in the southeast region, followed by 7.42 in the northeast region and the lowest frequency is 4.24 in the west-central region. Positive trends of it are found in the southeast and southwest regions and negative trends are in all other regions and the details are given in Table 3(iv).4.

Table 3(iv).4: Frequency distribution and trend values of categorized rainfall in June

Regions	Average frequency categorized rainfall				Trend of categorized rainfall			
	Moderately Heavy	Heavy	Very Heavy	Total heavy	Moderately heavy	Heavy	Very Heavy	Total Heavy
Country	3.30	2.20	0.98	6.17	-0.0008	+0.0044	+0.0014	+0.0092
NE region	3.92	2.44	1.14	7.42	-0.012	-0.012	-0.007	-0.014
NW region	3.03	1.69	0.63	5.02	+0.006	+0.001	-0.004	+0.015
EC region	3.10	2.03	0.75	5.68	-0.002	+0.000	-0.009	+0.013
WC region	2.66	1.27	0.35	4.24	+0.030	+0.001	-0.005	+0.026
SE region	3.78	3.18	1.85	8.37	+0.017	+0.018	+0.012	+0.056
SW region	3.22	1.97	0.72	5.76	+0.002	+0.014	+0.003	+0.039

3(iv).8.2. Analysis of heavy rainfall frequency of July

The frequency distribution of moderately heavy, heavy, very heavy and total heavy rainfall of July and their respective trend are given in Table 3(iv).5. Frequency of total heavy rainfall varies from year to year and has a sharp increasing trend. The highest frequency is 11.58 in 1987 followed by 9.70 in 1997 and the lowest frequency is 3.87 in 1972, but the country-averaged frequency of it is 6.57 during the observed period. The linear trend of it is positive with the rate of + 0.022/year during the observed period as shown in Fig. 3(iv).27(b). Regional analysis shows that the highest frequency of total heavy rainfall is 9.76 in the southeast region, followed by 6.70 in the northeast region and the lowest frequency is 4.37 in the west-central region. Frequency of total heavy rainfall shows negative trend in the east-central and west-central regions and positive trend in all other regions and the details are given in Table 3(iv).5.

Table 3(iv).5: Frequency distribution and trend values of categorized rainfall in July

Regions	Average frequency categorized rainfall				Trend of categorized rainfall			
	Moderately Heavy	Heavy	Very Heavy	Total heavy	Moderately heavy	Heavy	Very Heavy	Total Heavy
Country	3.54	2.31	1.02	6.57	+0.004	+0.0084	+0.0009	+0.022
NE region	3.86	2.12	0.87	6.70	-0.011	+0.002	+0.006	+0.004
NW region	2.91	1.97	0.63	5.26	-0.001	+0.023	-0.006	+0.063
EC region	3.49	2.21	0.77	6.17	-0.002	-0.018	-0.010	-0.007
WC region	2.74	1.27	0.44	4.37	-0.002	-0.003	-0.006	-0.013
SE region	4.13	3.63	2.15	9.76	-0.002	+0.014	+0.007	+0.039
SW region	3.95	2.00	0.52	6.19	+0.014	-0.008	-0.005	+0.008

3(iv).8.3. Heavy rainfall analysis of August

The frequency distribution of moderately heavy, heavy, very heavy and total heavy rainfall of August and their respective trend is given in Table 3(iv).6. The frequency of total heavy rainfall in August is lower compared to June and July with high inter-annual variability. The highest frequency is 8.48 observed in 1998, followed by 8.38 in 1969 and 7.89 in 1965 and the lowest frequency is 1.97 in 1989, but the country-averaged frequency is 5.54. It has positive linear trend with the rate of +0.005/year during the observed period as shown in Fig. 3(iv).27(c). The highest

frequency of total heavy rainfall is 8.31 in the southeast region, followed by 5.81 in the northeast region and 5.06 in southwest region and the lowest frequency is 3.34 in the west-central region. Regional trend analysis shows negative trend in west-central, east-central and northeast regions and positive trend in all other regions and the details are given in Table 3(iv).6.

Table 3(iv).6: Frequency distribution and trend values of categorized rainfall in August

Regions	Average frequency categorized rainfall				Trend of categorized rainfall			
	Moderately Heavy	Heavy	Very Heavy	Total heavy	Moderately heavy	Heavy	Very Heavy	Total Heavy
Country	3.08	1.82	0.68	5.54	+0.0019	+0.0041	-0.0011	+0.005
NE region	3.56	1.70	0.63	5.81	-0.013	+0.001	-0.006	-0.019
NW region	2.47	1.35	0.50	4.31	+0.007	+0.002	-0.002	+0.007
EC region	2.82	1.52	0.59	4.94	-0.002	+0.009	-0.018	-0.002
WC region	2.17	0.95	0.23	3.34	+0.009	-0.006	-0.004	-0.001
SE region	4.02	2.97	1.31	8.31	+0.009	+0.003	+0.007	+0.006
SW region	3.06	1.58	0.41	5.06	+0.016	+0.001	-0.011	+0.006

3(iv).8.4. Heavy rainfall analysis of September

The frequency distribution of moderately heavy, heavy, very heavy and total heavy rainfall of September and their respective trend is given in Table 3(iv).7. The frequency of total heavy rainfall of September has sharp increasing trend with inter-annual variability. The highest frequency of it is 6.78 in 2004, followed by 6.65 in 1991 and 6.42 in 1967 and the lowest frequency is 2.17 in 1972, but the country-averaged frequency of it is 4.29. The linear trend of + 0.0161/year is found during

the analyzed period and the details are shown in Fig. 3(iv).27(d). From the regional analysis, it is found that the highest frequency of it is 4.66 in the northeast region followed by 3.48 in the southeast region and the lowest frequency is 2.55 in the west-central region. It has negative trend in the southwest region and positive trend in all other regions and the details are given in Table 3(iv).7.

Table 3(iv).7: Frequency distribution and trend values of categorized rainfall in September

Regions	Average frequency categorized rainfall				Trend of categorized rainfall			
	Moderately Heavy	Heavy	Very Heavy	Total heavy	Moderately heavy	Heavy	Very Heavy	Total Heavy
Country	2.48	1.34	0.47	4.29	+0.0054	+0.0059	+0.0047	+0.0161
NE region	2.71	1.49	0.46	4.66	+0.0138	+0.0086	-0.0030	+0.0196
NW region	1.73	0.98	0.37	3.07	+0.0169	+0.0170	+0.0080	+0.0420
EC region	1.69	0.91	0.27	2.87	+0.0078	+0.0038	+0.0073	+0.0191
WC region	1.56	0.77	0.21	2.55	+0.0060	+0.0185	+0.0051	+0.0105
SE region	1.91	1.12	0.46	3.48	+0.0086	+0.0116	+0.0041	+0.0286
SW region	1.72	0.87	0.34	2.93	-0.0139	+0.0020	+0.0072	-0.0044

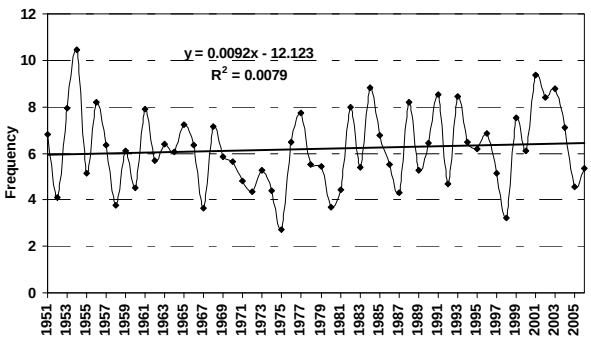


Fig. 3(iv).27(a): Annual variation of total heavy rainfall frequency of Bangladesh in June

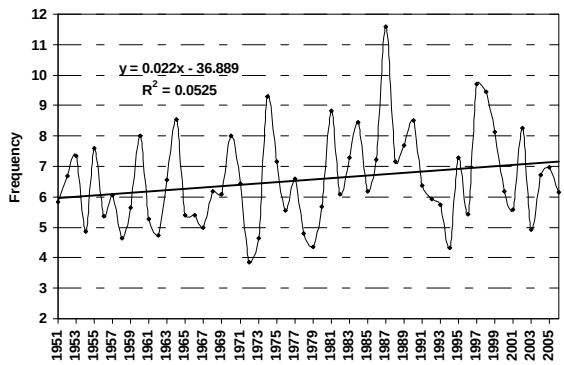


Fig.3(iv).27(b): Annual variation of total heavy rainfall frequency of Bangladesh in July

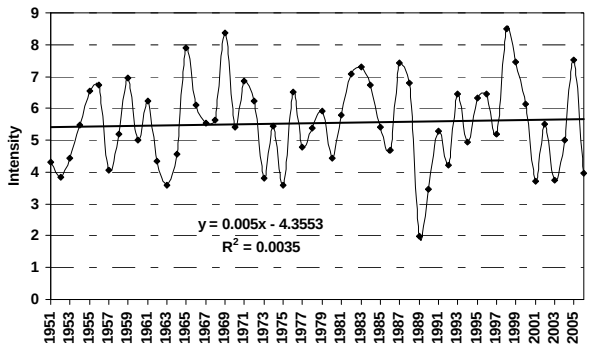


Fig. 3(iv).27(c): Annual variation of total heavy rainfall frequency of Bangladesh in August

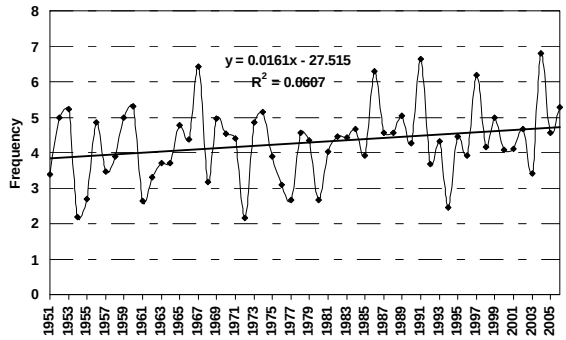


Fig. 3(iv).27(d): Annual variation of total heavy rainfall frequency of Bangladesh in September

3(iv).8.5. Spatial distribution of heavy rainfall frequency of June in Bangladesh

Analysis shows that the higher frequencies of total heavy rainfall are found in the southeastern part followed by northeastern part of the country and the lower frequencies are in west-central region. The highest frequency of it is 12.81 observed at Teknaf, followed by 10.57 at Sylhet and 10.29 at Cox's Bazar and the lowest frequency is 3.33 at Chuadanga as illustrated in Fig. 3(iv).28(a).

3(iv).8.6. Spatial distribution of heavy rainfall frequency of July in Bangladesh

Spatial distribution of total heavy rainfall frequencies in July is shown in Fig. 3(iv).28(b). The total heavy rainfall frequencies are mainly concentrated in the eastern half of the country, but the higher frequencies are observed in the southeastern part, followed by northeastern parts of the country. But the lower frequencies are mainly concentrated in the central and west central parts of the country. The highest frequency of it is 12.17 observed at Teknaf, followed by 11.57 at Cox's Bazar, 10.69 at Sandwip and 10.14 at Sylhet and the lowest frequency is 4.00 at Ishurdi.

3(iv).8.7. Spatial distribution of heavy rainfall frequency of August in Bangladesh

Spatial distribution total heavy rainfall frequencies in August is depicted in Fig. 3(iv).28(c). Distribution pattern of total heavy rainfall shows that the southeast and northeast regions of the country are very much susceptible for heavy rainfall. The highest frequency of it is 12.63 at Teknaf, followed by 9.57 at Sandwip, 9.52 at Cox's Bazar and the lowest frequency is 2.98 at Ishurdi.

3(iv).8.8. Spatial distribution of heavy rainfall frequency of September in Bangladesh

Spatial distributions of total heavy rainfall frequencies in September are shown in Fig. 3(iv).28(d). The frequency distribution pattern of total heavy rainfall shows higher frequencies in the southern parts, followed by northeastern tip. The highest frequency of it is 7.08 at Sylhet, followed by 6.24 at Hatiya and the lowest frequency is 3.04 at Ishurdi.

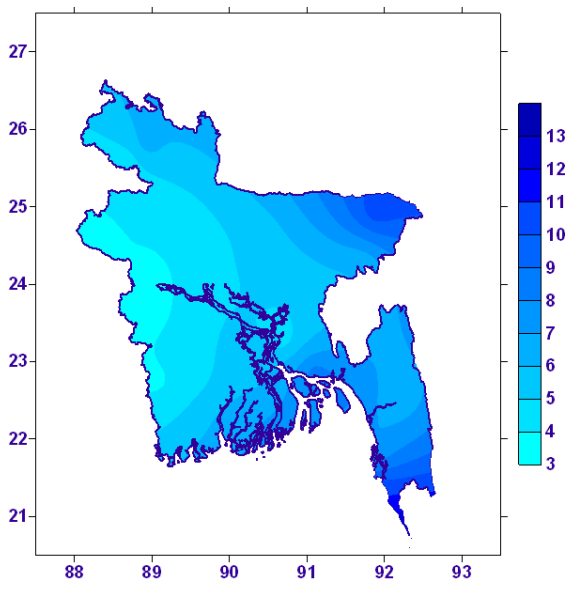


Fig.3(iv).28(a): Spatial Distribution of total heavy rainfall frequency of Bangladesh in June

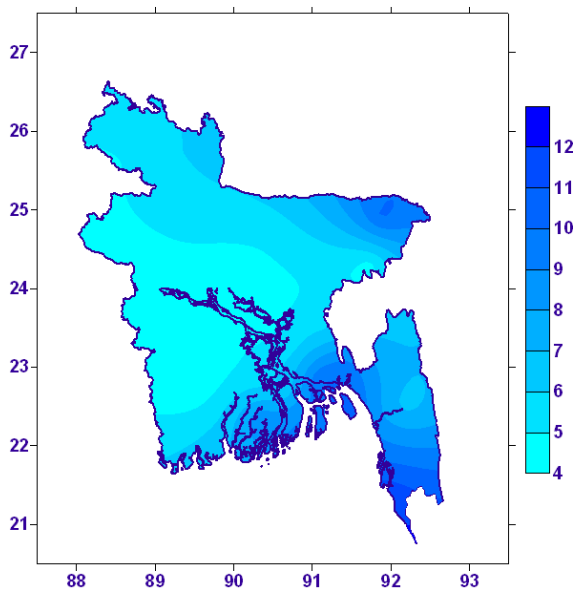


Fig. 3(iv).28(b): Spatial Distribution of total heavy rainfall frequency of Bangladesh in July

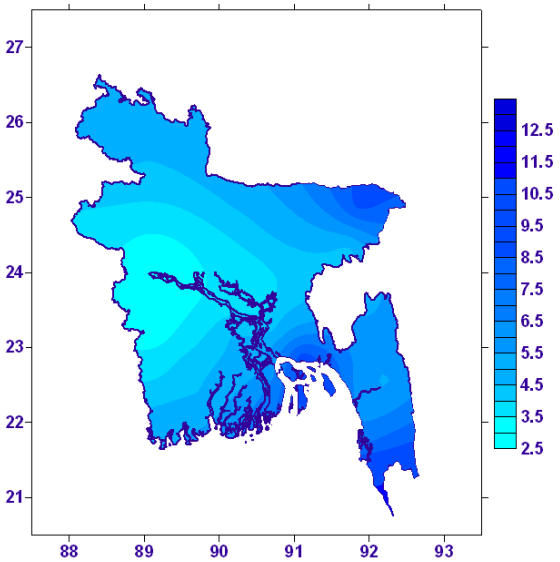
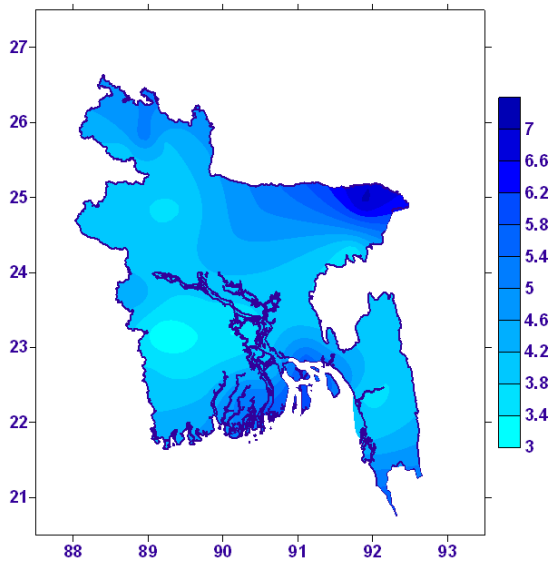


Fig. 3(iv).28(c): Spatial Distribution of total heavy rainfall frequency of Bangladesh in August



g. Fig.3(iv).28(d): Spatial Distribution of total heavy rainfall frequency of Bangladesh in September

3(iv).9. Forecasting System

a. Day to day forecasting

i. Conventional method of forecasting which depends on persistence and climatology, utilizing synoptic pattern and time series analysis of climatological data and information are utilized for issuance of forecast and warnings.

ii. Satellite & Radar image, Rawinsonde soundings and model forecast of ECMWF, JMA, NCMRWF and US Navy are now quite helpful for now-casting, warnings for extreme events, also helpful for short range (1-2 day) forecast very effectively.

iii. But the whole process is subjective having variations regarding the accuracy of forecast from person to person.

iv. To avoid this situation and also to acquire capability of deterministic forecast (1-5 days), especially to cater to the agriculture sector, BMD is implementing one project for the introduction of NWP technique in the forecasting system.

b. Seasonal Prediction in Bangladesh (Monthly)

An Expert level committee comprising experts from BMD and other related organizations is responsible for issuing forecast for the duration of a month / three months. The committee sits in the first three days at Bangladesh Meteorological Department (BMD). For issuing forecasts the following points are taken into consideration:

i. Previous and present surface Pressure, wind and Temperature distribution over Bangladesh, South Asian countries, Bay of Bengal and adjoining Bay areas.

ii. Previous and present Upper air temperature and wind distribution over Bangladesh, South Asian countries, Bay of Bengal and adjoining Bay areas.

iii. Upper air flow pattern over Bangladesh, South Asian countries, Bay of Bengal and adjoining Bay areas.

iv. Normal Surface Pressure, Wind and Temperature distribution over Bangladesh, South Asian countries, Bay of Bengal and adjoining Bay areas.

v. Normal Upper air temperature and wind distribution over Bangladesh, South Asian countries, Bay of Bengal and adjoining Bay areas.

vi. Normal Upper air flow pattern over Bangladesh, South Asian countries, Bay of Bengal and adjoining Bay areas.

vii. Analytical result from Climate Regression and Analog Model based on the hindcast data.

viii. Available Radar and Satellite Information

ix. ECMWF model prediction for the month

x. JMA model prediction for the month

xi. Previous and present El-Nino/La Nina condition and their relationship with the rainfall of Bangladesh

xii. APCC multi-model Ensemble prediction about Bangladesh and its surroundings

3(iv).10. Conclusions

i. The inter-annual variability of maximum temperature of monsoon season is low compared to other seasons in Bangladesh and its trend is positive (+0.0199°C/year) during the observed period (1950-2010).

ii. The inter-annual variability of minimum temperature of monsoon season is also low in Bangladesh and its trend is also positive (+0.0081°C/year).

iii. The mean temperature of Bangladesh varied between 25.2°C to 26.6°C during the observed period having positive trend of +0.0099°C/year.

iv. The tendencies of amount of rainfall in June, July, August and September are - 0.701 mm/year, +0.006 mm/year, -0.735 mm/year, +0.370 mm/year and -1.06 mm/year respectively during the observed period

v. Country-averaged frequency of total heavy rainfall of June is 6.17 and its linear trend is +0.0092/year during the observed period.

vi. In June, the higher and the secondary higher frequency of moderately heavy rainfall are found in the northeastern and southeastern parts of the country but in the case of heavy, very heavy and total heavy rainfall the higher and the secondary higher frequency are found in the southeastern and northeastern parts respectively. The lower frequencies of all categorized rainfall are mainly concentrated in the central and western parts of the country.

vii. Country-averaged frequency of total heavy rainfall of July is 6.57 and its linear trend is +0.022/year.

viii. The higher and the secondary higher frequency of moderately heavy rainfall in July are found in the northeastern and southeastern parts of the country but opposite situations are observed in the case of heavy, very heavy and total heavy rainfall. The lower frequencies of all categorized rainfall of this month are located in the central and western parts of the country.

ix. Country averaged frequency of total heavy rainfall of August is 5.54 and its linear trend is +0.005/year.

x. The higher and the secondary higher frequency of moderately heavy, heavy, very heavy and total heavy rainfall of August are found in the southeastern and northeastern parts of the country respectively but the lower frequencies of

all categorized rainfall are mainly concentrated in the central and western parts of the country.

xi. Country averaged frequency total heavy rainfall of September is 4.29 and its linear trend is + 0.0161/year respectively.

xii. In September, moderately heavy rainfall shows negative trend in the southwest region and positive trends in all other regions; heavy rainfall shows positive trend in all regions of the country; very heavy rainfall shows negative trend in northeast region and positive trend in all other regions; total heavy rainfall shows negative trend in southwest region and positive in all other regions.

xiii. The higher and the secondary higher frequency of moderately heavy, heavy, very heavy and total heavy rainfall of September are limited to northeastern and southeastern parts of the country.

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CHAPTER 3 (v)

MONSOON CLIMATE OF MYANMAR

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3(v).1. Introduction

Burma, is the largest country in mainland Southeast Asia. It lies between latitudes 9°N and 29°N, and longitudes 92°E and 102°E. Much of the country lies between the Tropic of Cancer and the Equator. It's positioned in southeast Asia and bordered by the Bay of Bengal, Andaman Sea, Gulf of Thailand and the countries of Bangladesh, India, China, Laos and Thailand.

3(v).2. Annual Mean Rainfall Distribution

It lies in the monsoon region of Asia, with its coastal regions receiving over 5,000 mm (196.9 in) of rain annually. Annual rainfall in the delta region is approximately 2,500 mm (98.4 in), while average annual rainfall in the Dry Zone, which is located in central Burma, is less than 1,000 mm (39.4 in). The normal annual rainfall over major stations of different parts of Myanmar (Fig. 3(v).1) is given in Table 3(v).1.

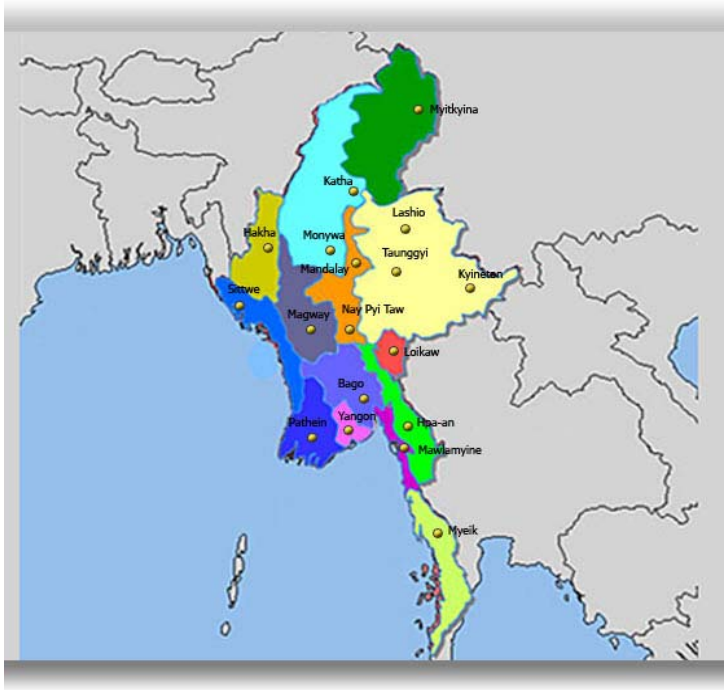


Fig. 3(v).1: Location of major cities of Myanmar

Sr. No.	Region / State	City	Annual Normal Rainfall (inch)
1	Kachin	Myitkyina	86.46
2	Eastern Shan	Kengtung	51.02
3	Southern Shan	Taunggyi	64.45
4	Northern Shan	Lashio	53.70
5	Chin	Hakha	70.55
6	Upper Sagaing	Homalin	92.80
7	Lower Sagaing	Monywa	31.38
8	Nay Pyi Taw	Nay Pyi Taw	47.68
9	Mandalay	Mandalay	32.80
10	Magway	Magway	31.85
11	Rakhine	Sittwe	183.98
12	Bago	Bago	131.22
13	Yangon	Yangon	106.30
14	Ayeyarwady	Pathein	103.66
15	Kayah	Loikaw	41.26
16	Kayin	Hpa-an	170.51
17	Mon	Mawlamyine	189.84
18	Taninthayi	Dawei	215.28

Table 3(v).1: Annual rainfall over Major cities of Myanmar

3(v).3. Monsoon onset and withdrawal over Myanmar

The onset of monsoon over northern Myanmar is gradually shifting one to two weeks later as seen from Fig. 3(v).2a. It is seen from Fig. 3(v).2a that after 1977, monsoon onset dates are later than normal date. Significantly delayed by about one to two weeks later than normal in 1992, 1997, 2003 and 2004.

However, the Fig. 3(v).2b shows that after 1976, monsoon withdrawal dates over southern Myanmar are earlier than normal date. Significantly earlier date of withdrawal of about 7 to 19 days is seen during 1977 to 2006.

Thus, because of delayed onset of monsoon over northern Myanmar and earlier withdrawal over southern Myanmar the duration of monsoon is becoming shorter after that of 1977 (Fig. 3(v).3). Significantly lowest shorter (of about 100 days) durations of monsoon are reported during 1979, 1998 and 2004. The shortening of the monsoon duration strongly impacts rainy paddy productions in Myanmar.

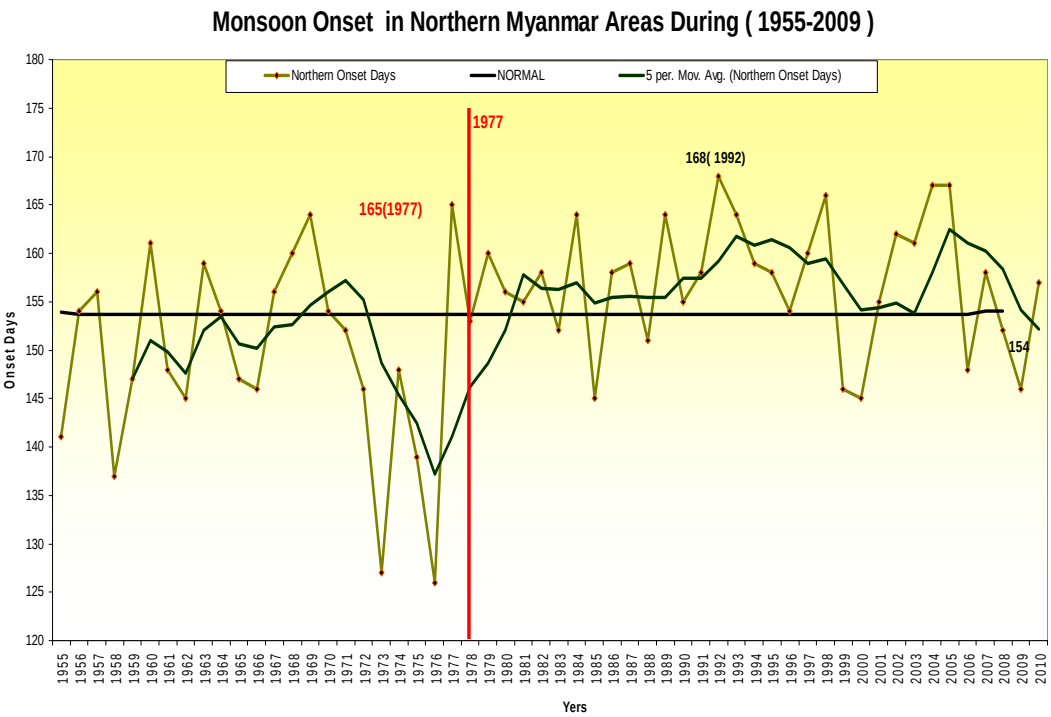


Fig. 3(v).2a: Monsoon onset dates over northern Myanmar

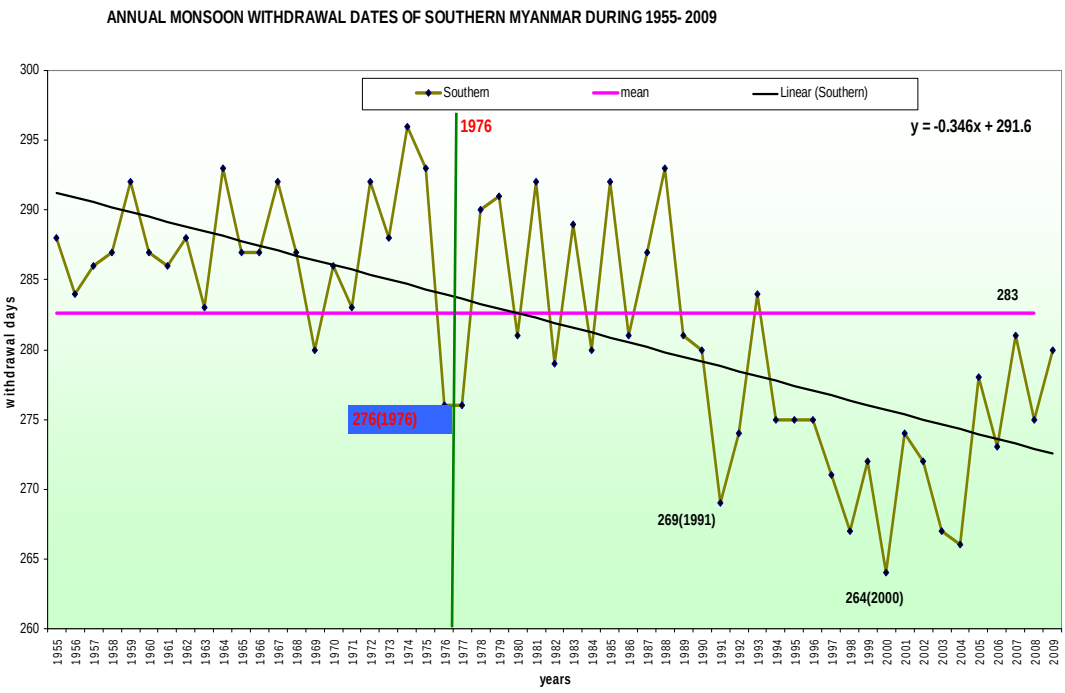


Fig. 3(v).2b: Monsoon withdrawal dates over southern Myanmar

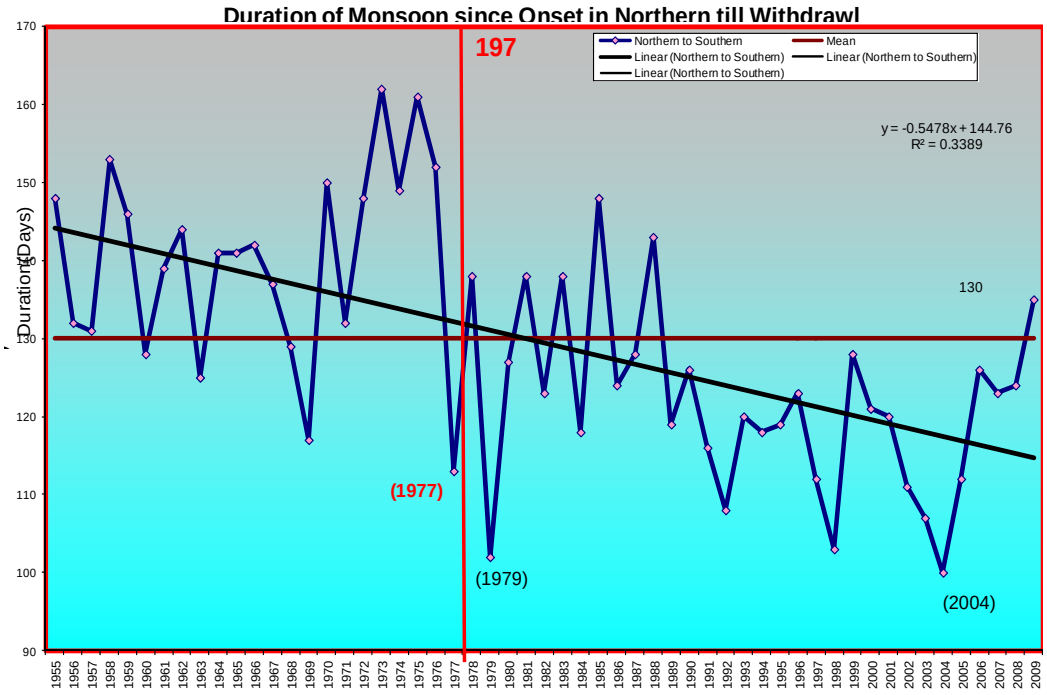


Fig. 3(v).3: Duration of monsoon since onset over northern Myanmar and withdrawal from Southern Myanmar.

3(v).4. Annual monsoon strength indices of Myanmar

Monsoon strength over Myanmar fluctuates gradually down trend from 1951 to 2009 by their linear trend(green color line) as seen in Fig. 3(v).4. Significantly near 25 years cycle (1953,1979, 2003) is observed in this down trend. Weak monsoon strength results in scanty rainfall over Myanmar.

The downtrend of rainfall shown in Fig. 3(v).5 is associated with downtrend of cyclone frequencies over Bay of Bengal as shown in Fig. 3(v).5. As the cyclonic disturbances over Bay of Bengal gives lot of rain over Myanmar, its decreasing frequency contributes to decreasing amount of rainfall over Myanmar.

The rainfall departure over two representative stations of Myanmar viz., (Taunggyi and Yangon) is shown in Fig. 3(v).6a and Fig.3(v).6b respectively.

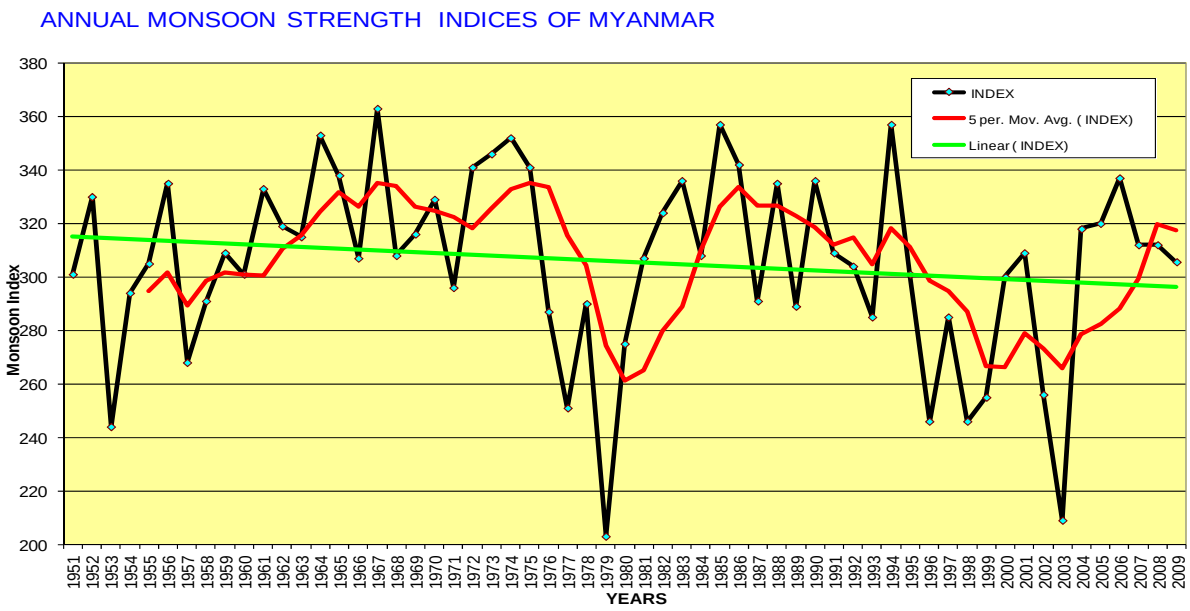


Fig. 3(v).4 : Annual monsoon strength indices of Myanmar

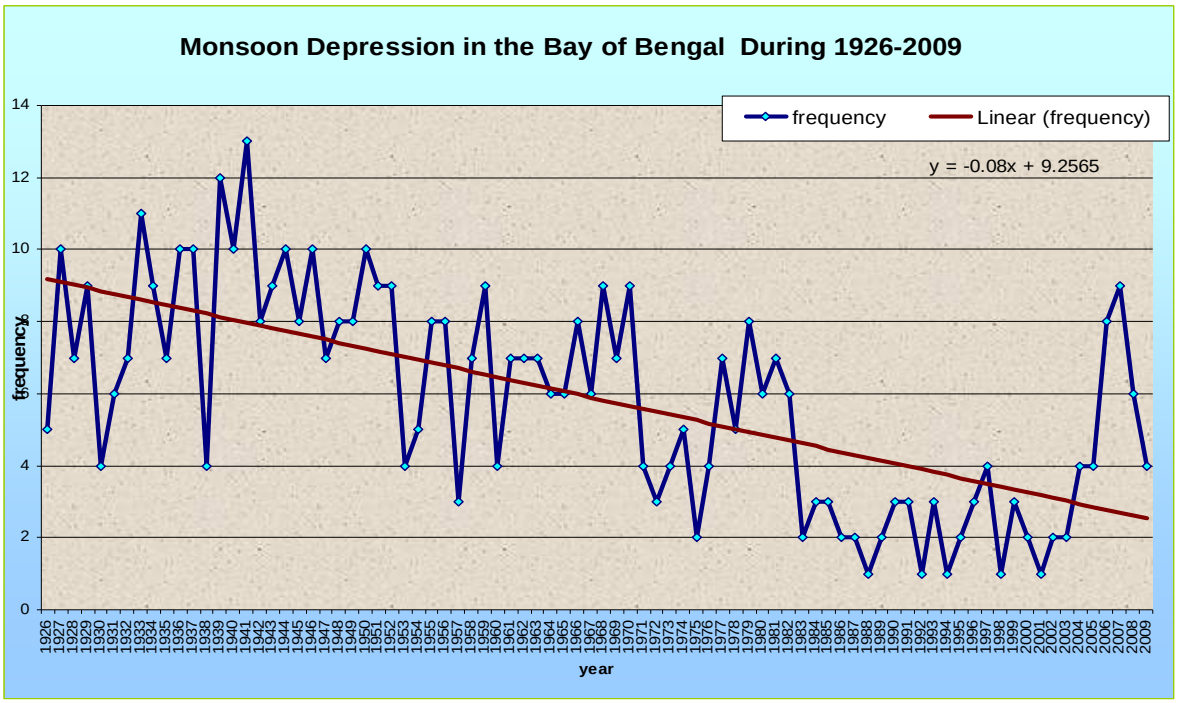
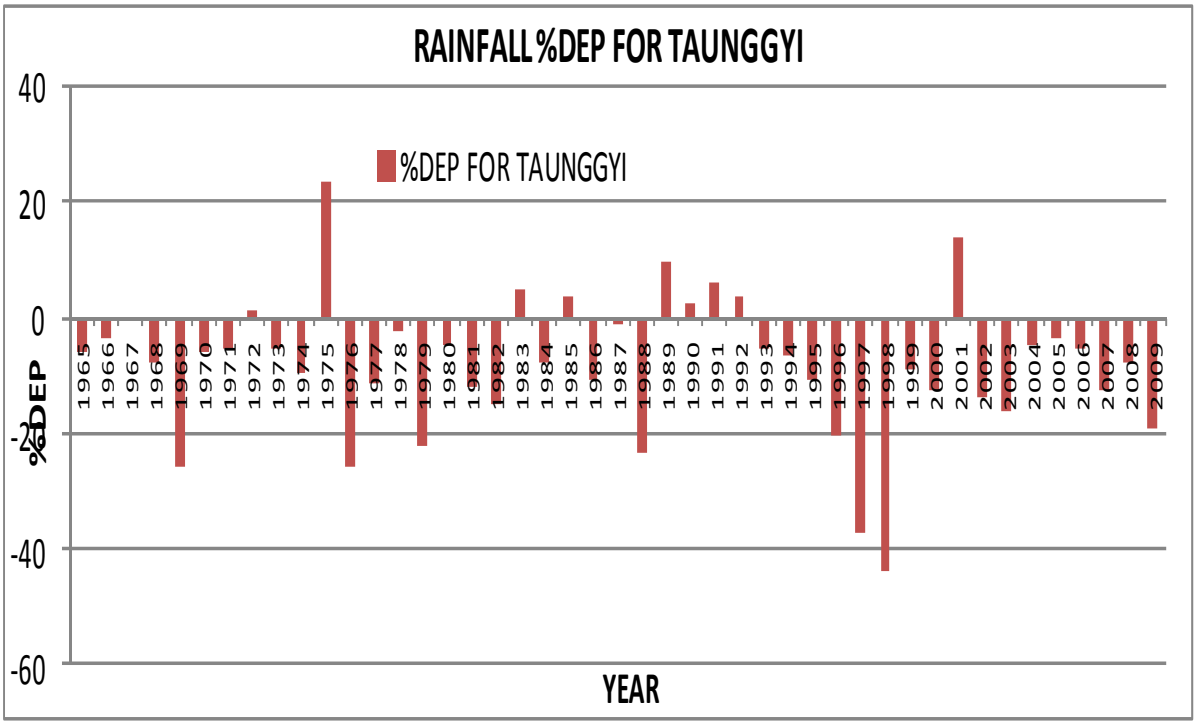
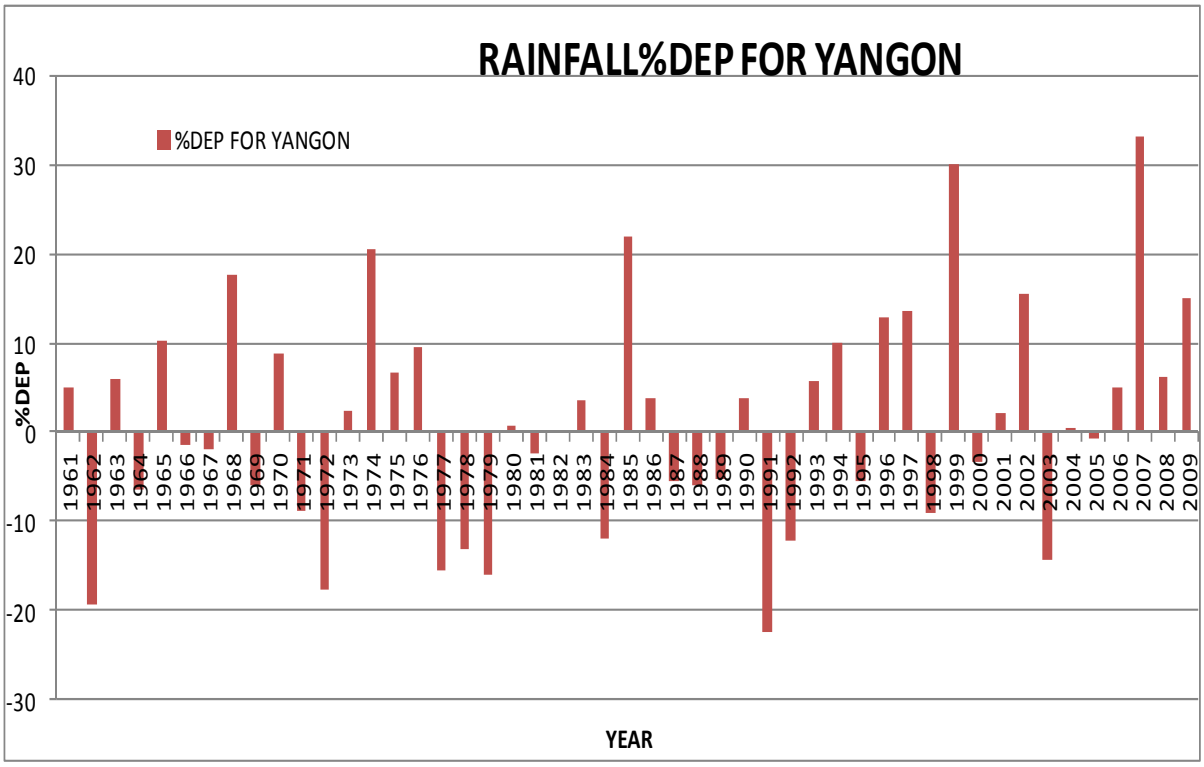


Fig. 3(v).5: Monsoon depression in the Bay of Bengal during 1926-2009

Fig. 3(v).6a and Fig. 3(v).6b: Rainfall departure over two individual stations (Taunggyi and Yangon)

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CHAPTER 3(vi)

MONSOON CLIMATOLOGY OF SRI LANKA

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3(vi).1. Introduction

Sri Lanka is an island in the Indian Ocean just south of the southern-most part of India and extend in latitude from approximately 06° N to 10° N and in longitude from approximately 80° E to 82.5^o E and has an extent of about 65,000 square kilometres. Hilly area of the country is in the central part.

The weather that Sri Lanka experiences could be broadly divided into monsoonal and inter monsoonal. The summer monsoon or southwest monsoon is from May to September and winter or northeast monsoon is from December to February. The inter monsoon periods are considered to be the periods of transition from one monsoon to the other. The first inter monsoon is from March to April whereas the second intermonsoon is from October to November. During the southwest monsoon weather is confined mainly to the southwest quarter and hilly regions of the country. Eastern and Northern regions receive rainfall during the northeast monsoon.

However, out of these two monsoons, southwest monsoon is more effective in Sri Lanka since it gives more rain over large parts of the Island and is experienced during nearly a half of the year. The farmers in the western and southern parts in the country mostly depend on southwest monsoon rainfall for their agriculture. Also the amount of rainfall which Sri Lanka receives during the southwest monsoon period, contribute much to the generation of hydro power electricity in the country.

The annual average rainfall varies from below 1000mm in the driest zones in the northwest and southeast of the island to over 5000mm in some areas on the western slopes of the central highlands.

Due to the oceanic influence and geographical location close to the tropics, the mean monthly temperatures in most parts of the island show only a small variation. For example, the mean temperature at Colombo during the cooler months from November to February is about 27°C, which is only about 03°C lower than that of the warmest months April/ May. The diurnal variation in the warmer months is about 06° C and in the cooler months it is about 11° C.

3(v).2. Southwest monsoon (Summer monsoon)

The southwest monsoon prevails from end of the May to September. The rain is possible at any time but mostly during the early hours of the day in the South-west quarter of the country (Western, Sabaragamuwa, Central provinces and Galle and Matara districts: Fig 3(v).5). The activity of the monsoon (Wind and rain) is higher from the start till about three week's period. Wet Zone (Fig 3(v).1) in the country has its main wet season from May to September, when the South-west Monsoon sweeps across the Arabian Sea like a massive wall of warm moist air. The higher slopes of the Central Highlands receive as much as 5,000 mm during this period, while even the coastal lowlands receive over 500mm (Fig 3(v).2). Average rainfall over the island during the southwest monsoon is around 556.0mm, and it is about 30% of the annual average rainfall of Sri Lanka.

Agriculture in the North and East suffers badly during the South-west Monsoon because the moisture bearing winds dry out as they descend over the Central Highlands, producing hot and often very strong winds. Thus June, July and August are almost totally rainless throughout the Dry Zone. For much of the time a strong, hot wind, called yalhulanga by the Sinhalese and kachchan by the Tamils (Fig. 3(v).3) prevails in this season.

3(v).2.1. Criteria to be satisfied to declare the onset of Southwest monsoon:

- At least 2.5 hpa pressure gradient across the island (From Colombo to Trincomalee) in the southwesterly direction.
- Extension of southwesterly winds from surface to at least upto 18000feet.
- Formation of surface low or tropospheric vortices in the vicinity of the island or on the Southwest Bay of Bengal.
- Occurrence of rain at least for two consecutive days in Galle, Colombo, Ratnapura and Nuwaraeliya districts (Districts in Southwest quarter).
- Appearance of tropical easterly jet (over 40kts) around Sri Lanka latitudes (Normally SW monsoon starts 5-10 days after the first appearance).

Average date of onset of Southwest monsoon is considered as 25th of May. In general, fairly heavy rain is experiences at the onset of the monsoon and during the months of June and July. Not much rain is experiences in the month of August, but again fairly heavy rain can be expected in the month of September. But in recent years this general pattern has changed significantly, may be due to Climate change. Sometimes, afternoon or evening thunder showers occurs over most parts of the island, and it is called break monsoon.

3(v).2.2. Criteria to be satisfied to declare the break of Southwest monsoon:

- Disintergration of lower tropospheric westerly monsoon flow.
- Pressure gradient across the island slackens and becomes weak.
- Weak monsoon flow with easterly flow above 10000 feet.
- Formation of mid tropospheric vortices between 700hpa and 300hpa pressure levels in the vicinity of Sri Lanka.
- Occurrence of afternoon or evening thunder showers over most parts of the island.
- Splitting of the tropical easterly jet at 200hpa pressure level, with one limb lying south of 10° N and the other around 22° N.

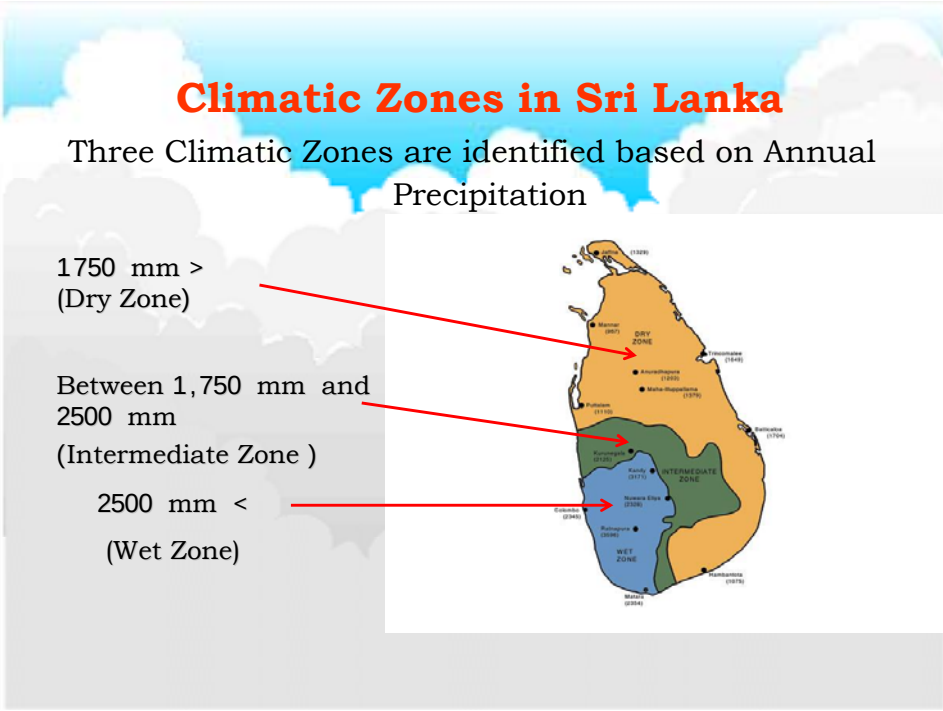

Fig. 3(v).1: Climate zones

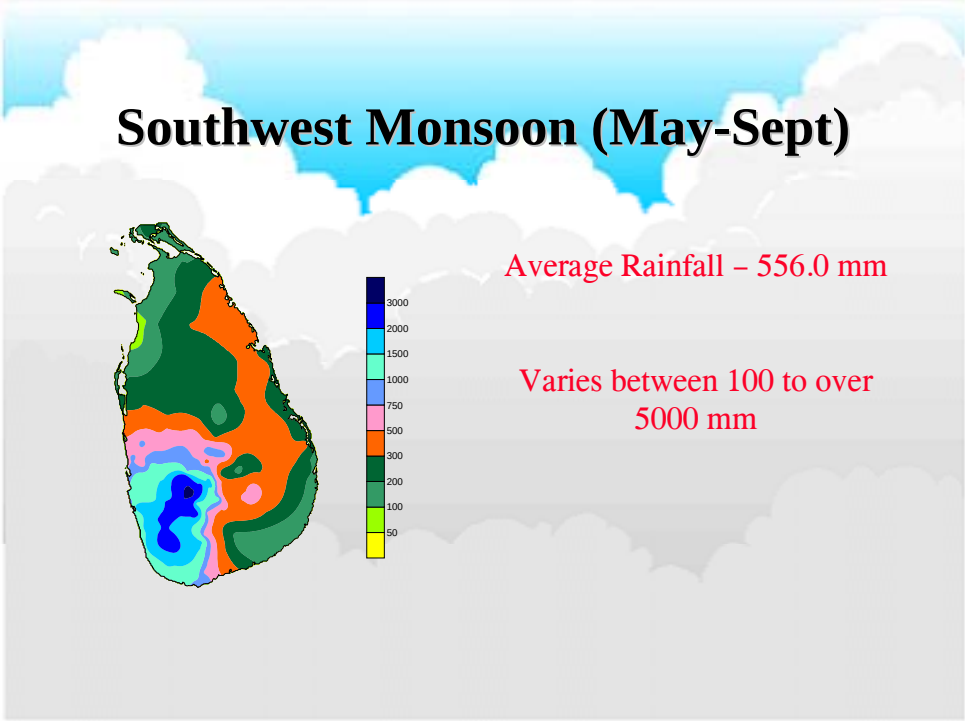

Fig. 3(v).2: Southwest monsoon rainfall distribution

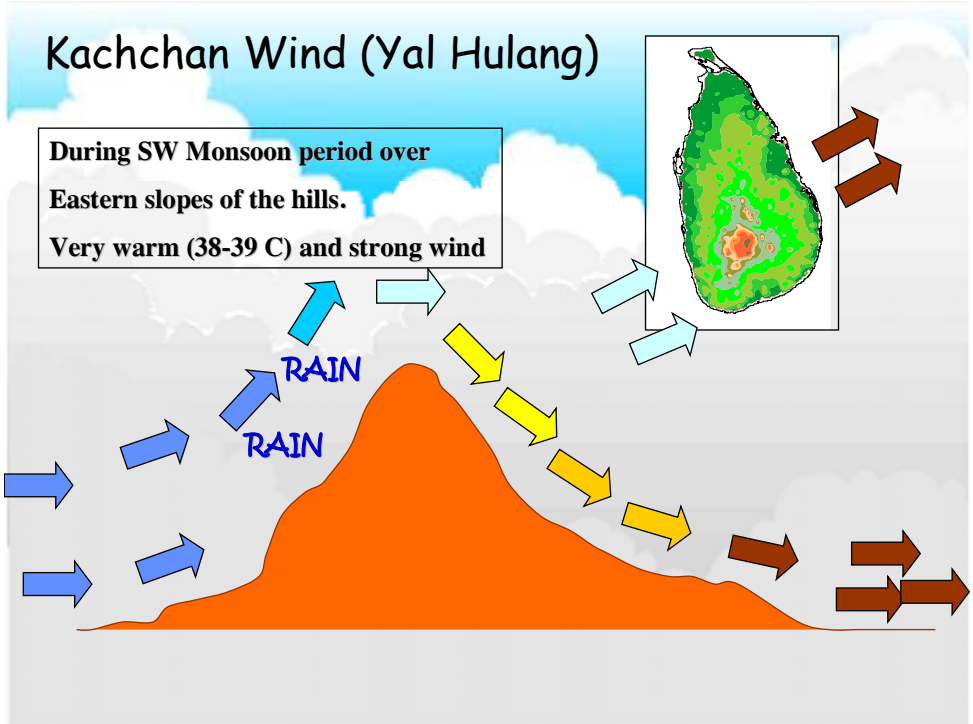

Fig. 3(v).3: Yal hulang (Kachchan Winds)

3(v).3. Northeast monsoon (winter monsoon)

The Northeast monsoon prevails from December to February next year. The rain is possible at any time of the day in the Northeast quarter of the island. Northeast quarter comprises Eastern, Northcentral provinces and some parts of the Northern province and Uva province. (Fig 3(v).4), and all the rest provinces are in the Dry zone of the country. Average rainfall during the northeast monsoon period generally varies between 100mm - 1,800 mm, and higher falls are over the eastern slopes of the central hills. Average annual rainfall over the island during the northeast monsoon is around 479.0mm, and it is about 26% of the annual average rainfall of Sri Lanka. Past rainfall data confirm that Northeastern parts of the island experiences heavy rain in the months of January and February during La Nina years.

The Wet Zone also receives some rain during this period, although the coastal regions of the South-west are in the rain shadow of the Central Highlands and are much drier. From October to December, cyclonic storms often develop in the Bay of Bengal. In general, Sri Lanka is not affected directly by most of these Cyclones, but

affected by indirect effects such as heavy rain, strong winds and sometimes storm surges, and can cause enormous damage and loss of life.

3(v).4. Criteria to be satisfied to declare the onset of Northeast monsoon:

- At least 1.5hpa pressure gradient across the island from Trincomalee to Colombo.
- Tropospheric easterlies decending to the surface.
- Occurrence of rain at least for two consecutive days in Batticaloa, Badulla and Trincomalee districts.
- Appearance of 200hpa ridge axis around Sri Lanka latitude.

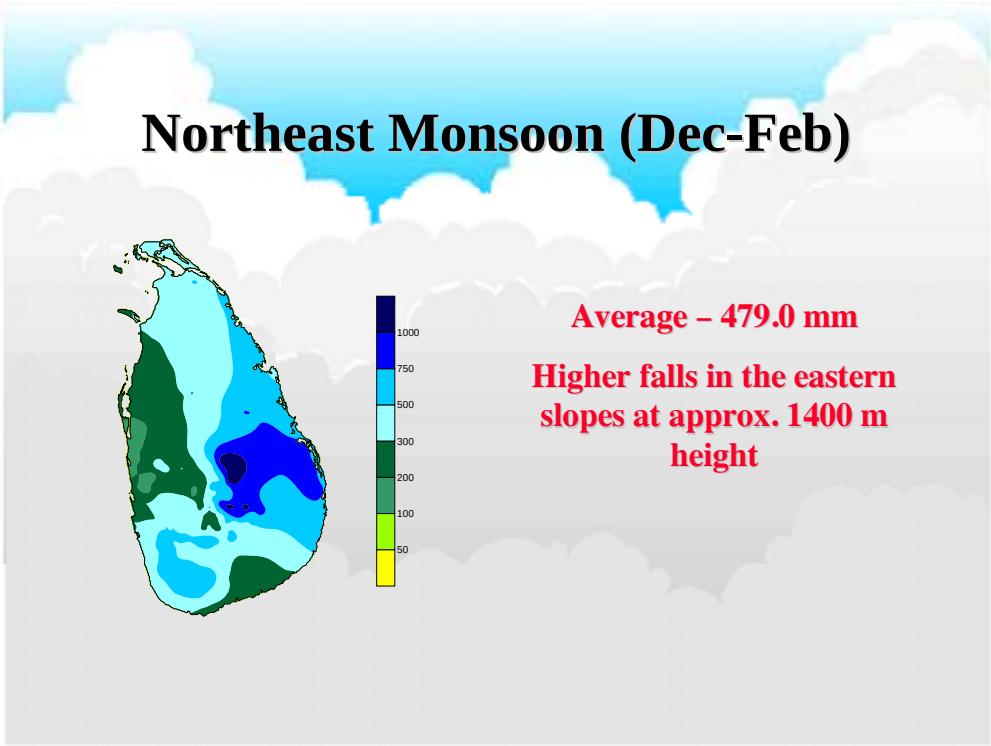

Fig. 3(v).4: Northeast monsoon rainfall distribution

Fig. 3(v).5: Sri Lankan map

CHAPTER 3 (VII)

SEASONAL RAINFALL CHARACTERISTICS AND ITS DISTRIBUTION IN THE MALDIVES

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3(vii).1. Introduction

Maldives is a group of about 1,200 islands, separated into a series of coral atolls, only 200 of the islands are inhabited. The chain of islands stretching north-south orientation lies in the southwest of India and south-southwest of Srilanka in the Indian Ocean. (Fig. 3(vii).1)

Fig. 3(vii).1)

Geographically Maldives is located between 72°32'30"E 73°45'54"E and 7°06'30"N to 0°41'48"S.

Total political area of the country is about 90,000 square km of which about 99% is water. The average elevation of the islands with flat topography is about 1m above mean sea level. The average size of these islands are 40 – 60 heactors, the largest being 500 hectors. None of these islands have either mountains or rivers. These low-lying islands are subjected to perennial beach erosion and stand at the mercy of any sealevel rise. Total population is about 300,000 of which 30% people live in the capital Male'.

The climate of Maldives is warm year round determined by two distinct seasons Northeast Monsoon (dry season) and Southwest Monsoon (wet season). Temperature varies very little with an annual average daily maximum of 30.4 ^{0}C and minimum at 25.9 ^{0}C.

3(vii).2. Data coverage

Data from 5 Meteorological stations for a period of 18 years (from 1992 – 2009) was used to analyze the rainfall characteristics and its distribution over the country.

Meteorological stations used for this study are:

S.No	Met Station Name	WMO location Index	Geographical location	Data period
1	HDh.Hanimaadhoo	43533	06°44′39″N 073°10′13″E	1992 – 2009
2	K.Hulhule	43555	04°11′30″N 073°31′45″E	1992 – 2009
3	L.Kadhdhoo	43577	01°51′33″N 073°31′19″E	1992 – 2009
4	GDh.Kaadedhdhoo	43588	00°29′17″N 072°59′49″E	1994 – 2009
5	S.Gan	43599	00°41′36″S 073°09′20″E	1992 – 2009

3(vii).3. Methodologies used

Monthly mean rainfall for the respective stations is plotted on a graph to find the seasonal distribution of rainfall for that particular station.

Following are the monthly rainfall graphs (Fig. 3(vii).2- a, b, c, d & e) for the five stations.

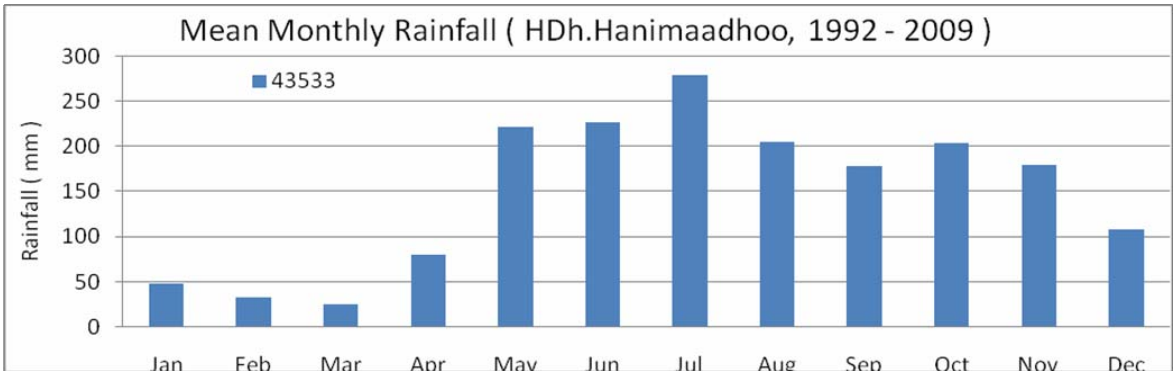

Fig: 3(vii).2a

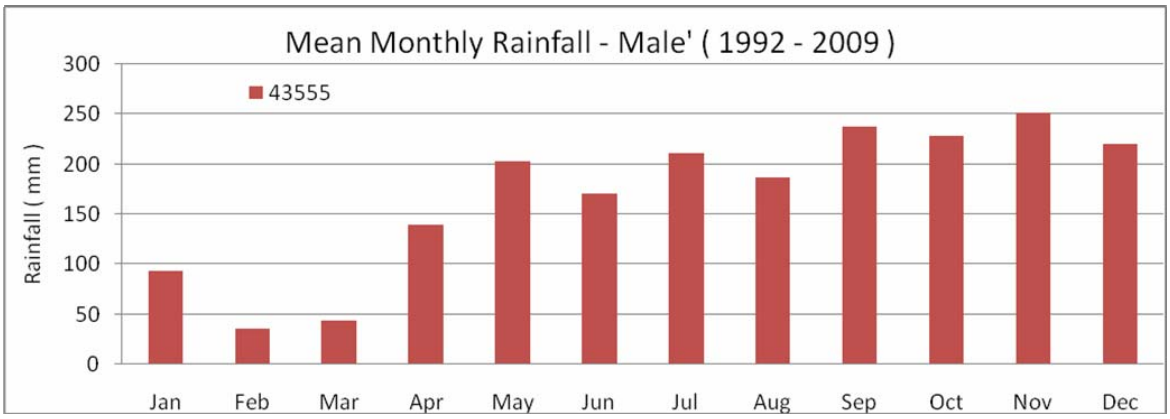

Fig: 3(vii).2b

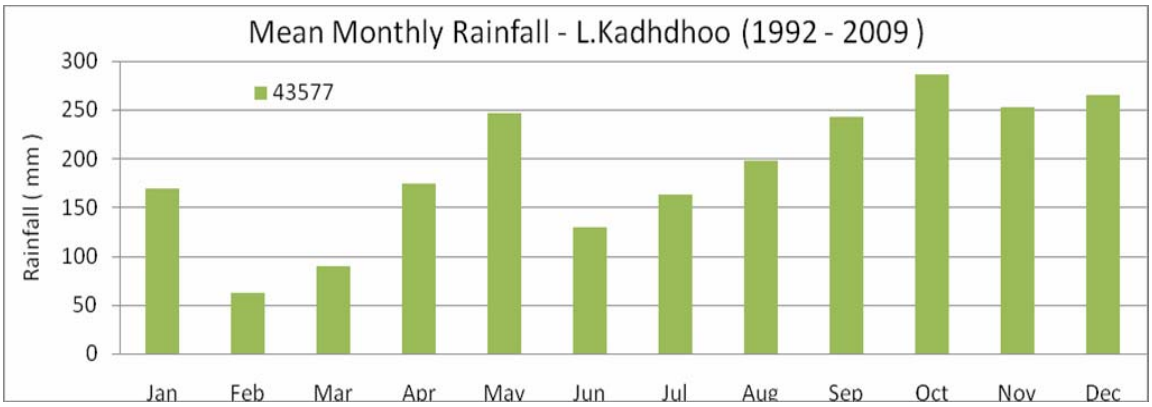

Fig: 3(vii).2c

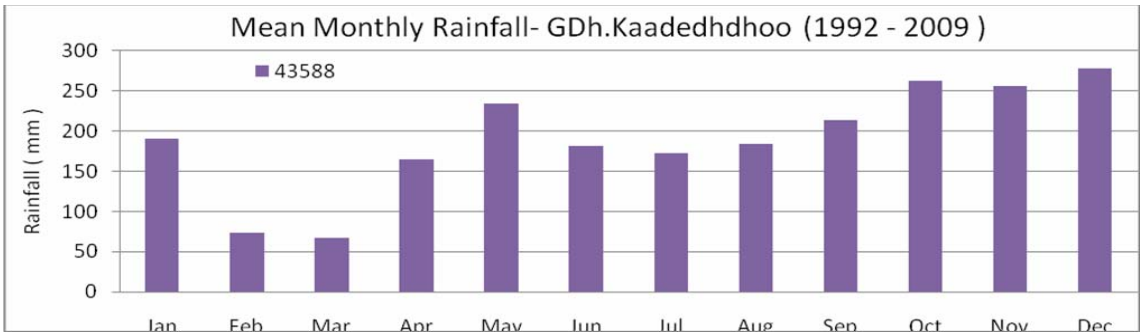

Fig: 3(vii).2d

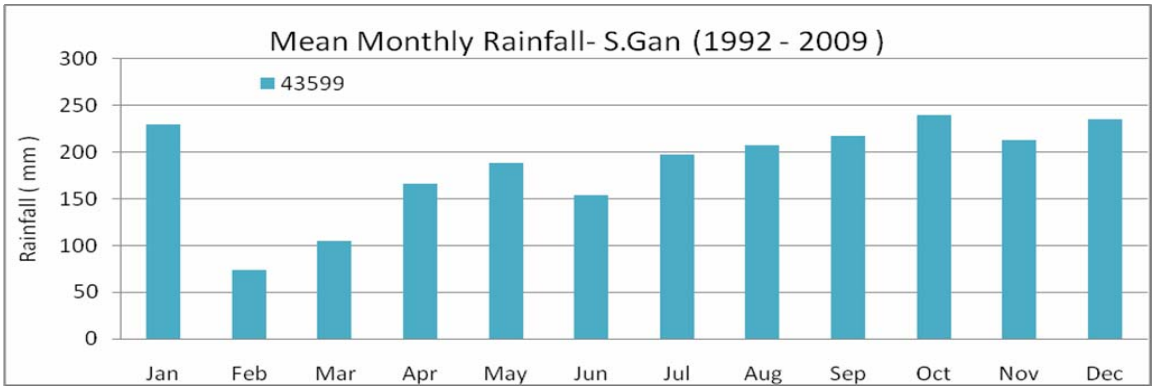

Fig: 3(vii).2e

3(vii).4. Results and discussion

The rainfall over Maldives is predominantly affected by the movement of the Inter Tropical Convergence Zone (ITCZ), which in turn, is dependent on the north-south movement of the sun. This result clearly indicates two distinct seasons in the Maldives, namely Northeast monsoon and Southwest monsoon. However, the duration of these two seasons differ from north to south. It shows that the influences of Northeast monsoon over southern atolls are very less as compared to northern atolls. The duration of Northeast monsoon (dry season) is relatively short or is not very significant in southern atolls as compared to that of northern atolls. The mean monthly rainfall graphs(fig.3(vii).2 a-e) show that the Northeast monsoon (dry season) over northern atolls extend from January to April for a period of about 4 months, but as we move from north towards southern atolls, the length of this dry period reduces to about 2 months from February to March. During Northeast monsoon northern atolls experiences significant amount of rainfall due to easterly wave activities passing westward across southern peninsular, India.

3(vii).5. Onset of Southwest monsoon

Due to geographical location of the country, the southwest monsoon gets onset earlier than any other countries of the South Asia. Normally, southwest monsoon (wet season) onset to southern atolls by 1st week of May and gradually progresses to north and gets fully established over the country by the last week of May. So the southwest monsoon is more pronounced and lengthier in southern atolls while southwest monsoon in northern atolls runs from mid-May to November, December being the transition period of two monsoons, brings a considerable amount of precipitation over the country.

The following are the onset dates of the Southwest monsoon for the country from 1990 – 2009.

Southwest Monsoon Onset

Years	Southern atolls	Central atolls	Northern atolls
2001	9th May	9th May	10th May
2002	2nd May	2nd May	7th May
2003	4th May	4th May	6th May
2004	2nd May	3rd May	4th May
2005	1st May	4th May	28th May
2006	8th May	15th May	28th May
2007	10th May	18th May	27th May
2008	5th May	15th May	26th May
2009	12th May	19th May	22nd May

Though there is significant intra-seasonal variability of rainfall, the inter-annual variability of rainfall is relatively small.

The graphs below (Fig. 3(vii).3 a, b, c, d, e & f) show that the total annual rainfall for the period from 1992 – 2009 for the selected stations and their departure from the mean for the given period.

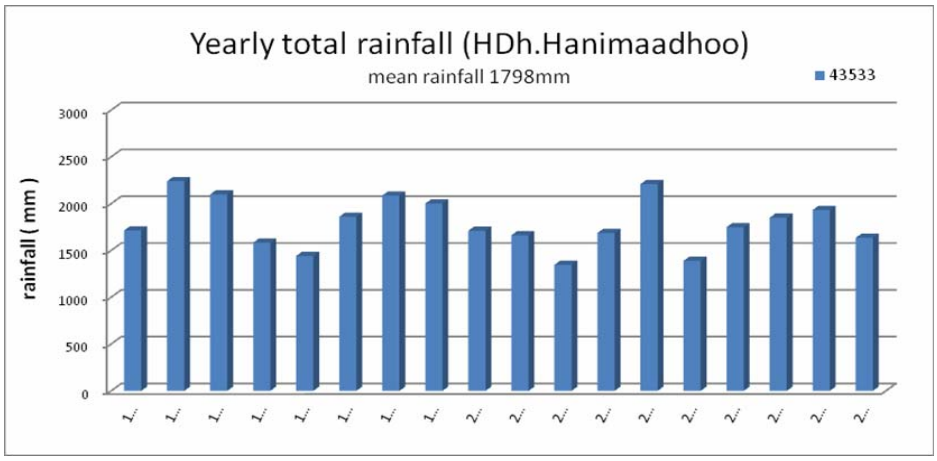

Fig: 3(vii).3a

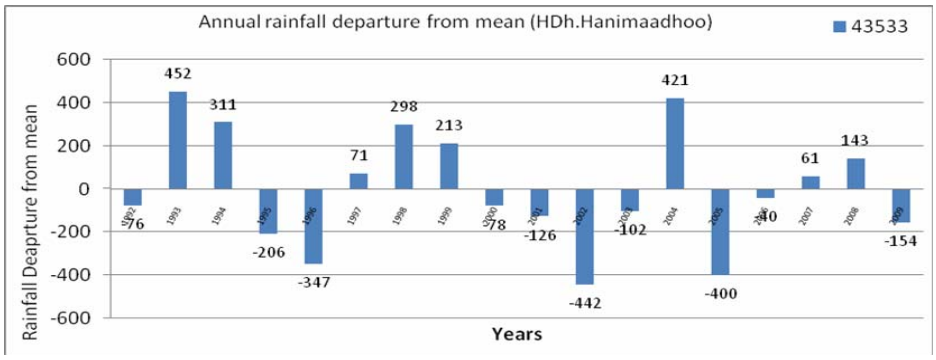

Fig: 3(vii).3b

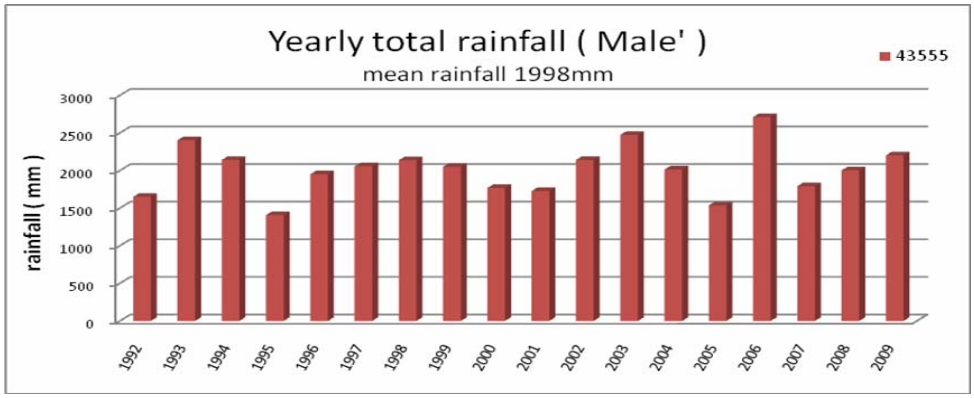

Fig: 3(vii).3c

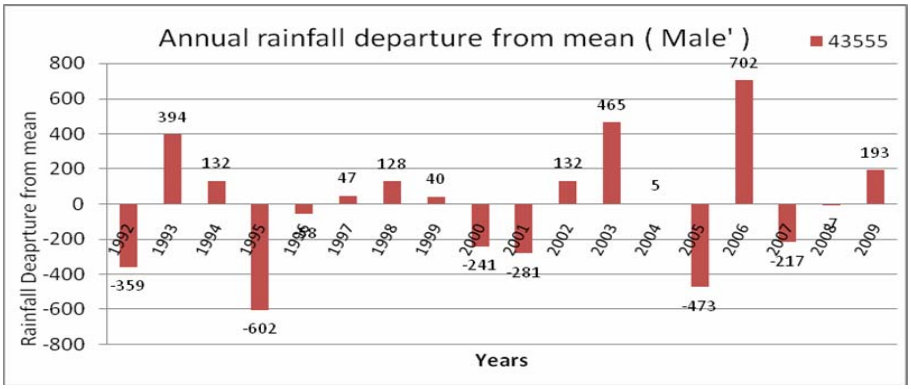

Fig: 3(vii).3d

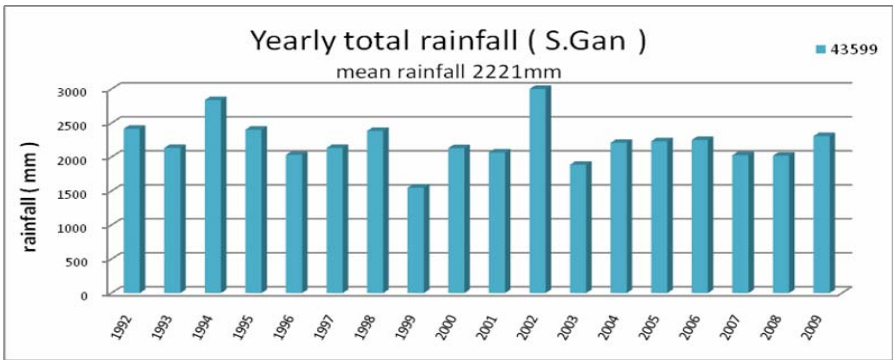

Fig: 3(vii).3e

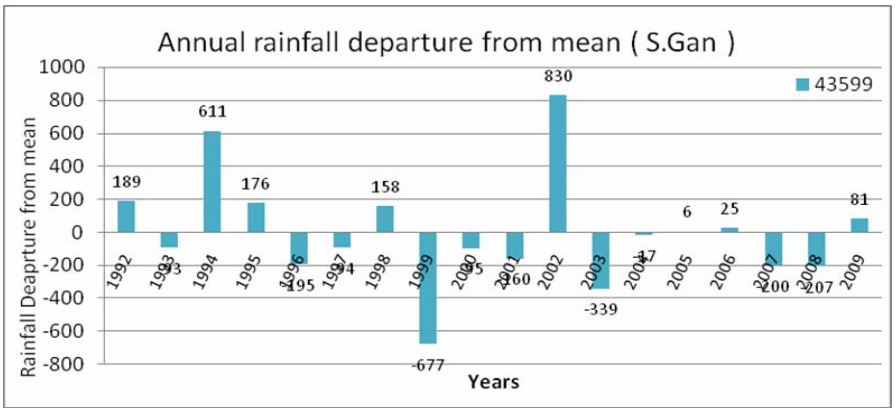

Fig: 3(vii).3f

The intra-seasonal variability of rainfall over the country, has very small impact on annual rainfall, as such the annual departure from mean seldom exceeds 700 mm.

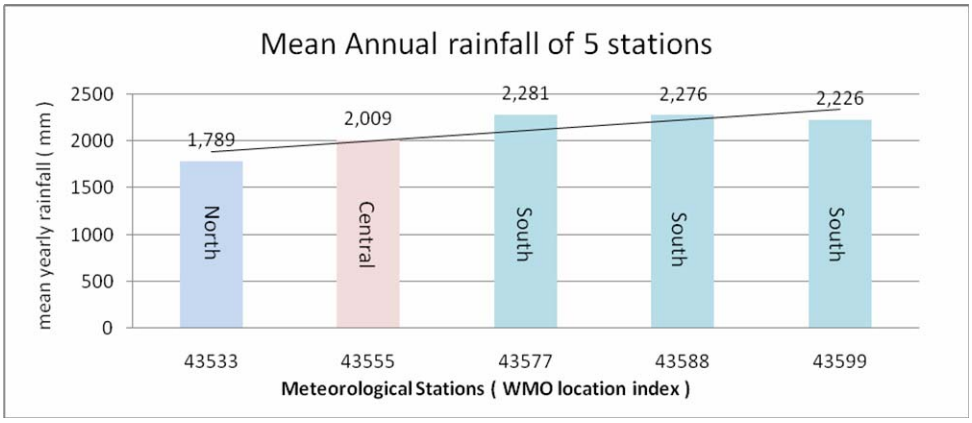

Fig: 3(vii).4

The fig. 3(vii).4 shows that the mean annual rainfall has an increasing trend from north to southern atolls.

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Country report – Drought 2009

Climate Report 2009, Climate Section, Maldives Meteorological Service.

CHAPTER 4

COMPONENTS AND SEMI-PERMANENT SYSTEMS DURING MONSOON SEASON

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4.1. Introduction

The Indian Summer Monsoon constitutes a part of Asian Summer Monsoon and is thus a unique feature of tropical circulation. The region of South Asia is characterized by unique geographical and physiographical features. This region is spread over vast continental area from equatorial to polar latitudes of northern hemisphere, surrounded by oceans to the south, near the equatorial latitudes. This geographical pattern faciliatates the development of centres of intense convection primarily due to differential heating of land and sea, resulting into concentration of rainfall within a particular epoch of the year. This phenomenon of occurrence of rainfall is a result of marked seasonal reversal of winds. The thermal contrast owing to peculiar distribution of land and oceans, coupled with the effect of Coriolis force, results into chracteristic patterns of meteorological parameters which show distinct seasonal reversal. These chracteristic patterns are obviously inter-related and are also related to tropical circulation pattern. Some of these patterns are semi-permanent in nature. They are observed throughout the summer monsoon season, though with variations in positions and intensity, on interannual and intra-seasonal scale. Such variations are closely linked with the intensity of monsoon circulation and rainfall distribution, both spatial and temporal, over India. The major semi-permanent features associated with the Indian Summer Monsoon are: 1. Mascerene HIGH, 2. Heat LOW, 3. Monsoon Trough (MT), 4. Low Level Jet (LLJ), 5. Tibetan HIGH (TH) and 6. Tropical Easterly Jet (TEJ). These features are schematically depicted in Fig. 4.1.

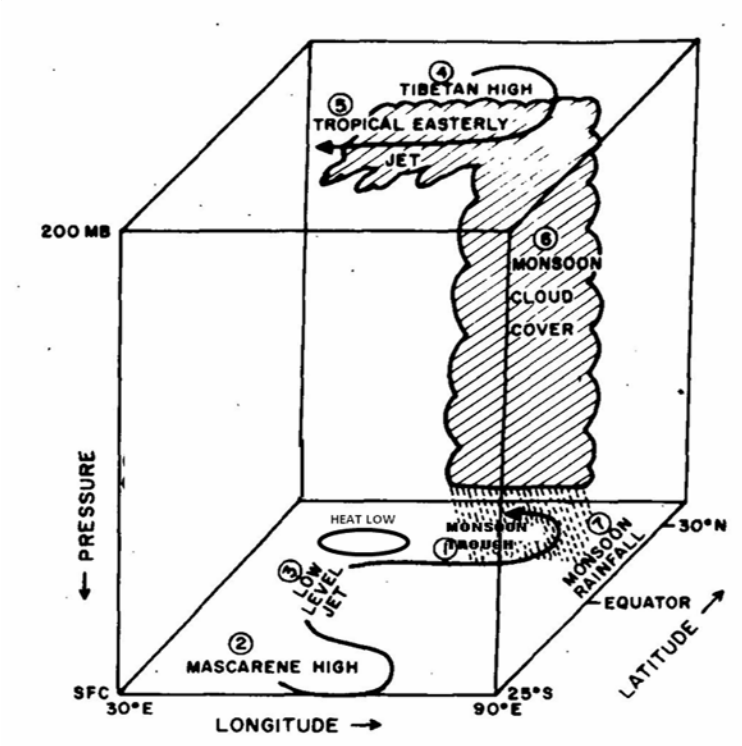


Fig. 4.1: Schematic view of the semi-permanent features during SW monsoon season (Krishnamurti and Bhalme (1976))

The characteristic features and their reltionship with the rainfall activity are discussed in the following sections.

4.2. Mascarene HIGH

As one of source regions of East Asian summer monsoon, the Southern Hemisphere circulation especially Mascarene HIGH and Australian HIGH, have important influence on East Asian summer monsoon. The summer monsoon currents over East Asia originate either from the north Pacific HIGH or from Mascarene HIGH (MH) over the subtropical Indian Ocean and Australian HIGH.

Thus, the Mascarene high is one of the important features of the tropical general circulation which has profound influence on South Asian climate and weather. It is the high-pressure area at sea level south of the equator in the Indian Ocean near Mascarene island (Fig. 4.2), with its centre located near 30°S, 50°E.

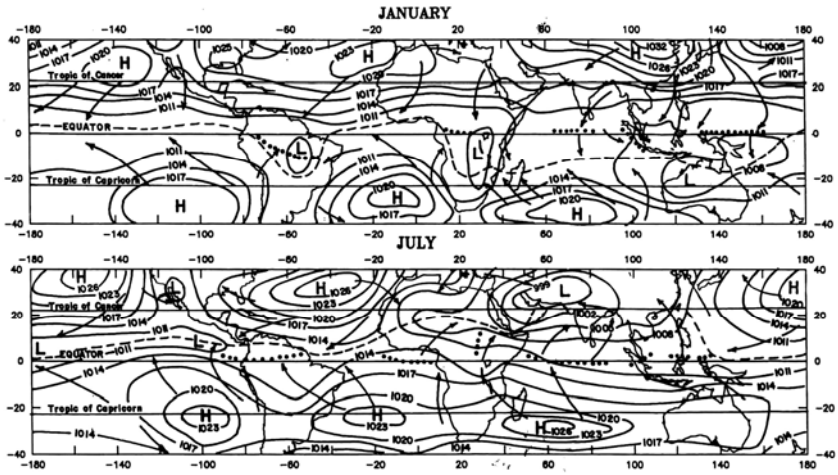


Fig. 4.2: The sea level pressure (hPa) and wind pattern over the tropics during January and July (Nieuvolt (1977))

The position and intensity of this high are considered to be closely linked to the South Asian summer monsoon activity. Although it is a semi-permanent circulation feature, it is more prominent during the northern hemispheric summer. The mean monthly value of the central pressure in the region of MH during June, July, August and September is about 1025 hPa (Ananthakrishnan, et. al. (1968)).

The Mascarene high undergoes short-period fluctuations in its intensity owing to the passage of extra-tropical westerly waves of the southern hemisphere. The intensification of the Mascarene high strengthens the cross-equatorial flow in the form of the East African Low-Level jet and the corresponding monsoon current over the Arabian Sea (Sikka and Gray, 1981). The intensity of the Mascarene HIGH is also found to be associated with the onset of the monsoon over India as well as the subsequent fluctuations in its activity (Okoola and Asnani, 1981). As a major element of the Indian summer monsoon system, oscillation in the strength of the Mascarene HIGH (MH) is linked to variability of monsoon rainfall, a minimum in monsoon rainfall lags about 9 days behind the maximum intensity of MH (Krishnamurti and Bhalme, 1976).

The MH plays a crucial role in the interactions of general atmospheric circulation between the two hemispheres (Feng, et. al. (2003)). The study shows that with the intensification of MH, the Somali Low-Level jet is significantly enhanced

together with the summer monsoon circulation in the tropical Asia and western Pacific region.

The Antarctic Oscillation (AAO) dominates the interannual variability of MH; the intensity of MH is enhanced when the circumpolar low in high latitudes of the Southern Hemisphere deepens (Feng. et. al. (2003)).

Based on the reanalysis data from NCEP/NCAR and other observational data, Xue, et. al. (2004) has examined the interannual variability of the Mascarene high (MH) and Australian high (AH) during boreal summer from 1970 to 1999. The study reveals that the Antarctic Oscillation (AAO) dominates interannual variability of MH, and MH tends to be intensified with the development of the circumpolar lows in high southern latitudes. It is also observed that, with the intensification of MH, the Somali jet, and Indian monsoon westerlies, tends to be strengthened. With the intensification of MH during boreal spring through summer, the Meiyu/Baiu rainfall from the Yangtze River valley to the Japan Islands tends to increase, while less rainfall is observed outside of this region.

4.2. Heat LOW

The land low over south Asia during the pre-monsoon season (during March to May) is a part of the global scale low pressure belt extending from the Sahara to central Asia across Arabia, Iran, Afghanistan, Pakistan, northwest India, and even up to Myanmar (Fig. 4.3).

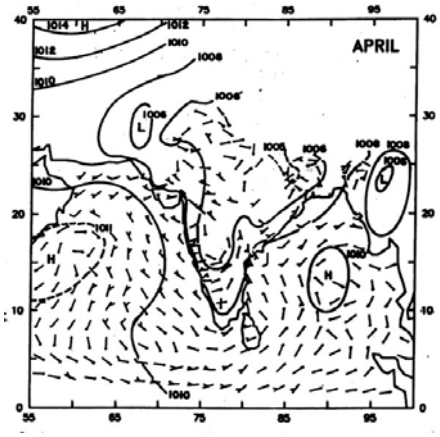


Fig. 4.3: Mean sea level pressure (hPa) and surface wind pattern overSouth Asia during April (Rao and Ramamurty (1968)

The Himalayan barrier has an important effect on the development of Heat LOW during the summer monsoon season over India. During the monsoon months, the Himalaya forces the rain-bearing winds up the slopes to release most of their moisture on the southern side of the mountains. This results in the location of the heat low over Sindh and adjoining areas and the monsoon trough across northern India. The location of these low-pressure centres is determined by the mountain and hill configurations. The mountains in general influence the atmospheric flows as a result of mechanical lifting and also through heating due to condensation, which in turn is caused by air ascending over the mountain barriers.

During the summer months, the centre of the seasonal low, popularly known as the 'heat low', is located over the region, where maximum temperatures of 45^0C or more are experienced. The resulting steep horizontal pressure gradients generate strong surface winds which, together with loose sand, produce dust and sand storms. The boundary layer, consisting of approximately the lowest 1.5 km over the region, experiences a strong temperature inversion (Ramage (1971), Das (1962)). The inversion suppresses the convective activity and ascent of the moist air, thereby suppressing the cloudiness and precipitation. The dust content in the atmosphere is relatively much higher over these regions and it extends vertically almost to a height of 10 km (Bryson and Swain (1981)). The radiational effect of the dust content is such that it induces cooling in the upper layers which results in large-scale sinking (Das (1962)), thus enhancing the conditions for a drier climate.

By May, the continental low-pressure areas completely dominate North Africa and Asia. The main centre over South Asia lies over northwest India and adjoining Pakistan, with an extension as a trough into peninsular India, which becomes more marked in June. The centres of the low in different months are – over east Madhya Pradesh in April, Punjab in May and Pakistan in June and July. The centres of the heat low over land areas are located near regions of maximum heating out of reach of maritime air mass. The development of heat low causes the circulation, which brings cooler air mass in favourable area offsetting the effect of solar heating. The centres of low-pressure area develop depending upon the balance between these two factors. The lowest pressure for the whole belt from Africa to Asia is over the

central parts of Pakistan. This is not the place of the highest temperature for the lowest pressure to develop. Sahara records much higher temperature. Banerji (1930) considered that the configuration of mountain ranges in the northwest corner of the sub-continent and the adjustment of wind field and consequently the pressure pattern throws up the low pressure centre over the central parts of Pakistan.

The tapering shape of peninsular India, the Himalayan barrier to the north and the Assam hills to the northeast, which make the maritime air mass pervade over most of South Asia, displace the centre of the low to the extreme northwest of the sub-continent. The blocking of cold air incursions from the north by the Himalayas in the lower troposphere also makes the heat low more intense in this region. The existence of the lowest pressures of the Afro-Asian low in this region, in spite of the relatively higher temperatures in the Sahara, is the result of the topographical configuration and the associated adjustments in the wind field.

When the western disturbances, which bring cold air in their association, move to the north with the advance of summer, heating continues uninterruptedly in May in the absence of any rain or clouding over the area of the heat low.

The progressive development of a heat low over the Indian sub-continent and its location over central parts of Pakistan in July is perhaps the most important causative factor of the Indian Summer Monsoon circulation.

Monsoon activity reaches a peak in July when this low-pressure area extending from North Africa to northeast Siberia is most intense, with its main centre over north Baluchistan (Fig. 4.4).

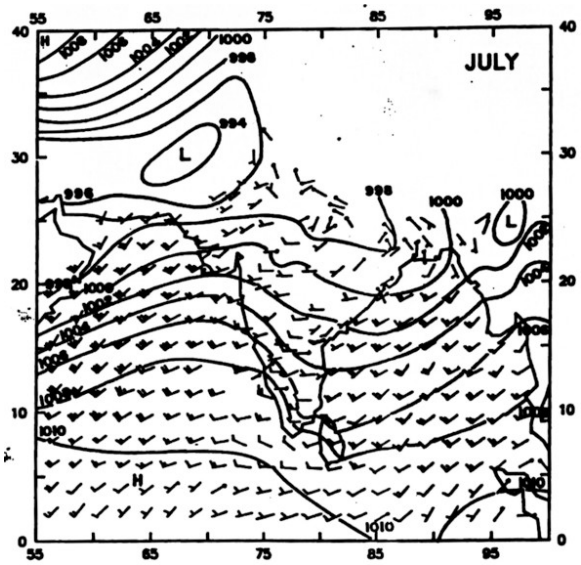


Fig. 4.4: Mean sea level pressure (hPa) and surface wind pattern over South Asia During July (Rao and Ramamurty (1968)).

The mean monthly values of the sea level pressure values in the Heat LOW region are 998 hPa in June, 996 hPa in July, 998 hPa in August and 1000 hPa in September. During the summer monsoon season, on individual charts, the lowest pressure in the Heat LOW region can be as low as 985 hPa (Ananthakrishnan, et. al. (1968)).

A trough lies over north India with its axis running from northwest India to Head Bay (north coast of the Bay of Bengal), which is known as the ‘monsoon trough’. As the monsoon becomes established over South Asia, the subtropical anticyclones over the Indian Ocean are strengthened, thereby creating strong pressure gradients north of the subtropical highs up to the monsoon trough. This facilitates large-scale cross-equatorial flow into the South Asian region. The pressure patterns over South Asia, south of the monsoon trough, are characterized by weak ridges in the Arabian Sea off the west coast of India and in the Bay of Bengal off the Tennaserim coast and over Myanmar, a weak trough over eastern peninsular India off the east coast. The intensity of the Afro-Asian low starts decreasing in August, with a consequent decrease in the pressure gradient south of the monsoon trough. Towards the end of the monsoon season, the trough over northern India shifts to the Bay of Bengal and the pressure field becomes flat over almost all South Asia.

The heat low is shallow, extending only up to about 1.5 km above sea level and is overlain by a well-marked ridge extending to the upper troposphere, which is part of the subtropical high-pressure belt. There is frictional convergence in the heat low and weak ascent in it while subsidence is likely in the ridge. A more stable layer may occur in the lower part of the column of subsidence due to vertical shrinking.

The role of the upper ridge in maintaining the surface heat low is not clearly established. As a ridge overlies the heat low above 1.5 km, the question arises whether the ridge has a role in maintaining the surface low. As the axis of the ridge is close to the axis of the heat low, the pressure distribution of the ridge is unfavourable for the maintenance of the low. While it may be argued the subsidence in the upper ridge is maintaining the clear skies required for the heat low, this is not the case in the months before the monsoon sets in. In April and May, the sub-tropical ridge is well outside the heat lows over the sub-continent. (Rao, 1976).

The desert area of the Indian sub-continent comes under the heat low with the sub-tropical ridge aloft. Bryson and Baerries (1967) have pointed out that the desert coincided with the extent of the divergent sinking of air at 700 hPa. As pointed out by Desai (1968) the shallowness of the monsoon air over the desert is due to the continental air brought in by the sub-tropical ridge.

Even over the Arabian Sea, west of 68°E, monsoon air mass is only 1 to 1.5 km deep, with continental air aloft, brought by the anticyclonic flow which cannot be ascribed to any surface dust, unless the argument is built around advection of dust from neighbouring land areas.

The intensity of the heat low has been correlated with monsoon activity. Negative pressure anomalies in the heat low region and positive anomalies over peninsular India are regarded as favourable conditions for good monsoon activity. Pressure gradient would then be strong over the Peninsula, which is conducive to monsoon rains. The heat low may also strengthen when the ridge aloft weakens under the influence of westerly troughs moving further north. Some westerly troughs

can cause formation of weak lows over northern India and lead to increase of rainfall.

Ramage (1971) has shown that the surface pressure at Jacobabad, in the heat low, is inversely related to the intensity of monsoon rains over a strip of the sub-continent between 18^0 and 27^0N. He has quoted the observation of Dixit and Jones (1965 prepublished) that upper tropospheric temperatures over the heat low are warmer by 2^0-6^0 C during the epochs of rains than during the epochs of monsoon lull and concludes that subsidence aloft raises temperature of middle and upper troposphere and reduces surface pressure in the heat low. The rainfall in this strip is not only influenced by tropical systems approaching from the east but also by the trailing edge of the troughs in westerlies in mid-and upper troposphere. Both the systems even interact in causing rains. The effect of such systems would be to weaken the mid-and upper tropospheric ridge above the surface heat low and show that the strengthening of the heat low is positively correlated to increase of monsoon rains.

4.3. Monsoon Trough

The monsoon trough, also considered as the Equatorial Trough of the northern summer season or Intertropical Convergence Zone (ITCZ) over India during the summer monsoon season, is depicted by a line on a weather map showing the locations of minimum sea level pressure within the monsoon region. In this season, a weaker trough persists within five degree south of the equator. Riehl (1954) has shown that the pressure profile about the equatorial trough (in summer hemisphere) averaged over the globe is quite symmetrical in northern and southern summer. In the Indian monsoon trough, the pressure gradient equator-wards of the trough is higher than that averaged for the whole globe. This is not so much due to up glide along the slope of the trough or convergence in it, as due to the different synoptic systems that prevail with the different positions of the trough.

Monsoon Trough is also a convergence zone between the wind patterns of the southern and northern hemispheres. As such, westerly monsoon winds lie in its equatorward portion while easterly trade winds exist poleward of the trough. Right

along its axis, heavy rains can be found which mark the beginning of the peak of a location's respective rainy season.

Monsoon depressions form in the vicinity of the monsoon trough, with each capable of producing a year's worth of rainfall in a relatively short time frame. The migration of the monsoon trough into a landmass such as Asia, Australia, North America, or across Africa heralds the beginning of their annual rainy season during their summer months.

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

During the Indian summer monsoon season, Monsoon Trough is the most conspicuous signature of the monsoon circulation at the surface. The monsoon trough is a major semi-permanent feature of the summer monsoon circulation in the lower troposphere and exerts considerable influence on the summer monsoon activity in South Asia. The trough line runs at surface level from Ganganagar to Kolkata through Allahabad, with west to southwest winds to south and easterlies to the north of the trough line. The air mass to the south of the trough line is the Arabian Sea monsoon while the air to the north may have some travel over the Bay of Bengal. The mean surface wind at Kolkata is mainly from south in spite of the depression, which forms in the North Bay of Bengal. In the daily weather charts, the trough can be located up to 500 hPa level (about 6 km) during periods when the monsoon has advanced into the northern parts of India. This trough is seen in the upper levels of the atmosphere, up to about 6 km above sea level (a.s.l.), the trough line sloping southwards with height. At 4 km (a.s.l.), it runs from Bombay to

Sambalpur. The monsoon trough indicates the separation between air masses of northern and southern hemispheric origin.

The trough line tilts southward with height; the slope in the western sector is steep in the lower tropospheric and less marked in the middle troposphere, while in the eastern sector the slope is moderate in the lower troposphere and increases appreciably in the middle troposphere (Srinivasavn, et. al. (1971)). The temperature to the north of the trough is on average about 2^0C higher than that to the south, which explains the southward slope of the trough axis with height (Joshi and Desai, 1985). The features such as, about 2^0C higher temperature to the north, and the slope of the trough line in the right direction, indicates resemblance to a quasi-stationary front. The mean position of the monsoon trough during the month of July is depicted in Fig. 4.4. The trough axis exhibits considerable day-to-day variation in its position, which has vital bearing upon the monsoon rainfall distribution in the region. No other semi-permanent feature has such a control on monsoon activity. This arises mainly from the different synoptic systems that prevail in association with different positions of the trough.

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

Riehl (1954) discusses the distribution of rainfall relative to the equatorial trough. Heavy rainfall and equatorial trough coincide seasonally as well as in the annual mean over the oceans. The rainfall maxima over land lies equator ward of the trough wherever its position deviates markedly from the equator.

When the rainfall relative to the equatorial trough is averaged, symmetrical profiles over the oceans are obtained in southern summer, the continental peak (in rainfall) lies 2^0 latitude north of the trough; in the northern summer, it departs much farther.

The mean rainfall is higher to the south of the trough on account of the heavy rains in the southwest quadrant of depressions which travel west-northwestwards, a little to the south of the mean position of the monsoon trough line. Along the mean position of the trough itself, rainfall is minimum and again increases towards the foot of the Himalayas.

Normally some rain occurs near the surface trough and to south on account of the convergence in it and southward slope with height. When the trough rapidly shifts north or south, which can be even 5^0 latitude in a day, monsoon activity is enhanced in that area. The latent instability of the air mass enhances rainfall in the trough when it is over Bihar and east Uttar Pradesh, even without a low.

It is observed that normal rainfall is not maximum near the mean position of the trough. The increase from July to August in the frequency of the position to the north in the western and central parts is interesting, as the breaks in monsoon are more in August. The monsoon trough is not so well defined at the beginning of the season, while towards the end the western end becomes diffuse. In September, the eastern end shows wide fluctuations in position and orientation.

Ramage (1971) explains the position of the monsoon trough and the less rains in its mean position as due to subsidence. The anabatic winds on the southern slopes of the Himalayas are part of the large-scale summer circulation as made out by Flohn (1968). Though strongest in the afternoon, they prevail throughout the twenty-four hours in a day. They coincide with the northern rainfall maximum and presumably form a part of a local vertical circulation in which air returns southwestward in the middle troposphere and tends to sink over the northern plains. The associated subsidence warming diminishes surface pressure. However, the southward slope of the trough and the existence of a weak Hadley cell between the trough and the Himalayas with ascent in the trough may not support such a picture.

Since the monsoon trough is an area of convergence in the wind pattern, and an elongated area of low pressure at the surface, the trough converges low level

moisture and is defined by one or more elongated bands of thunderstorms when viewing satellite imagery.

During the monsoon season, there are occasions when the monsoon activity abruptly weakens and dry weather prevails over most parts of South Asia. These situations are called `break monsoon periods' and are characterized by certain distinct circulation features. The position of trough line close to the foot-hills is referred as 'break in monsoon' on account of drastic decrease in rains over the country, though the Himalayan mountain belt and parts of northern India experience heavy falls which can cause floods in the rivers originating there. This is sometimes associated with an increase in rainfall in the southernmost parts of the Indian peninsula and Sri Lanka. During this period, situations characterized by cessation of rain and the disappearance of monsoon like synoptic conditions, may occur. A swing of the trough to the central parts of India generally occurs with monsoon depressions from the North Bay of Bengal moving west to west northwest across the country. The monsoon depressions, which originate over the Head Bay of Bengal, move across the Gangetic plains in the northwesterly direction along the trough line. The run of the trough line cannot be uniquely identified when such systems are embedded.

The breaks occur with varying intensities and durations and are mostly associated with the north-south movements of the monsoon trough.

The number of break monsoon days in the months of July and August for the period 1888-1968 are shown in Table 4.1 (Ramamurthy, 1969). Ramamurthy (1969) adopted the following criteria for defining the occurrence of a break:

(i) the monsoon trough is absent near the surface up to about the 850 hPa level;

(ii) the duration of such a synoptic situation is at least 2 days.

Table 4.1: Statistics of break monsoon situations during July and August for the Period 1888-1967 (data from Ramamurthy, 1969)

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

It is observed that the middle 10 day period of August has witnessed the most frequent occurrence of break monsoon situations, which are also of longer duration than in July.

De, et. al .(1998) have studied the breaks during the period 1968 to 1997, based on the criterion by Ramamurty (1969). This study has revealed that, during the period 1968 to 1997, there were 193 break days with 33 break situations, of which most have occurred in July and August.

Rajeevan et. al., (2008) have suggested a criteria for identification of active and break events of the Indian summer monsoon on the basis of the high resolution daily gridded rainfall data set over India (1951-2007). They have identified the active and break events from the average rainfall data over a critical area, called the core monsoon zone within which the monsoon trough/ Continental Tropical Convergence Zone (CTCZ) normally fluctuates in the peak monsoon months of July and August. Active and break events are defined as periods in which the normalized anomaly of the rainfall over the monsoon zone exceeds 1 or is less than -1.0 respectively, provided the criterion is satisfied for at least three consecutive days. With this criterion, it has been observed that:

i. On an average, there are 7 days of active and break events during the period July and August.
ii. Breaks tend to have a longer life-span than active spells. While, almost 80% of the active spells lasted 3-4 days only 40% of the break spells were of such short duration.

iii. A small fraction (9%) of active spells and 32% of break spells lasted for a week or longer, of these, almost 30% break spells persisted for more than 10 days. Active events are more common than breaks.

iv. While active events occurring almost every year, not a single break occurred in 26% of the years.

v. There are no significant trends in either the days of active events or break events during the monsoon season.

Table 4.2 shows the normal positions of the monsoon trough. It is observed that the trough is in the normal position only 30-47% of the time, indicating its high intra-seasonal variability.

Table 4.2: Mean positions of monsoon trough during 1961-69 (after Srinivasan and Ramakrishnan, 1970)

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

It is also seen that the western part of the trough is more often towards the north than towards the south, while the eastern part shows the opposite trend. This is due to the influence of the mid-latitude disturbances on the northwestern parts of the subcontinent and the formation of monsoon depressions over the north Bay of Bengal (Srinivasan and Ramakrishnan 1970). The energy balance studies in the monsoon trough region indicate that the vertical structure of the trough has a large convergence zone in the lower levels and a compensating outflow higher up (Anjaneyulu, 1969).

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

The increases in the relative vorticity, within the monsoon trough, are normally a product of increased wind convergence within the convergence zone of the monsoon trough. Wind surges can lead to this increase in convergence. A strengthening or equatorward movement in the subtropical ridge can cause a strengthening of a monsoon trough as a wind surge moves towards the location of the monsoon trough. As fronts move through the subtropics and tropics of one hemisphere during their winter, normally as shear lines when their temperature gradient becomes minimal, wind surges can cross the equator in oceanic regions and enhance a monsoon trough in the other hemisphere's summer. A characteristic way of detecting whether a wind surge has reached a monsoon trough is the formation of a burst of thunderstorms within the monsoon trough.

If a circulation forms within the monsoon trough, it is able to compete with the neighbouring thermal low over the continent, and a wind surge occurs at its periphery. Such a circulation which is broad in nature within a monsoon trough is known as a monsoon depression. Monsoon depressions are generally asymmetric, and tend to have their strongest winds on their eastern periphery.

Alexander et. al. (1978) have examined the monsoon rainfall activity in terms of the charts of anomaly of pentad winds at lower, middle and upper troposphere, associated with strong and break monsoon conditions. It has been brought out by this study that:

i. A couple of days prior to the beginning of break monsoon, an anomaly ridge extends from peninsular India to Malaysia in the lower and middle troposphere. In the upper troposphere, an anomaly trough is observed to the north west of India.
ii. A couple of days prior to the beginning of strong monsoon, a marked anomaly trough extends from peninsular India to Thailand in the lower

and middle troposphere. In the upper troposphere, an anomaly ridge is observed to the north west of India.

The weekly mean synoptic flow patterns for strong and weak monsoon conditions exhibit contrasting features (Fig.4.5).

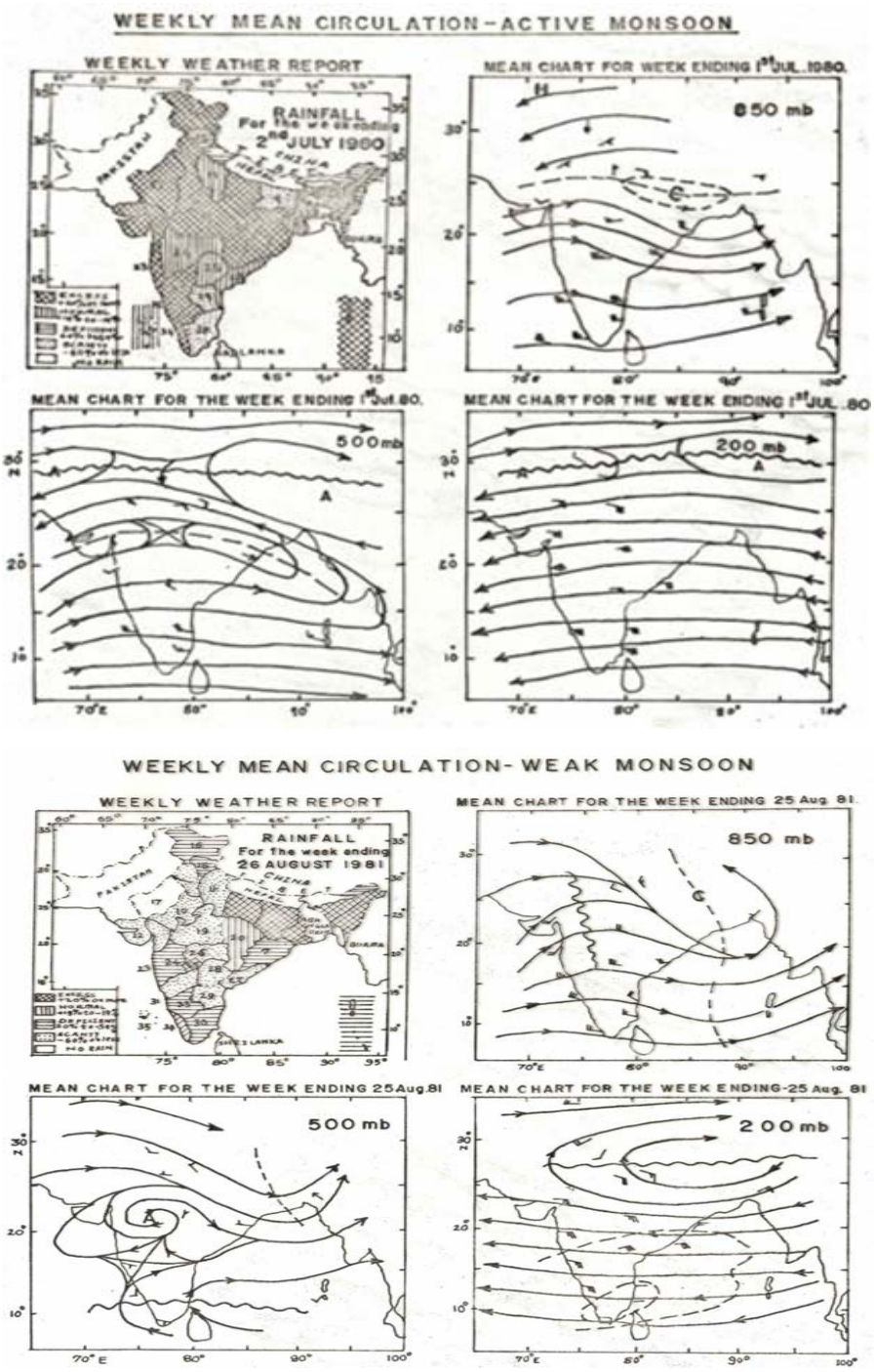


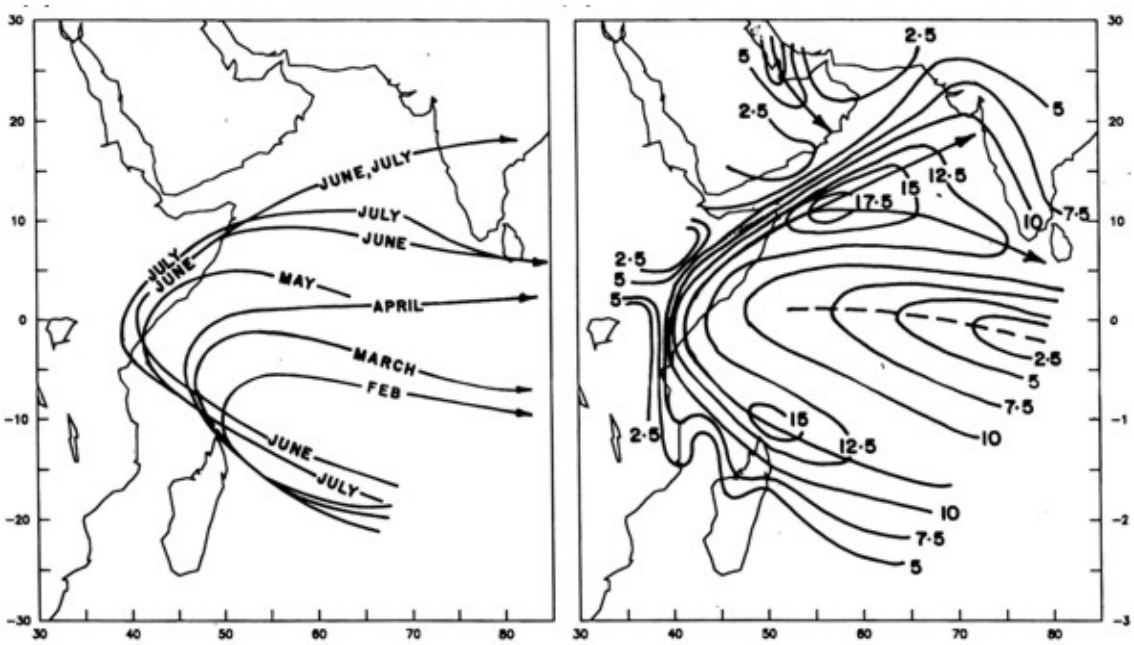
Fig. 4.5: Weekly Mean Circulation corresponding to active and weak monsoon conditions

4.4. Low-Level Jet

The Low-level jet streams, which are found in several parts of the world, refer to strong low-level currents on a monthly or seasonal basis, but they are much weaker and smaller than the planetary-scale upper-tropospheric jet streams. These are generally located in the lowest 1-2 km of the troposphere and are strongly influenced by local factors such as orography, friction and the diurnal cycle of heating. Some of the locations favourable to the generation of low-level jet streams are slopes of mountains parallel to the anticyclonic flow around the subtropical anticyclones, north south oriented coasts near cross-equatorial flow and narrow mountain gaps.

Although the low-level jets are weak and limited in spatial extent, they constitute an important part of the regional circulation owing to their role in large-scale moisture and momentum transport. Findlater (1966a and b) discovered the cross-equatorial East African low-level jet (also known as low-level cross-equatorial jet, Findlater's jet or Somali jet) from the analysis of aircraft and pilot balloon observations, and it has subsequently been recognized as a major feature of the lower-tropospheric circulation over the western Indian Ocean during the Indian summer monsoon season. This jet constitutes the strongest cross-equatorial flow in the lower troposphere at any level and forms part of a major low-level air current in the monsoon system flowing from the vicinity of Mauritius, through Kenya, Somalia and across the Arabian Sea to the west coast of the Indian Peninsula and further eastwards. It is estimated that this jet stream, situated near the western periphery of the monsoon regime, could account for about 50% of the total cross-equatorial transport of air in the lower troposphere in the month of July.

During April, the Low-Level Jet (LLJ) flows across Malagasy, and as the season advances it gradually penetrates into eastern African and swings across the equator into the Arabian Sea towards the west coast of India by July (Fig. 4.6 (a); the jet stream attains its maximum intensity during the period June to August.



(a) (b)

Fig. 4.6: The Low Level Jet Stream over Indian Ocean (a) Mean Monthly Position and (b) Mean Monthly Velocity during July (Findlater (1971))

Many of its characteristics of the LLJ closely follow those of the surface circulation, such as speed maxima around 10°S off the northern tip of Madagascar and over the western Arabian Sea, as well as an apparent branching of the flow over the eastern Arabian Sea (Fig.4.6(b)). Its seasonal evolution is also closely linked to the seasonal changes in the surface circulation.

Over the Gulf of Aden, the Southwest Monsoon usually sets in towards the end of May or early in June, shortly after it has become fully established over the western Arabian Sea. Once established, conditions persist throughout June, July and August. Near the eastern entrance, SSW winds prevail and the wind speed increases very rapidly as the entrance is approached from the west. In July, typical conditions are: 11-16 kt over the Gulf and eastward to about 52^0E, becoming 22-27 kt in the area of 52-54^0E, and further increasing to 28-33 kt in the vicinity of 56-60^0E. A marked increase in wave and swell heights are also experienced as one passes eastward out of the Gulf of Aden into the western Arabian Sea.

The Low Level Jet known as ‘‘Findlater Jet’’ or "Somali Jet" is thus a relatively narrow wind stream along the East African Coast off the coast of Somalia over the

Indian Ocean and peninsular India during summer monsoon season (Joseph and Raman (1966), Findlater (1969 a and b)) and is part of the larger Southwest Monsoon circulation pattern. The Somali Jet is one of the strongest and most sustained low-level wind systems on earth. It is normally strongest in July and August when core maximum speeds up to 100 kt have been observed. The core is usually centered at an elevation of about 5000 ft. There are generally three local speed maxima, north of Madagascar, off the coast of Kenya, and to the east of Socotra Island. These are semi- permanent low-level wind features during the Southwest Monsoon.

The Low Level Jet is the northern branch of a cross-equatorial flow, giving rise to a major supply of moisture in support of the Asian summer monsoon. The LLJ has two main functions. It is a channel carrying the moisture generated by the trade winds over the south Indian Ocean and the evaporative flux from the Arabian Sea to the areas of monsoon rainfall production over south Asia. In addition, the cyclonic vorticity north of the LLJ axis in the atmospheric boundary layer is a dynamic forcing for the generation of vertical upward air motion and rainfall and for the genesis of monsoon depressions over the North Bay of Bengal.

The LLJ is best manifested at 850 hPa with a core speed of 20-30 m/s which are occasionally found to reach a maximum core speed of 50 m/s near Madagascar and off the Somali coast. Strong horizontal and vertical shears characterizes the LLJ. It emanates from the Mascarene High, crosses the equator, runs close to east African high lands and then turns to the Arabian Sea and extends further eastwards crossing India. The axis of the LLJ is observed to be quite stable over the western Indian Ocean. However, it is subjected to north-south oscillations over India and eastern Indian Ocean. The displacements of convective heating zones are responsible for the latitudinal oscillations of the LLJ and the associated active/ break spells in the monsoon circulation and rainfall (Sikka and Gadgil (1980)). During the phase of active monsoon, the LLJ passes through the peninsular India and this region receives higher rainfall as compared to the rest of India. However, during the break monsoon phase, the LLJ shows a branching out over eastern Arabian Sea; one heading towards north-east (into the monsoon trough) and the other turning towards south-east, into another convective zone formed in the equatorial region

(Joseph and Simon, 2005). A study by Joseph and Sijikumar (2004) has shown that during the active monsoon, the core of the LLJ passes eastward through peninsular India around latitude 15^0N. During the break monsoon, the LLJ moves south eastwards from the central Arabian sea and by-passing India passes eastward close to Sri Lanka equator to 10^0N. (Fig.4.7). There is often seen at this time a weaker LLJ axis through north India around latitude 25^0N.

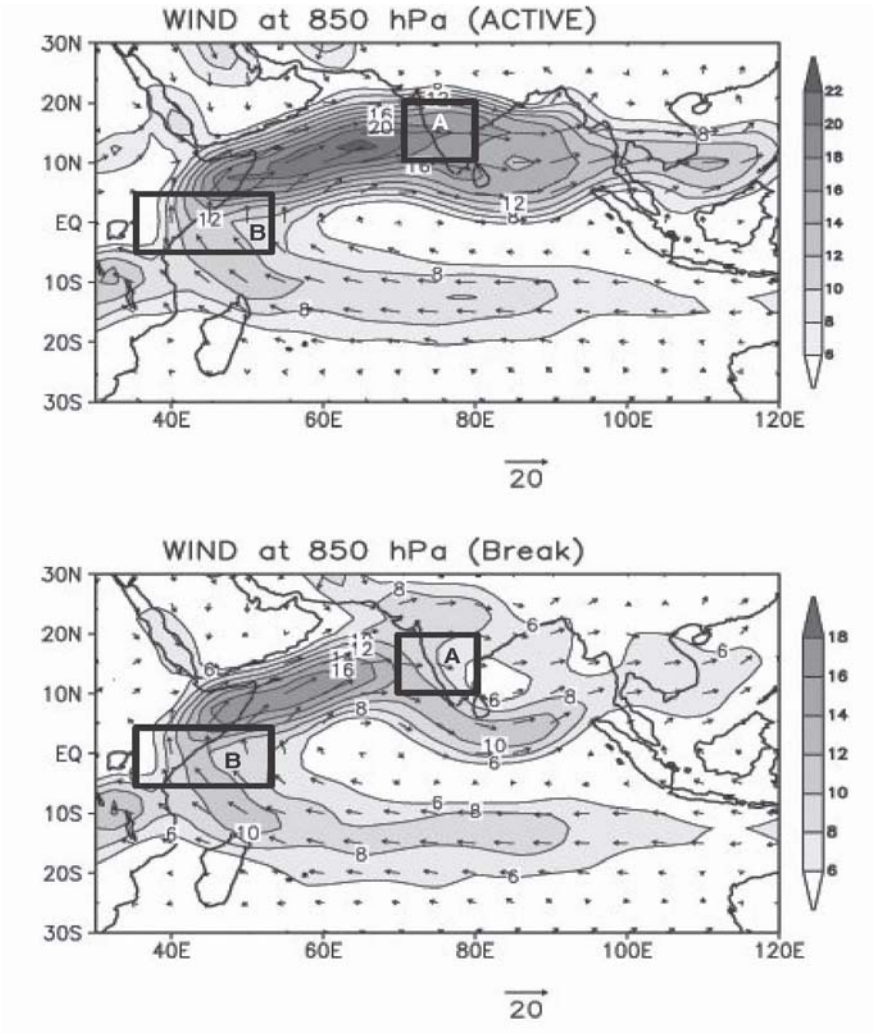


Fig. 4.7: Composites for active and break monsoon days during June to August of 1979 to 1990 extracted from Joseph and Sijikumar (2004). Top figure gives the wind flow at 850hPa in m/s for active spells and bottom figure the wind flow at 850hPa for break spells.

A study by Halpern and Woiceshyn (2000) has shown that, the interannual variations of the Somali Jet in the Arabian Sea during 1988–99 were linked to El Nino and La Nina episodes and to India west coast rainfall. The average date of

Somali Jet onset was two days later in El Nino events in comparison with La Nina conditions. The monthly mean strength of the Somali Jet was 0.4 m s^{-1} weaker during El Nino episodes than during La Nina intervals. When the monthly mean intensity of the Somali Jet was above (below) normal, there was an excess (deficit) of rainfall along the Indian west coast of India.

As the core of the LLJ is located within the planetary boundary layer, where the atmosphere-ocean coupling takes place, the LLJ induced variability should be seen in the surface winds.

In association with the LLJ, strong winds of the order of 30 m/s at 1 km above mean sea level are observed over the southwestern Arabian Sea during summer monsoon months. During the course of jet development, wind stress causes oceanic upwelling along the East African coast, which, in turn, influences the atmospheric thickness pattern in the lower atmosphere and thus the jet itself (Saha, 1974). The aerological structure of the jet indicates cool and moist air in the lower levels on the east, the warm and dry air at higher levels on the west. The level of maximum wind slopes upward to the east. The jet core is predominantly cloudy, with an axis of maximum cloudiness over land and to the west (Findlater, 1977).

When the Somali Jet intensifies, the Southwest Monsoon flow over the Arabian Sea (and clouds/rain over western India) intensifies 1 to 2 days later. If the maximum surface pressure gradient over the Arabian Sea area occurs north of 23^0N (over land), the monsoon flow over the Arabian Sea weakens. If the maximum gradient is over water (13 to 21^0 N.), then the monsoon activity over the Arabian Sea strengthens.

4.5. Tibetan HIGH

A remarkable aspect of the large-scale circulation during the summer monsoon season over South Asia is the upper-tropospheric anticyclone situated over the Tibetan plateau. The Tibetan plateau gives heat to the atmosphere, with the maximum in late spring and early summer. The inter-annual variability of the strength

of this heat source considerably influences the monsoon activity over a large part of South Asia.

The Tibetan plateau, located more than 4500 m above sea level with a length of about 2000 km and width of about 600 km in the west and about 1000 km in the east, is considered to be one of the key factors in the development of monsoon circulation in the region. The atmospheric pressure on the surface of the plateau varies between 700 and 500 hPa. The Tibetan plateau exerts its influence as a mechanical barrier in the atmospheric flow as well as a high-level heat source (Murakami (1987)). An anticyclone appears in the upper troposphere over Tibet during the Indian summer monsoon season, primarily due to latent and sensible heating over the plateau.

The Tibetan anticyclone is thus a warm high located over the Tibetan plateau in the middle or upper troposphere during the monsoon season and having the highest amplitude near 200 hPa (Fig. 4.8).

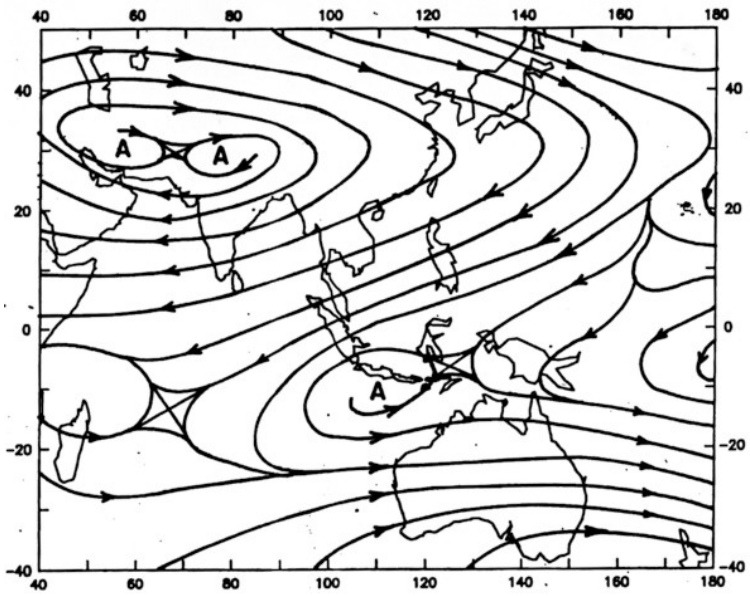


Fig. 4.8: The mean position of the Tibetan Anticyclone at 200 hPa (Pant and Rupa Kumar (1997)).

Generally, an anticyclone appears in the middle and upper troposphere, over Southeast Asia in May, and then moves northwestward, reaching the Tibetan plateau during the summer monsoon season. From about September, the

anticyclone migrates southeastward again towards Indonesia and loses definition after about October.

In July, at about 700 hPa and aloft, a ridge lies over Pakistan and northwest India to the west of about 75°E, with its axis along 30°N. Another high appears to the east of 80°E at 500 hPa with axis near about 28°N. According to Ramage and Raman (1972) and Chin and Lai (1974), this high has its centre at 28°N, 98°E, distinct from the Pacific High at 140°E. At this level, the high covers the Tibetan Plateau while the centre of the high is at its eastern periphery. It is more marked at 300 hPa, and has its extent between 70°E and 110°E, with centre near 30°E, 90°E, while the Pacific High has weakened very much and is centred near 120°E. At 200 hPa (Fig. 4.9), the only centre is at 30°N, 88°E and the high extends from 78°E to at least 140°E.

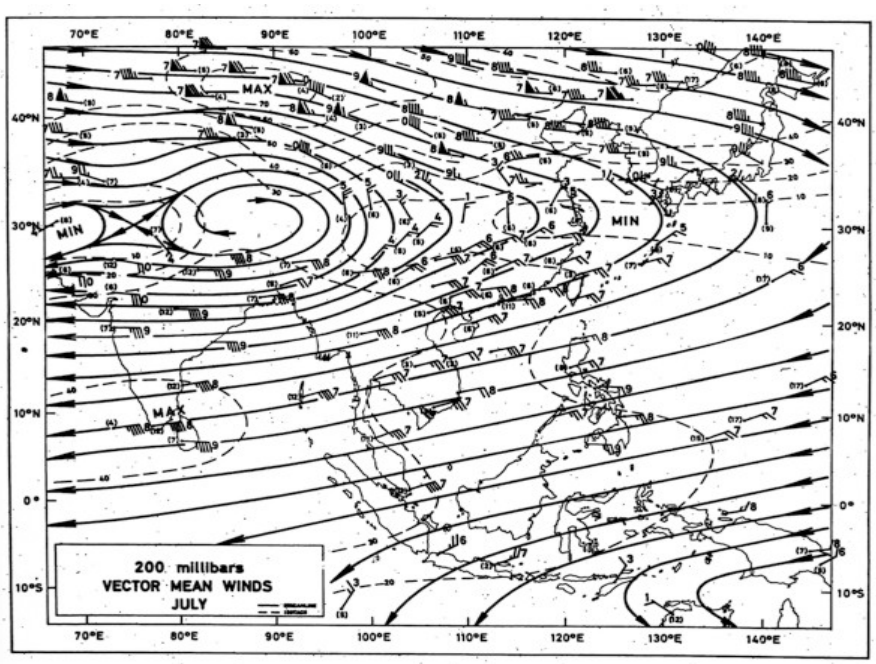


Fig.4.9: Mean Vector Winds at 200 hPa during July over SE Asia (Chin and Lai (1974)).

There is only a broad high-pressure belt at 100 hPa from even 30°E to 150°E with its axis along 35°N over Indian region. Interestingly, there is also a ridge at 700 hPa with its axis along 40°N, between 75°E and 95°E, just north of the Tibetan massif. The high over Tibet from 500 hPa upward centred near Tibet at 500 hPa and

over Tibet at 300 hPa and 200 hPa, is known as the 'Tibetan High'. Flohn (1950) first described the formation of this permanent warm anticyclone. In June, the axis of the anticyclonic belt is at about 25°N, at 300 and 200 hPa, and near 30°N only at 100 hPa, at the southern periphery of Tibet. August is similar to July, but the anticyclone is a little more to the north and slightly more intense. In September, the anticyclonic belt is near about 26°N up to 200 hPa and 30°N at 100 hPa. Thus, in all these months, the 100 hPa position is at 30°N or further pole ward but extending well outside the limits of the Tibetan Plateau. Between 500 hPa and 200 hPa, the high-pressure belt is well to the south of Tibet in June and September but over Tibet in the other two months in between. The anticyclonic shear north of Tibetan High is higher than the shear on the equator ward side of the high (Krishnamurthy, 1971). The issue of whether the heating of the plateau by the incoming solar radiation is responsible for the development of anticyclonic cell over Tibet between 500 and 200 hPa has not been yet clearly been resolved. For development of a 'high', colder air should occur at some higher level rather than warmer air below. The core of the easterly jet is at least 15° south of the centre of the Tibetan High and easterlies are weaker nearer Tibet. Hence, the role of this system in causing higher speeds in the easterly jet is a debatable point. However, the heating over Tibet at mid-tropospheric level, accentuates the seasonal north south temperature gradient. The decrease in heights of 500 and 200 hPa surfaces from Tibet is more towards east in January and to the west in July. The radiation balance over the Himalayan ranges and Tibet is important for the dynamics of the Tibetan High.

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

The variations in the intensity and position of this high and its orientation are closely related to the monsoon circulation over South Asia. Ramaswamy (1965) points out that well–distributed rainfall over India is associated with well-pronounced and east-to-west oriented anticyclone over Tibet at 500 and 300 hPa levels, and a pronounced high index circulation over Siberia, Mongolia and north China. The

Tibetan ‘High’ may sometimes shift much to the west of its usual position. In such a situation, the monsoon may extend further westward into Pakistan and in extreme cases into north Iran, though such a westward position of the Tibetan ‘High’ would be against its having origin in the heating effect of the Tibetan Plateau. A southward shift of this high from its normal position as a result of the protrusions of the mid-latitude westerlies is seen accompanied with reduction in monsoon activity over India and neighbourhood.

It has also been observed that the shifting of the Tibetan anticyclone eastwards to southern China and northern Myanmar allows extra-tropical westerlies to penetrate into the monsoon regime of South Asia. Such situations, leading to increased meridional flow, usually result in the so-called `break monsoon’ conditions with an abrupt weakening of the monsoon activity over most of India. Krishnamurti and Bhalme (1976) suggest that warm hydrostatic tropospheric columns accompany the combination of the Tibetan high in the upper troposphere and the monsoon trough at sea level over north India and the foothills of the mountains.

At the time of the onset of monsoon over Kerala, Tibetan ‘High’ is not established. The position of the sub-tropical high is to the south and even the mid-latitude westerlies may be prevailing in northern India in upper troposphere.

Krishnamurthy (1971) regards planetary scale ascent and divergence of air in the ridges can account for generation of negative relative vorticity and maintenance of anticyclonic systems.

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

The role of the Himalaya and Tibetan plateau mountain systems in the generation and maintenance of the South Asian monsoon is demonstrated in numerical simulation models (Godbole, 1973; Hahn and Manabe, 1975), in which the monsoon circulation is practically absent when the mountain influence is removed.

4.6. Tropical Easterly Jet

The Tropical Easterly Jet (TEJ) is the meteorological term which refers to an upper level easterly wind that starts in late June and continues until early September. It is an important upper tropospheric feature of summer monsoon circulation over south Asia, indicative of horizontal temperature gradient in the troposphere. The north-south movement as well as the structure of the Tropical Easterly Jet is observed to be closely linked to the monsoon activity. This strong flow of air that develops in the upper atmosphere during the Asian monsoon is centred around 15°N, 50-80°E and extends from South-East Asia to Africa. The strongest development of the jet is at about 15 km above the earth's surface with wind speeds of exceeding 40 m/s over the Indian Ocean.

The Tropical Easterly Jet appears as a band of strong easterlies extending from Southeast Asia across the Indian Ocean and Africa to the Atlantic, generally at a height of about 14 km. Koteswaram (1958a and 1958b) studied in detail the easterly jet stream over India. Over the region to the south of the sub-tropical ridge over Asia, the easterly flow concentrates into jet stream centred near about the latitude of Chennai at 100 hPa in July. The jet stream runs from the east coast of Vietnam to the west coast of Africa. The location of the jet stream over Africa is at 10°N. In general, the jet is at an accelerating stage from the South China Sea to south India and it decelerates thereafter. Consequent upper divergence is regarded as favourable for convection upstream of 70°E and subsidence downstream. On the anticyclonic side of the easterly jet stream, constancy of absolute angular momentum is observed, quite similar to the anticyclonic side of the westerly jet stream.

From the development to the decay, the jet axis remains close to about 14°N with core speeds exceeding 40 m/s (Koteswaram, 1958 a and b). However, later analyses have indicated that it is generally situated around 9°N (Mokashi, 1974). The jet axis lies between levels of 200 and 100 hPa, sloping towards the north.

The development of the tropical easterly jet stream is related to the thermal wind pattern during the northern hemispheric summer. It owes its origin to the fact that the maximum heating and tropospheric thickness pattern in summer is located in the subtropics rather than near the equator, and is aided by the presence of land surfaces in the subtropics of Eurasia and Africa, especially the elevated heating surfaces of the Central Asian mountain massifs.

The evolution of the Tropical Easterly Jet stream proceeds concurrently with the build-up and decay of the upper-tropospheric ridge in the subtropics (Hastenrath, 1991).

Along the South Asian longitudes, the presence of sea near the equator and the high Tibetan plateau near 40°N creates intense pressure and temperature gradients between the subtropical and equatorial latitudes in this region. The subtropics have high temperatures and the equatorial region has low temperatures, throughout the troposphere. As a result, there is low-pressure area in the subtropics and a high-pressure area in the near-equatorial latitudes in the lower troposphere, while the situation is reversed in the middle and upper troposphere. Due to the low value of the Coriolis parameter, even a slight gradient of pressure is able to create considerable zonal flow in these latitudes. This is one of the reasons for the easterly jet stream being particularly strong over South Asia (Asnani, (1993), (2005)).

The tropical easterly jet is present over the south Indian peninsula from June to August in latitudinal belts between 12^0 and 15^0N (Fig.4.10) and disappears by September. The core of the Tropical Easterly Jet is observed near Trivandrum in June and September and between Trivandrum and Chennai in July and August.

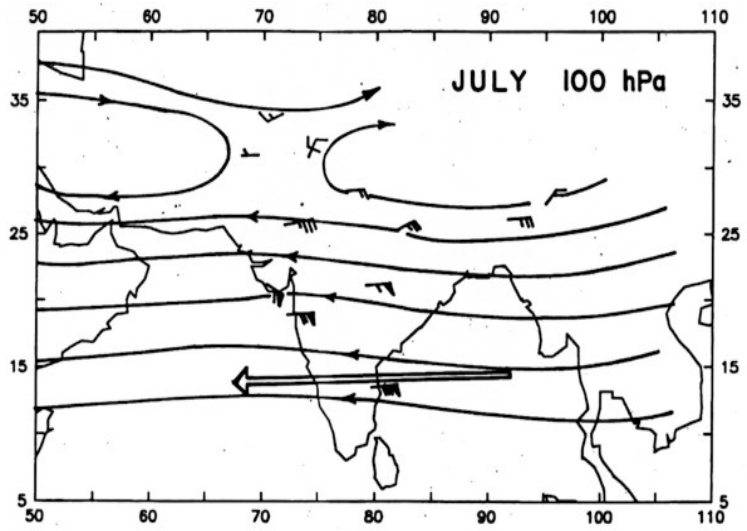


Fig. 4.10: Distribution of Mean Monthly upper winds at 100 hPa over South Asia (Rao and Ramamurty (1968)).

During the south west monsoon season over India, as the temperatures decrease to the south from 25^0N, easterlies strengthen over the Peninsula to 40 to 50 kt by 200 hPa. At 200 hPa, the temperature decrease southwards starts at 30^0N itself. At 140 and 100 hPa, the temperature gradient south of 20^0N is very flat but southward decrease persists to the north. This northward increase of temperature extends almost up to the Pole. This is a global feature. However, the gradient is steep between 20^0 and 50^0N. Due to the effect of the thermal wind, the easterlies increase with height up to about 150/100 hPa over the Peninsula at which level they are maximum.

Even at 100 hPa, the strong temperature gradient prevails over northern India. Hence easterly winds increase with height even above 100 hPa. The easterly jet stream over the Peninsula is, thus, an effect of the southward decrease of temperature over the region in the entire troposphere. This thermal gradient is effective in reversing the moderately strong westerlies of the lower troposphere and building speeds of 80 kt by 100 hPa. In fact, this is the effect of the sun's position north of 20^0N at this time of the year. The heating of the Tibetan plateau and transfer of heat to free atmosphere may have relatively minor influence as no strong temperature gradient develops in its vicinity to south until the 200 hPa level or

higher. At 150 and 100 hPa, a strong decrease in temperature towards south is observed even in January on account of the polar and tropical tropopauses.

The horizontal shear at 100 hPa to the south of the TEJ between Chennai and Thiruvananthapuram is 0.09 hr^{-1}. To the north, between Chennai and Visakhapatnam, it may be of the order 0.07 hr^{-1}. The Easterly Jet seems to have a very sharp vertical profile. The vertical shear between 9 and 16 Km varies between 15 and 20 hr^{-1}, the maximum shear being about 5 Km below the core. Above the core, between 16 and 18 Km the shear is about 22 hr^{-1} (Rao and Ramamurty (1968)).

The position and speed of the jet have appreciable spatial and temporal fluctuations. The strongest portion of the jet stream is over Indian peninsular region, with a maximum speed of about 75m/s (Koteswaram, 1958 a and b). The rainfall distribution related to the tropical easterly jet system is indicative of the vertical motion patterns in the lower troposphere in its association. In the entrance region of the jet over Asia, abundant rainfall is found to the north of the jet axis, and in the exit region over West Africa reverse pattern is observed (Koteswaram 1958a and b).

Tropical Easterly Jet is generally weaker during the summers of warm events in Southern Oscillation when anomalously warm waters appear over the central and eastern Pacific and drought occurs over the Indian sub-continent. It is observed that divergence (convergence) exists on the upstream (downstream) side of TEJ. The tropical divergent circulations (east-west Walker and local Hadley circulations) during such summers are weakened and shifted eastward. As such, divergence anomalies appear in the upper troposphere over equatorial Africa or the east coast of Africa, while convergence anomalies exist over the Indian sub-continent or the Arabian Sea. These changes of the tropical divergent circulations may cause the change in the energetics of TEJ. The kinetic energy generation and destruction associated with TEJ are less in during dry summers (Chen and Loon (1987)).

By analyzing 12-year (1979-1990) 200 hPa wind data from National Centers for Environmental Prediction-National Center for Atmospheric Research reanalysis, Sathiyamoorthy, et. al. (2007) has demonstrated that the intraseasonal time scale (30-60 days) variability of the Tropical Easterly Jet (TEJ) reported in individual case studies occurs during most years. In the entrance region (east of ~70° E), axis of the TEJ at 200 hPa is found along the near equatorial latitudes during monsoon onset/monsoon revivals and propagates northward as the monsoon advances over India. This axis is found along ~5^0 N and ~15^0N during active monsoon and break monsoon conditions, respectively. The examination of the European Centre for Medium Range Weather Forecasts reanalysis wind data also confirms the northward propagation of the TEJ on intraseasonal time scales.

During the intraseasonal northward propagations, axis of the TEJ is found about 10^0-15^0 latitudes south of the well-known intraseasonally northward propagating monsoon convective belts. Because of this 10^0-15^0 displacement, axis of the TEJ arrives over a location about two weeks after the arrival of the monsoon convection. The systematic shifting of the locations by convection, low level monsoon flow and TEJ in a collective way during different phases of the monsoon suggests that they all may be inter-related.

According to Rao, et. al. (2008), Tropical Easterly Jet of summer monsoon over the north Indian Ocean is weakening in recent years. The absolute easterly shear shows a strong negative correlation (significant at 99.9% level) with the number of severe storms suggesting that a decrease in easterly shear is favourable for the formation of more severe tropical storms. For the first time in recorded history, a category 5 Hurricane formed in June 2007 together with two more severe tropical storms over the north Indian Ocean. Thus, if the present decreasing trend of TEJ intensity continues there is a strong likelihood of the formation of tropical cyclones of hurricane intensity even during the summer monsoon. Presently, these intense systems are known to form only in the pre and post monsoon seasons, when the vertical wind shear is low.

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CHAPTER 5

CLIMATIC PATTERN

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5.1. Introduction

The 'Southwest Monsoon' is an enigma – a complex quasi periodic change in the tropical circulation. Every year after the scorching summer, winds emanating from high pressure cells over the southern Indian Ocean travel far northeastwards to enter into the Indian sub-continent where pressure is low, due to land heating and the ITCZ has migrated as the monsoon trough. The word 'Monsoon' has Arabic root and it refers to 'Season'. However, 'monsoon' is usually meant as the seasonal reversal of wind and associated changes in rainfall pattern. The southwesterly winds are usually quite strong in lower levels over the south / central Arabian Sea and adjoining peninsula with the maxima at the 850 hPa level, turns over the Bay of Bengal and enters as easterly / south easterly currents into northern/ northeastern region to the north of the monsoon trough. The monsoon air mass is maritime and moist up to a great depth .Temperatures in the troposphere decrease to the north in winter, while during the monsoon season, this gradient is reversed. The westerly jet stream, aloft, of the previous season is replaced by the easterly jet stream.

In this chapter the characteristic features of important climatological elements during the monsoon mainly over the Indian sub-continent and the adjoining Indian Seas, are described.

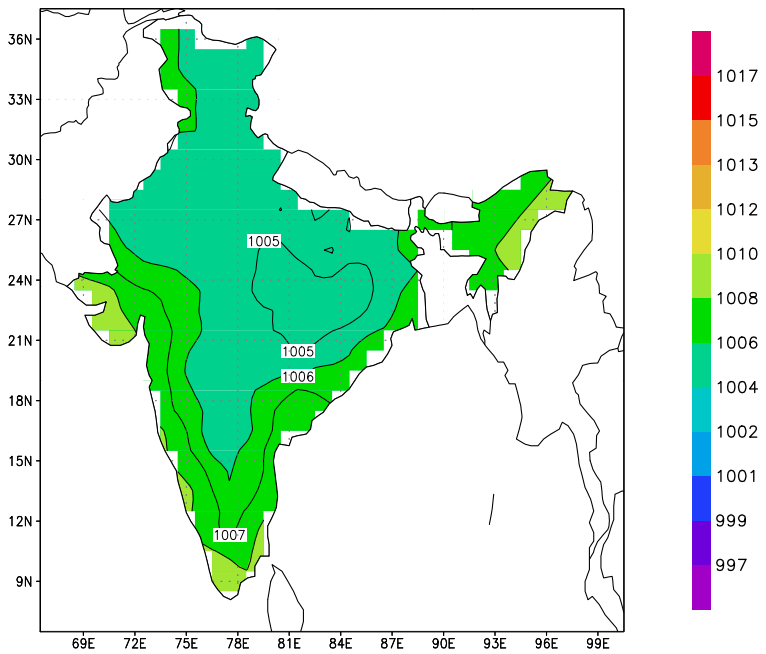
5.2. Sea–level Pressure Pattern

The chief features of the surface pressure distribution in the monsoon season are the heat low over Pakistan and adjoining extreme northwestern parts, the monsoon trough thence to the northwest Bay and the strong pressure gradient to the south.

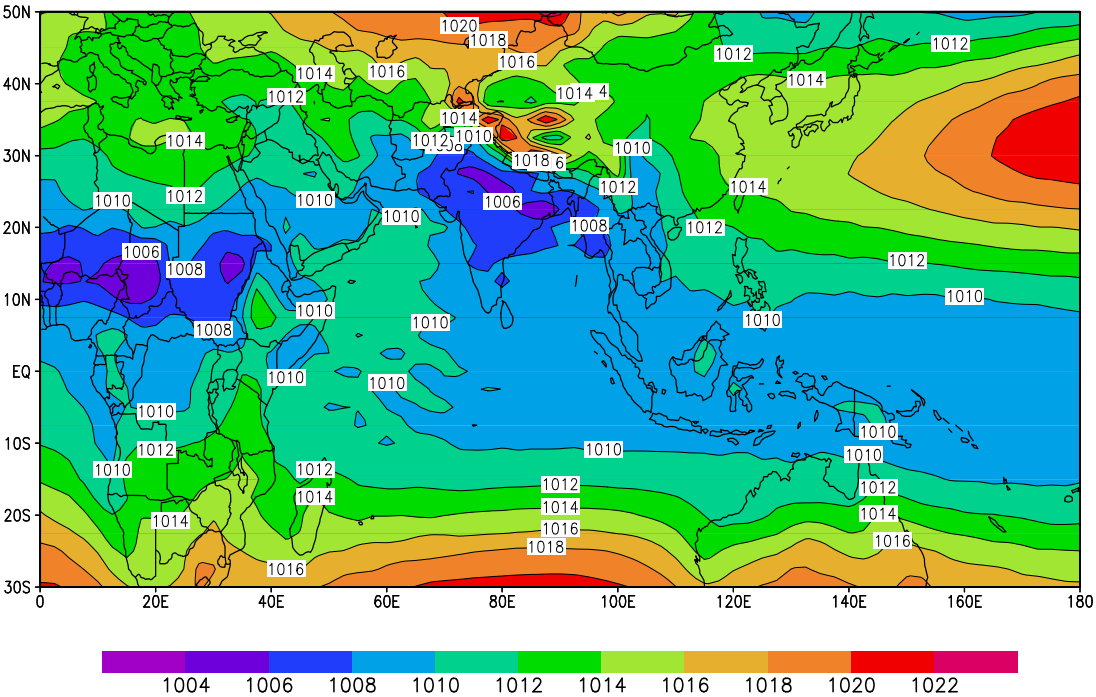
Due to heating, a thermal heat low starts appearing over the south Peninsular parts of India in March itself and it moves northwards (along the march of sun) in subsequent months.. In April, the heat low could be seen over east central India with sea level pressure around 1005 hPa (Fig 5.1a). Relatively higher-pressure values over the Arabian Sea and the Bay of Bengal are observed. Such heat lows are established globally and over North Africa a heat low could be also be along around 10° N (Fig 5.1b). During May, main centre of low pressure over India is near 24° N, 80° E and the low is more intense (1001 hPa, Fig 5.2a).

The heat low is more marked in June (997 hPa, Fig 5.3a) with the main centre over extreme northwestern parts of the country and adjoining Pakistan with a small secondary centre of low around east central parts. In July, which is the rainiest month, the low pressure area around the extreme northwestern parts is equally intense (Fig 5.3b). A trough lies over north India with axis from Sriganganagar to the Head Bay, which is termed as the 'monsoon trough'. Pressure gradient is generally more to south of this trough and sometimes in active monsoon conditions, more than six isobars could be seen between Thiruvanathapuram to Mumbai.

Intensity of low is substantially decreased in August (1000 hPa, Fig 5.4a) and September (1004 hPa, Fig 5.4b). Similarly, the pressure gradient south of monsoon trough, which is maximum in June and July and is slightly less in August, decreases significantly in September.



(a)



(b)

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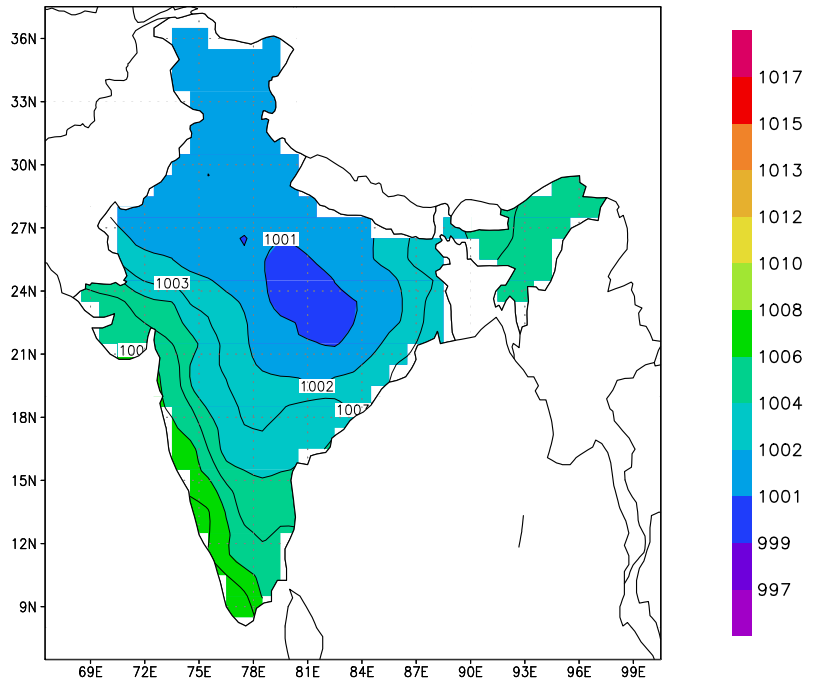
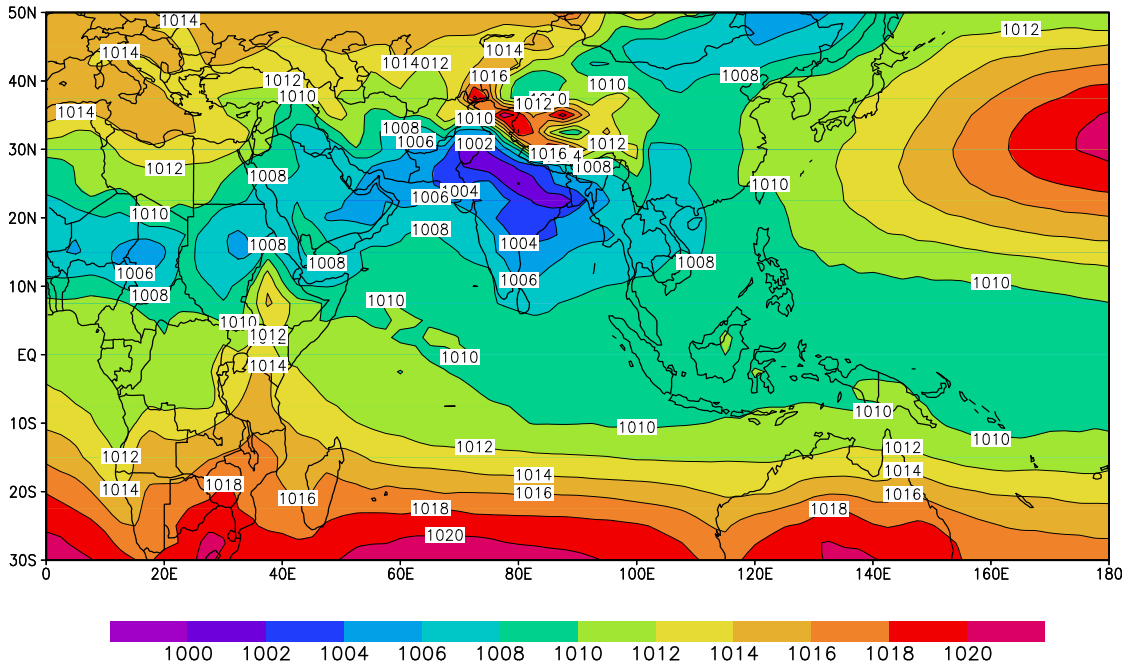
Fig. 5.1: Sea level pressure pattern over, a) India and b) eastern half of the Globe during April.



(a)



(b)

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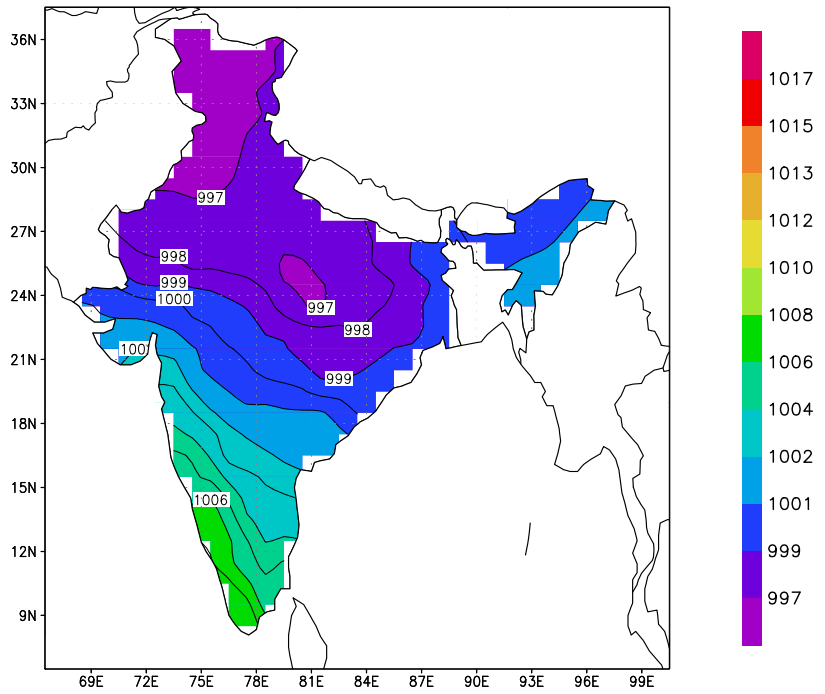
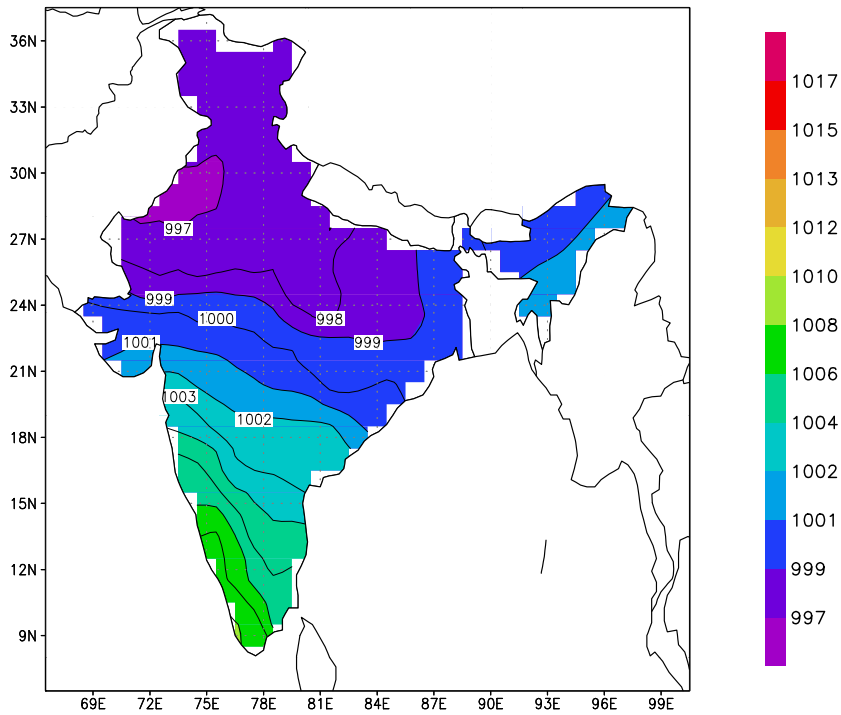
Fig. 5.2: Sea level pressure pattern over, a) India and b) eastern half of the Globe during May

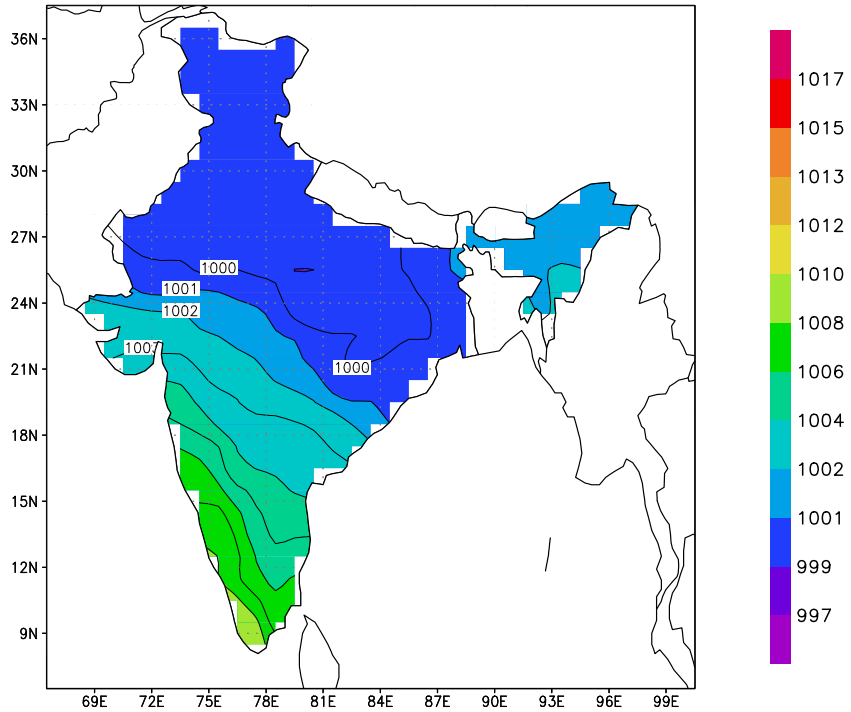


(a)

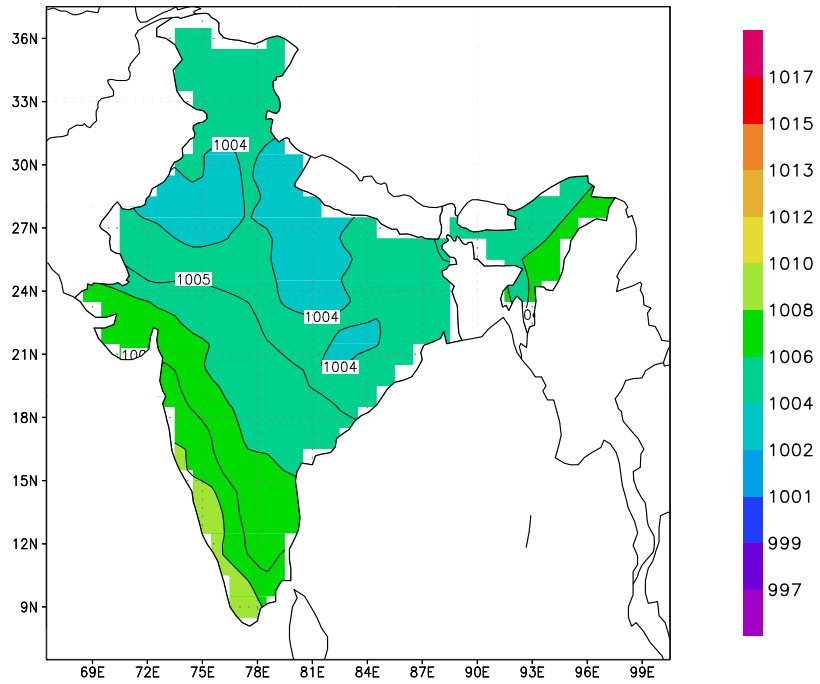


(b)

Fig. 5.3: Sea level pressure pattern over India during a) June and b) July.



(a)



(b)

Fig. 5.4: Sea level pressure pattern over India during a) August and b) September.

5.3. Surface and Upper Air Winds

In June, surface winds (Fig 5.5a) are from east, to the north of the trough line running through centre of the heat low to the North Bay of Bengal. Over west Bengal winds are mainly from south while over the rest of India, winds blow from west to southwest, more westerly off the west coast of the Peninsula and practically southwesterly in the Bay. Speeds usually range between 10 to 15 m/s in the Arabian Sea. In July, also, similar wind pattern with greater intensity and a slight southward shift in the monsoon trough may be seen (Fig 5.5b). Wind pattern in August (Fig 5.5c) is similar to July. By September (Fig 5.5d), there is a weakening of pressure gradient and winds weaken, particularly over the sea areas. Seasonal surface winds are shown in fig. 5.6.

Seasonal upper winds at 850 h Pa level (Fig. 5.7) are generally strong over the south Arabian Sea to the south peninsula (exceeding 15 m/s). However, strength of westerly/ southwesterly wind decreases with height. Fig. 5.8 shows vector wind at 500 hPa. At 200 hPa (Fig. 5.9) onwards, easterlies (of the order of 10 to 20 m/s) from the Tibetan anticyclone prevail throughout India.

An important feature of the upper winds is the easterly jet stream (wind speed generally of the order of 30 m/s) with its core at about 100 hPa located along 13^0N (Fig. 5.10). This is known as the Tropical Easterly Jet Stream (TEJ) and extends approximately from 40^0E to 100^0E across the Afro Asian region during the summer monsoon season.

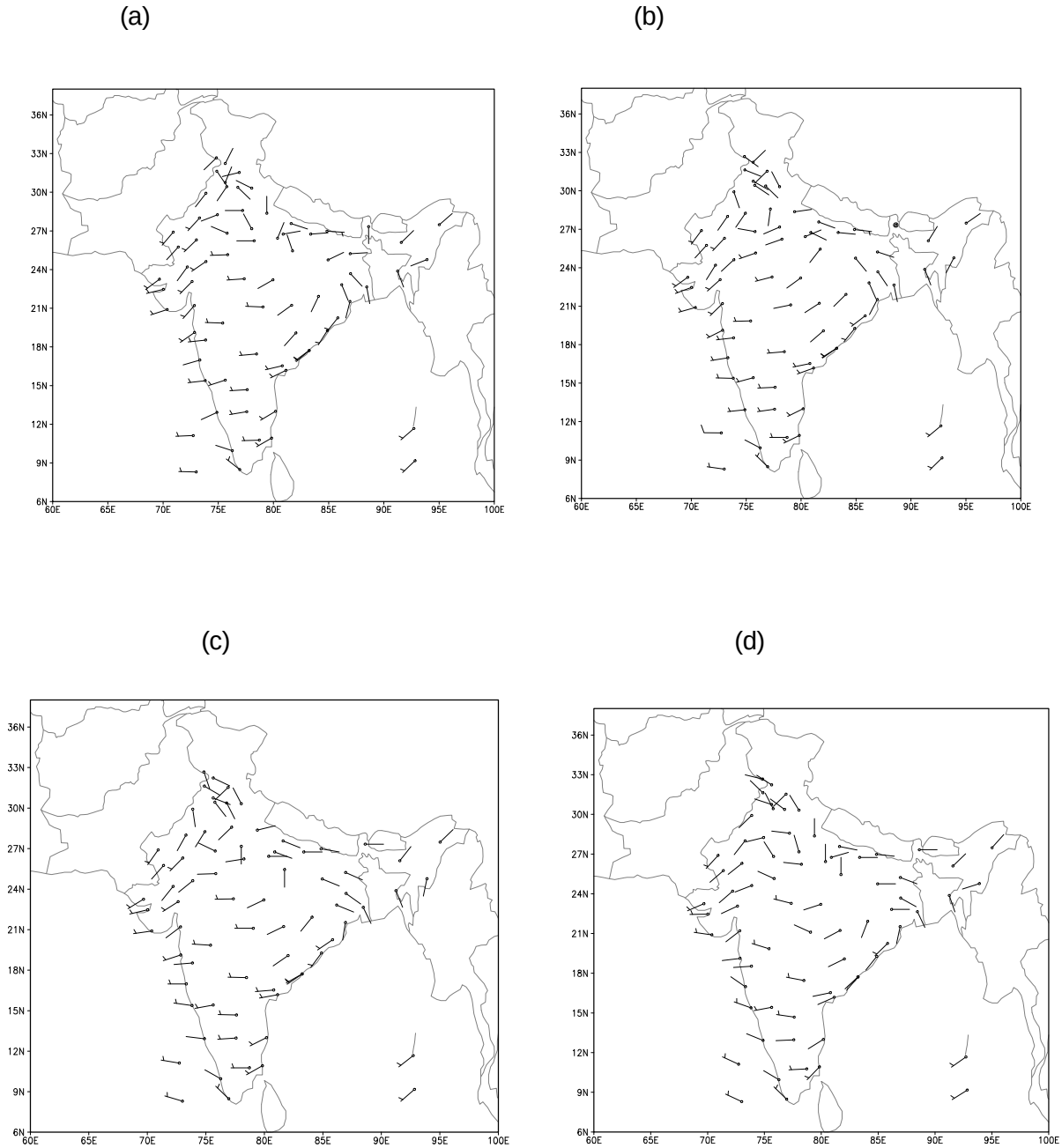
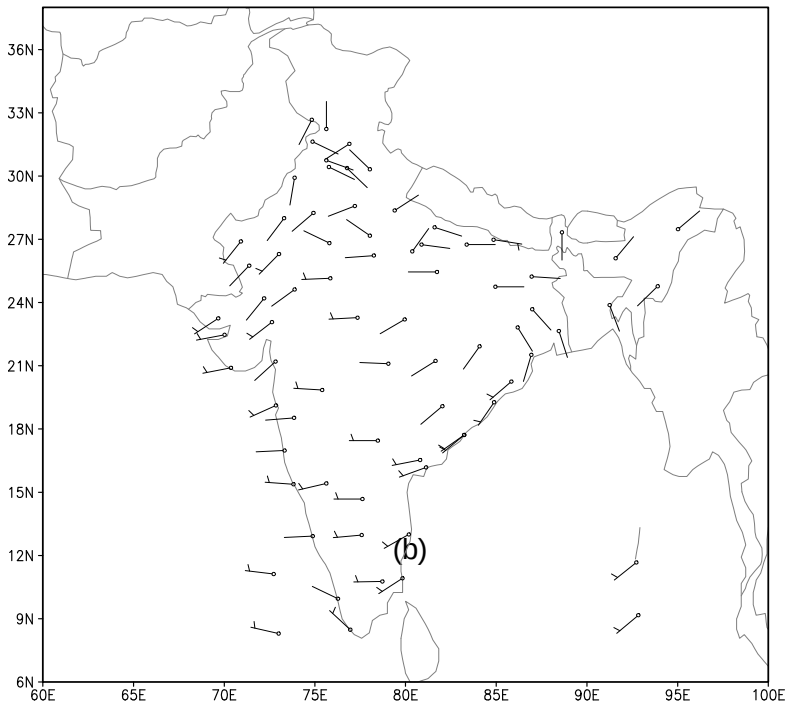


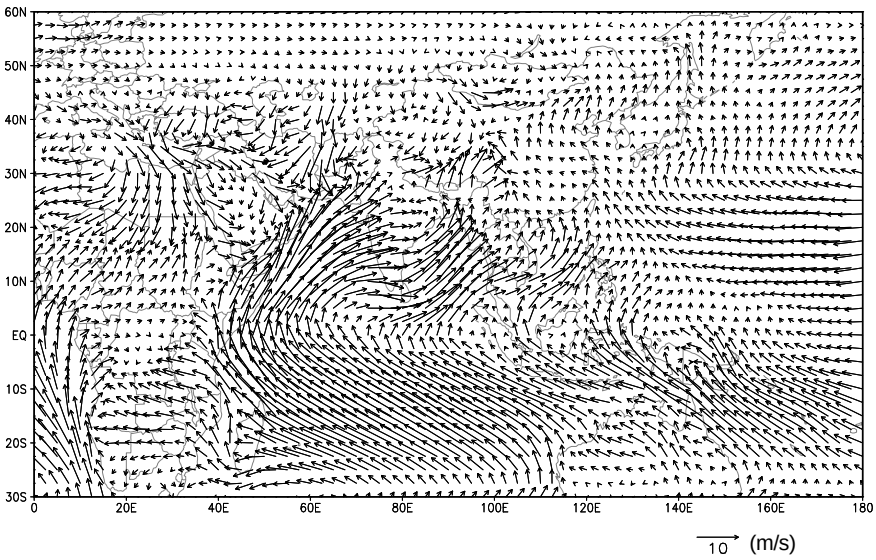
Fig.5.5: Surface wind (m/s) for a) June, b) July, c) August and d) September

(a)



(b)

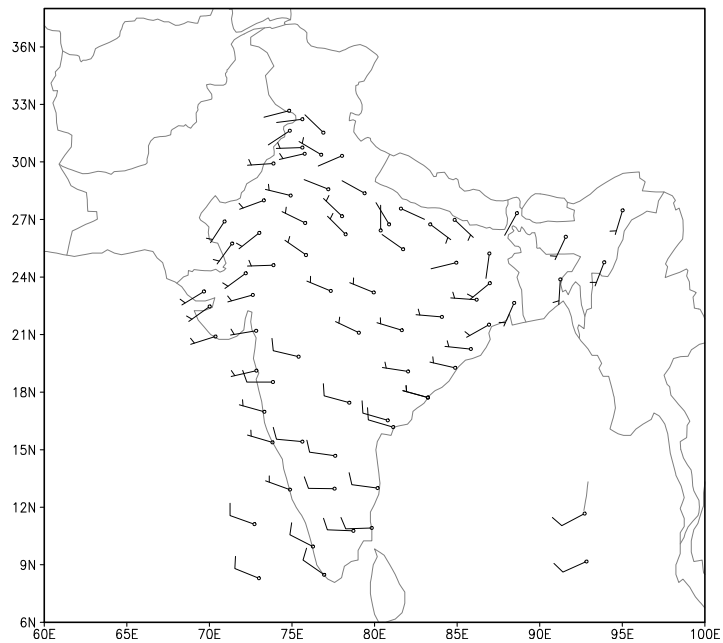
(b)



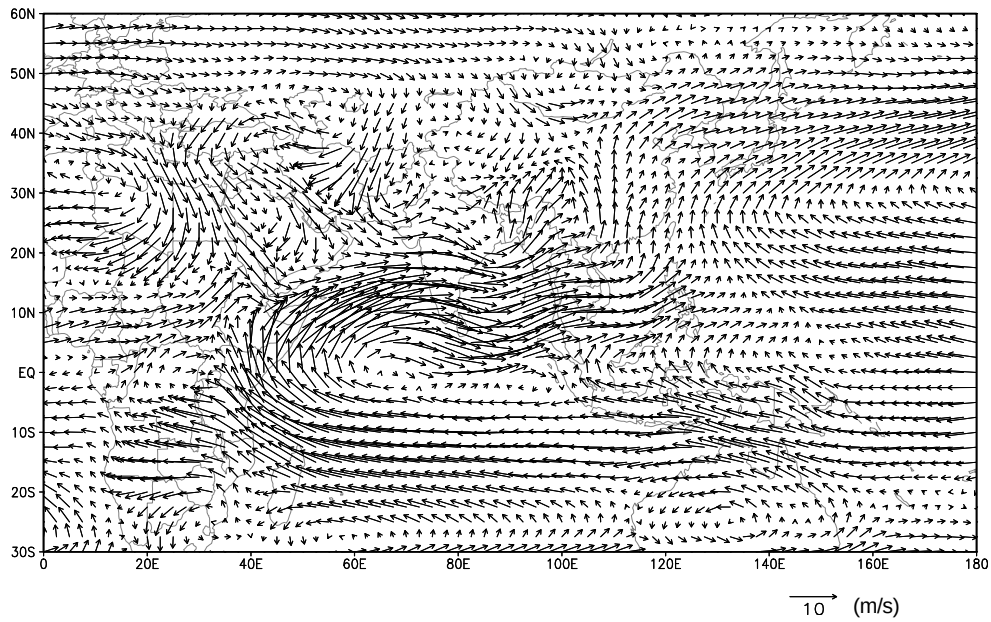
(Source: www.esrl.noaa.gov)

Fig.5.6: Seasonal (June- September) surface wind (m/s) over a) India and b) eastern half of the globe

(a)



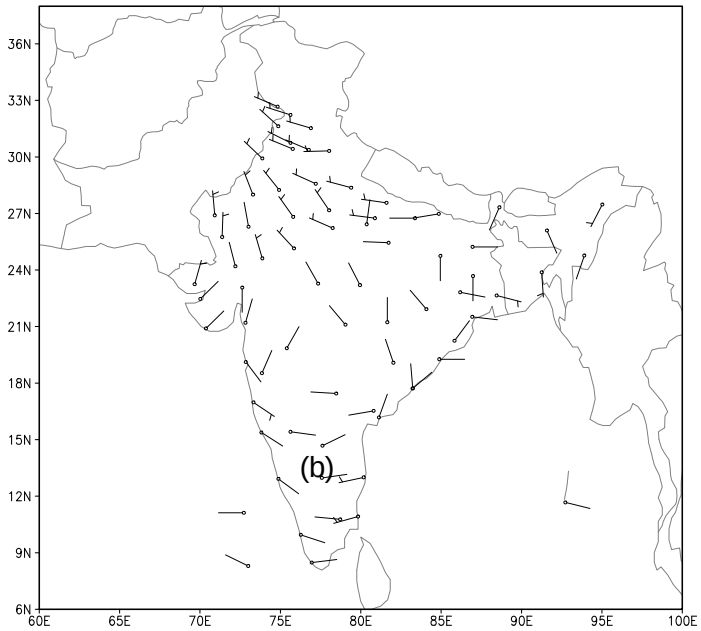
(b)



(Source: www.esrl.noaa.gov)

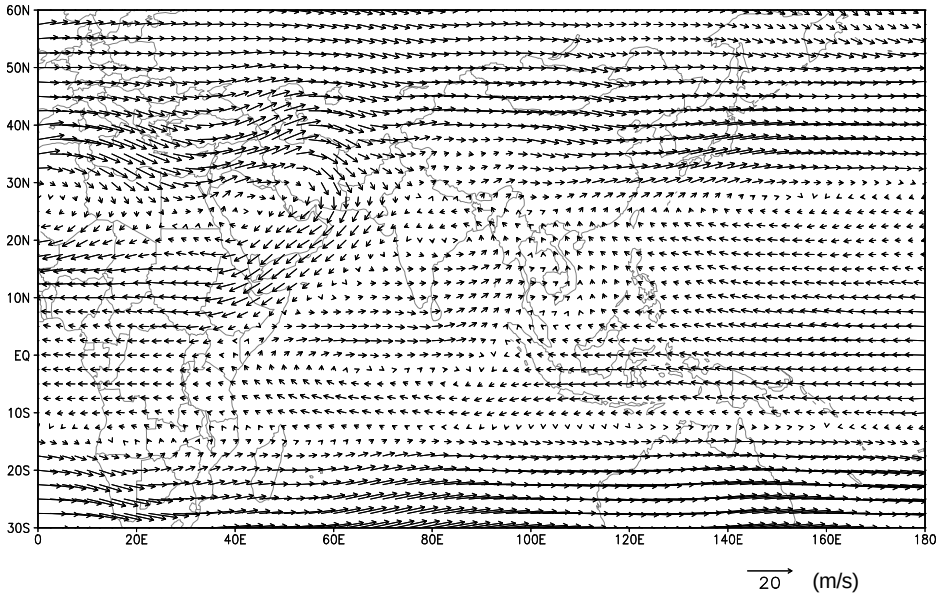
Fig. 5.7: Seasonal (June- September) 850 hPa wind (m/s) over a) India and b) eastern half of the globe.

(a)



(b)

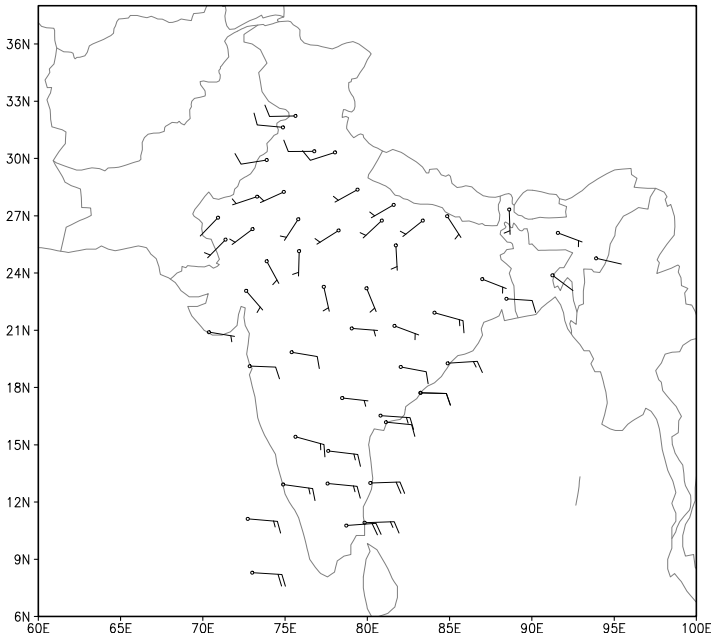
(b)



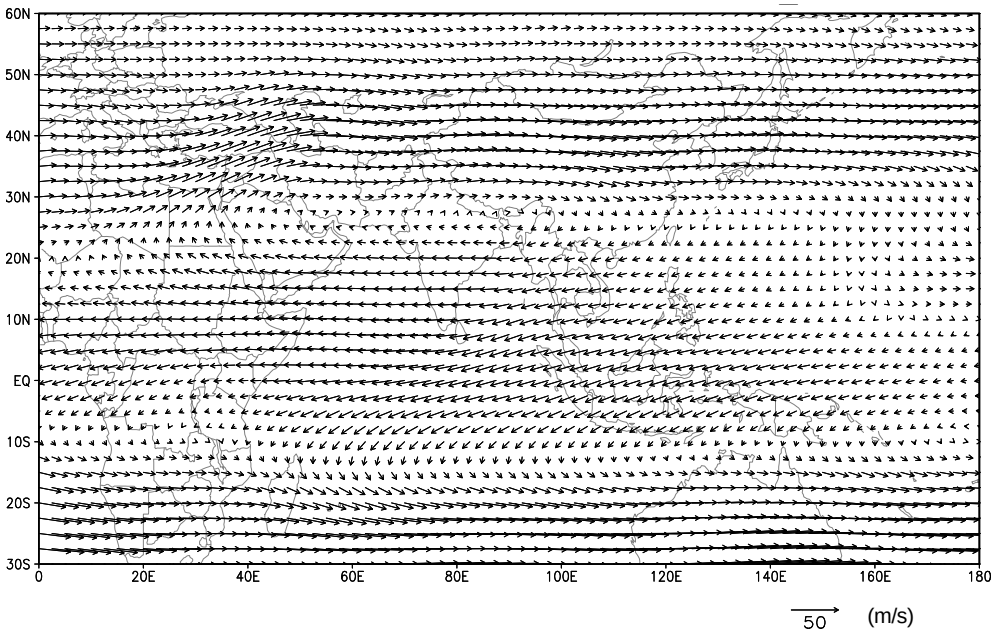
Source: (www.esrl.noaa.gov)

Fig. 5.8: Seasonal (June- September) 500 hPa wind (m/s) over a) India and b) eastern half of the globe.

(a)



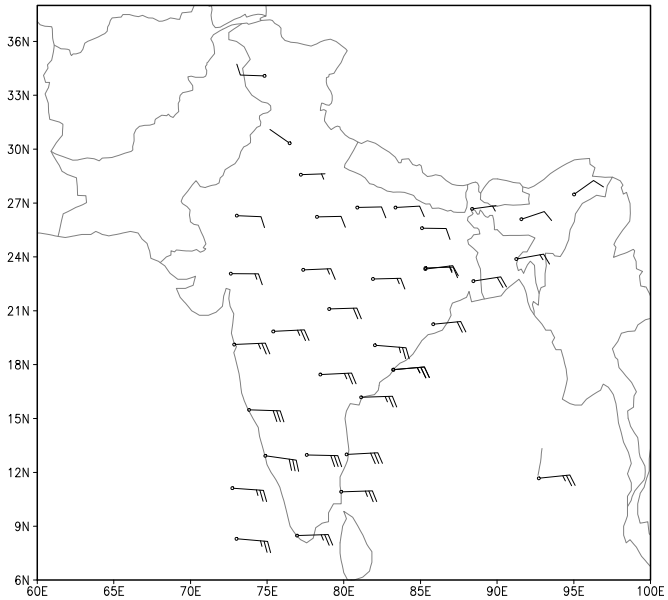
(b)



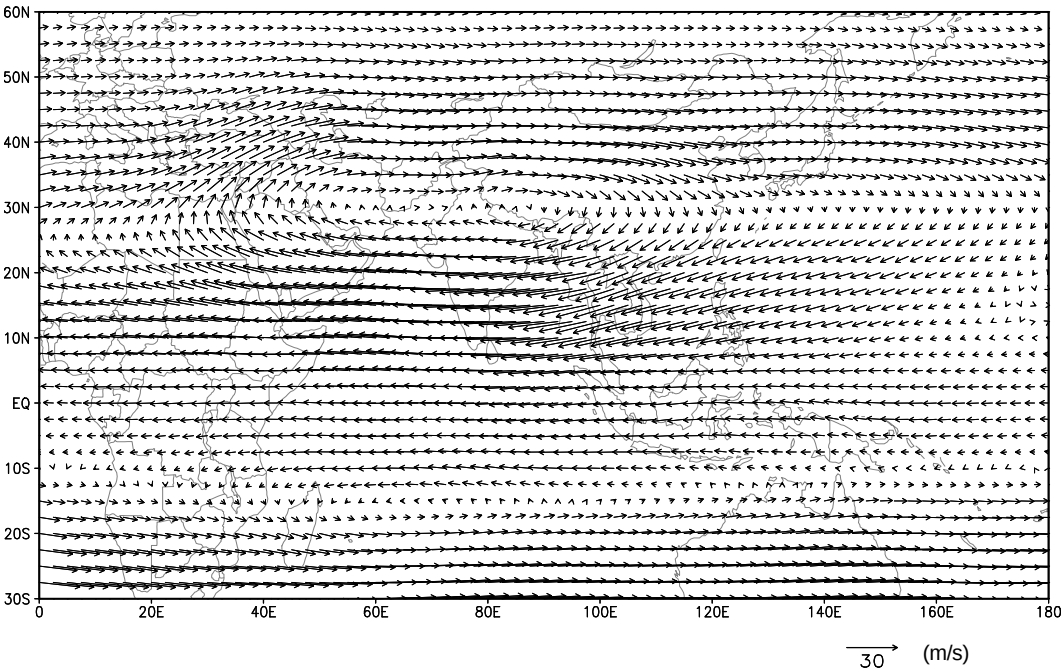
(source: www.esrl.noaa.gov)

Fig. 5.9: Seasonal (June- September) 200 hPa wind (m/s) over a) India and b) eastern half of the globe

(a)



(b)



(Source: www.esrl.noaa.gov)

Fig. 5.10: Seasonal (June- September) 100 hPa wind (m/s) over a) India and b) eastern half of the globe.

5.4. Rainfall

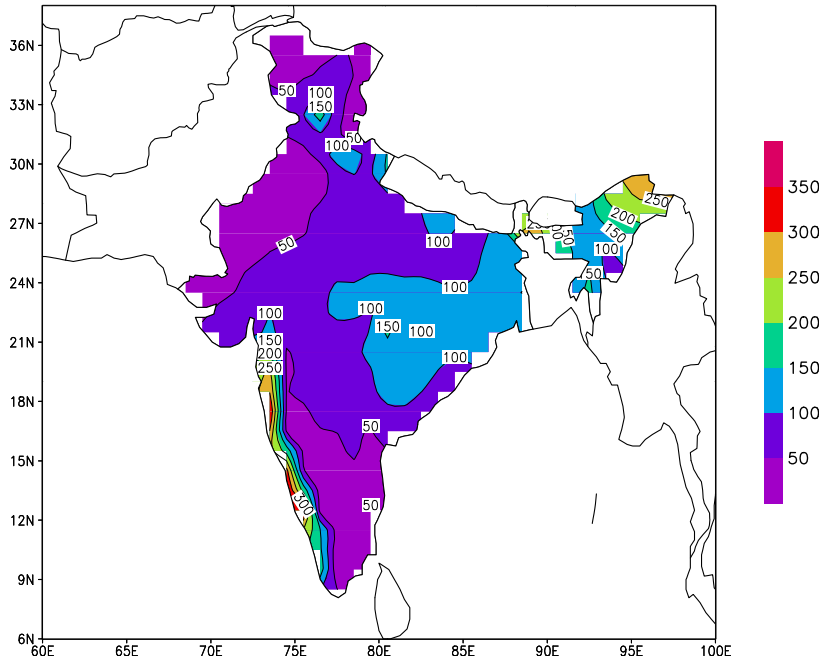
The rainfall distribution in the southwest monsoon period, lasting from June to September and its coefficient of variation exhibit remarkable spatial variation. The prevailing winds blow almost at right angles against the Western Ghats and the Khasi–Jaintia hills and, therefore, orographic influence is dominant in the distribution of rainfall in this season. Windward side of mountains receives very heavy rainfall and it decreases rapidly towards the leeward side. The west coast and the northeastern region receive maximum amount of rainfall the order of 250—300 cm. Rainfall over most parts of Rajasthan, southeast peninsula and Jammu & Kashmir is generally less than 50 cm (Fig. 5.11a).

The coefficient of variation of rainfall ranges from 60 per cent over western desert, to 30 per cent or less in the rainiest areas (Fig. 5.11b), This is mainly due to large variation in the average rainfall. In the southeastern parts of the peninsula, rain shadow regions of Maharashtra, Jammu & Kashmir, it is more than 100 per cent.

It may again be mentioned that the mean monthly rainfall amounts are not uniform during this period. Broadly, rainfall increase with the setting of the monsoon, reach a maximum in July and then decrease (Fig. 5.12). Coefficient of variation of rainfall of individual months is more than that for the whole season.

The summer monsoon rainfall shows large inter annual variability and intra seasonal variations. In some epochs the rainfall pattern changes from the normal pattern with decrease in rainfall over the west coast and central parts, together with rise in rainfall in the foothill regions in north and Tamil Nadu. This is termed as break in the monsoon and is associated with changes in pressure pattern and circulation as well. Prolonged breaks result in drought in the major parts of India coupled with, floods in the NE India.

(a)



(b)

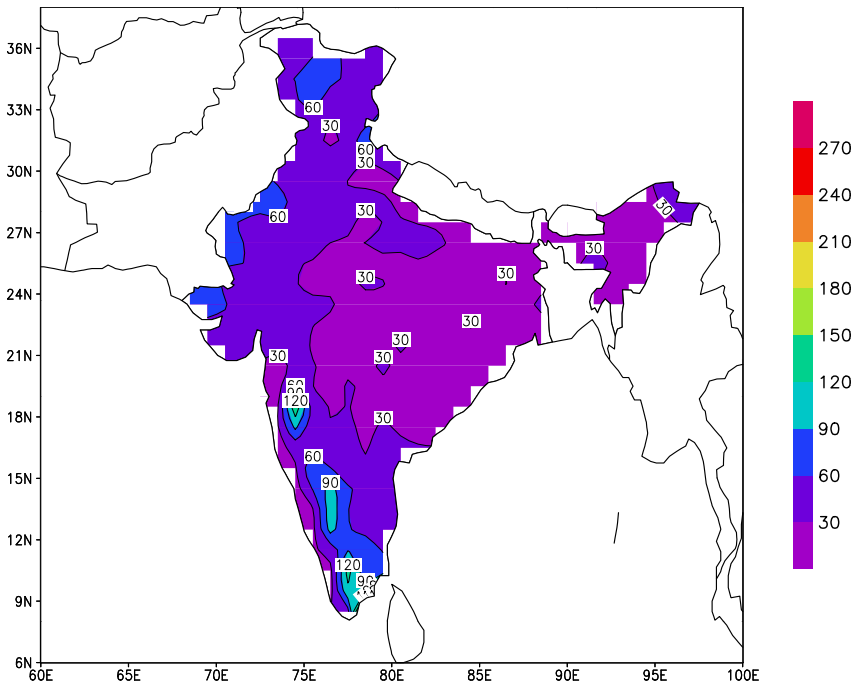


Fig. 5.11: Monsoon seasonal (a) rainfall and (b) coefficient of variation.

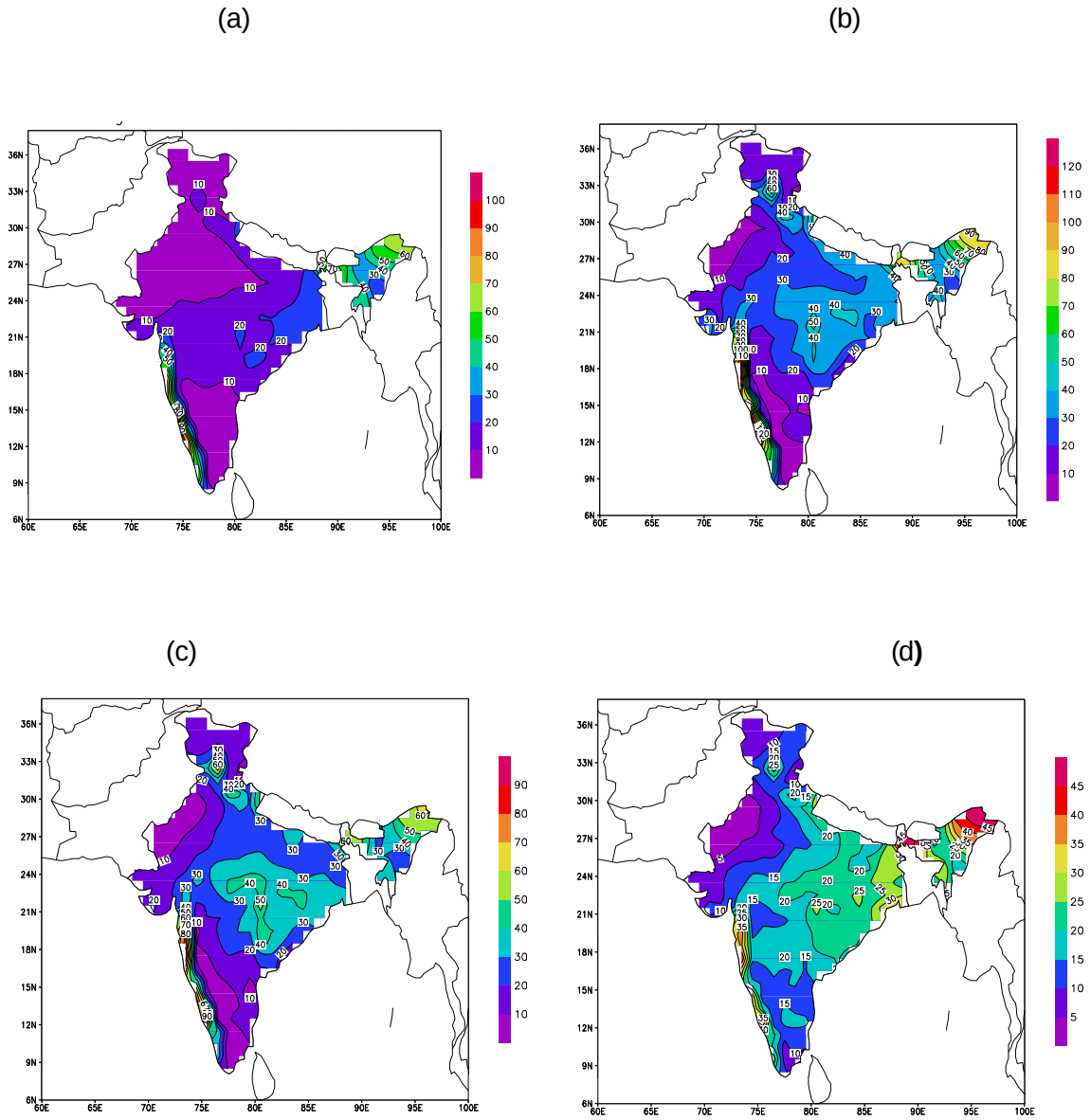


Fig. 5.12: Mean monthly rainfall for a) June, b) July c) August and d) September months.

5.5. Surface Temperatures

With the march of the sun in the northern hemisphere, land gets increasingly heated after January and by April (Fig. 5.13), mean surface temperatures are generally more than 30° C over most of the Peninsula and some parts of central India. Temperatures more than 32° C over some peninsular parts of the country may also be observed. Temperatures along the coasts are between 28° C and 30° C. However,

mean temperatures of the order of 16^0- 18^0C may be observed over the extreme northern parts of the country. In May also similar pattern in mean temperature (Fig. 5.14), with extension of high temperature zone towards northwestern region could be seen.

In June, the highest mean temperatures are confined to the northwestern region (west Rajasthan) and adjoining central parts of the country. Over southern parts of the west coast and northeastern region mean temperatures are relatively much less mainly due to persistent cloudiness. During the other monsoon months similar patterns could be observed, however highest mean temperature decreases from 34^0 C from June to less than 30^0 C in September. Fig. 5.15 shows mean monthly surface temperature for monsoon months.

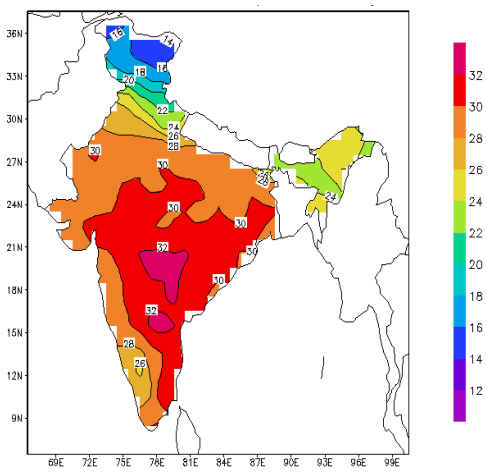


Fig. 5.13: Mean surface temperature during April

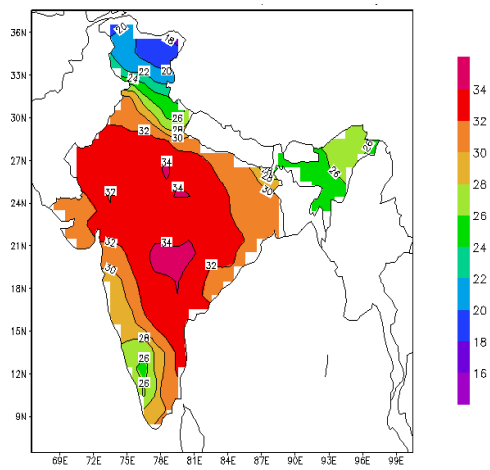


Fig. 5.14: Mean surface temperature during May

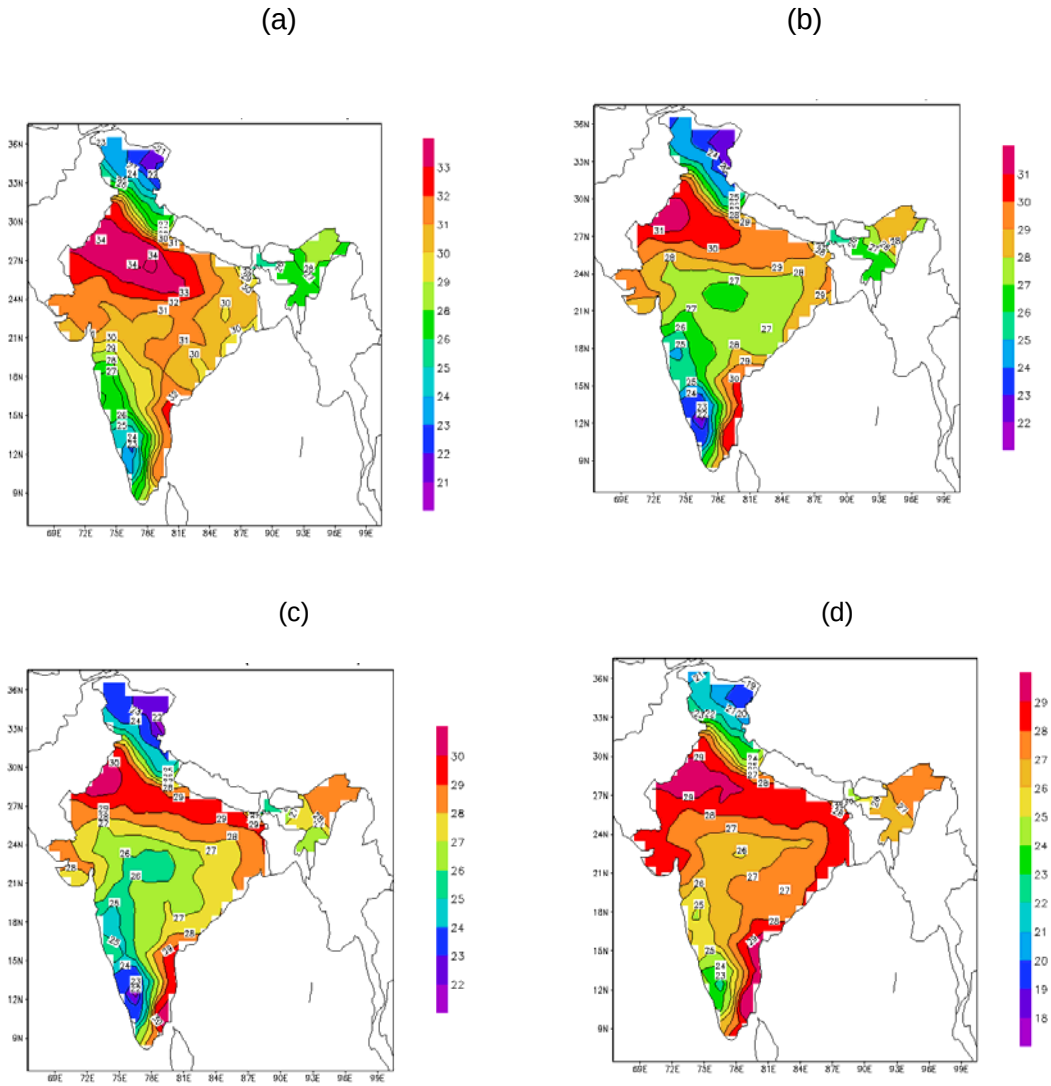


Fig. 5.15: Mean surface temperature during a) June, b) July, c) August and d) September.

5.5 Upper Air Temperatures

During the pre-monsoon months, at 850 hPa a thermal high with temperature more than 22^0 C develops over India with centre near north peninsula and neighbourhood. Temperature decreases in all directions from this centre. However, at 700 hpa, the thermal high (10^0 C) shifts southwards to the east peninsula. At 500 h Pa, a weak thermal high is over the south Peninsula with temperature around -06^0 C. At 200 hPa, the zone of maximum temperature shifts to the eastern and northeastern region with temperature around $-.51^0$ C. It may be observed that temperature gradient over lower and middle levels is around 6^0 C, but it is only 3^0 C at 200 hPa. These temperature patterns are shown in fig. 5.16.

In the monsoon season, the thermal high at 850 hPa shifts to the extreme north/ northwestern parts of the country with temperature almost same (22^0 C) as observed in the pre-monsoon season. At 700 h pa, the thermal high shifts slightly southwards with centre near western Gujarat (12^0 C). At 500 h Pa, a weak thermal high (-02^0 C) is over eastern parts viz: Uttar Pradesh and Bihar. At 200 h Pa, zone of maximum temperature (- 47^0 C) shifts again to the northern/northwestern region. It may be again observed that temperature gradient over 850 h Pa and 200 h Pa is around 6^0 C, but it is only 3^0 C at 700 hPa and 500 hpa levels.

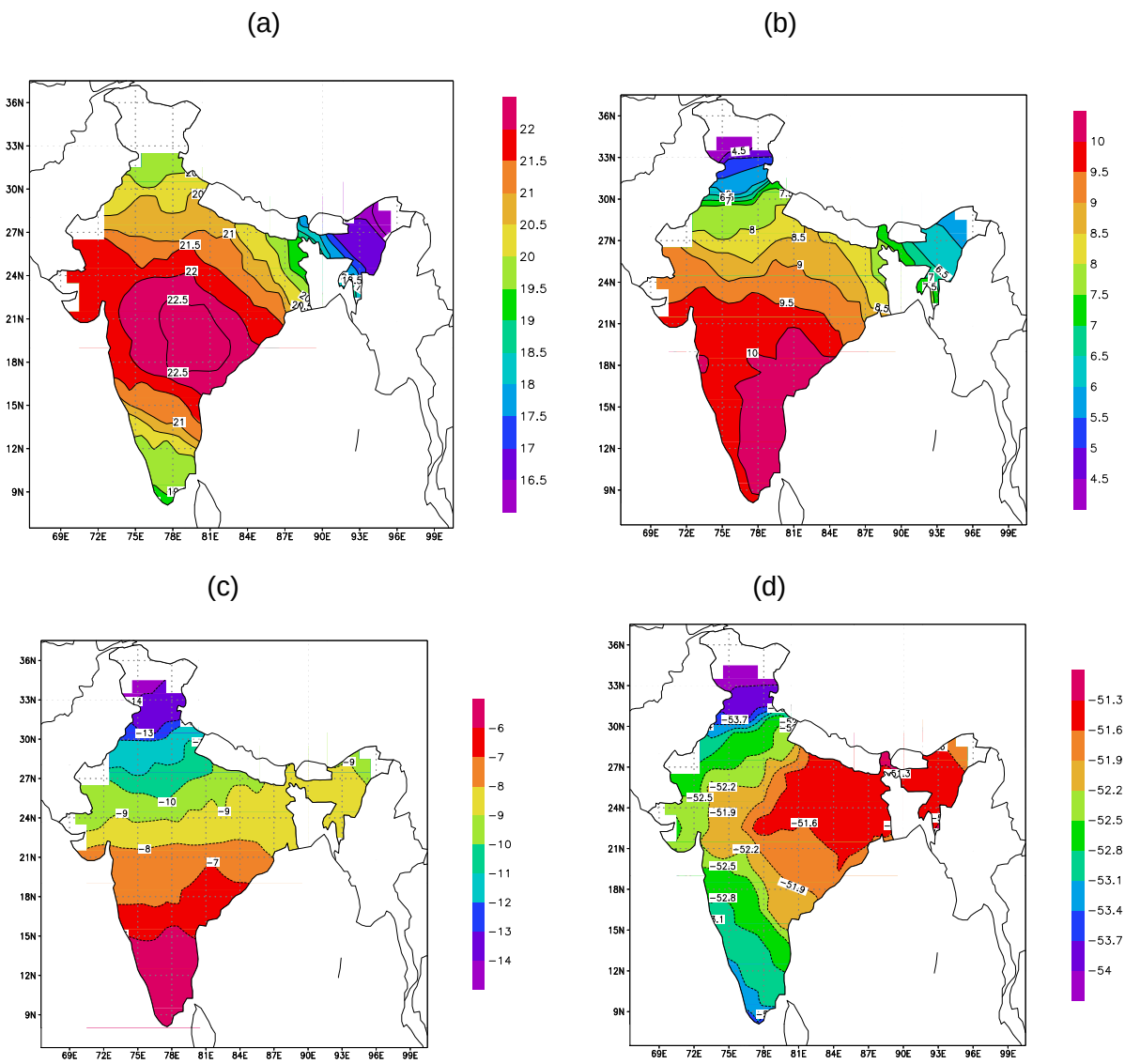


Fig. 5.16: Temperature pattern during pre-monsoon season over, a) 850 hPa, b) 700 hPa, c) 500 hPa and d) 200 hPa levels.

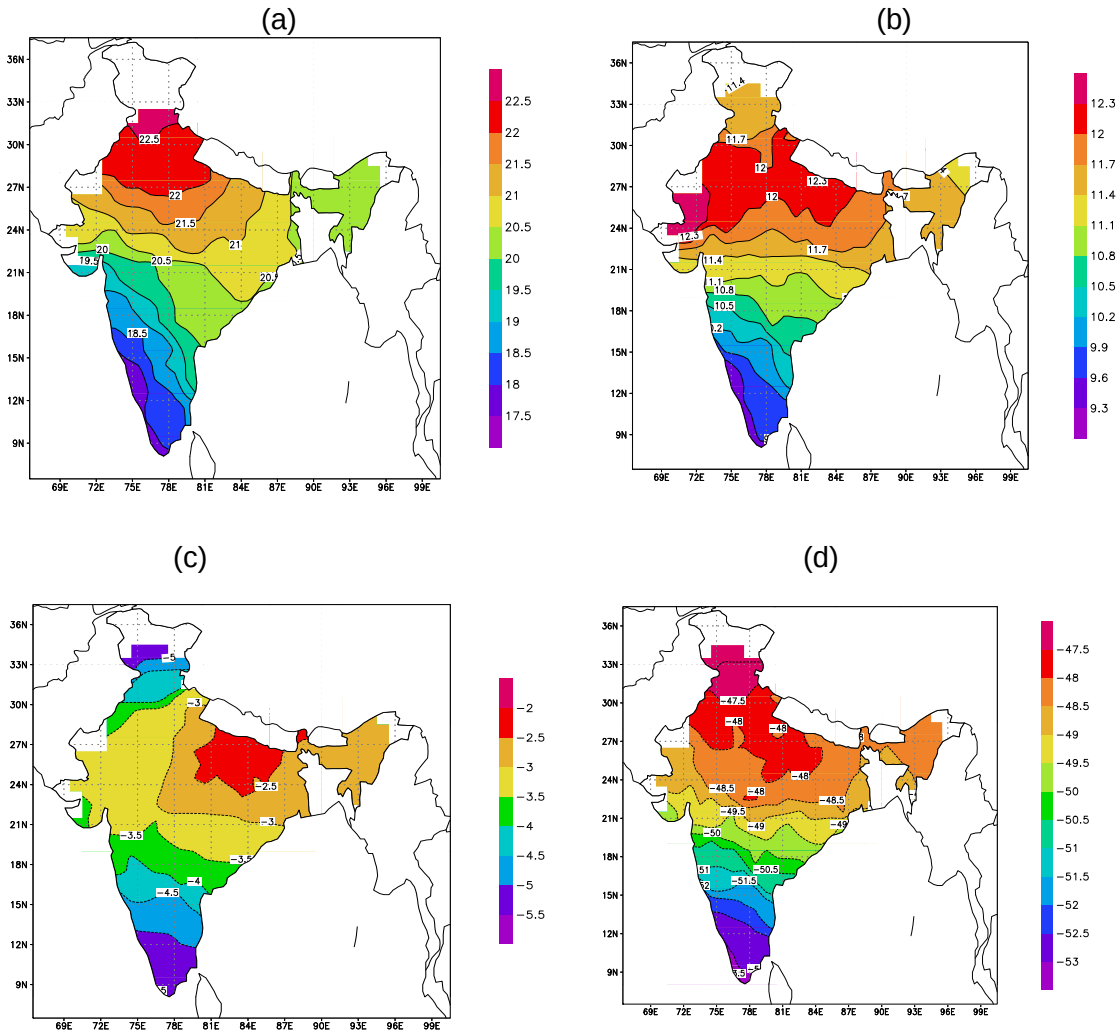


Fig. 5.17: Temperature pattern during monsoon season over a) 850 hPa, b) 700 hPa, c) 500 hPa and d) 200 hPa levels.

5.6 Summary of some important studies

During 1950s, Koteswaram and George (1958) made pioneering contribution in the study of monsoon structure and the formation of monsoon disturbances over the Bay of Bengal. Pisharoty and Asnani(1957) reported their work on preferential distribution of heavy rainfall in south sector of a monsoon depression from dynamical point of view. Another important contribution by George (1956), dealt with the effects of off-shore vortices on the monsoon rainfall along the west coast of India. During 1960s, Rao (1962) reported his work on meridional circulation associated with Indian summer monsoon.

The work of Joseph and Raman (1966) on the Low Level Jet (LLJ) over Peninsular India during July led to the discovery of an important component of summer monsoon. Colon (1964) and Desai (1968) reported the findings of their works on the interaction of summer monsoon current and the sea surface over the Arabian Sea. In 1970 Desai made significant contribution on the nature of low level inversions over the Arabian Sea during monsoon. Another important contribution was made by Sikka and Gadgil (1978) on the relationship between large-scale rainfall rainfall over India during summer monsoon and lower and upper tropospheric vorticity.

One of the important areas of monsoon research during 1970-1990 was the study of monsoon depressions. Sastry and Souza (1970, 1971**)** made important studies about the thermal structure, stratification and circulation over the Arabian Sea during the monsoon season.

Rao and Rajamani (1970), Sarkar and Chowdhary (1988), Mooley and Shukla (1989) and others made important contributions on structure and vertical motion fields associated with monsoon depressions. Several important findings on intra-seasonal oscillation (Singh and Kripalani, 1990) were reported. There was a notable research on the intra seasonal variations of the Asiatic summer monsoon by Chandrasekhara R. Kondragunta (1990)**.**

Pant (1964) reported that the onset of monsoon over India and adjoining seas, south of 15°N is associated with the disappearance of the pre-monsoon high over central parts of the country and the formation of the monsoon trough near 90°E at the 700-mb level.

Rao and Desai (1973) highlighted that the heat-low over Pakistan and the air from across the equator drawn north and northeastwards are considered the primary causes responsible for the setting up of monsoon circulation up to about 600 mb.

Analyses by Rajeevan (1991) revealed that years of droughts in India are associated with cyclonic circulation anomalies and cold thermal anomalies in the troposphere between 500 hPa and 200 hPa over northwestern India.

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CHAPTER 6

ONSET, ADVANCE AND WITHDRAWAL OF MONSOON

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6.1. Introduction

The date of onset of southwest monsoon over Kerala, the southernmost state of India is a crucial date for India as it marks the beginning of a four month (June to September) period of rains in India which accounts for over 70% of its annual rainfall. Normal dates of the onset (withdrawal) of monsoon rains over different parts of India, as given by India Meteorological Department (1943) are shown in Fig.6.1a & 6.1b. These figures are based on the long-term average pentad (five day non-overlapping) rainfall graphs prepared for several observatory stations. The middle date of the pentad, which shows an abrupt increase (decrease) in rainfall, was taken as the monsoon onset (withdrawal) date for each station. Climatologically, the monsoon sets in over the extreme southern part of peninsular India (Kerala) by the end of May. It sets in over the south-east Bay of Bengal and the adjoining Myanmar about two pentads earlier. Over peninsular India south of latitude 23°N, the advance of monsoon rains is from south to north; this takes place in about 3 pentads following the Monsoon Onset over Kerala (MOK). North of latitude 23°N, the monsoon rains advance from east to west. By the end of June monsoon covers more than 90% of the area of India. By mid-July the whole of India is covered by the monsoon. In early September summer (south-west) monsoon rains begin to withdraw from north-west India. By mid-October the SW monsoon ends and the north-east monsoon rains begin over the southern peninsular India. Joseph et al (1994) has given a review and large number of references on monsoon onset over Kerala.

During the monsoon onset over India dramatic changes are known to occur in the large-scale atmospheric structure over the monsoon region. Some of the well known ones associated with the onset are a rapid increase in the daily rain rate, an increase in the vertically integrated moisture and an increase in the strength of the low level monsoon flow (kinetic energy). The date of MOK is operationally (real time) announced by India Meteorological Department (IMD) each year. These were subjective estimates till 2005, prepared by several forecasters over the years. Regarding this the following is quoted from Ananthakrishnan and Soman (1988). “Although the onset of monsoon is associated with changes in the circulation features in the lower and upper troposphere, a sustained increase in the rainfall at the observatory stations over Kerala and the island stations over the south-east Arabian sea is an essential feature of the monsoon onset. It is difficult to quantify this precisely and so the experience of the forecaster plays a key role in declaring the date of monsoon onset in individual years”.

The dates of MOK declared by IMD for each year during the 100 years 1901 to 2000 are shown in Fig.6.2a. These are subjective estimates based primarily on the daily rainfall reports from rain gauge stations of the synoptic network. At MOK rainfall has to be widespread spatially over Kerala and persistent for a few days. Accompanying such rainfall, the lower tropospheric westerly wind (the monsoon current) over Kerala is strong and deep and the relative humidity of the air is high from the surface to at least 500 hPa (Rao, 1976). For several decades till the 1980s IMD was taking all these factors into consideration in a subjective way to fix the date of MOK. But during the 1990s and upto 2005, IMD declared monsoon onset when after 10 May a large percent of the synoptic stations of Kerala and the island stations of southeast Arabian sea reported rainfall for two consecutive days as per the rain only criteria for monsoon onset given by Ananthakrishnan et al (1967).

There is considerable inter-annual variability in the dates of MOK as may be seen in Fig.6.2a. The 100-year mean date of MOK is 01 June and its standard deviation is 7.4 days. The earliest onset was on 11 May in 1918 and the most delayed onset was 18 June in 1972. The spread of the dates of MOK is shown by the histogram in Fig. 6.2b. The 7-year moving average marked in Fig.6.2a shows the climate change in the date of MOK. At the beginning of the 20th century MOK was

mostly in the second week of June. Dates of MOK became earlier as decades passed and during the 1950s and 1960s monsoon set in Kerala in many years by mid-May. Since the 1960s the dates of MOK have moved towards their long-term mean date of 01 June.

6.2. PMRP and Bogus Monsoon Onset

Studying the weekly rainfall of the 4 meteorological sub-divisions of India south of latitude 13°N (Lakshadweep, Kerala, Tamilnadu and Andaman and Nicobar Islands) during March to May of the years 1960 to 1984, Joseph and Pillai (1988) found that a peak in rainfall activity occurs in the east-west belt containing these 4 sub-divisions about six weeks before the date of MOK. This flare up of rainfall activity about 40 days or an intra-seasonal cycle before MOK (as may be seen from Sikka and Gadgil, 1980) was called by them Pre Monsoon Rain Peak (PMRP). Fig 6.3(a) shows the location of these 4 sub-divisions lying between latitudes 8°N and 13°N and longitudes 70°E and 90°E. A superposed epoch diagram with the 0-week containing the date of MOK during each of the 25 years 1960 to 1984 shows the composite weekly rainfall anomaly over the area from 10 weeks prior to MOK to 3 weeks after it as shown in Fig.6.3(b). The area has a negative rain anomaly 3 weeks before MOK. If x is the day of the rain peak (PMRP) in the period 01 April to 10 May counted from 01 April as day 1 and y is the day of MOK as declared by IMD also counted from 01 April, the straight line of best fit by the least square deviations method is,

$$Y = 0.75x + 46.19$$

The linear correlation coefficient between x and y is 0.87. From the regression equation it is seen that if the PMRP occurs on 01 April (x=1), the estimated monsoon onset is after 46 days (on 17 May) and when PMRP is on 10 May (x=40), the estimated monsoon onset is after 36 days (on 15 June). Since (y-x) is dependant on the calendar date, it is apparently controlled by the seasonal changes in the atmosphere and the ocean (higher SST and vertically integrated moisture in the atmosphere as the season advances). Ramesh Kumar et al (2004) did a similar study but using the GPCP pentad rainfall data of a later period 1979 to 2001 and

obtained a linear correlation between x and y as 0.64 and the linear regression equation between x and y as

$$Y = 0.45x + 47.5$$

PMRP has been chosen as one of the predictors in the statistical multiple regression model currently in use by IMD for the long range prediction of the date of MOK (Pai and Rajeevan, 2007)

Flatau et al (2001) describes “Bogus Monsoon Onset”, a phenomenon which they found generally occurring in May in late monsoon onset years. They have reported six such occurrences in 1967, 1972, 1979, 1986, 1995 and 1997 during the period 1965 to 1997. A recent case in 2002 has also been studied (Flatau et al, 2003). Bogus monsoon onset is nothing but the PMRP of Joseph and Pillai (1988) which occurs every year. Since it has been found as occurring in association with delayed monsoon onsets, the bogus onsets occur in May when the atmosphere has much more moisture than in April (SST also is higher) and so is associated with much more rainfall and stronger low level winds than when it occurs in April. In 1972 IMD had declared monsoon onset over Kerala in their official bulletins on 16 May in association with such a spell of heavy rainfall and strong monsoon like westerly low-level winds, but soon issued a correction calling it as temporary monsoon onset. Later that year IMD declared monsoon onset on 18 June. This temporary monsoon onset of 1972 is one of the bogus monsoon onsets reported by Flatau et al (2001).

6.3. Processes in Atmosphere and Ocean leading to MOK

To understand the monsoon onset process, the temporal and spatial evolution of deep convection in association with MOK has to be understood. We have in the recent literature, a few studies on these using different compositing methods.

(a) Joseph et al (1994) used OLR data of the ten years 1975, 1976, 1977, 1980, 1981,1982,1984,1985, 1986 and 1987. For these ten years, the IMD date of MOK lay between 28 May and 04 June with a mean of 31 May and standard deviation of only two days. The period two days earlier to MOK to two days after it has been

called the zero pentad (0p). The composites for -8p to +1p are given in Fig.6.4. OLR contours of 240 watts / m^2 and less at intervals of 20 watts / m^2 are shown and are considered to be broadly indicative of convection. At -8p there is organized deep convection in a band around the equator east of about longitude 70^oE and extending into the west Pacific Ocean. By -7p the convection in the west Pacific has decreased and that in the Indian Ocean has organized into a super-cluster and moved slightly northwards. At -6p the distribution of convection shows the possible existence of a pair of cyclones, one on either side of the equator in the Indian Ocean. By -5p the Indian Ocean is practically free of convection but there is plenty of convection over Southeast Asia. At -7p the western end of the convective cloud cluster gives rain in Kerala. This is the PMRP discussed earlier which occurs even in years of non-delayed MOK. At -4p an elongated narrow band of convection forms close to the equator in the Arabian Sea longitudes. This convective band grows rapidly in area and intensity and moves north and culminates in MOK at 0p. The rapid break up of convection over the western Pacific ocean at -2p and -1p is a characteristic feature associated with MOK. The intense zone of deep convection extends from the south Arabian Sea to south China at MOK. The monsoon onset isochrones given by Tao and Chen (Fig. 6.5) shows that monsoon sets in over Kerala and south China at the same time in climatic means.

(b) Joseph et al (2006) used a different method for the compositing. The time difference between PMRP and MOK is taken as the period of an intra-seasonal oscillation (ISO). It is found to vary considerably from year to year. They used OLR and 850 hPa wind data of years with ISO period between 30 and 40 days (mean 35 days). They composited data of 9 such years 1979, 1982, 1984, 1986, 1987,1990,1993,1995 and 1996. These contain a two-week delayed MOK year of 1979 and also a two-week early MOK year of 1990. The composites in OLR are very similar to Fig.6.4. Composites made using 850 hPa wind data (NCEP/NCAR reanalysis) are given in Fig.6.6. At -8p an area of convection forms near the equator south of the Bay of Bengal between longitudes 70^oE and 110^oE. In association with this convection, a band of strong westerlies are seen near the equator. This is like the Gill (1980) model of wind response to an equatorial deep convective heat source with two cyclonic circulations one to the north and the other to the south of the equator. Convection and 850 hPa wind at -7p are like that of PMRP or bogus

monsoon onset (rain and westerly winds in Kerala as at MOK). By -5p convection has covered south-east Asia and the south China Sea along with strong monsoon westerlies over those areas. These westerlies replace the easterlies associated with the Western Pacific Sub-tropical High seen in earlier pentads. We may consider this phase as the onset of the South China Sea Monsoon (SCSM) in some years, which as an average of 1948 to 2000 occurs on 21 May as may be seen from Wang et al (2004) who used a 850 hPa wind based definition for onset. It may be noted that the correlation between the date of monsoon onset over Kerala and the date of the SCSM onset using data of the period 1948 to 2000 is -0.06. The time difference between these onset dates (MOK-SCSM) has varied between -14 and +41 days. MOK is linked to a major planetary scale phenomenon namely the Low Level Jet-stream as described later. According to Wang et al (2004) defining SCSM monsoon onset in individual years is noticeably controversial. In view of the lack of a relation between MOK and the SCSM onset, it is difficult to accept the description of the Asian Summer Monsoon onset as occurring in two stages, the first transition and the second transition related to the monsoon onsets over South China Sea and later over India – eg Lau et al (1998) and Hsu et al (1999). We may however say that the monsoon onset process over south Asia has two stages, the first and second in terms of the associated convection in the Bay of Bengal and eastwards from -8p to -5p and convection over the Arabian sea and eastwards from -3p to 0p. At -3p a fresh area of convection forms close to the equator and south of the Arabian Sea between longitudes 50°E and 75°E. During the following pentads, this area of convection grows in area and intensity and moves north to bring in MOK. There is a major difference between the 850 hPa wind response to convection associated with PMRP (-7p) and MOK (0p). At MOK the trade wind easterlies of south Indian ocean cross the equator near the east African coast (about 40°E), but at PMRP (-7p) the trade winds are found to cross the equator further east between longitudes 60°E and 75°E. The longitudinal locations of the areas of convection are also very different. These properties are to be made use of to eliminate possible PMRPs while trying to recognize the date of the true MOK.

(c) Wang et al (2009) studied the intra-seasonal process related to the Indian summer monsoon onset. They examined the two dimensional evolution of convection and circulation features before, during and after MOK on ISO time scales.

For this purpose a sequence of composite pentad mean of OLR and 850 hPa wind anomalies were plotted from -4p to +3p (with respect to MOK as 0p). Climatic means were obtained using data of 29 years of daily OLR (NOAA) and 60 years of 850 hPa wind (NCEP/NCAR reanalysis). The daily climatology for finding anomalies was the annual mean plus the first three harmonics with periods 360, 180 and 120 days. About 4 pentads before MOK, suppressed convection controls southern India and the south-east Arabian Sea, which in subsequent two pentads advances northeastwards to cover most of the Indian peninsula. This dry phase of ISO prior to MOK, cause serious heat waves in India. At -2p, positive convection anomalies are formed over the western Indian Ocean that move northeastward. When the enhanced convection reaches the southern tip of peninsular India, MOK occurs. Throughout the evolution of the ISO, the 850-hPa wind anomalies moved in concert with the OLR (convection) anomalies. These results are similar to those of Joseph et al (1994) and Joseph et al (2006) who made composites in two different ways using data of 10 and 9 years respectively. Wang et al (2009) used data of all the years available.

To understand the physical processes involved in the monsoon onset, we have also to study the temporal and spatial changes in the vertically Integrated Water Vapour (IWV) from 1000 hPa to 300 hPa, the monsoon current or the cross equatorial Low Level Jetstream (LLJ) and the Sea Surface Temperature (SST) of the warm pool area from -8p to the MOK.

(i) Integrated Water Vapour (IWV): In Fig.6.7 are given the changes in the daily OLR and IWV averaged over the inner box around Kerala (marked in the top part of the figure) and their anomalies from –10p to 0p of the 9 year composite of Joseph et al (2006). There is increased convection and IWV around PMRP and MOK. That at PMRP IWV has a maximum has been shown recently in a study by Rameshkumar et al (2009) using HOAPS-3 data set. There is suppressed convection and drying of the atmosphere during –5p to –2p. This is the time when peninsular India is affected by heat waves as reported by Flatau et al (2002) in their study of Bogus Monsoon Onset years. The outer box shown covers a much larger area (the monsoon area), at different parts of which organized convection occurred during the 14 pentads prior to MOK. The changes in IWV in this large area from –14p to 0p (MOK) due to the

pumping up of moisture by the deep convection and its horizontal spreading and mixing are shown in Fig.4.8. It is seen that IWV of the outer box grows and reaches very large values at MOK of the order of 45 Kg/m^2. This growth of IWV to large values in the monsoon area is needed for organizing MOK when convection, rainfall and the monsoon current strength reach very large values and also cover large areas (on the planetary scale). The growth of IWV in the final 3 pentads before MOK is very rapid as shown in Fig.6.8 and also in the study by Pearce and Mohanty (1984) of four monsoon onsets, who suggests that there is a positive feed-back between convection and the low level monsoon current during this phase. It may also be seen from Fig.6.8 that IWV remains steady (no growth) in the monsoon area during the pentads –5 and –4 when the convection in the Indian Ocean is minimum (suppressed convection phase).

(ii) Low Level Jetstream (LLJ): From Fig.6.6 it is seen that a cross-equatorial Low Level Jetstream (LLJ) of the Findlater (1969) type forms at –3p and strengthens rapidly during the following three pentads. At MOK it is a well-developed LLJ with its axis passing just south of Kerala and crossing the equator along the east African coast (near longitude 40°E). The jet core lies close to the 850 hPa level. The 60-day average has the 0p zonal wind and OLR of 12 years 1979-1990 (5 days around the IMD defined MOK of each year) is shown in Fig.6.9 (Joseph and Sijikumar, 2004). The east-west band of intense convection is on the cyclonic shear side of the LLJ just to the north of the LLJ axis, as may also be seen from the Hovmuller diagram showing the changes in OLR and 850 hPa zonal wind averaged over the longitudes 70°E to 85°E in Fig.6.10, based on the 9 year composites of Joseph et al (2006). Convection is induced and maintained by the cyclonic vorticity in the atmospheric boundary layer north of the LLJ axis and the consequent vertical motion of the moist monsoon air (Ekman pumping) in an environment with large convective available potential energy. LLJ also serves as a conduit for transporting the large amounts of moisture required for the monsoon rains from its main source region in the tropical Indian Ocean south of the equator. A positive feedback during the period -3p to 0p between the LLJ and the convective rain band can be inferred from Fig.6.10. Joseph and Sijikumar (2004) finds that the linear correlation between the convection and the LLJ zonal wind in north Indian ocean is very high and significant at a lag of 2-3 days, convection leading. It is clear from these studies that MOK occurs on the first day of

the formation of a well developed, strong and deep LLJ as described in Findlater (1969). Such an LLJ, which is a planetary scale phenomenon, has its genesis at the time of MOK.

In the wind composite of Joseph et al (2006) for 0p showing a fully developed LLJ, we do not see an onset vortex at 850 hPa in the southeast Arabian Sea but only large shear cyclonic vorticity north of the LLJ axis. In the composite wind charts for 700 hPa given in Soman and Krishnakumar (1993) also there is no sign of an onset vortex in the southeast Arabian sea at 0p (see Fig.6.3a of that paper). Following the description of an "onset vortex" in southeast Arabian sea in association with MOK of the FGGE MONEX year 1979 by Krishnamurti et al (1981), several monsoon researchers have been searching for an onset vortex as a trigger for the monsoon onset. Climatologically May-June is a period favourable for the genesis of depressions and tropical cyclones in southeast Arabian Sea and it is possible that in some years with the availability of large cyclonic shear vorticity in association with the LLJ at the time of MOK, a depression or cyclonic storm takes genesis there in some years. Ananthakrishnan et al have done a detailed a study on this examining MOK of the years 1901 to 1968. Summarising their results Rao (1976) states "there is a prounced tendency for the formation of low pressure systems at the leading edge of the monsoon current (LLJ). In 45 percent of the years, a trough of low pressure or a more intense system (cyclonic storm in 8% of MOK) is present in the Arabian Sea at the time of onset of monsoon along the west coast. The monsoon may also advance along the west coast with a disturbance in the Bay of Bengal. Still 25% of the monsoon onset is without any surface low pressure system". Krishnamurti et al (1981) in Table 6.2 of their paper has given Ananthakrisnan et al's (1968) summary of the conditions existing at the time of MOK in each of the years 1901 to 1968.

(iii) Sea Surface Temperature (SST): Analysing TMI SST data of several recent years Joseph et al (2006) found that the centre of the oceanic warm pool is located at -8p to -6p in the Bay of Bengal between latitudes 10°N and 15°N with maximum SST more than 31°C. At this time Arabian sea SST is much lower. During the period -8p to -6p a large area of convection formed south of the axis of the warm pool in the Bay of Bengal in the area of large SST gradient, but with SST close to 29°C. The

origin of this convection is likely to be by the mechanism suggested by Lindzen and Nigam (1987), which is particularly effective in low latitudes. This area of convection moves northeastwards across the Bay of Bengal and it is associated with strong westerlies at the surface. The convective clouds and the surface winds destroy the warm pool in the Bay of Bengal by -5P and replace it with colder surface waters. In the mean time the SST of the Arabian sea increases under cloud free conditions and a warm pool with SST more than 31°C forms in the Arabian sea in the latitude belt 10°N to 15°N by -3P. A large area of convection now forms and grows in the low latitudes of the Arabian Sea south of the axis of the warm pool during the pentad -3p. This area of convection moves northeast and brings about the MOK. The Arabian Sea warm pool is destroyed as the monsoon current and convection advance north. Fig 6.11 (a to d) shows the changes in the warm pool SST in the Bay of Bengal and the Arabian sea prior to the MOK of 2003.

6.4. Objective methods for Date of Monsoon Onset

Ananthakrishnan and Soman (1988): Apart from the small number of synoptic observatory stations, Kerala has a dense network of rain gauge stations having a very long period of records. Ananthakrishnan and Soman (1988) derived dates of monsoon onset using an objective criteria based on rainfall only. An objective monsoon onset data set is needed particularly for research. They derived dates of monsoon onset separately for south and north Kerala, the dividing line being the latitude of 10°N. South Kerala had 44 raingauge stations and north Kerala 31 stations having daily rainfall measurements from 1901. For the spatially averaged daily rainfall of south and north Kerala a limit of 10 mm per day was arbitrarily chosen as demarcation between light and heavy rain spells after examining long period rain records of May and June. The distinction in the nature of the two types of rain spells is important. Heavy rain spells imply deep convection in association with synoptic systems or the ITCZ; light spells are mostly associated with isolated pre-monsoon thunderstorm activity not associated with weather systems. The date of monsoon onset is taken as the first day of transition from light to heavy rain spell category, with the proviso that the average daily rainfall during the first five days after the transition should not be less that 10 mm. The five-day persistence is incorporated in the onset criteria to exclude rain produced by short duration synoptic

weather systems. The rain system associated with monsoon onset has a much larger spatial scale and longer time duration. According to Ananthakrishnan and Soman (1988), the rain spell that heralds the onset of monsoon over Kerala has a mean duration of 15 days and the associated daily mean rainfall is 26 mm. These have large variations in individual years.

To highlight the abrupt nature of the rainfall transition during the onset, the daily mean rainfall of south and north Kerala were composited by the superposed epoch method. The composite for the 80 year period 1901 to 1980 for one of the two regions (South Kerala) is given in Fig. 6.12 for the period 70 days before the monsoon onset to 70 days after it, the 0-date corresponding to the date of monsoon onset in each year. The monsoon onset dates derived for south and north Kerala and the IMD dates of MOK are given in Ananthakrishnan and Soman (1988) for each year of the period 1901 to 1980. The statistical properties of these three onset series are given in Table 6.1. The linear correlation coefficients between the dates of MOK (by IMD) and the dates of onset over South and North Kerala by Ananthakrishnan and Soman (1988) are 0.83 and 0.81 respectively.

Joseph, Sooraj and Rajan (2006): The OLR and 850 hPa wind composites made by them have already been discussed. They further found that the westerlies associated with the LLJ becomes not only strong but deep also at the time of MOK. Fig 6.13 shows the vertical structure of the average zonal wind in a box just south of Kerala 70-85°E and 5-10°N at -2p, -1p, 0p and +1p of the nine year composite. Between -1p and 0p the depth of the monsoon current (zonal westerly winds) has deepened to 400 hPa and strength increased to 10 m/s with the level of maximum wind close to 850 hPa. The depth of westerlies has not further increased from 0p to +1p. An objective method for deriving the date of MOK, which can be operationally used, has been developed by Joseph et al (2006). It has three steps.

Step-1: The daily strength of the average zonal wind in the box just south of Kerala (70-85°E, 5-10°N) is to be monitored daily for the lower and middle tropospheric levels from an objective gridded analysis done operationally beginning on 5 May. At MOK the area mean zonal wind of the box should reach 6 m/s at 600 hPa. Fig 6.14 shows such an analysis for the delayed

onset years 1972 and 1979, the normal onset year of 1987 and an early onset year of 1990.

Step-2: If a possible MOK is found by step-1 during the period 5 May to 25 May, check whether it is a PMRP or MOK by examining the spatial patterns of OLR and 850 hPa wind fields.

Step-3: The slow and steady movement of organized convection (rainfall) from the equatorial area to the latitudes of Kerala to bring about MOK is checked in a Hovmuller diagram (similar to Fig.6.10) averaging OLR between longitudes 65°E and 80°E.

Using this 3-step method Joseph et al (2006) derived dates of MOK for the years 1971 to 2003. This objective method with slight variations was adopted by India Meteorological Department for the declaration of the date of MOK operationally (real time) with effect from the monsoon of 2006. Pai and Rajeevan (2007) have described in detail the criteria adopted by IMD for declaring the date of MOK based on rainfall, wind field and OLR. They are:

(1) If after 10 May, 60% of the synoptic rainguage stations of Kerala and Lakshadweep report rainfall of 2.5 mm or more for two consecutive days, MOK may be declared on the second day, provided the following criteria are also satisfied

(2) Depth of westerlies should be maintained upto 600 hPa, in the box equator to latitude 10°N and longitude 70-80°E. The zonal wind speed over the area bounded by latitudes 5-10°N and longitudes 70-80°E should be of the order of 15-20 knots at 925 hPa.

(3) OLR should be below 200 watts/m^2 in the box confined by latitudes 5-10°N and longitudes 70-75°E.

Using these criteria they have derived dates of MOK for the years 1971 to 2007.

Wang, Ding and Joseph (2009): Using only the 850 hPa wind data (NCEP/NCAR reanalysis) they have given an objective method for the date of MOK. They made use of the fact that at MOK the Findlater (1969) type of LLJ becomes well organized and passes through the South Arabian Sea (SAS) box defined by longitudes 40°E

and 80°E and latitudes 5°N and 15°N. They found that the rapid establishment of steady SAS westerlies in this box is in excellent agreement with the time of abrupt commencement of the monsoon rains over Kerala. An Ocean Circulation Index (OCI) for the Indian summer monsoon was defined as the daily average 850 hPa zonal wind in the SAS box. The OCI as an average of the years 1948 to 2007 on the climatic date of MOK (01 June) was obtained as 6.2 m/s. The date of MOK in an individual year is defined as the first day when OCI exceeds 6.2 m/s, with the proviso that OCI in the following six consecutive days also exceed 6.2 m/s. By a correlation method it was shown that rain in Kerala and also in the adjoining southeast Arabian Sea are having significant correlation with the 850 hPa zonal wind in the SAS box (Fig.6.15). Thus Wang et al (2009) has given an objective method for MOK using a single parameter (850 hPa zonal wind) which is easy to use with NWP products for the determination of the date of MOK and also in research studies on the inter-annual and decadal variability of MOK. The 60 year (1948-2007) onset dates derived by the OCI method have a high correlation of 0.81 with the subjective IMD onset dates except that in six years there are large differences between the two.

Fasullo and Webster (2003): To diagnose onset and withdrawal of monsoon in India (not Kerala) they have used Vertically Integrated Moisture Transport (VIMT) instead of rainfall.

$$VIMT = \int_{Surface}^{300hPa} qUdp$$

Where q is the specific humidity and U is the wind vector. They find that its variability particularly over the Arabian Sea is substantial during both onset and withdrawal of monsoon. An index named the Hydrological Onset and Withdrawal Index (HOWI) is derived from VIMT which is used to determine the date of monsoon onset / withdrawal over India. The authors claim that this index is indicative of the transition in the large scale monsoon circulation over India rather than being highly sensitive to synoptic variability and that it is not vitiated by bogus monsoon onsets and spatial variations of the monsoon circulation. They also find that the monsoon onset date by this method is well correlated with the June to September rainfall of India whereas the IMD derived date of MOK is not (Dhar et al, 1980). Fasullo and Webster (2003)

have given in their paper the monsoon onset and withdrawal dates using their HOWI index for the years 1948 to 2000.

Xavier, Marzin and Goswami (2007): Monsoon onsets (withdrawal) have been defined by Xavier et al (2007) as the day when the tropospheric heat source shifts from south to north (north to south). Please also see Goswami and Xavier (2005). The establishment of the tropical convergence zone around latitude 10oN delineates the physical monsoon season. Xavier et al (2007) found that the meridional gradient of the tropospheric temperature (averaged between 600 and 200 hPa) is proportional to the meridional gradient of deep tropospheric heating and could lead to acceleration of the deep tropospheric circulation. (a large number of studies on related areas are referred to in their paper, particularly studies by He et al, 1987 and Li and Yanai, 1996). Their objective definition of the large scale monsoon onset (over India) is based on the reversal of GrTT, (Gradient in Tropospheric Temperature as average of 600 to 200 hPa) between a northern box (40-100°E, 5-35°N) and a southern box (40-100°E, 15°S-5°N) denoted by GrTT. The onset date (GrTT onset) is defined as the date when GrTT changes sign from negative to positive and withdrawal date when the change is from positive to negative. The authors claim that the onset and withdrawal dates thus defined are physically based. A characteristic of the large scale onset of the Indian summer monsoon is an abrupt increase in the kinetic energy (KE) of the low level monsoon flow (Krishnamurti, 1985**)**. The day of abrupt increase in the KE of 850 hPa winds averaged over a large region (40-100°E, 5-15°N) above a threshold value of 40 m^2s^{-2} and persisting for 5 consecutive days is taken as the KE onset. Xavier et al (2007) have derived onset days by both GrTT and KE methods for the period 1950 to 2003. There is a strong linear correlation of 0.77 between these two onset days.

Table 6.2 gives the onset dates for Kerala derived by the old subjective method of IMD and the new objective method of IMD (Pai and Rajeevan, 2007), by the objective methods of Joseph et al (2006) and Wang et al (2009) and the objective onset dates for India derived by Fasullo and Webster (2003) and the onset dates by the GrTT method of Xavier et al (2007) for the 30 year period 1971 to 2000. While the mean onset dates for Kerala by the four methods are between 1 and 3 June, the mean onset date for India by Fasullo and Webster is 6 June and the mean

onset date by Xavier et al is 29 May. Table- 6.3 gives the matrix of linear correlations between pairs of these onset dates. Salient features of these correlations are: The correlation between IMD (subjective) and IMD (objective) is 0.91. Equally high (0.90) is the correlation between IMD (objective) and Joseph et al (2006). The correlations between the onset dates by the wind only definition of Wang et al (2009) on one side and IMD (subjective), IMD (objective) and Joseph et al (2006) on the other side are close to 0.80. The correlations between Wang et al (2009) or Xavier et al (2007) with the other dates of onset are all low and in the range 0.56 to 0.69 except that the correlation between Wang et al (2009) and Xavier et al (2007) is a high value of 0.84 possibly because both these depend on the strengthening of the large scale 850 hPa wind and tropospheric heating fields which are closely related as shown by Joseph and Sijikumar (2004).

A question arises as to whether the GrTT onset of monsoon derived by Xavier et al (2007) is for onset over India or a much larger or different area of south Asia. A comparison with the objective IMD onset dates for Kerala shows that in the years 1972, 1979, 1983, 1986, 1995 and 1997 when MOK (IMD) was delayed by 10 to 19 days (mean delay of 12 days from the long term mean date of MOK), the onset by Xavier et al was two pentads earlier than the IMD's objectively derived dates of MOK. At minus 2p the area of organized convection and rainfall is south of India and over southeast Asia (see Fig.6.4 and also the OLR composite of Joseph et al, 2006) and there is no convection (rainfall) over India. In the years 1985, 1990 and 1999 when IMD onset was two weeks earlier than normal, the onset dates by Xavier et al (2007) have very little difference from the IMD dates. Possibly the onset dates derived by Xavier et al define the beginning of strong convective heating in some part of the large monsoon area of south Asia which can also increase the kinetic energy of the monsoon flows through the area for the KE onset chosen by Xavier et al (2007). The area chosen by Wang et al (2009) for their zonal wind averaging to derive onset dates is mostly overlapping with the area for the KE chosen by Xavier et al (2007) and that may be the reason for the high correlation between their onset dates. In this connection the study by Krishnamurti and Ramanathan (1982) is relevant.

6.5. Interannual Variability of Monsoon Onset over Kerala (MOK).

Examining the MOK data for 120 years (1870 to 1989) Joseph et al (1994) found that there were 22 years when MOK suffered large delays of 8 days or more (one standard deviation or more). 16 of those 22 years are associated with a moderate or strong El Nino. There were 13 strong El Ninos during this period and 9 of them are associated with moderate or large delays of MOK in their 0 or +1 years. Delays of MOK are mostly associated with the +1 Year of the El Nino; a Chi^2 test has shown a significant association of delayed onset with the +1 year of El Nino (eg delayed onset years like 1900, 1915, 1942, 1958, 1983). There are a few 0-year cases of El Nino associated with large delay of MOK like the year 1972 with the MOK occurring 18 days late. Studying the association between Sea Surface Temperature (SST) and the delays in MOK, Joseph et al (1994) found that the SST of Indian and Pacific oceans during the previous winter and pre-monsoon seasons are related to the inter-annual variability of the date of MOK as shown by the correlation maps given in Fig.6.11 of Joseph et al (1994). Associated with delayed MOK, equatorial areas of Indian and central Pacific oceans have warm SST anomalies and tropical north Indian ocean and sub-tropical west Pacific ocean have cold SST anomalies persisting from the previous winter. The SST (HadIsst) - MOK (IMD objective) correlations using data of the period 1970 to 2002 are given in Fig.6.16 for January, February, March and April months separately, showing the persistence of the correlations through the winter and pre-monsoon months prior to MOK. The highest correlation in all these four months is over the subtropical western Pacific Ocean. Joseph et al gave a hypothesis that with such a distribution of SST anomaly over the Indian and Pacific oceans, the cloud band associated with the southern hemisphere ITCZ of boreal winter continues south of the equator longer than normal and its crossing the equator and establishing itself in the northern hemisphere is delayed, resulting in the delay of MOK. In the years when MOK has large delays, the tropical cyclone season of west Pacific ocean and the adjoining south Indian ocean gets extended into the April to June period, showing indirectly the prolonged residence of the ITCZ in the southern hemisphere. That this happens has been shown by Joseph et al (1994) analyzing the cyclone data of 1910 to 1980. The hypothesis proposed by Joseph et al (1994) relating SST anomalies of Indian and Pacific oceans with the variability in the dates of MOK has been verified in

numerical modeling studies by Ju and Slingo (1995), Soman and Slingo (1997) and Annamalai et al (2005). In 1983 MOK was delayed by two weeks. Fig 6.17a shows the SST anomaly of April 1983. The OLR anomaly of the same month is given in Fig 6.17b. In the southern hemisphere tropics there is positive convection anomaly (negative OLR anomaly) and there is suppressed convection in the northern hemisphere tropics. The east-west extent of the SST and OLR anomalies is global, showing the involvement of the global ITCZ in the inter-annual variability of the date of MOK.

6.6. Role of Intra Seasonal Oscillation (ISO) in Monsoon Onset

Joseph showed it and Pillai (1988) that Monsoon Onset follows PMRP over Kerala after 30-50 days, a manifestation of an intra-seasonal monsoon mode linked with the warm pools and SST gradients in the Bay of Bengal and Arabian Sea. It is not clear whether this mode is the same as the northward propagating low frequency mode in convective clouds and 850 hPa wind anomalies of period 30-50 days during the monsoon season shown by Yasunari (1980), Sikka and Gadgil (1980) and Krishnamurti and Subramanyam (1982).

That the convective cloudiness represented by OLR has an oscillation between tropical areas of Indian and west Pacific oceans with a 30-50 day period was shown by Lau and Chan (1986). Joseph (1990a) examined whether this east – west oscillation influences the timing of MOK. They examined the tropical cyclone activity in western North Pacific Ocean in relation to the convective activity in the Indian Ocean at PMRP and MOK. Tropical west Pacific Ocean north of the equator generates 26 out of the 79 tropical cyclones produced by global oceans in a year (Gray, 1978). About 60 percent of these occur in the monsoon period. Joseph (1990a) examined two groups of years one with Excess Monsoon Rains (EMR) over India in the four monsoon months June to September and the other with Deficient Monsoon Rains (DMR). The following broad inferences were drawn from the study. Depressions, tropical cyclones and Typhoons do not occur in the northwest Pacific Ocean around the dates of PMRP and MOK, when the Indian Ocean is convectively active. The 30 – 50 day oscillation in cyclonic systems in northwest Pacific Ocean is particularly prominent in DMR years. Fig-6.18 shows this for a period 90 days before

MOK to 45 days after it. In this composite of 5 DMR years, no depression, cyclonic storm or typhoon existed in the western Pacific Ocean north of the equator during the period of 10 days before MOK to 15 days after MOK. Similarly there is absence of cyclonic systems over a 20-day period around the PMRP.

We thus find that the timing of MOK is influenced not only by the timing of the annual cycle in the ITCZ (south to north movement) as shown in the previous section-4.5 but also by the east-west oscillation (30-50 day mode) in convection between the Pacific and Indian oceans and by the 30-50 day mode of the Indian ocean. We do not know whether there is a relation between these oscillations and the equatorial Madden Julian Oscillation of period 30-50 days – Madden and Julian (1994). Detailed studies are also needed on the role of the middle latitude westerlies and associated weather systems north of India in the variability of the date of MOK.

6.7. Advance and withdrawal phases of the monsoon over India.

Deshpande et al (1986) have given dates of monsoon onsets over Kerala and Mumbai for each year of the period 1901 to 1984 as obtained from the weather reports published by IMD. The average travel time of monsoon from Kerala to Mumbai is 8 days (about 1° latitude per day) and its standard deviation is 7 days; the extreme values are 29 and -7 days. (A negative value means that the monsoon has set in over Mumbai earlier than over Kerala. Such a situation occurred in 3 other years of the 84 year record and for each of them it is -1 day). Analysing this data set, Joseph (1990b) found that the linear correlation between the date of MOK (x) and the date of onset over Mumbai (y), days counted from May 01 as day 01, is 0.48 for the period 1901 to 1984. What is interesting is that the linear correlation between x and (y-x), the travel time of monsoon from Kerala to Mumbai is -0.63, which is very large and highly statistically significant. A linear regression relation is obtained as

$$(y-x) = 27.9 – 0.6\ x$$

From this relation it is seen that when monsoon sets in Kerala on 10 May, 30 May and 19 June, the estimated onset dates at Mumbai are after 22, 10 and -2 days respectively. Joseph (1990b) has given an explanation for this feature of monsoon

advance. In the north Indian Ocean around India the axis of the climatological monthly mean SST maximum lies close to the latitude of Kerala in May but shifts to the latitude of Mumbai in June. In a year when MOK is late, say by mid June, the monsoon cloud band after its formation and bringing rains to Kerala, quickly adjusts to the June position of the SST maximum axis which is close to Mumbai. This also explains why in some years monsoon has set in Mumbai earlier than in Kerala.

As mentioned in the introduction the average advance time of the monsoon from its onset in Kerala on 01 June to cover the whole country (by 15 July) is 45 days. The duration of the advance phase of the monsoon has considerable inter-annual variability as seen from the analysis by Khole (2009) of the data of the period 1945 to 2007 (see fig. 6.19). The mean and standard deviation of the advance phase duration are 38 and 11 days respectively. The shortest advance phase was in 1953 when the monsoon covered the whole of India in 17 days after MOK; the longest duration of the advance phase was in 2002, a period of 78 days.

According to Rao (1976) monsoon activity often weakens after an advance through about 500 Kms and a fresh surge is needed to make the forward edge of monsoon rains to move further. In the monsoon of 2002, monsoon advance stagnated or halted on three occasions (a) 13 to 19 June (b) 5 to 18 July and (c) 20 July to 14 August. These are called ‘hiatus’ in the progression of the monsoon. Fig. 6.20 gives the total number of stagnation or hiatus days in the monsoons of 1960 to 2002 taken from Kalsi et al (2004). The hiatus duration in a monsoon has a mean of 19 days and it has a standard deviation of 9 days. Khole (2009) analyzing the long hiatuses of the recent years 2002, 2004 and 2006 find that there had been an anomalously high influence of the middle latitude westerly systems during the monsoon advance period in those years. It is well known that tropical cyclone activity in the west Pacific Ocean affects the Indian monsoon. It is suggested that we examine the role of these tropical cyclones in causing monsoon stagnation or hiatus.

As the monsoon advances over India, major changes occur to the Sub Tropical Jetstream (STJ) which in the winter and pre-monsoon seasons has its core south of the Himalayas and is very strong. The STJ shifts to the north of Tibet with

the advance of the monsoon. According to Ananthakrishnan et al (1968), in years of normal or delayed MOK there is a sudden weakening of the upper tropospheric westerlies and STJ over north India at the time of MOK. In years of early MOK, westerlies over north India persisted in strength for about a fortnight after MOK. Soman and Kumar (1993) find a shift of the sub-tropical ridge at 150 hPa from about 17°N latitude on day -20 (in a composite of several years with MOK as 0 day) to about 26°N on day +15. The rapid shift of the latitude of the ridge occurs only after MOK. The easterlies at 150 hPa spread northward with the shifting of the ridge resulting in the establishment of the Tropical Easterly Jetstream (TEJ) through peninsular India.

The climatological withdrawal of monsoon from northwest India begins in the first fortnight of September. Its timing has considerable inter-annual variability. The withdrawal of monsoon and its gradual equator ward movement and the deceleration of the low level westerly flow is heralded by the seasonal cooling of the Asian continent (Ramage, 1971). At the time of onset and advance of monsoon the rain belt associated with the tropical convergence zone (TCZ) and the associated Maximum Cloud Zone of Sikka and Gadgil (1980) moves north. But monsoon withdrawal is associated with the inability of the TCZ and its associated rain band to advance to higher latitudes in the 30 – 50 day oscillation. So a rain based definition for the withdrawal of monsoon is difficult to be framed. In the literature very few studies are available on the withdrawal phase of the monsoon. Rao (1976) has associated the withdrawal of monsoon with the southward displacement of the surface pressure trough, the establishment of dry continental air and the development of anti-cyclonic flow over north and central India. Based on these characteristics, Syroka and Toumi (2004) have defined an index for monsoon withdrawal based on the characteristic low level circulation associated with the summer monsoon. A Daily Circulation Index (DCI) is defined as the difference in average 850 hPa zonal winds between a southern box (5-15°N, 50-80°E) and a northern box (20-30°N, 60-90°E). DCI captures the position and intensity (vorticity) of the monsoon trough. DCI has been used by Syroka and Toumi (2004) to define monsoon withdrawal from north and central India (north of latitude 15°N) which can be taken as the end of the southwest monsoon season. Using NCEP/NCAR reanalysis data they have derived the withdrawal dates of the Indian summer

monsoon (SW monsoon) for the years 1958 to 2000 (fig.6.21). DCI is smoothed by a seven day running mean. The date of monsoon withdrawal is taken as the first of seven consecutive days for which the smoothed DCI is negative. The mean date of withdrawal is 19 October with a standard deviation of 14 days (almost double that of MOK). The earliest withdrawal date is 23 September 1994 and the most delayed one is 23 November 1998.

India Meteorological Department at present uses following guidelines for declaring the withdrawal of southwest monsoon on operational basis:

a) Withdrawal from extreme Northwestern parts of the country should not be attempted before 1st September.

b) *After 1st September:*

The following major synoptic features should be considered for the first withdrawal from the western parts of NW India.

i) Cessation of rainfall activity over the area for continuous 5 days.

ii) Establishment of anticyclone in the lower troposphere (850 hPa and below)

iii) Considerable reduction in moisture content as inferred from satellite water vapour imageries and tephigrams.

c) *Further withdrawal from the country:*

i) Further withdrawal from the country may be declared, keeping the spatial continuity, reduction in moisture as seen in the water vapour imageries and prevalence of dry weather for 5 days.

ii) SW monsoon should be withdrawn from the southern peninsula and hence from the entire country only after 1st October, when the circulation pattern indicates a change over from the southwesterly wind regime.

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Table 6.1: Statistical properties of dates of Monsoon onset over Kerala by IMD, and over South Kerala and North Kerala by AS (Ananthakrishnan and Soman, 1988) for the period 1901-1980

Parameter	South Kerala (AS)	North Kerala (AS)	Kerala (IMD)
Mean Date	30 May	01 June	2 June
Standard Deviation	8.8 days	9.2 days	7.8 days
Earliest onset date	7 May 1918	8 May 1918	11 May 1918
Latest onset date	22 June 1972	22 June 1972	18 June 1972

Table 6.2: Dates of onset of monsoon during 1971-2000 - See description of onsets in the text (Dates are with respect to 01 May. 15 means 15 May, 40 means 09 June)

Year	IMD (Subjective)	Dates by Objective Methods				
		IMD (Objective)	Joseph et al (2006)	Wang et al (2009)	Fasullo and Webster (2003)	Xavier et al (2007)
1971	27	27	27	25	27	29
1972	49	50	50	46	47	41
1973	35	36	38	36	36	28
1974	26	23	23	27	49	28
1975	31	33	33	31	38	26
1976	31	30	27	29	30	27
1977	30	29	41	38	39	36
1978	28	29	37	29	35	22
1979	44	43	44	44	44	36
1980	32	34	35	33	32	33
1981	30	30	31	33	40	25
1982	32	30	27	32	37	32
1983	44	43	45	45	45	35
1984	31	32	30	29	31	26
1985	28	24	23	23	27	24
1986	35	43	38	37	40	30
1987	33	32	32	33	33	33
1988	26	33	35	32	38	20
1989	34	35	35	21	34	19
1990	19	18	16	17	19	16
1991	33	33	33	33	36	31
1992	36	36	37	39	43	41
1993	28	34	35	29	36	30
1994	28	28	28	30	34	29
1995	36	41	37	39	42	32
1996	34	40	42	33	36	26
1997	40	43	52	46	51	36
1998	33	34	34	37	40	36
1999	25	22	22	19	43	22
2000	32	32	32	15	29	18
Mean	32.33	33.23	33.96	32	37.03	28.9
S. D	6.12	7.13	8.1	8.11	7	6.48

Table 6.3: Linear Correlation between monsoon onset dates (1971-2000) by different methods (See text)

IMD (Subjective)	1					
IMD (Objective)	0.91	1				
Joseph et al (2006)	0.83	0.90	1			
Wang et al (2009)	0.79	0.79	0.81	1		
Fasullo and Webster (2003)	0.59	0.56	0.61	0.69	1	
Xavier et al (2007)	0.68	0.59	0.59	0.84	0.59	1
	IMD (Subjective)	IMD (Objective)	Joseph et al (2006)	Wang et al (2009)	Fasullo and Webster (2003)	Xavier et al (2007)

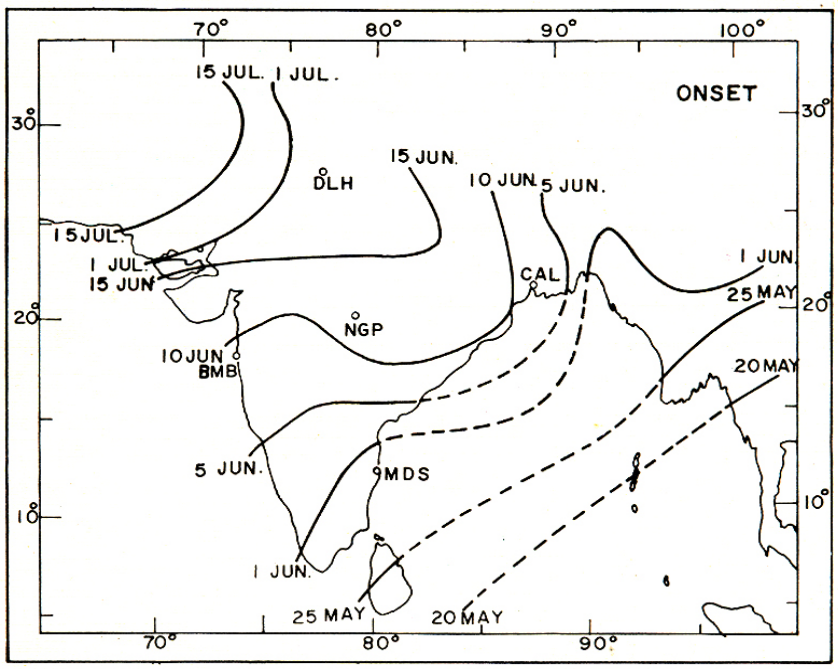


Fig. 6.1a: Dates of onset of southwest monsoon over India derived using long Period pentad mean rainfall of observatory stations (IMD, 1943)

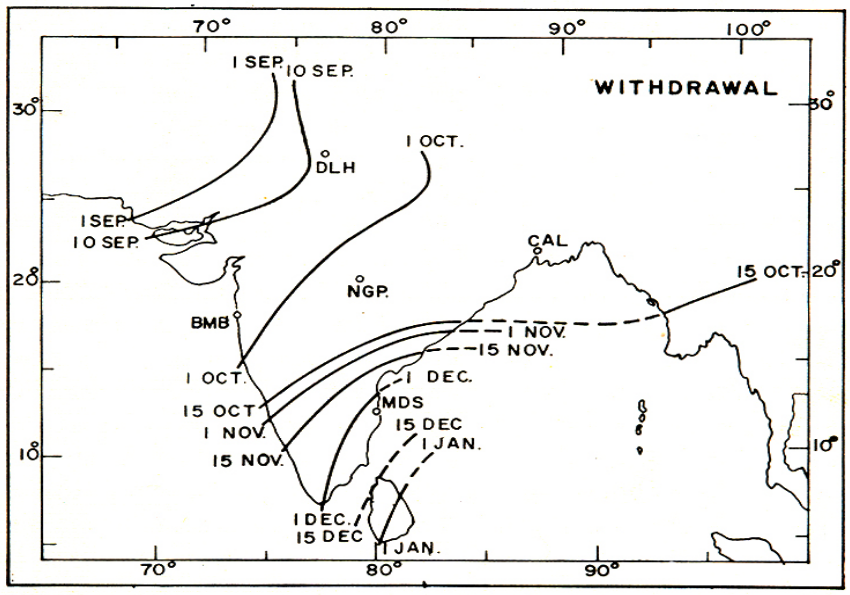


Fig. 6.1b: Dates of withdrawal of southwest monsoon over India derived using long period pentad mean rainfall of observatory stations (IMD, 1943)

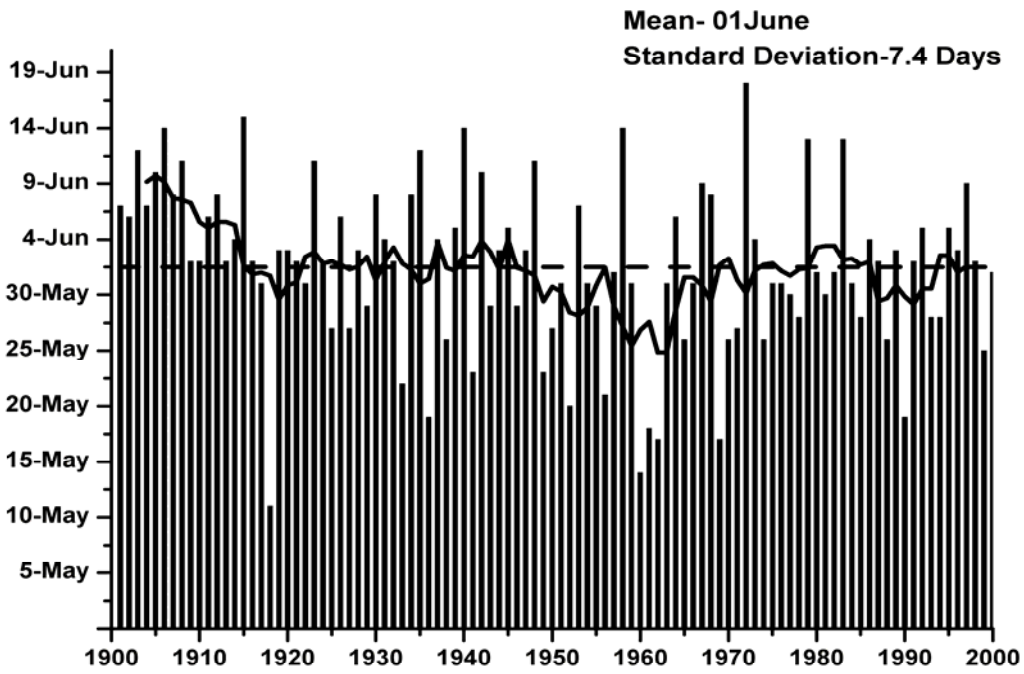


Fig. 6.2a: Subjectively determined IMD dates of onset of monsoon over Kerala 1901 to 2000. 100-year mean is shown by the broken line and the seven year moving average by the continuous line

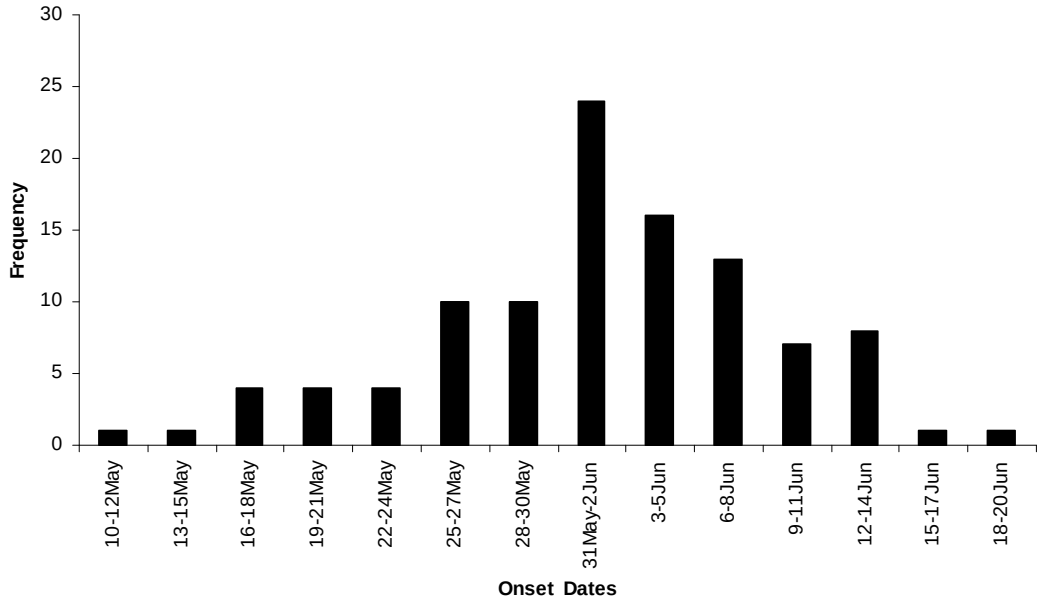


Fig. 6.2b: Histogram of the dates of monsoon onset given in Fig. 6.1a

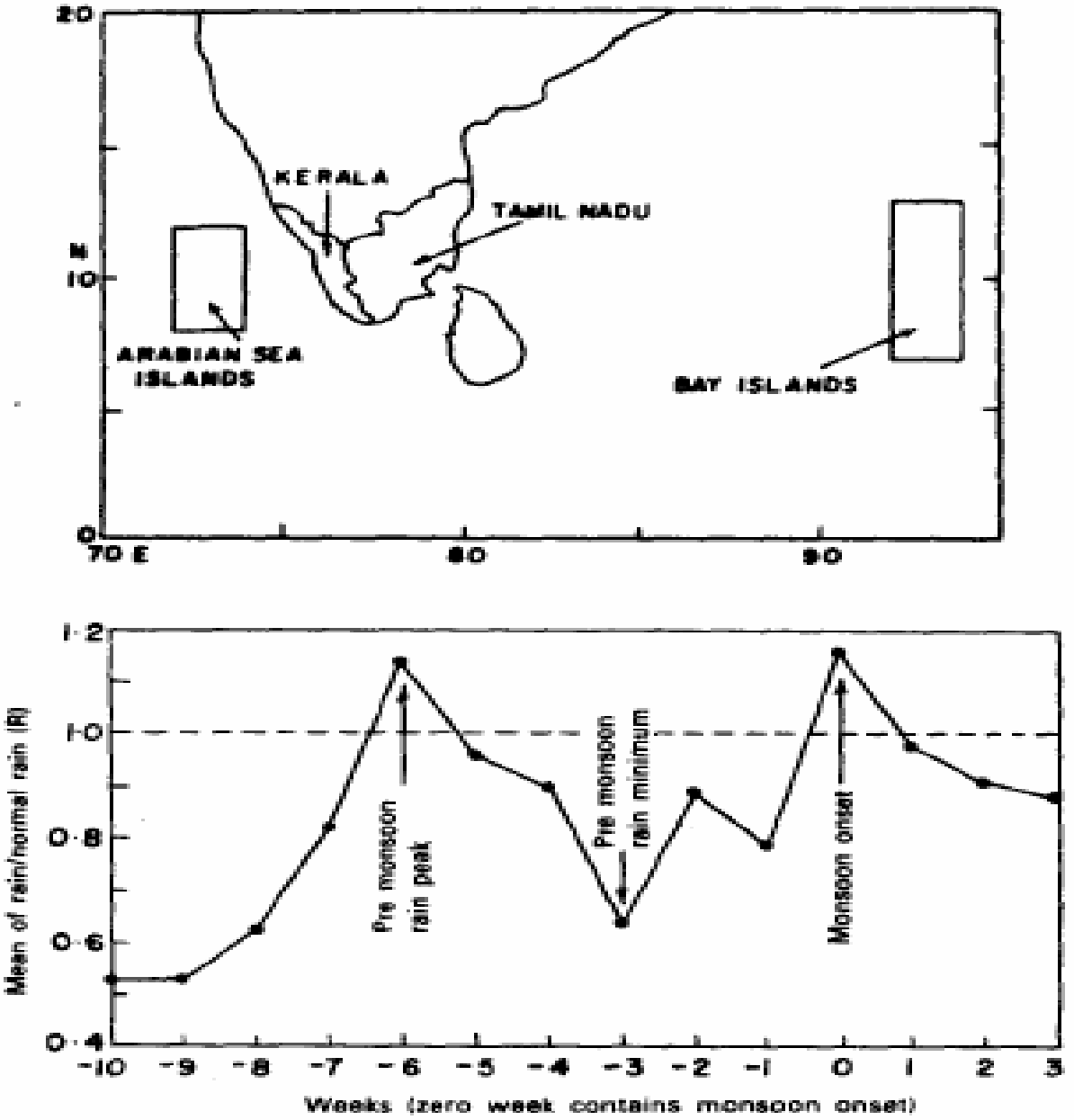


Fig. 6.3: The 4 met subdivisions of India in the area 8-13oN and 70-90oE are shown at top. At the bottom is given the super-posed epoch diagram of the weekly rainfall anomaly as mean of the 4 sub-divisions using data of 1960-1984 for the period 10 weeks prior to MOK to 3 weeks after it (MOK is in the 0-week) – from Joseph and Pillai (1988) and Joseph et al (1994)

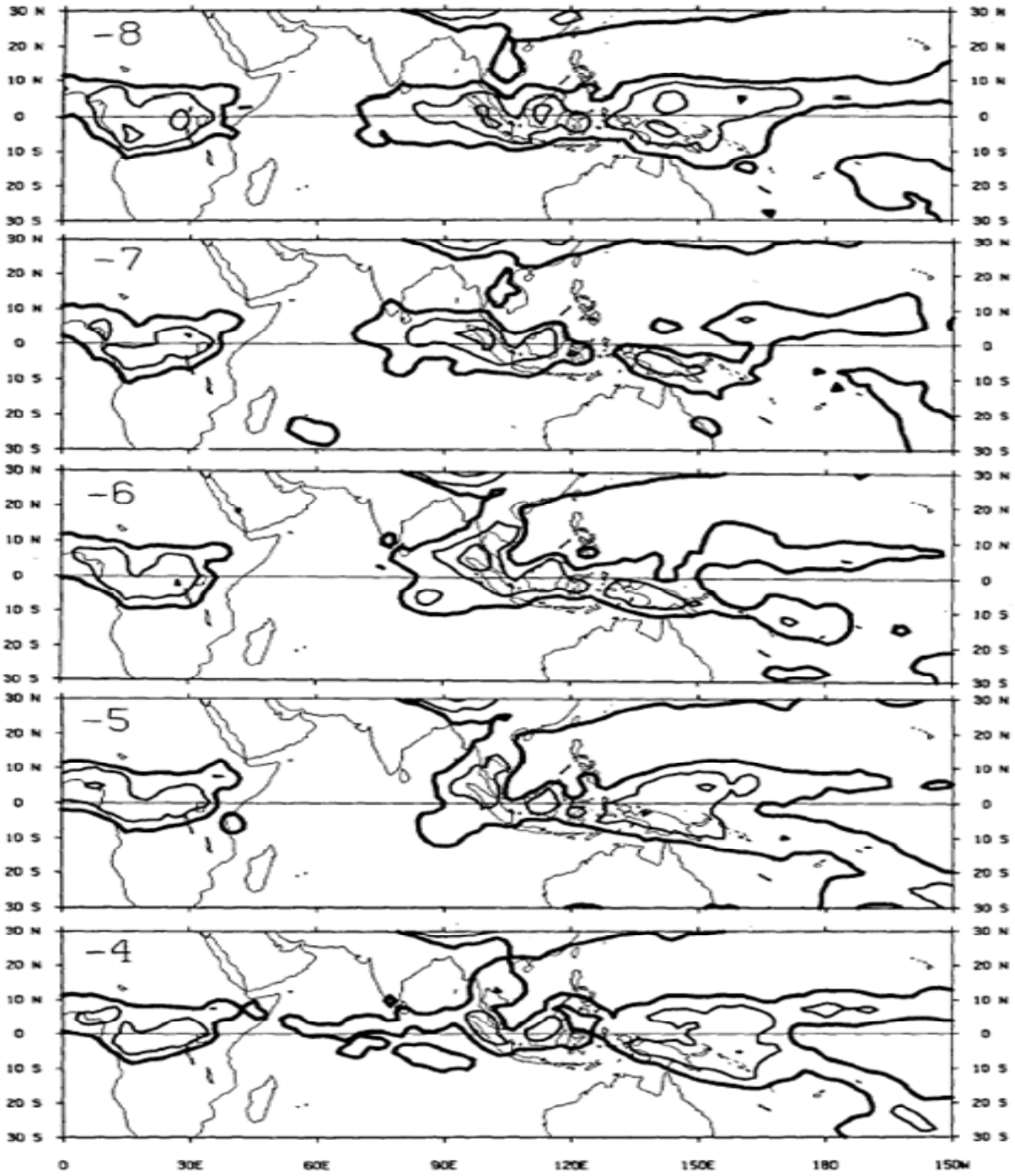


Fig.6.4: 10 year Composite OLR for pentads -8 to +1 (marked at top left). MOK is in the 0 pentad. Only contours of 240 Wm-2 (thick line) and less at 20 Wm-2 intervals marked. (from Joseph et al 1994)

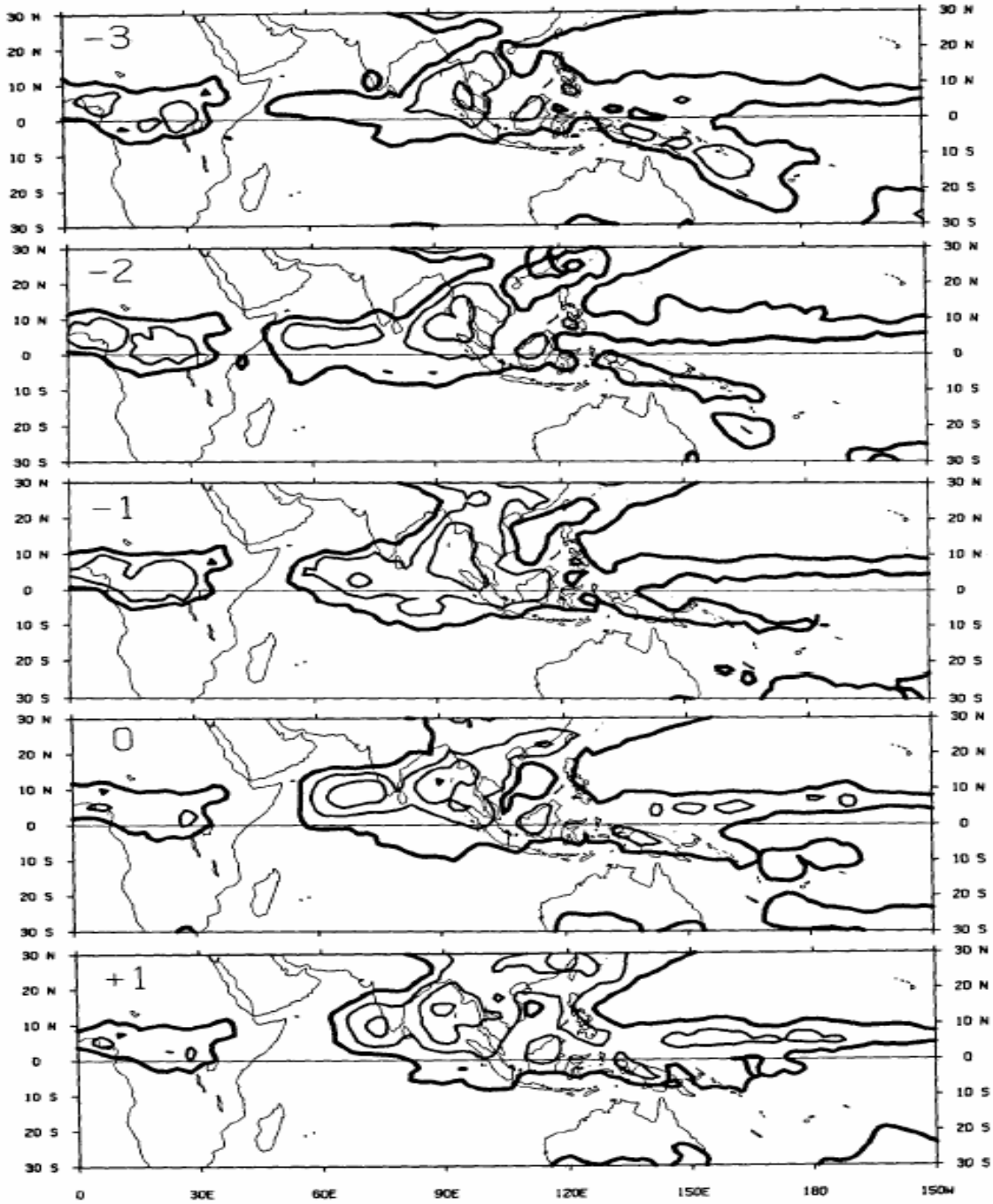


Fig. 6.4 (contd.)

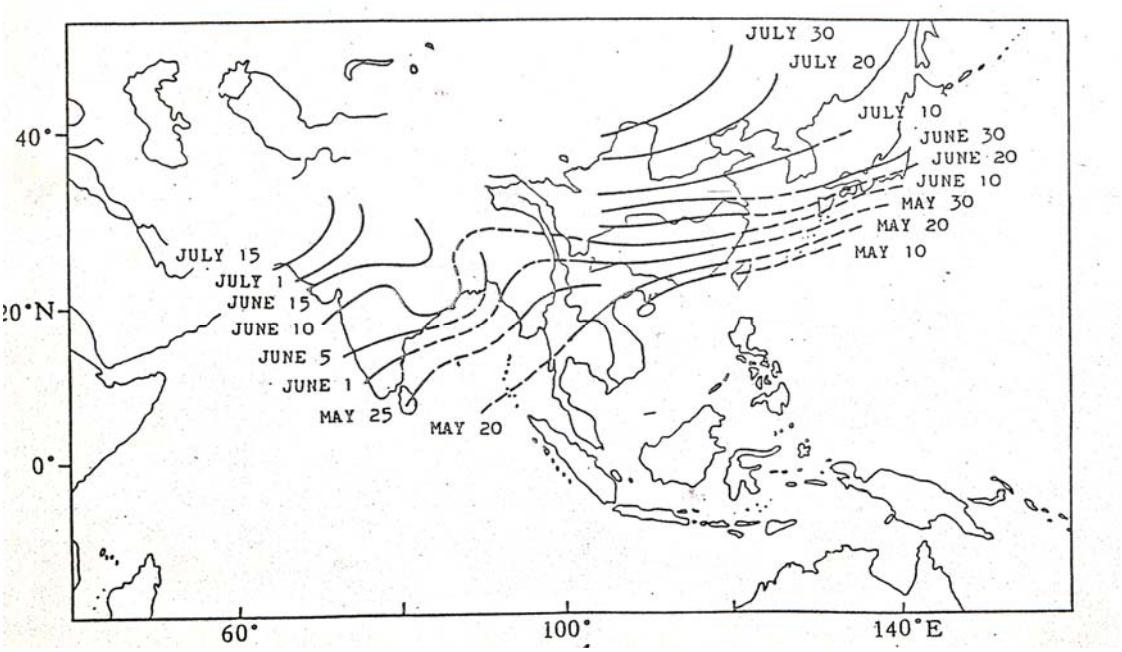


Fig. 6.5: Isochrones of mean onset dates of Asian summer monsoon similar to Fig-6.1a determined from long term rainfall records (from Tao and Chen, 1987)

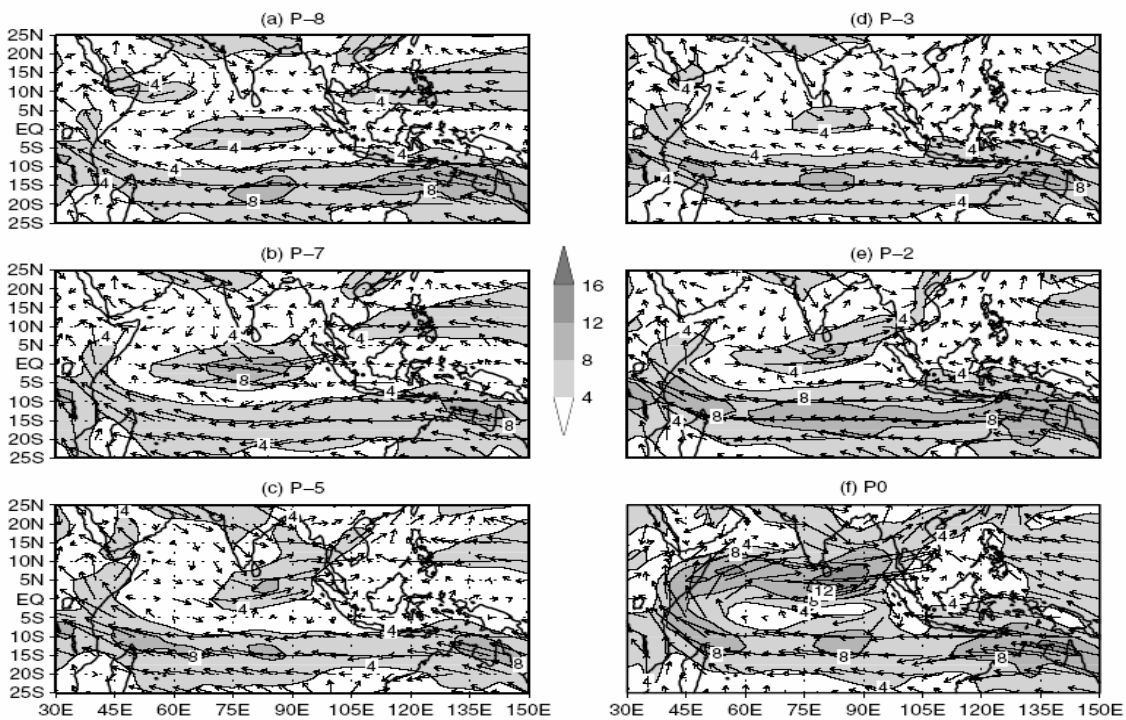


Fig. 6.6: 9 year composite mean 850 hPa winds (m/s) of selected pentads -8 to 0 (MOK at pentad 0) taken from Joseph et al (2006). Contour interval 4 m/s. Shaded area has winds 4 m/s and more

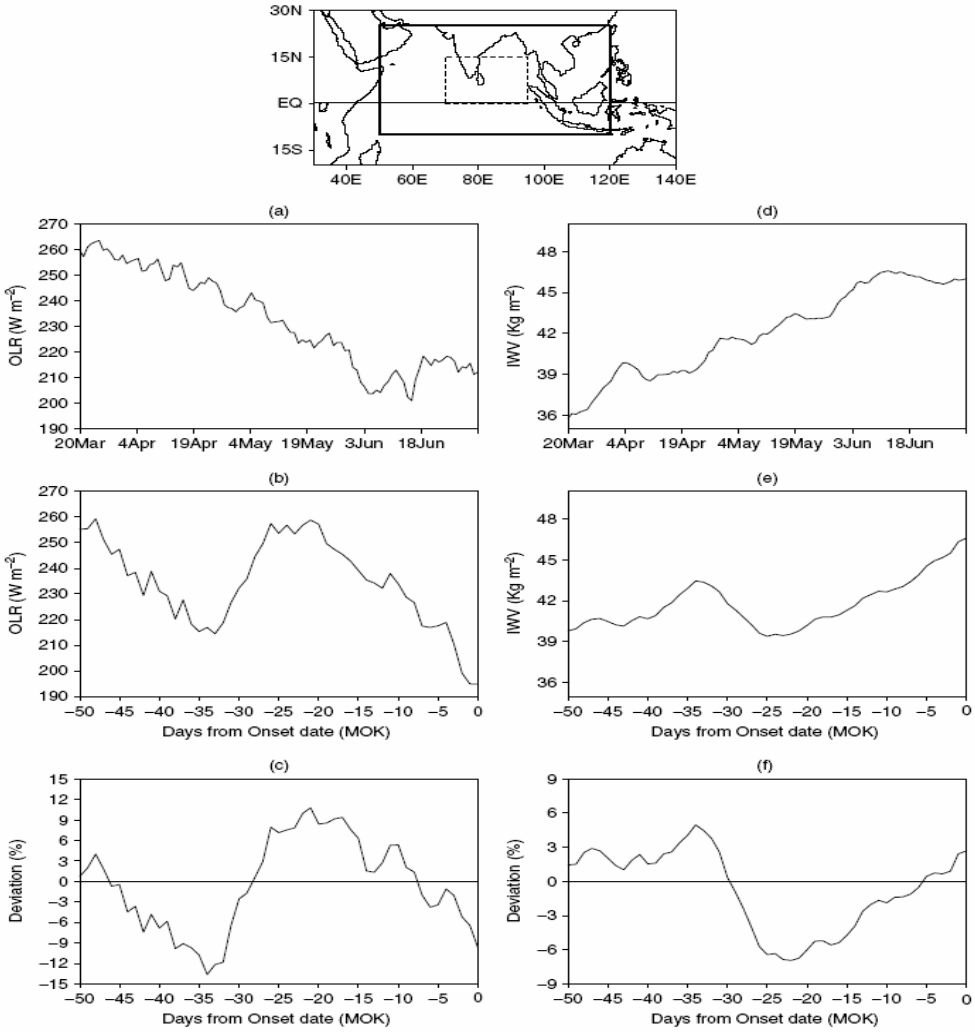


Fig. 6.7: 20 year (1979-1998) average by calendar date of (a) OLR and (d) IWV. Composite mean (b) OLR and (e) IWV of 9 years and their anomaly from the 20 year mean in (c) and (f). All are for the inner box shown by the dotted line in the fig at the top (from Joseph et al, 2006)

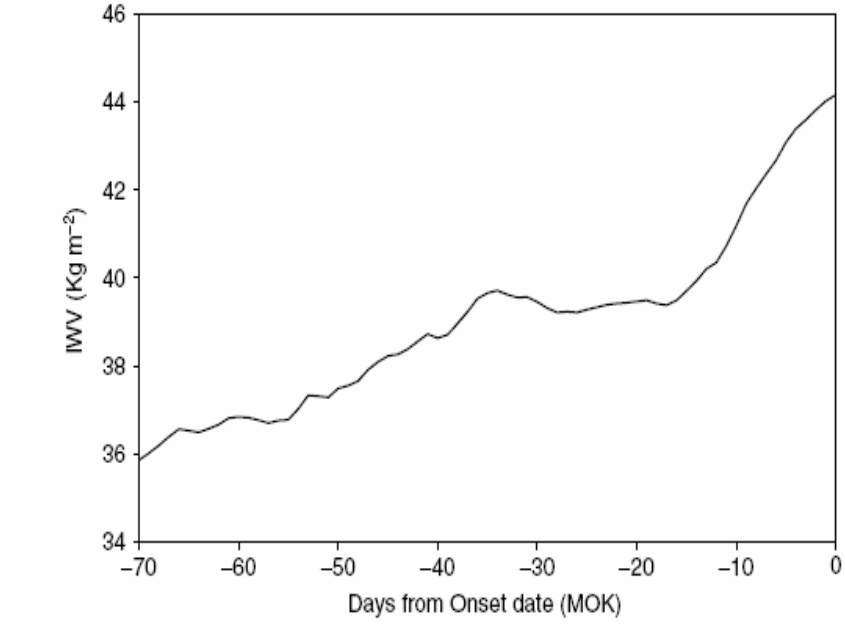


Fig. 6.8: Composite mean IWV in the outer box (of fig-7 top) 50oE-120oE, 10oS-25oN for 9 years (from Joseph et al. 2006)

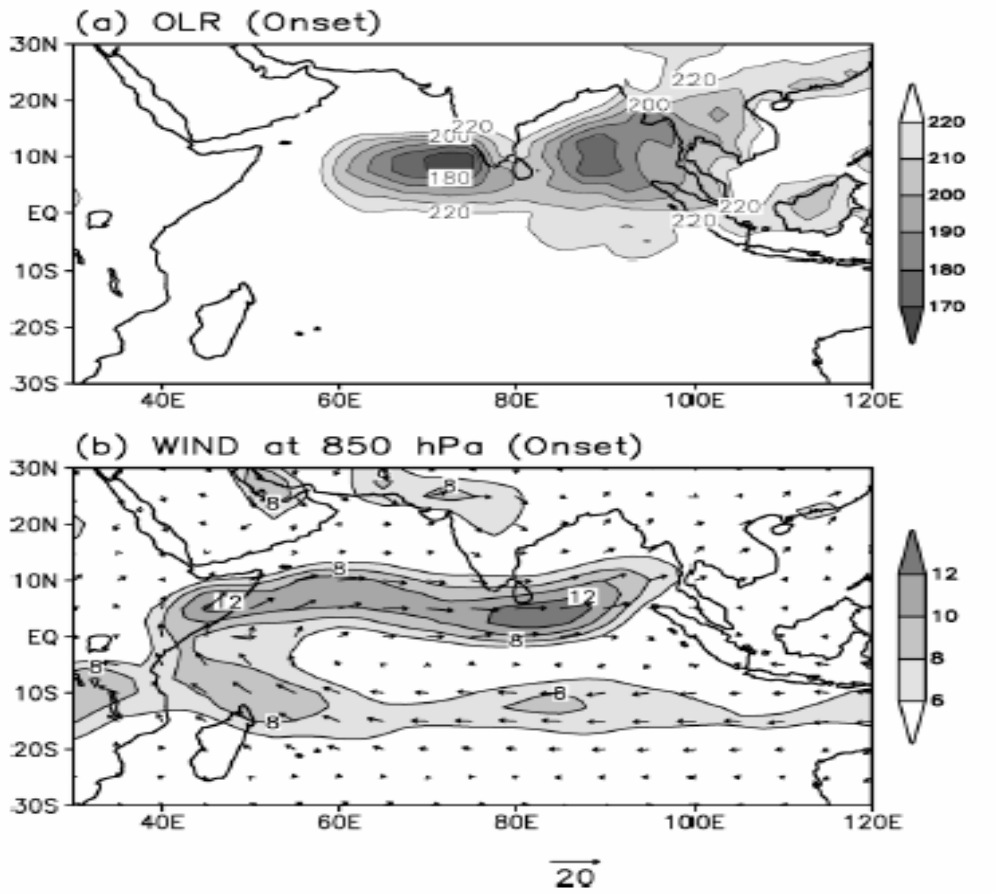


Fig. 6.9: Composites for the pentad around MOK of the 12 years 1979-1990 in (a)OLR with isolines 220 Wm-2 and lower at intervals of 10 Wm-2 and (b) 850 hPa wind vectors and isolines of the wind magnitudes, 6 m/s and more at 2 m/s intervals (from Joseph and Sijikumar, 2004)

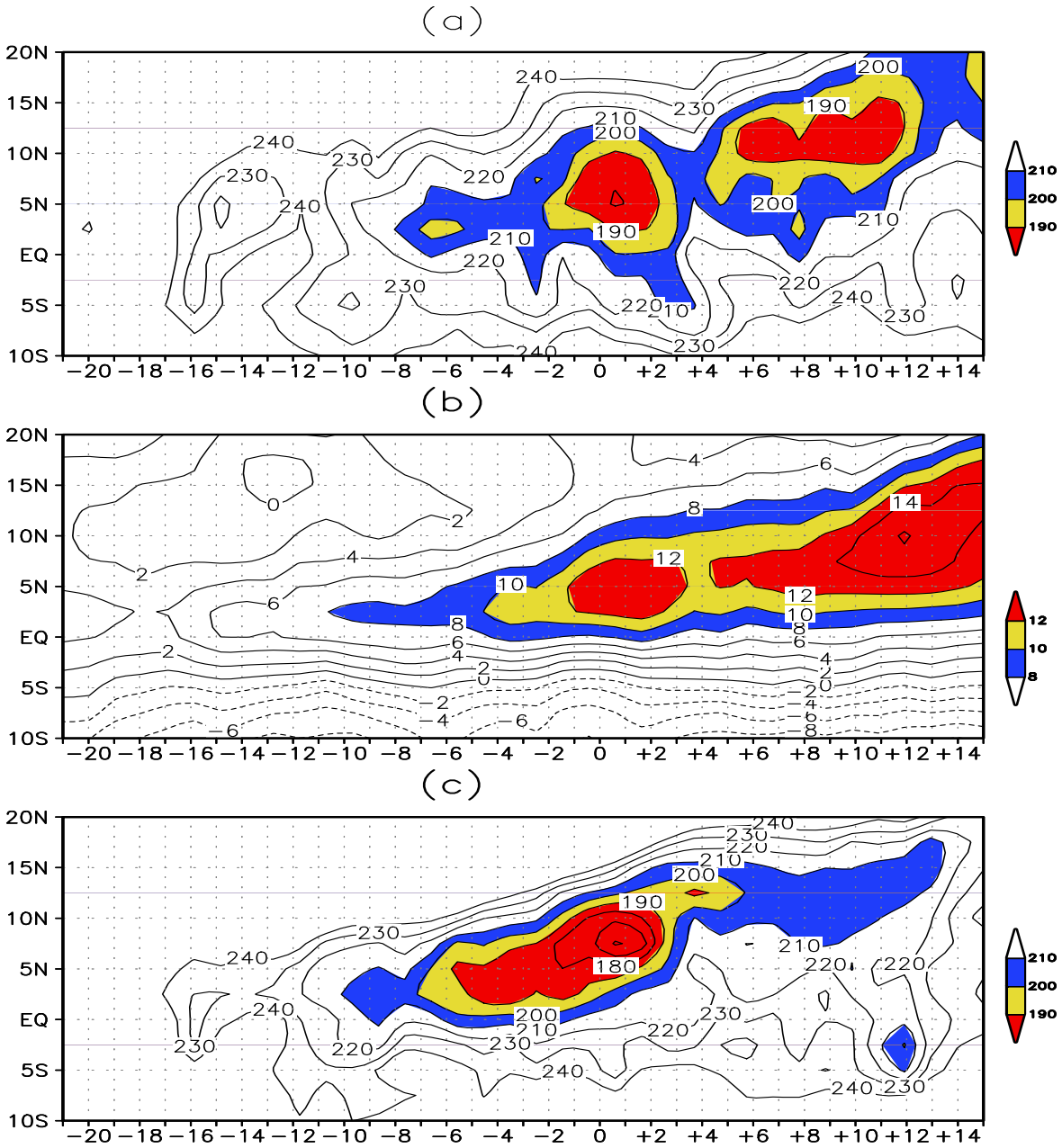


Fig. 6.10: 9 year composite Hovmuller diagrams of daily (a) OLR in Wm-2 and (b) 850 hPa zonal wind in m/s both averaged over 70-85oE and (c) OLR in Wm-2 averaged over 60-70oE. The decrease in convection due to the Arabian sea cooling from one pentad after MOK is seen in (c). The x-axis shows days with respect to MOK as 0 day. (from Joseph et al, 2006)

Composite TMI SST at P-7 (2-6May 2003)

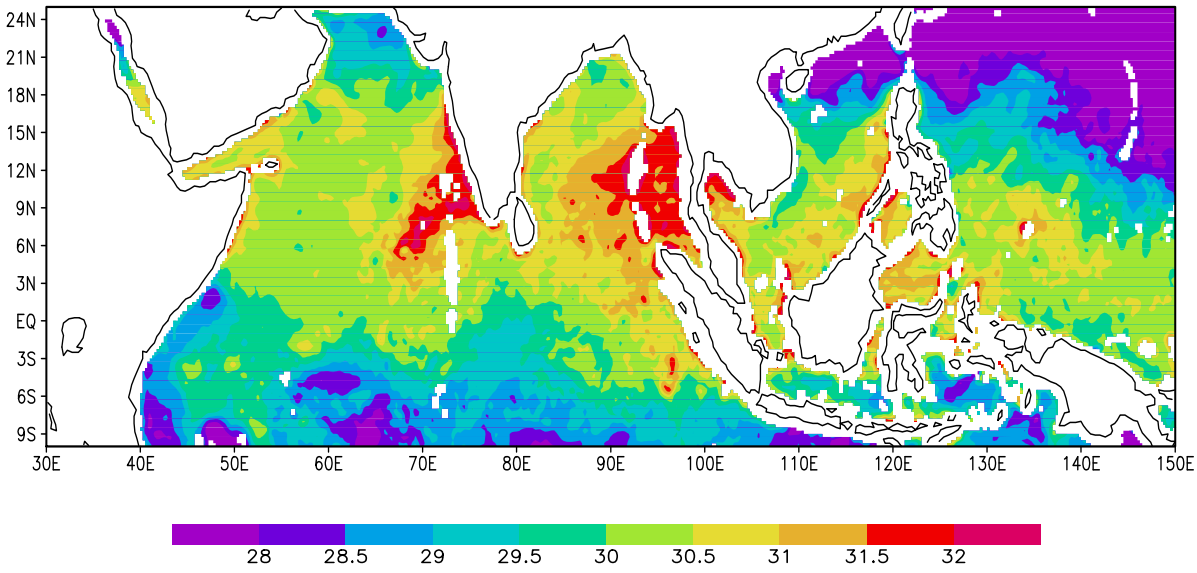
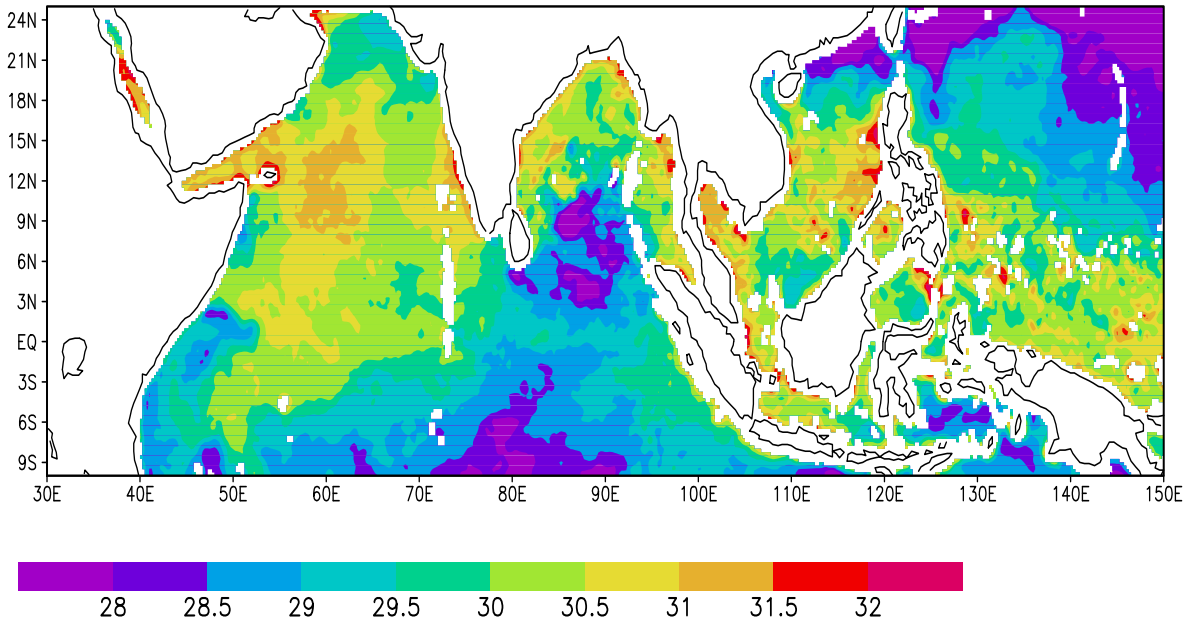
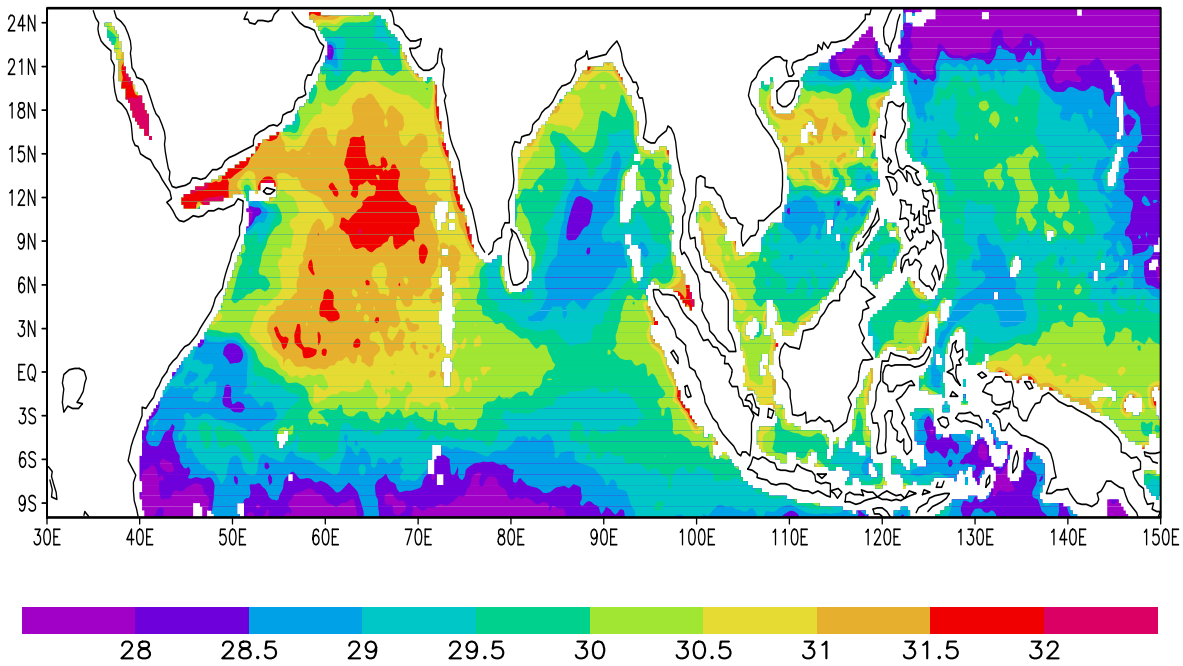


Fig. 6.11 (a)



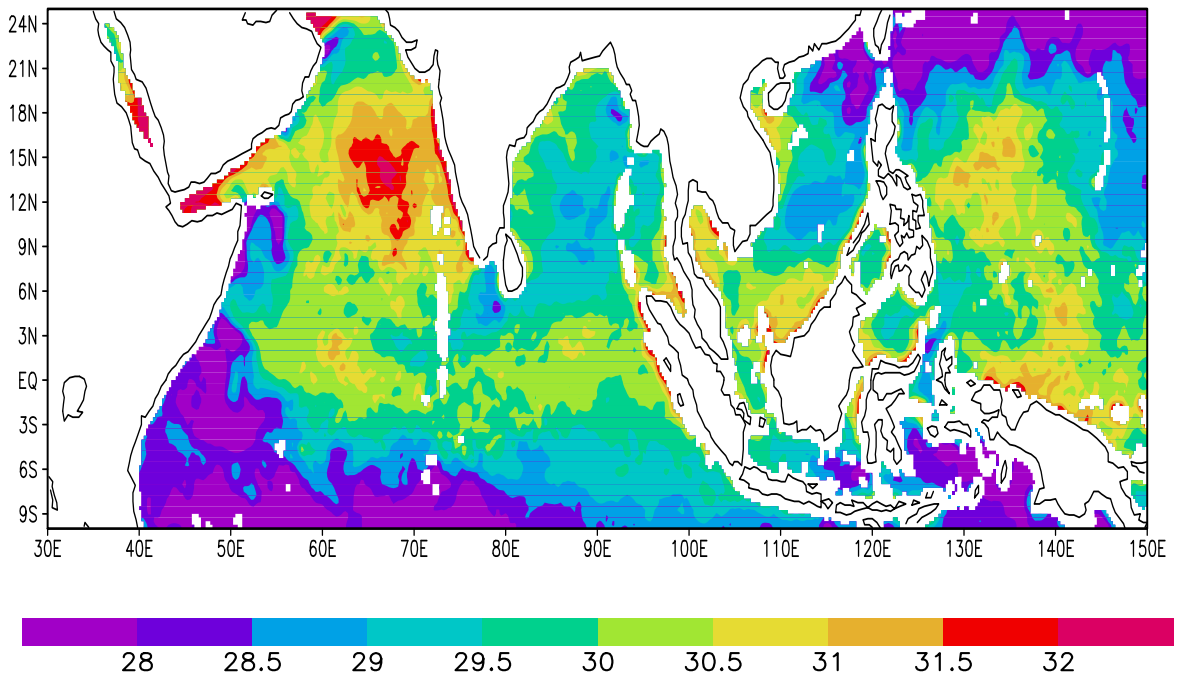
Composite TMI SST at P-5 (12-16May 2003)

Fig. 6.11 (b)



Composite TMI SST at P-2 (27-31May 2003)

Fig. 6.11 (c)



Composite TMI SST at P0 (6-10 June 2003)

Fig. 6.11 (d)

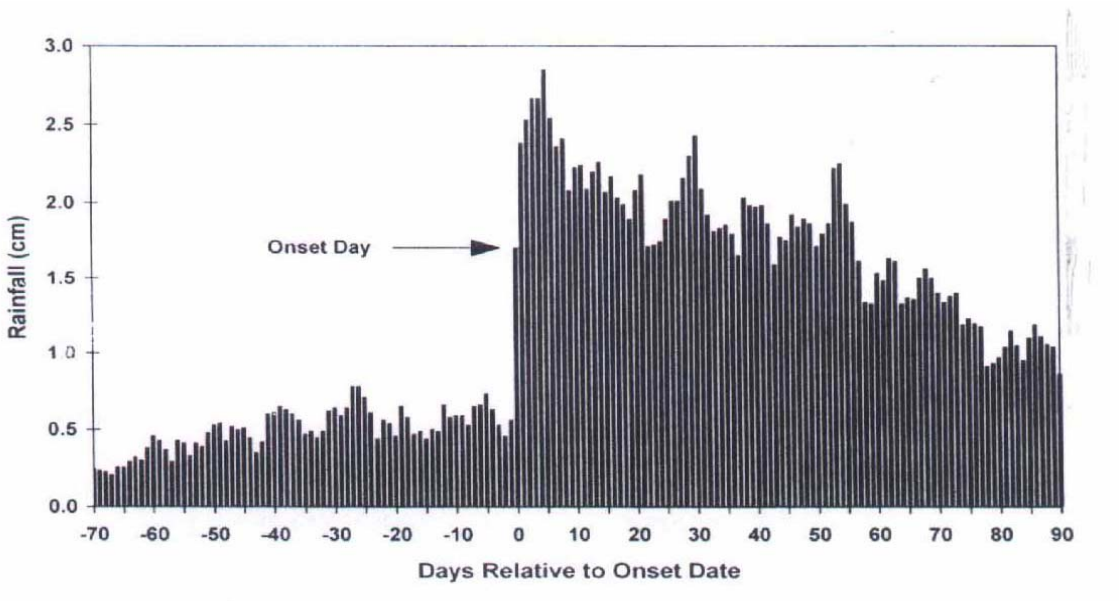


Fig. 6.12: Mean daily rainfall (1901 to 1980) of south Kerala with respect to MOK as 0 day (from Ananthakrishnan and Soman, 1988)

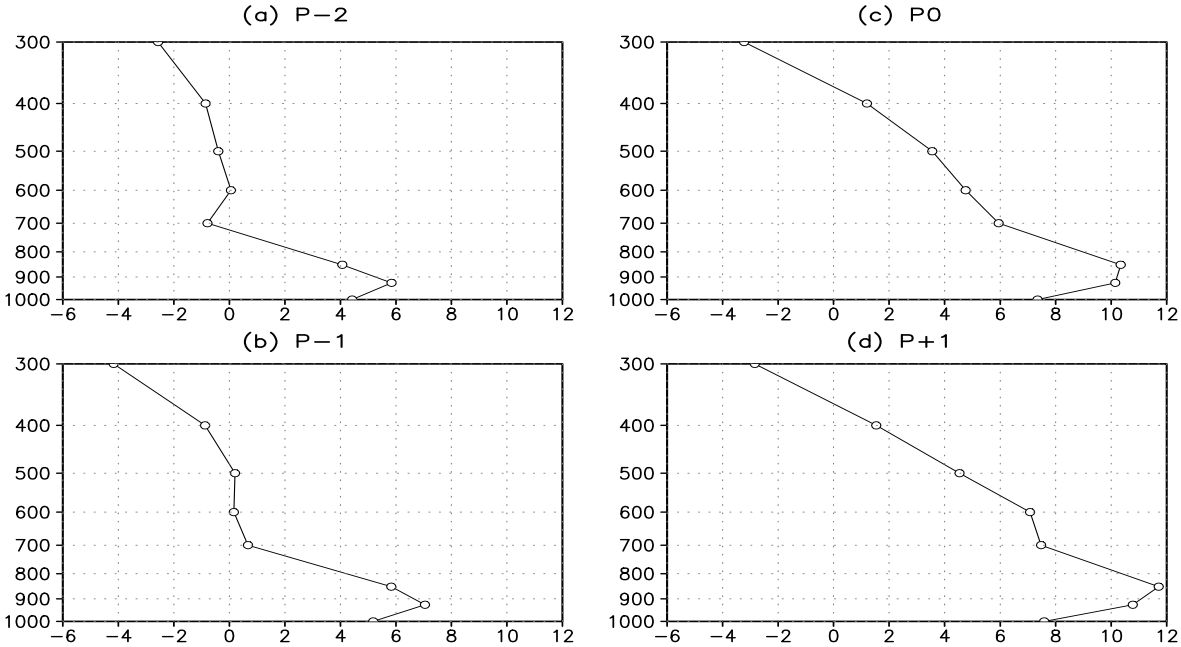


Fig. 6.13: 9-Year Composite vertical structure of pentad U-wind averaged over the area bounded by 70-85oE and 5-10o.(from Joseph et al, 2006)

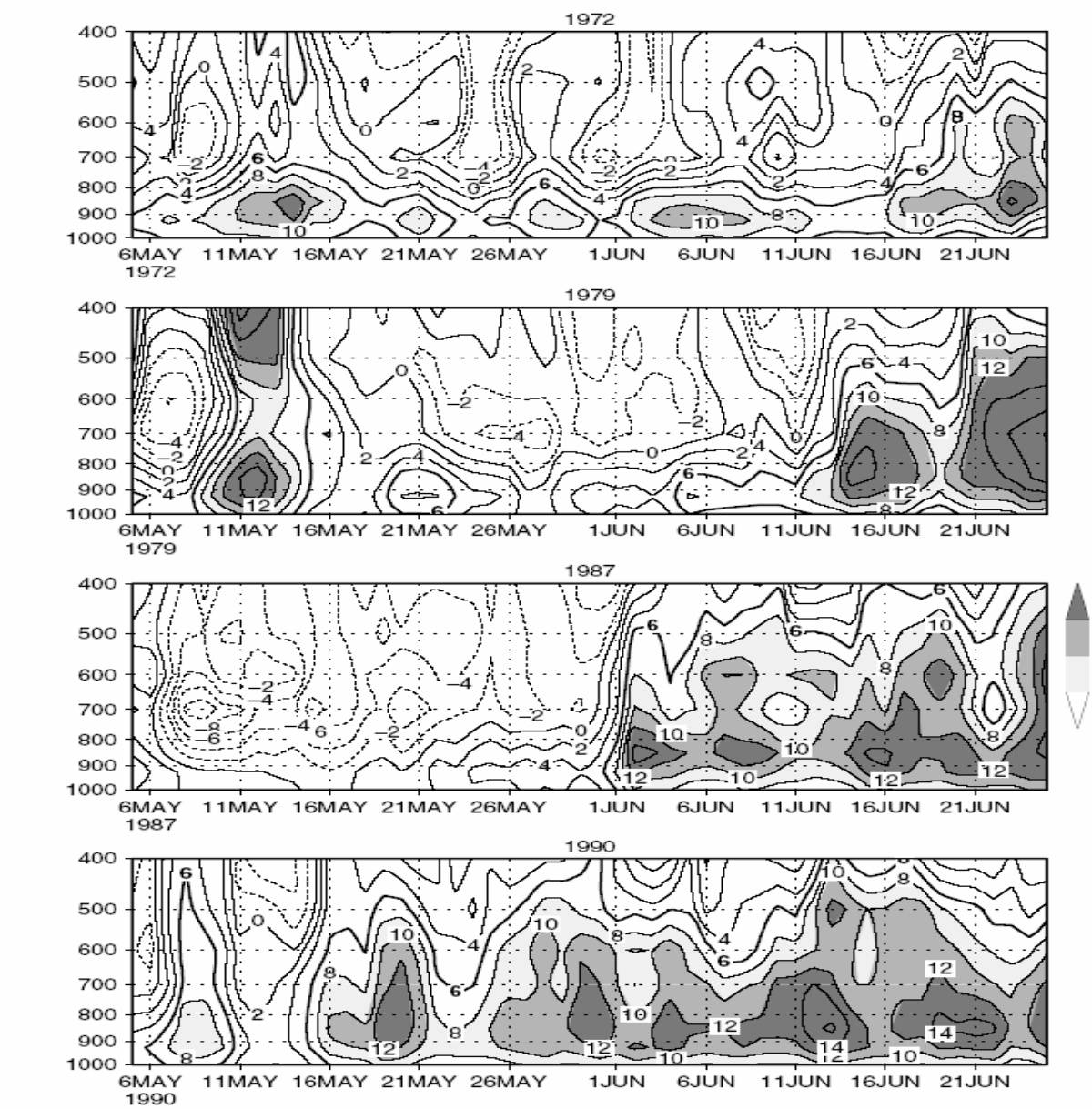


Fig. 6.14: Hovmuller diagram of the daily zonal wind averaged over the box 70-85oE and 5-10oN. Contours are at intervals of 2 m/s. The thick line is the 6 m/s contour. Contours 10 m/s and more are shaded (from Joseph et al, 2006)

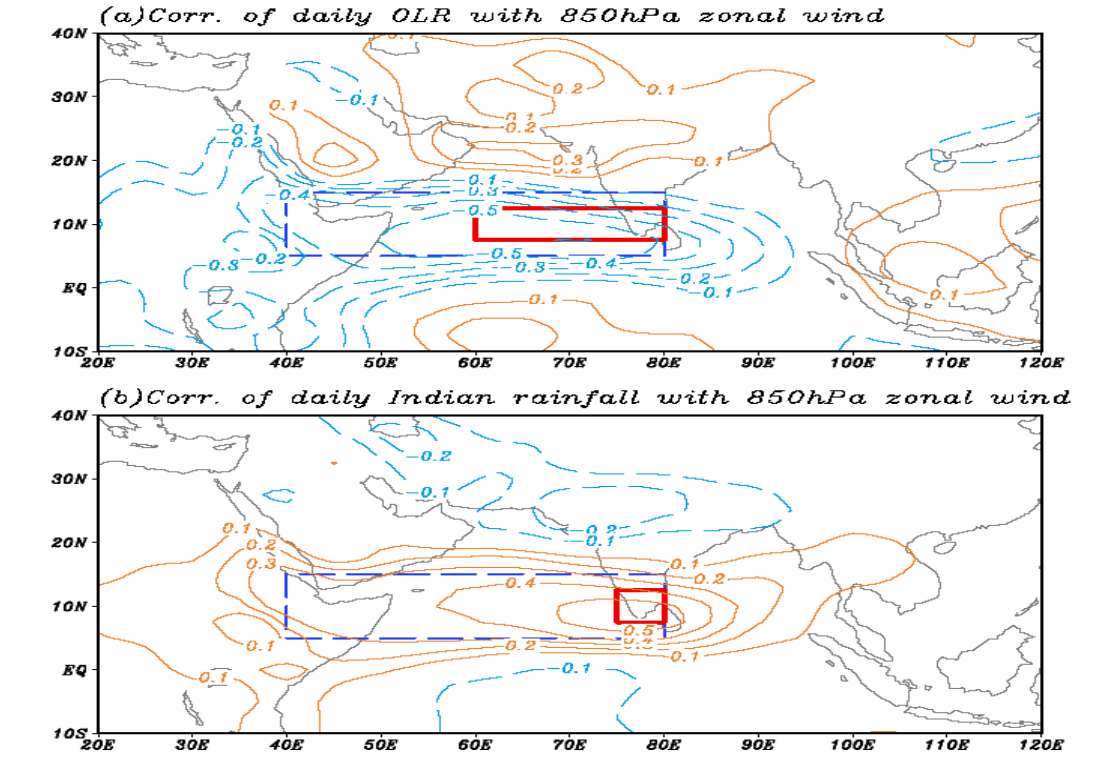


Fig. 6.15: Correlation Coefficient (cc) of (a) daily mean OLR in the red box and (b) daily mean of 1ox1o gridded rainfall in the red box with 850 hPa zonal winds at grid points. CCs are significant in the box with blue broken lines. The mean 850 hPa wind of this box is used to define monsoon onset (Wang et al, 2009)

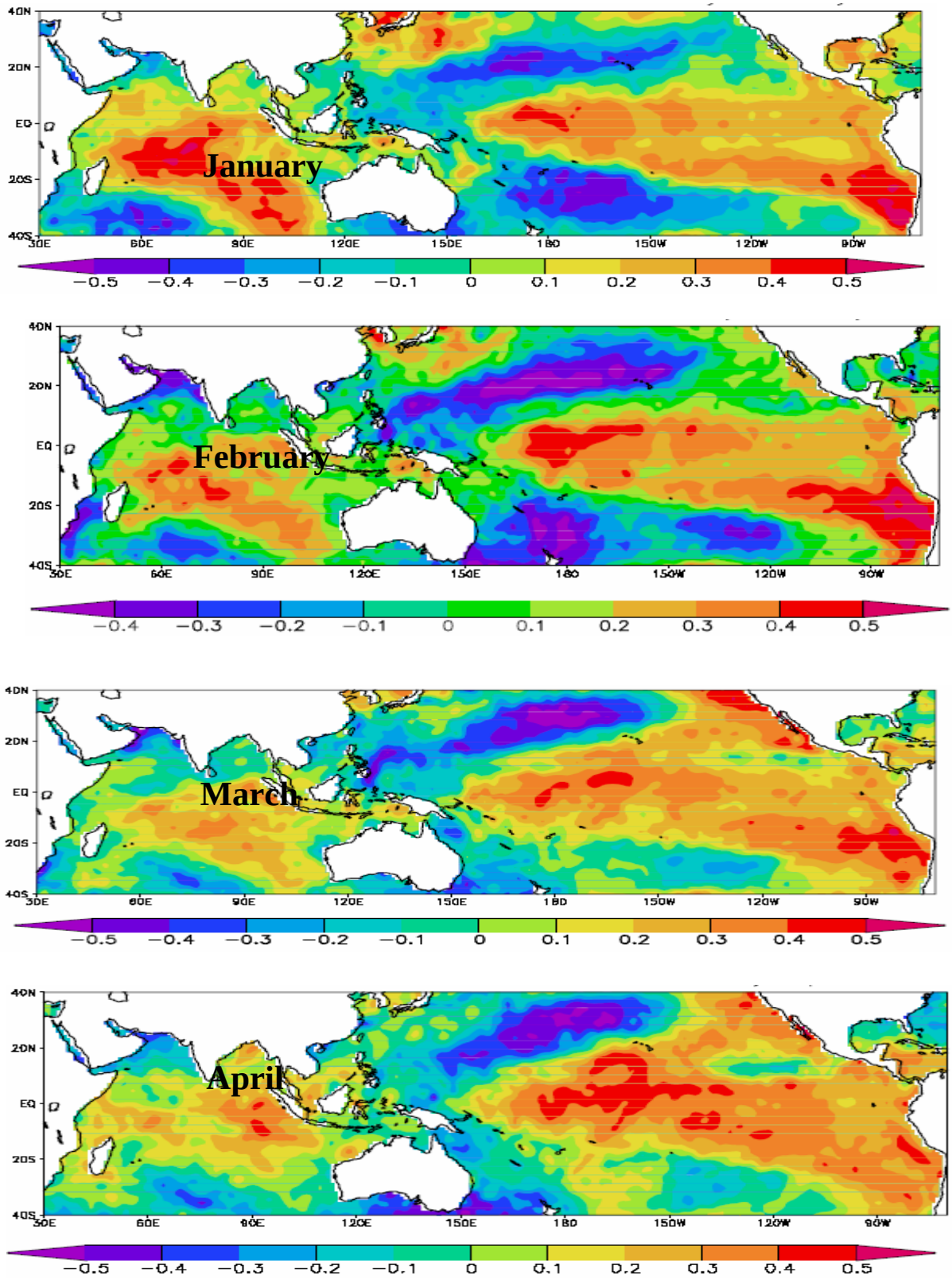
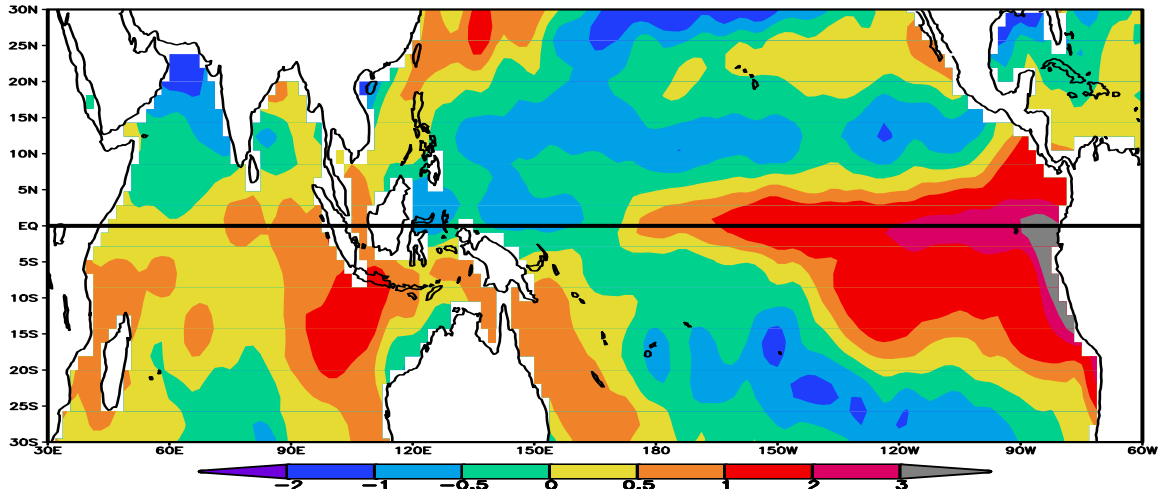


Fig. 6.16: Correlation coefficient between HadISST and IMD's objective dates of MOK (1971- 2002)



SST Anomaly April 1983 – Late Onset 13th June

Fig. 6.17 (a)

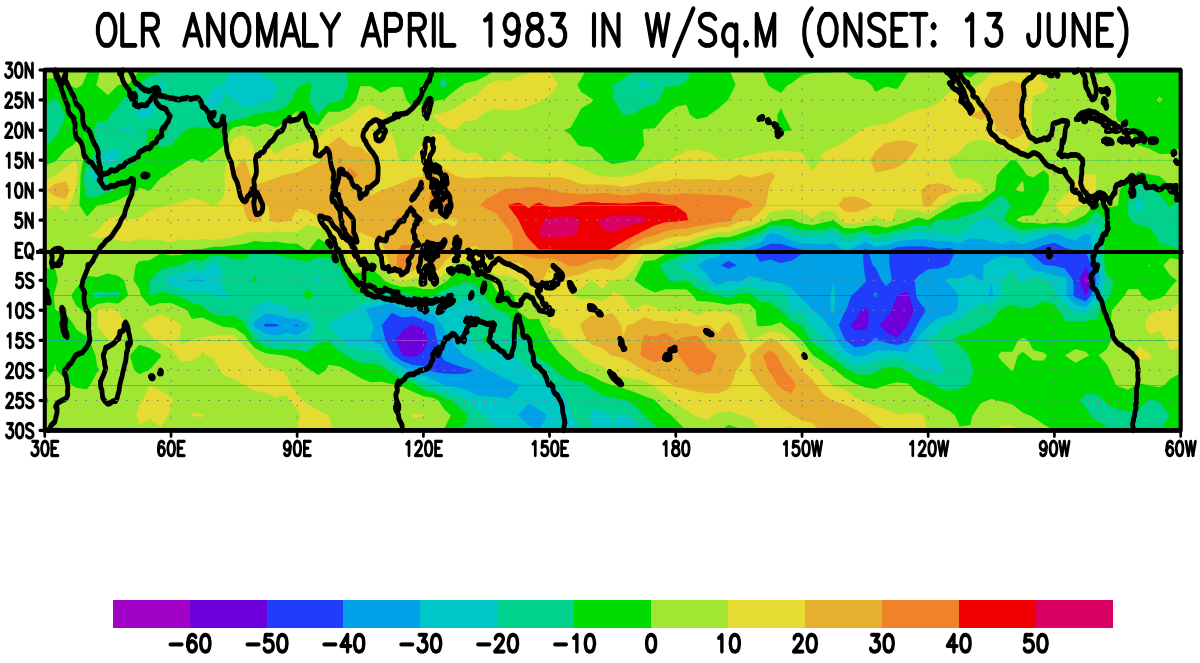


Fig. 6.17 (b)

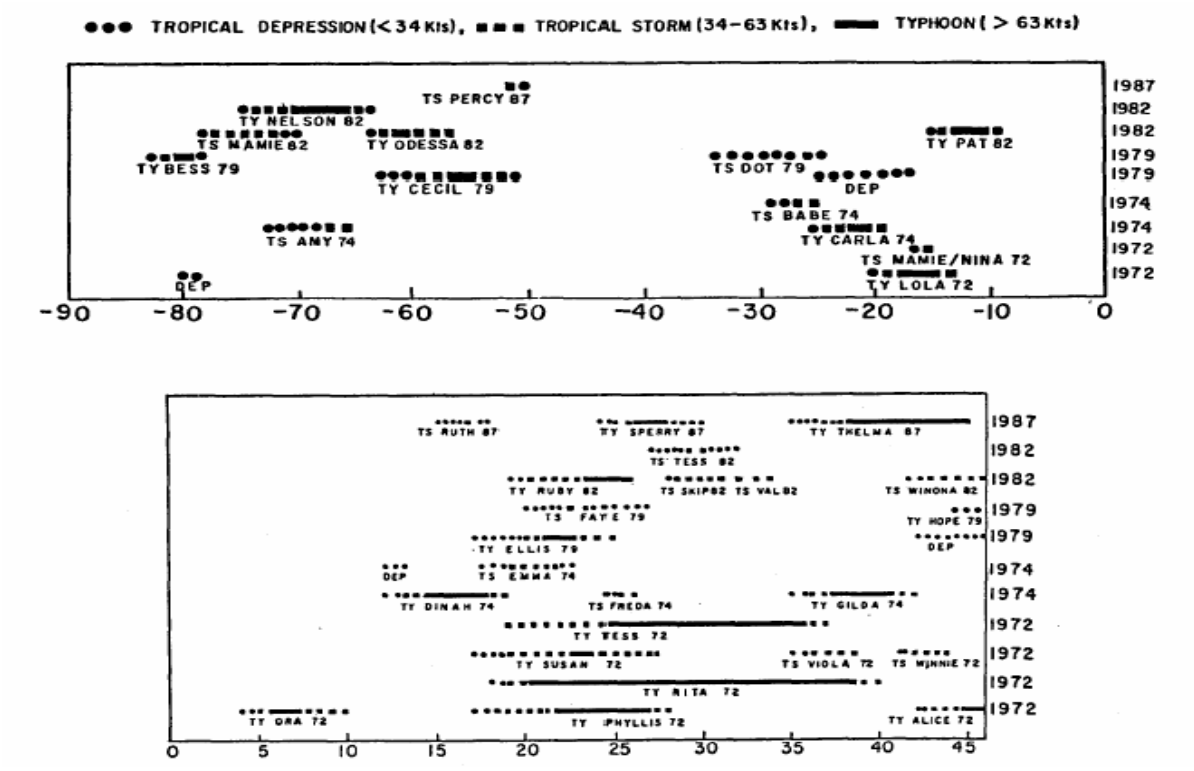


Fig. 6.18: Duration of depression, tropical cyclone and typhoon over western north Pacific ocean with respect to the date of MOK taken as the 0 day from -90 days to +45 days during years of deficient monsoon rainfall of India (from Joseph, 1990a and Joseph et al, 1994)

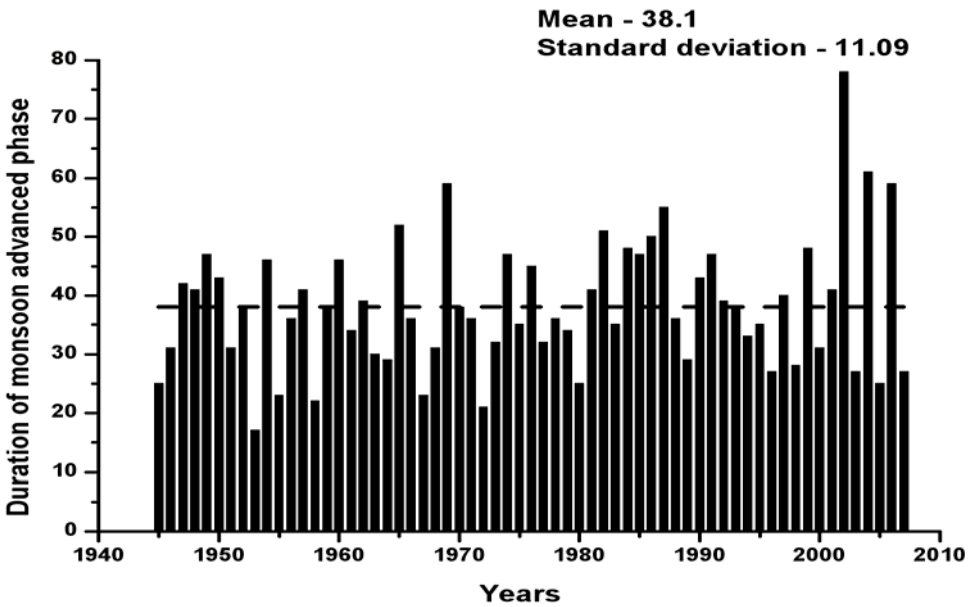


Fig. 6.19: Duration in days of the period from MOK till the monsoon covered the whole of India (advance phase) of the years 1945 to 2007 (Khole, 2009)

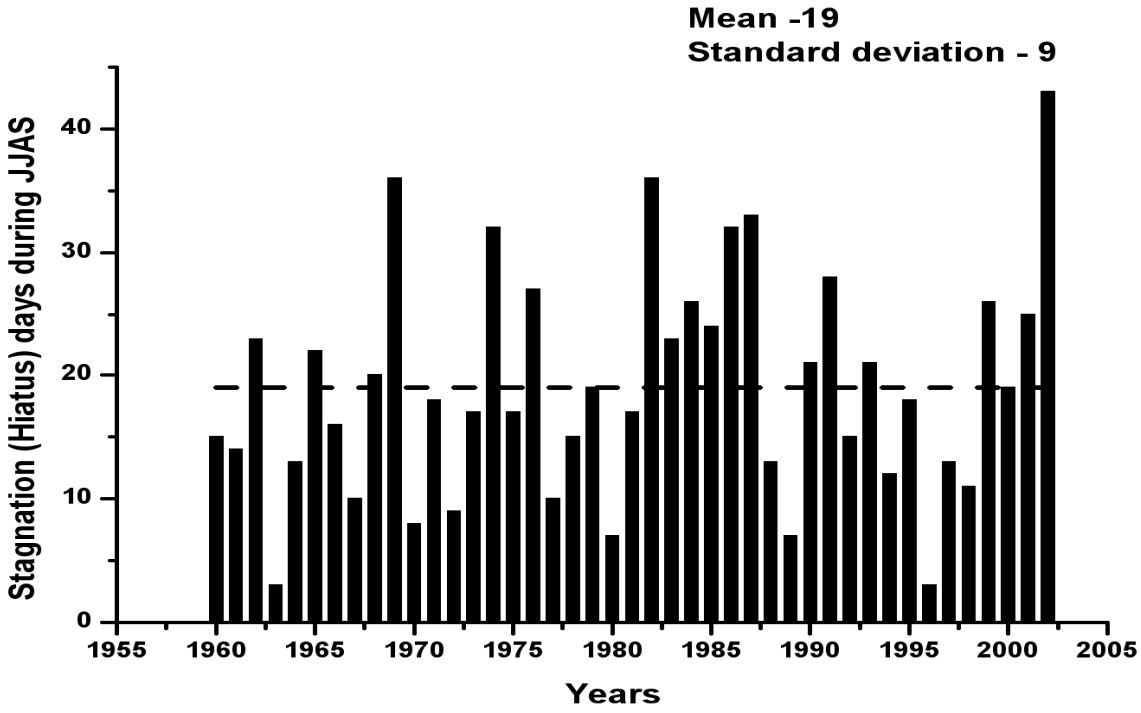


Fig. 6.20: Number of stagnation or hiatus days in the monsoon seasons of 1960 to 2002 (Kalsi et al, 2004)

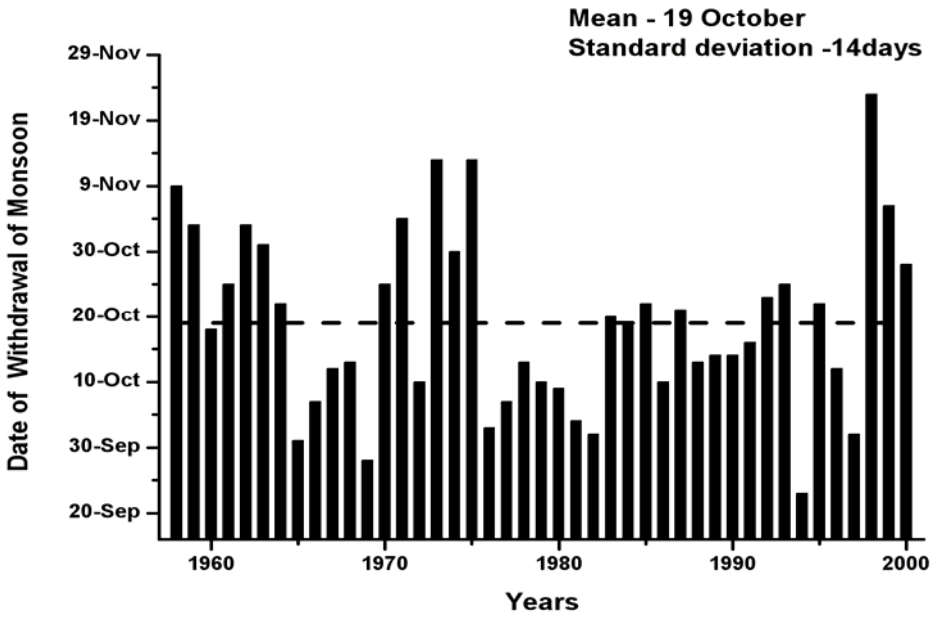


Fig. 6.21: Dates of withdrawal of southwest monsoon from India during 1958 to 2000. (Syroka and Toumi, 2004)

CHAPTER 7

OPERATIONAL WEATHER FORECASTING DURING MONSOON SEASON-TECHNICAL ASPECTS

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7.1. Introduction

Weather plays an important role in almost all aspects of the life on earth. Hence, its accurate and timely forecasting has got wide implications ranging from increasing the Agricultural Production to reducing the damage to life and property.

To monitor and provide weather forecasts on operational basis is one of the principal mandates of India Meteorological Department (IMD). Currently, IMD renders forecasting services to farmers, fishermen, shipping, air navigation etc., apart from the general public. The main objective is to forewarn people so as to reduce the number of deaths and damages from impending natural hazards like floods, cyclonic storms and other inclement weather. The services rendered by forecasting offices are of three types: (i) Monitoring, (ii) Preparation of Climatology and (iii) Forecasting.

The Indian Summer Monsoon is the most important component of Indian weather and climate system. The Indian summer monsoon is characterized by large interannual and intraseasonal variability as regards its onset, withdrawal and spatio-temporal distribution of rainfall. As such, it becomes very pertinent to meticulously monitor the weather developments during the summer monsoon season on an operational basis and describe the same in various forecast bulletins issued by IMD to different user agencies. This facilitates the effective communication of forecasts

and warnings to the users (particularly for disaster management agencies), so as to enable them to initiate appropriate actions for the benefits of the public at large.

In order to maintain the required standards and to bring about uniformity among all forecasting offices, various procedures have been formulated by IMD. These procedures also facilitate the meteorological research community to assess various actions adopted by IMD in an objective manner. These procedures include various criteria for onset and withdrawal of Indian Summer Monsoon, description of spatio-temporal distribution of rainfall during the season, formats for issuance of intensity and activity of weather forecasts and warnings, etc. These operational procedures, which are to be followed by all forecasting offices of IMD during the Indian summer monsoon season, are discussed in the following sections.

7.2. Forecasting Organisation

For the purpose of issuing of weather forecasts, India is divided into 36 regions which are known as meteorological subdivisions. Fig. 7.1 depicts these subdivisions.

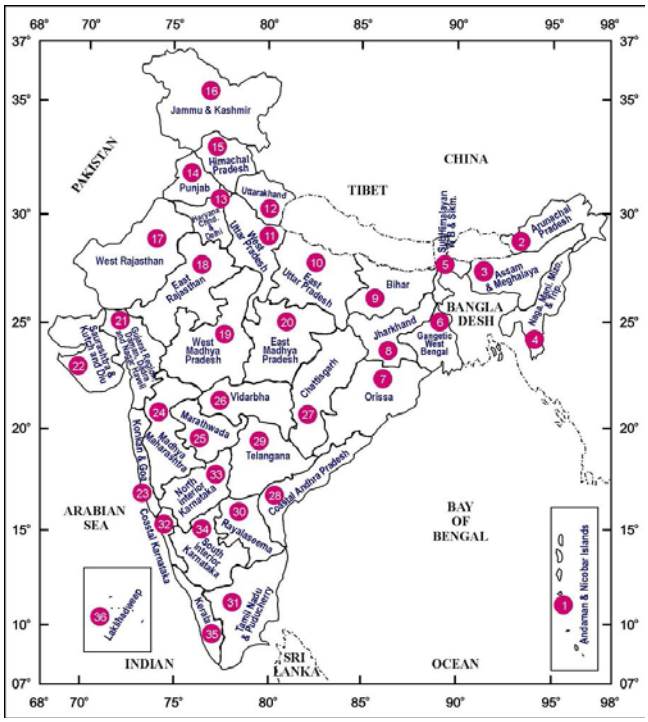

Fig. 7.1: Map showing the 36 meteorological sub-divisions of India

The organisation for providing different types of non-aviation forecasts and warnings is given in the table below:-

S. No.	Category/ Meteorological Offices issuing Weather Forecasting	Details of Service	User(s)
1.	**Marine** RSMC New Delhi	(i) Tropical weather outlooks and Tropical cyclone advisories	Countries in the WMO/ESCAP Panel region bordering the Bay of Bengal and the Arabian Sea
	Cyclone Warning Division, New Delhi Area Cyclone Warning Centres (ACWCs) at Kolkata (Alipore), Mumbai (Colaba), Chennai (Nungambakkam), Cyclone Warning Centres (CWCs) Ahmedabad, Bhubaneswar & Visakhapatnam	(i) Forecasts for Bay of Bengal & Arabian Sea. (ii) Coastal forecasts (iii) Cyclone Warnings (iv) Port Warnings	Ships Ships, Govt. Deptts. Maritime State & Public Ports
	Area Cyclone Warning Centre (ACWCs) at Mumbai, Kolkata & INOSHAC, Pune	Fleet Forecast twice a day, frequency of bulletins increases to four during tropical storm period for Arabian Sea, Bay of Bengal and Indian Ocean upto 10°S (60°E to 100°E). Issuance of sector wise fleet forecast of wind speed, wind direction and weather over the regions 5°N - 10°S, 60°E-100°E	Indian Navy
	Global Maritime Distress and Safety System (GMDSS), ACWCs Mumbai, Kolkata & Indian Ocean and southern hemispheric Analysis Centre, Pune.	Bulletins twice a day are issued for Meteorological area VIII N for Arabian Sea, Bay of Bengal and Indian Ocean.To north of Equator, the frequency increases to six during tropical storm period	All ships.

2.	**Fisheries** Area Cyclone Warnings Centres (ACWCs) at Kolkata (Alipore), Mumbai (Colaba), Chennai (Nungambakkam), Cyclone Warning Centres (CWCs) Ahmedabad, Bhubaneswar, Visakhapatnam	(i) Adverse weather along coast, 4 times daily (including nil warnings) (ii) Cyclone Warning	(i) Fishing Craft through A.I.R. (ii) Fisheries

7.3. Forecasts and bulletins issued by various meteorological centres

The responsibility of issuing forecasts, bulletins and warnings by various meteorological offices differ widely. The ACWCs and CWCs involved in the cyclone warning work have an additional set of bulletins and forecasts to be issued by them, which are elaborately described in Chapter 9 of the Cyclone Manual, viz., 'Bulletins and Warnings'. Similarly, those meteorological offices attached to the aerodromes are to follow an entirely different routine of Bulletins and Warnings to ensure the safety, regularity and efficiency of air navigation.

The forecasting responsibilities of all other Meteorological Centres (MCs), Regional Meteorological Centres (RMCs), Weather Central, Pune and NHAC, New Delhi, are also well defined and differ from one another. All these centres are to issue forecasts twice a day based on the morning and evening charts. Weather Central, Pune issues the India Weather Bulletin (IWB) twice a day and All India Weather Summary on a daily basis. Northern hemispheric Analysis Centre (NHAC), New Delhi, considering inputs from all the regional inferences and IWB, issues the National air news to the media twice daily. Various such additional tools as the numerical model outputs, satellite and RADAR products, Automatic Weather Stations' data, etc., are utilized for issuing all such forecasts.

7.4. Weather Warnings

7.4.1. Warnings for the designated/registered users (Album Page Warnees)

Each RMC/MC has a list of designated/registered users (previously known as Album Page Warnees) with specified limits of meteorological parameters like, rainfall or temperature. If the rainfall or temperature is forecast to be beyond the range of these specified limits, then the RMC/ MC is to issue corresponding warning to the concerned users/ Album Page Warnees. This list needs to be kept updated and registers should be maintained for the warnings issued.

A) Text of warning for heavy to very heavy rainfall issued to Album Page Warnees.

The Album Page Warnings are to mention the actual limits in the text of warning for heavy to very heavy rainfall and other parameters for which warnings are issued to the Album page warnees.

Example of Heavy Rainfall warning for Album Page Warnee whose requirement for rainfall is 5 cm or more is:-

Rather Heavy / Heavy rain/ very heavy rain may occur at one or two/ few/ many/ almost at all the places in your district when rainfall will exceed 5 cm or more till_____ hrs IST of _____ date".

B) Inclusion of date and time of cessation of heavy rainfall warning.

The Album Page Warnings are to mention the ending time and date in the text of the heavy rainfall warnings.

Example :- Heavy rain/ very heavy rain may occur at one or two/ few/ many/ almost at all the places in _____ districts/ subdivisions till _____ hrs. IST of date".

7.4.2. Farmers' Weather Bulletins (FWBs)

A second Farmers' Weather Bulletin is issued to the AIR stations at night based on 1200 UTC charts for broadcast early in the next morning (the first FWB is based on 0300 UTC charts). The full text of the second Farmers' Weather Bulletin is issued by about 2200 IST based on 1200 UTC charts (or later based on 1800 UTC charts, if special night watch is kept on any particular day), to all the AIR stations which receive the FWB at present. The second FWB is issued daily during the monsoon months (June to September) and selected other months which are considered important for agriculturists in the different regions. The Meteorologists-in-charge discuss with the farm and Home Service Unit of the local AIR and decide on the months outside the monsoon season during which the second FWB will be necessary and issue the second bulletin accordingly. The major points related to this are summarized below:

1. A second Farmers' Weather Bulletin is issued during the monsoon months by all the forecasting offices.
2. The second bulletin is based on the 1730 hrs. IST charts and is issued by the concerned Meteorological offices by about 2200 hrs. IST. It is broadcast early next morning at a time depending upon the working hours of the different AIR stations so that maximum use of this bulletin could be made by the agriculturists/ farmers, who hear them before they begin their day's work.
3. Further, as there are periods other than the monsoon months which are important in different parts of India for agricultural operations, the second Farmers' Weather Bulletin is issued during the significant months other than the southwest monsoon season [e.g. (i) October – December in south India when considerable rainfall occurs in these months, (ii) The winter months over northwest India when rains and cold waves occur in these parts of the country]. The period is decided through mutual consultations between the representatives of the local Regional Meteorological Centres and AIR authorities at those places.
4. The second Farmers' Weather bulletin has the preamble "Farmers' Weather Bulletin for ……(area) valid until the evening of …….(date). For

example, the FWB issued on the night of 3rd October for broadcast on 4th morning will be valid up to 5th evening.

5. M.Cs/ RMCs which keep regular night watch, base the second FWB on 1800 Z charts but ensure that the FWB will reach the A.I.R. stations in time for the morning broadcast. MCs/RMCs which keep special night watch during periods of disturbed weather also issue the second FWB based on 1800 Z charts on those occasions.

7.4.3. Preparation of Weekly Weather Reports

1. Weather Central, Pune and Northern Hemispheric Analysis Center (NHAC), New Delhi, prepare the All India Weekly Weather Reports. Also, all Meteorological Centres prepare district wise Weekly Weather Reports for each state.
2. The description of the week's and season's rainfall is done on district scale by the MCs and for this purpose, isopleths of week's rainfall departure and season's rainfall departure (both in percentage) are prepared on the special rainfall charts by RMCs and MCs.
3. Under the heading "Chief Synoptic Features and associated Weather", spatial distribution of rainfall in the different Meteorological sub-divisions of the state during the week are also briefly summarised. The activity of the monsoon, floods and damages caused are also included under this heading.
4. On page 2 of the report, the stations are arranged alphabetically against each district under the different Meteorological/ Revenue sub-divisions. The stations on page 2 of the report includes all the observatories in the state and state rain gauge stations from which daily rainfall data are received and for which daily accumulated rainfall normal are available.
5. For computing the week's rainfall for all the stations, the daily rainfall data of all the stations received are tabulated in suitable registers. The season's rainfall total for each station is updated at the beginning of every month on receipt of the monthly total for the previous month from each station as well as for each week on receipt of late data of the previous week.
6. On the occasions when significant rainfall occurs, a few copies of Weekly Weather Reports along with coloured maps showing district wise weekly

rainfall departure from normal are prepared and supplied to a few VIPs like the Governor, the Chief Minister, Minister for Agriculture, etc. These maps are also published in the local press so that the public in general becomes aware of the rainfall situation, particularly during the monsoon season.

7.4.4. Free supply of Regional Daily Weather Reports (RDWRs)

The following broad guide lines have been drawn up for regulating the free supply of Regional Daily Weather Reports, particularly to Govt. Officials (non-departmental).

1. Free supply of RDWR to Government Officials (Non-departmental)

1.1 Fresh free supply may be undertaken only on receipt of a specific request from the party.

1.2 An exception may be made in the case of dignitaries like the President of the Indian Union, Governors, Prime Minister, Chief Ministers, Ministers of the Central and State Governments, who are directly or indirectly interested in Meteorology.

1.3 Head of the concerned Postal Circle: According to existing rules of registration of newspapers for postal concession, a copy of the news paper has to be submitted to the head of the Postal circle.

1.4 National libraries, to which all publications are supplied free.

1.5 Officials/ Offices who are entitled to get weather information for operational reference/research purposes. Under this category, the Army, Navy and Air Force establishments including the HQrs of the three services and other Government Officials, who require weather data daily for their work are included. The Regional Daily Weather Reports would be a convenient medium for supplying the weather information which they are normally entitled to receive. However, no two Officials/ Offices of the same Ministry/ Department stationed at the same place are normally entitled for free supply.

1.6 Offices/ Officials to be supplied on exchange basis: This category would include mainly the meteorological services of the neighbouring countries. As meteorological information from these countries will be useful for the

department, RDWR will have to be supplied to them in exchange for their reports.

1.7 The agencies from whom the department gets free services, such as the heads of departments in charge of rain recording stations in states as the rainfall reports from the states are received by this department free of cost. These agencies may be supplied with the RDWR of the concerned region free of cost on request. Also, Honorary Superintendents of part-time observatories maintained by this department may be supplied on request, free of cost a copy of the RDWR in recognition of the free services rendered by them to the Department.

1.8 The recipient may be supplied with the report of the region of his interest.

2. Free supply of RDWR to news papers and news agencies

2.1 Free supply of the RDWR of the concerned region may be made to such news papers and news agencies, who are on the list of accredited news papers/ news agencies of the State Governments concerned, for providing publicity to weather reports. A free supply may be made on the understanding that the parties concerned make their own arrangements for collection of the report at no extra cost to the department. The news papers getting a copy of the report should also supply a complementary copy of the paper to the office concerned for keeping a check on the weather data published.

2.2 The supply to a news paper/ news agency may cease when it stops publication of the report.

2.3. All fresh supplies will have to be undertaken only with the approval of the DDGM (WF), IMD, Pune.

2.4. The distribution list of RDWR as on the 1st day of the year need not be sent to HQrs office. However, this return and the quarterly changes need to be sent to DDGM (WF), Pune. Also the returns about Regional Weather Summaries/ FWB/ AIR schedules need to be sent to HQrs office as well as to DDGM (WF), Pune.

7.4.5. Compactness of Weather Bulletins supplied to All India Radio

In order to make the bulletins brief and more appealing, the following guide lines are to be adopted:

(a) Messages should be edited in such a manner as to bring out the important features. A newspaper type featuring should be our guide in this respect.

(b) The terms like rain or thundershowers should be avoided. In such cases, the phenomenon is to be described as rain.

(c) In describing spatial distribution, if differences between some sub-divisions are only marginal, they can be suitably combined. It has to be remembered that spatial distribution determined from the few departmental observatories may differ to some extent from the state rain gauge data.

(d) When a large number of stations have reported heavy rainfall in the region, the lower limit of rainfall amounts may be restricted so that a large number of stations need not be listed in the summary, ensuring however, that rainfall figures from all the Meteorological subdivisions are given. On occasions when heavy rainfall reports are few, discretion may be used to select one or two significant amounts per meteorological sub-division, depending upon the number of such sub-divisions in the concerned RMC.

(e) In the summary of observations, the information in respect of changes or departures of maximum and minimum temperatures may be restricted to plus or minus 3^0C or above only.

(f) When bulletins are becoming rather lengthy, insignificant features, such as isolated rainfall in monsoon season, can be omitted.

(g) Whenever monsoon is described as strong or vigorous in an area, it is not necessary again to mention spatial distribution of rainfall for that area.

(h) The descriptions of temperature and rainfall distribution in the summary may be confined to the Meteorological sub-divisions without further dividing the sub-divisions for this purpose except when there is marked difference between parts of the same sub-division.

7.4.6. Use of Uniform Normal figures for all the meteorological elements

In order to bring in more objectivity and transparency, a unified normal period (viz. 1941 – 1990; the climatological period as per the WMO norms) is used for rainfall and a normal period of 1961 – 1990 for the meteorological elements like maximum temperature, minimum temperature and rainfall is adopted.

7.5. Dissemination of Bulletins and Warnings:

This is the most important part of weather forecasting services. The forecasts should reach the public and user agencies in time and without any distortion. For this purpose, forecasting offices may make use of all the available means including mobile phone, internet services, etc. as the situation demands.

7.6. Onset/ Withdrawal of SW Monsoon

The arrival of monsoon current over the Andaman Seas, its onset over the mainland and further advance need to be monitored closely. During the onset/ withdrawal phase, bulletins are issued by Weather Central, Pune, RMCs and MCs which are posted on their respective websites as well as disseminated by the media. To avoid contradictory bulletins to the public from various meteorological offices, Weather Central, Pune, will issue an advisory bulletin [known as Tentative Bulletin] at about 1130 hrs IST by Fax/ AMSS to RMCs/MCs regarding the advance/ withdrawal of monsoon pertaining to different meteorological subdivisions as and when occasion arises. This will be prefixed by the wordings, 'Tentative Bulletin dated -……..' and will be based on the 0000 UTC and 0300 UTC charts as well as telephonic discussions among Director, Weather Central, Director-in-Charge of the concerned RMCs/ MCs and NHAC, New Delhi.

7.6.1. Onset Phase

The monitoring commences from 10th May onwards, every year. MC Thiruvananthapuram, SATMET (New Delhi) and NHAC (New Delhi) are required to send the specified data, as given in the forecasting circular No. 1/2006 (Appendix

5.1.2) to the O/o DDGM (WF) via e-mail/Fax on a regular basis, till the onset is declared over Kerala.

The guidelines to be followed for declaring the onset of monsoon over Kerala and its further advance over the country are enlisted below:

I. Onset over Kerala

a) Rainfall

If after 10th May, 60% of the available 14 stations enlisted*, viz. Minicoy, Amini, Thiruvananthapuram, Punalur, Kollam, Allapuzha, Kottayam, Kochi, Trissur, Kozhikode, Talassery, Cannur, Kasargode and Mangalore report rainfall of 2.5 mm or more for two consecutive days, the onset over Kerala be declared on the 2nd day, provided the following criteria are also in concurrence. (*Station Locations are shown in Fig. 7.2; also subject to modification considering the regularity of availability of observations on operational mode).

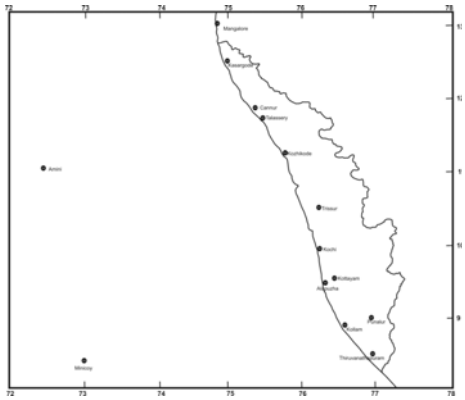

Fig. 7.2: Locations of the stations for monitoring rainfall for declaration of onset of SW monsoon over Kerala

b) Wind field

Depth of westerlies should be maintained up to 600 hPa, in the box equator to Lat. 10^{0}N and Long. 55^{0}E to 80^{0}E. The zonal wind speed over the area bounded by Lat. 5-10^{0}N, Long. 70-80^{0}E should be of the order of 15 – 20 kts. at 925 hPa. The source of data can be RSMC wind analysis/satellite derived winds.

C) OLR

INSAT derived OLR values should be below 200 Wm^{-2} in the box confined by Lat. 5-10^{0}N and Long. 70-75^{0}E.

II. Further advance of monsoon over the country.

a) Further advance is declared based on the occurrence of rainfall over parts/sectors of the sub-divisions and maintaining the spatial continuity of the northern limit of monsoon.

The following auxiliary features may also be looked into.

b) Along the west coast, position of maximum cloud zone, as inferred from the satellite imageries may be taken into account.

c) The satellite water vapour imageries may be monitored to assess the extent of moisture incursion.

7.6.2. Withdrawal of SW Monsoon

The guidelines to be followed for declaring the withdrawal of southwest monsoon are enlisted below:

a) Withdrawal from extreme Northwestern parts of the country should not be attempted before 1st September.

b) *After 1st September:*

The following major synoptic features should be considered for the first withdrawal from the western parts of NW India.

i) Cessation of rainfall activity over the area for continuous 5 days.

ii) Establishment of anticyclone in the lower troposphere (850 hPa and below)

iii) Considerable reduction in moisture content as inferred from satellite water vapour imageries and tephigrams.

c) *Further withdrawal from the country:*

i) Further withdrawal from the country may be declared, keeping the spatial continuity, reduction in moisture as seen in the water vapour imageries and prevalence of dry weather for 5 days.

ii) SW monsoon should be withdrawn from the southern peninsula and hence from the entire country only after 1st October, when the circulation pattern indicates a change over from the southwesterly wind regime.

7.7. Commencement / cessation criteria for NE monsoon

7.7.1. Commencement of NE Monsoon Rains

For declaring onset of Northeast Monsoon following criteria may be considered.

i. Withdrawal of southwest Monsoon up to 15^0 N.
ii. Onset of persistent surface easterlies over Tamil Nadu coast.
iii. Depth of easterlies up to 850 hPa over Tamil Nadu coast.
iv. Fairly widespread rainfall over the coastal Tamil Nadu, south coastal Andhra Pradesh and adjoining areas.
v. Onset is not to be declared before 10th October even, if the conditions described above exist.

7.7.2. Specifications for describing the activity/ strength of Northeast Monsoon.

Weak Monsoon: Rainfall less than half the normal in any meteorological sub-division over the NE monsoon regime.

Normal Monsoon: Rainfall half to less than one and a half (1½) times the normal over any meteorological sub-division in the NE monsoon regime.

Active Monsoon: i) Rainfall 1½ to 4 times the normal.

ii) Rainfall in at least two stations should be 3 cm in coastal Tamil Nadu and south coastal Andhra Pradesh and 2 cm elsewhere over any meteorological sub-division in the NE monsoon regime.

iii) Rainfall in that sub-division should be fairly widespread or widespread.

Vigorous Monsoon: i) Rainfall exceeding 4 times the normal.

ii) Rainfall in at least two stations should be 5 cm in coastal Tamil Nadu and south coastal Andhra Pradesh and 3 cm elsewhere in any meteorological sub-division in the NE monsoon regime.

iii) Rainfall in that sub-division should be fairly widespread or widespread.

7.8. Strength and performance of monsoon

The procedures for describing the activity of monsoon, mentioning the temperature conditions when the situation arises and the criteria followed in declaring the withdrawal etc. are provided in the respective forecasting circulars.

ACTIVITY

7.8.1. Strength of Monsoon

The following criteria are adopted for describing the activity of the monsoon:

(a) DESCRIPTIVE TERM OVER THE SEA.

Descriptive Term	***Wind speed (in knots) reported or inferred to be existing***
	Up to 12 knots
	13 to 22 knots
	23 to 32 knots
	33 knots and above.

(b) DESCRIPTIVE TERM OVER LAND AREA

SPECIFICATION

Weak monsoon Rainfall less than half the normal.

Normal monsoon Rainfall half to less than 1½ times the normal

(mention of ‘normal’ monsoon may not be necessary in general).

Active/ strong monsoon i) Rainfall 1 ½ to 4 times the normal.

	ii) The rainfall in at least two stations should be 5 cm, if that sub-division is along the west coast and 3 cm, if it is else where. iii) Rainfall in that sub-division should be fairly widespread to widespread.
Vigorous monsoon	i) Rainfall more than 4 times the normal. ii) The rainfall in at least two stations should be 8 cm if the sub-division is along the west coast and 5 cm if it is elsewhere. iii) Rainfall in that sub-division should be fairly widespread or widespread.

While describing the activity of the monsoon,

i) The normal of stations, whenever available should be used.

ii) Till normal for all the stations are available the following procedure should be adopted:

Number of stations in a sub-division with normal: a

Normal for these stations: b

Average normal for the sub-division: b/a

Total number of stations reporting rainfall: c

Actual total rainfall reported by these stations: d

Therefore, the average rainfall for the sub-division: d/c

Compare d/c with b/a and describe the activity of the monsoon accordingly, other conditions being fulfilled.

(3) i) In the sub-divisions, where the percentage of hill stations is high, the hill stations are also taken into account for describing the activity of the monsoon. In other sub-divisions, the hill stations are excluded.

ii) The monsoon activity is described in all the sub-divisions of northeast India as is done for sub-divisions of other regions.

iii) The monsoon activity is not described over the Bay Islands and the Arabian Sea Islands.

7.8.2. Monsoon activity in Daily Weather Bulletins (All India Weather Summary)

7.8.2.1 Subdued Monsoon Activity

The criterion for describing the subdued monsoon activity during monsoon season is as follows:

1. Whenever, a countrywide dry spell continues for more than 3 days and is expected to continue for at least 2 more days, the description of monsoon should be given as 'weak or subdued monsoon for the country as a whole'. RMC level bulletins may also include this information. The information on prolonged dry or wet spells over individual pockets is included in the All India Weather Reports.
2. Information on prolonged dry spells/ subdued activity over the meteorological subdivisions also should be included in the RMC/ MC level bulletins if such a spell is persisting for 2 or more days and is expected to continue.
3. Monsoon activity for a sub-division might be considered subdued if the rainfall realised is less than ½ the normal.

In order to describe the subdued rainfall activity on a particular day, the following criteria be observed.

(i) Spatial distribution of rainfall remains mainly dry, isolated or scattered for two consecutive days.

(ii) Mean actual rainfall of that particular sub-division remains less than the normal for the consecutive two days.

(iii) The Forecast issued for the next 48 hrs on the particular day for the sub-division is also mainly dry, isolated or scattered.

Upon satisfying all the above criteria simultaneously, monsoon activity be described as subdued on the second day.

7.8.2.2. All India Drought Year

Based on rainfall deficiency, the meteorological drought is defined on a sub-division scale. The meteorological droughts are classified into two following categories:

(a) Moderate drought: when seasonal rainfall deficiency falls between 26 to 50%

(b) Severe drought: when seasonal rainfall deficiency exceeds 50%.

The criterion for defining all India drought year is as follows:

When the rainfall deficiency is more than 10% and when 20 to 40% of the country is under drought conditions, then the year is termed as ***All India Drought Year*** and when the spatial coverage of drought is more than 40% it is called as ***All India Severe Drought Year.***

7.9. High temperature epochs which need mentioning in the weather bulletins within the monsoon season

Quite often, the weak or break monsoon conditions or some anomalous circulation features give rise to persistent high temperatures over some parts of the country which can have adverse influence on day to day activities.

These need to be foreseen and described in the weather reports/ bulletins, as per the following guidelines:

a) Whenever there is a high temperature epoch (expected to prevail for 3 days or more) within the monsoon period (i.e. after the advance and before the withdrawal) over a particular meteorological sub–division, with the temperature values reaching 40^0C or more (anomaly $+5^0$ C or more) for inland region, it should be suitably mentioned in terms of the comfort (heat) index in weather reports.

b) The threshold value may be 35^0C (anomaly $+5^0$ C or more) for the hilly regions and coastal areas.

c) The phenomenon being an area specific one, two or more stations should satisfy the above condition, in order that a particular sub–division or part of it will qualify to be mentioned in this regard.

d) The practice of declaring the heat wave, hot day etc. is continued as such, only prior to the advance and after the withdrawal of monsoon from the particular region.

7.10. Terminology:

7.10.1. Rainfall

7.10.1.1 *Phraseology used for describing spatial distribution of rainfall in weather bulletins and the limit of minimum reported rainfall for describing the spatial distribution in Regional Daily Weather Reports.*

The term 'isolated' is used in "Farmers Weather Bulletins" as in the case of General Weather Bulletins.

Distribution of rainfall amounts of less than 2.5 mm and more than "trace" are also to be described in non-monsoon months as indicated below,

Distribution Description

i. Widespread/ fairly widespread very light rain plus isolated light rain.
ii. Widespread/ fairly widespread very light to light rain.
iii. Widespread/ fairly widespread very light/ light rain plus isolated moderate rain.
iv. Moderate rain has been isolated in (name of sub-division). Very light/ light rain has also been widespread/ fairly widespread.
v. In the case of a scattered distribution of light or higher intensities of rainfall, the distribution of very light rain in the same sub-division will not be described as it may not be significant.

7.10.1.2. *Definition of the terms heavy, very heavy and exceptionally heavy rain*

When record rainfall has occurred, it is highlighted suitably. The statistics of extreme values of rainfall are available and these are useful in describing the rainfall as exceptional or otherwise. O/o ADGM (R) Pune, has prepared isopleths of highest rainfall. These maps are helpful in making such descriptions.

7.10.1.3. *Specifications for description of rainfall, temperatures and strength of monsoon*

Rainfall and temperatures are mentioned in metric units in all the weather bulletins, reports, forecasts and Farmers' Weather Bulletins issued by the India Meteorological Department.

7.10.1.4. *Introduction of extremely heavy rainfall terminology*

The following terminologies are used for indicating rainfall amounts:

Rainfall amount in mms	**Plotted on charts as**	**Descriptive term used**
0.0		No rain*
0.1 to 2.4	. . .	Very light rain
2.5 to 7.5	-	Light rain
7.6 to 15.5	1	Moderate rain
15.6 to 24.4	2	
24.5 to 35.5	3	
35.6 to 44.4	4	Rather heavy
44.5 to 55.5	5	
55.6 to 64.4	6	
64.5 to 75.5	7	Heavy rain
75.6 to 84.4	8	
84.5 to 95.5	9	
95.6 to 104.4	10	
104.5 to 115.5	11	
115.6 to 124.4	12	
≥ 124.5 to 244.4	13 to 24	Very heavy rain
≥ 244.5	≥ 25	Extremely heavy rain

when the amount is a value near about the highest recorded rainfall at or near the station for the month or season. However, this term will be used only when the actual rainfall amount exceeds 12 cm.	Exceptionally heavy rain.

Exceptionally heavy rain:

The amounts for heavy and untimely rainfall warnings are determined by the requirements of the individual warnees. In cases where the requirements of warnees are available in inches, corresponding whole cms may be taken.

“No rain” should strictly relate to occasions of zero rainfall at all reporting stations.

7.10.1.5. *Norms for weekly/ seasonal rainfall distribution.*

The description presented in the monsoon summary of short (Weekly) as well as long term (seasonal) rainfall distribution over a meteorological sub-division has only one norm.

The norms followed for describing weekly as well as seasonal rainfall distribution in all the summaries are as follows:

1. Excess (E) + 20% and above.
2. Normal (N) + 19% to – 19%
3. Deficient (D) - 20% to – 59%
4. Scanty (Sc) - 60% or less
5. No rain - 100%

7.10.1.6. *Terminology of rainfall distribution*

Existing New Hindi Version

New	**New Hindi Version**
At most places	**YkxHkx lHkh LFkkuks aij**
At many places	**vusd LFkkuksa ij**
At a few places	**dqN LFkkuks aij**
Isolated	**dgha dgha**

All the above terminologies mentioned above are used in Indian Daily Weather Summary, Indian Daily Weather Reports, Weather Bulletins, Regional Daily Weather Reports, and Weekly Weather Reports.

7.11. Action by ACWCs / CWCs

7.11.1. In the sea area bulletins issued by ACWCs/ CWCs, Part II usually contains 'Weather Seasonal ', when there is no synoptic system in the area. However, during the monsoon season the strength of monsoon is described in part II as per the following specifications:

Strength of monsoon	**Corresponding wind speed over the area**
Week	Upto 12 kts
Moderate	13 – 22 kts
Strong	23 – 32 kts
Vigorous	33 kts and above

7.11.2. Hoisting LC – III in association with Monsoon

The Local Cautionary signal, LC-III, over the sea area is hoisted,

(a) when squally weather is expected in the port due to the first advance of monsoon or,

(b) Whenever, after the monsoon has established, it is expected to strengthen markedly following a period of weak or moderate monsoon and cause markedly squally weather at the port.

The signal remains hoisted for such time as the associated threat of squally weather at the port remains. The criterion followed for hoisted LC-III under condition (b) above is that the expected wind speed should be 30 kts or more. This minimum limit of 30 kts has been adopted with a view to restrict the number of occasions on which LC-III will have to be hoisted. The term *'markedly squally weather'* is always used in all such messages.

7.12. Action towards the maintenance of rainfall statistics

7.12.1. O/o DGM (Hydrology) is to include the updated rainfall statistics of the season ending the previous week along with the current weekly and cumulative seasonal statistics in the rainfall statistics messages. An additional column providing updated cumulative values till the previous week is added in routine weekly rainfall reports issued by DGM (Hydrology), New Delhi.

7.12.2. All RMCs/MCs are to send the rainfall data as text file attachments through e-mail regularly to NHAC (Computer Division), DGM (Hydrology), DDGM(WF) and ADGM(R), Pune. O/o DGM (Hydrology) is to make use of all these data for description of rainfall.

7.13. Accounting the weather aberrations

Any significant deviation from the normal pattern of weather during the season, such as abnormally high/ low rainfall amounts, temperatures and other special weather phenomena may be reported immediately with suitable meteorological explanations,

facts and figures in the form of tables and pictures to the O/o DDGM(WF), Pune, under the title 'Significant Weather Events' by email : ws@imdpune.gov.in. The O/o DDGM(WF), in turn, is to compile these inputs for sending to DGM (NHAC) New Delhi for posting on the website for public consumption.

7.14. Issuance of local forecast 4 times a day

All forecasting offices are to issue local forecasts 4 times a day based on 00, 06, 12 & 18 UTC observations and update the same within 2½ hrs. of the respective observations in the website and disseminate to other users.

The following schedule is to be strictly adhered to regarding the issuance and updating of the forecasts:

Forecast	Based on Observations of	Websites to be updated by
1st	0000 UTC	0230 UTC (0800 hrs. IST)
2nd	0600 UTC	0830 UTC (1400 hrs. IST)
3rd	1200 UTC	1430 UTC (2000 hrs. IST)
4th	1800 UTC	2030 UTC (0200 hrs. IST)

7.15. Forecast verification procedures

The following procedure is followed by all forecasting offices for preparation of the forecast verification reports:

7.15.1. Verification of Local Forecast:

The local forecast is verified in terms of the following three parameters:

(a) **Temperature:** The forecasts for maximum and minimum temperature are issued throughout the year. The same is verified, with the existing criteria; in addition Root Mean Square Error (RMSE) also be computed. The temperature forecast issued based on 0300 UTC charts, for communication to Headquarters for the display in the IMD, New Delhi website and forecast

based on 1200 UTC charts, which is communicated to the press, be considered for verification. The verification may be carried out for 24 hrs. and 48 hrs. forecasts, separately.

(b) **Rainfall and Significant/Special Weather Events:** The schedule for issuance of forecast for rainfall and significant / special weather events if expected be the same as that for the issuance of temperature forecast, as given above.

(i) **Rainfall:** Rainfall verification is made in terms of its occurrence/non-occurrence (yes/no) using 2 X 2 Contingency table and computing various skill scores as in Annexure I.

(ii) **Significant/ Special weather events:** Forecasts of Fog, thunderstorm, dust storm, hail storm, haze, mist, smog, heat wave, cold wave, hot day, cold day, squall, heavy rain and sky conditions be verified based on the procedures mentioned above. Also in this connection, in order to bring about uniformity, the forecasts for IMD website, WMO website and that for IVRS need to be disseminated by 1130 hrs IST.

7.15.2. Categorical forecast

The performance of categorical forecasts like QPF, issued by FMOs for different river catchments and spatial distribution and intensity of rainfall over different meteorological subdivisions be verified using various Contingency tables.

The QPF issued for different river catchments be verified by computing Percentage Correct (PC), Heidke Skill Score (HSS) and Critical Success Index (CSI), from 6 X 6 Contingency table. The detailed procedure of this forecast verification technique is given in Annexure II.

7.15.3. Verification of sub-division wise forecasts

(a) The present verification method be continued in case of spatial distribution and intensity verification (Annexure III)

Spatial distribution forecast without any intensity specification might be understood as that of light, moderate and rather heavy category.

(b) The verification is reported on subdivisional level. The RMCs communicate the forecast verification data on sub–divisional level to the O/o DDGM (WF), Pune.

Annexure I

Deterministic forecast

(e.g. forecast verification of heavy rainfall (≥ 7 cm)

Over at least two stations in a sub-division

Observed	**Forecast**	
	Yes	**No**
Yes	**A**	**B**
No	**C**	**D**

PROBABILITY OF DETECTION (POD) = A/ (A+B)

FALSE ALARM RATE (FAR) = C/ (C+A)

MISSING RATE (MR) = B/ (B+A)

CORRECT NON-OCCURRENCE (C-NON) =D/(C+D)

CRITICAL SUCCESS INDEX (CSI) = THREAT SCORE = A/ (A+B+C)

BIAS FOR OCCURRENCE (BIAS)= (A+C)/(A+B)

PERCENTAGE CORRECT (PC) = (A+D)/(A+B+C+D)*100 = HIT RATE * 100

TRUE SKILL SCORE (TSS) = A/(A+B) + D/(C+D)-1

HEIDKE SKILL SCORE (HSS) = 2(AD-BC)/ ((B2+C2+2AD+(B+C)(A+D))

FOR BEST/ PERFECT FORECAST, POD=1, FAR=0, MR=0,

C-NON=1, BIAS=1, CSI=1, TSS=1, HSS=1 AND PC=100%

The warning for heavy, very heavy and extremely heavy rainfall, cold wave, heat wave etc. also be verified using the above 2 X 2 Contingency table.

Annexure II

CATEGORICAL FORECAST (QPF)

Observed range (mm)	Forecast range (mm)						
	0	1-10	11-25	26-50	51-100	>100	Total
0	a	b	c	d	e	f	A
1-10	g	h	i	j	k	l	B
11-25	m	n	o	p	q	r	C
26-50	s	t	u	v	w	x	D
51-100	y	z	aa	ab	ac	ad	E
>100	ae	af	ag	ah	ai	aj	F
Total	G	H	I	J	K	L	T

PC = ((a+h+o+v+ac+aj)/T)*100

CSI = a/(A+G-a), h/(B+H-h), o/(C+I-o), v/(D+J-v), ac/(E+K-ac), aj/(F+L-aj)

HSS=(a+h+o+v+ac+aj)-(AG+BH+CI+DJ+EK+FL)/T)/(T-(AG+BH+CI+DJ+EK+FL)/T)

The POD, FAR, MR, CSI, BIAS, PC, TSS and HSS, etc, for each category be calculated by reducing the above 6x6 Contingency Table into 2 x 2 Contingency table for Yes / No forecast. The final skill score be the average of these values.

Annexure III

VERIFICATION OF SPATIAL DISTRIBUTION FORECAST

Observed Range	Forecast range					
	Dry	Isol	Scatt	Fairly widespread	Widespread	Total
Dry	a	b	c	d	e	J
Isolated	f	g	h	i	j	K
Scattered	k	l	m	n	o	L
Fairly widespread	p	q	r	s	t	M
Wide spread	u	v	w	x	y	N
Total	O	P	Q	R	S	T

PC = ((a+g+m+s+y)/T)*100

CSI = a/(J+O-a), g/(K+P-g), m/(L+Q-m), s/(M+R-s), y/(N+S-y)

HSS= (a+g+m+s+y - (JO+KP+LQ+MR+NS)/T)/ (T-(JO+KP+LQ+MR+NS)/T)

Conversion of categorical forecast into deterministic forecast:

The POD, FAR, MR, CSI, BIAS, PC, TSS and HSS, etc., for each category are calculated by reducing the above 6x6 Contingency Table into 2x2 Contingency Table for yes/no forecast. The final skill score is the average of these values.

Verification of Intensity of Rainfall

Observed range (mm)	Forecast range (mm)						
	Light	Moderate	Rather Heavy	Heavy	Very heavy	Extremely heavy	Total
Light	a	B	C	d	e	f	A
Moderate	g	H	I	j	k	l	B
Rather heavy	m	N	O	p	q	r	C
Heavy	s	T	U	v	w	x	D
Very heavy	y	Z	Aa	ab	ac	ad	E
Extremely heavy	ae	Af	Ag	ah	ai	aj	F
Total	G	H	I	J	K	L	T

PC = ((a+h+o+v+ac+aj)/T)*100

CSI = a/(A+G-a), h/(B+H-h), o/(C+I-o), v/(D+J-v), ac/(E+K-ac), aj/(F+L-aj)

HSS=(a+h+o+v+ac+aj)-(AG+BH+CI+DJ+EK+FL)/T)/(T-(AG+BH+CI+DJ+EK+FL)/T)

The POD, FAR, MR, CSI, BIAS, PC, TSS and HSS, etc, for each category be calculated by reducing the above 6x6 Contingency Table into 2x2 Contingency Table for yes/no forecast. The final skill score is the average of these values.

7.16. System intensity over land

1. When there is a low pressure system as intense as or higher than that of a depression over land, the concerned RMCs track the system in consultation with O/o DDGM (WF), Pune and HQrs New Delhi, until it weakens into a low pressure area.
2. The RMCs advise the concerned state government either directly or through respective MCs for special actions to be taken, if any.
3. Special bulletins are issued on the basis of 03, 09, 12 & 18 UTC charts – 4 bulletins per day – till the weakening of the system into a low pressure area.
4. The system intensity over land may be fixed as per the procedures listed below:

(i) The upper level winds at 0.9 km a.s.l., central pressure, pressure departure, and satellite or radar data are not always consistent with the system intensity.

(ii) The system intensity over land is to be defined solely based on the number of closed isobars at the intervals of 2 hPa within 3^0 radius from the centre of the low pressure system as classified below:

(a) Low pressure area	1
(b) Depression	2
(c) Deep Depression	3 – 4
(d) Cyclonic Storm	5 and above

(iii) As regards the other land systems, viz., western disturbances, the criteria for classifying the system are that, when two or more closed isobars at 2hPa intervals are discernible on the sea level chart, the disturbance may be described as a Western Depression.

Each forecasting office is to prepare a checklist on all these procedures to be followed along with a list of warnees (and phone numbers of disaster mitigation agencies) and keep it handy prior to the starting of the season.

Reference:

"Forecasters Guide", India Meteorological Department, 2008.

CHAPTER 8 (i)

OBSERVATIONAL AIDS AND MONITORING OF MONSOON

(i) SATELLITE MONITORING OF MONSOON

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8(i).1. Introduction

In the pre-satellite era, understanding of monsoon was based on the meteorological data collected by land-based observatories and ships. The launch of first satellite TIROS-I dedicated to the study of weather and climate on 1st April, 1960 marked the beginning of satellite era in the meteorology. Since then more than 300 satellites have been launched to monitor the weather and climate of the earth.

The role played by the oceans in the onset and sustenance of monsoon was not well understood in the pre-satellite era due to lack of observations from the oceanic regions. It was hypothesized that the monsoon circulation was driven by the differences in the surface temperatures over land and adjacent ocean. One of the serious limitations of this theory arises from the fact that after the onset of monsoon there is hardly any difference in surface temperatures over land and adjoining oceanic regions. The sustenance of the monsoon for about four months after the onset can not be explained from the land-ocean contrast theory. Webster and Fasullo (2003) have suggested that the role of land-ocean contrast in surface temperature seems to be limited to the monsoon onset processes only. After the onset, monsoon system is substantially driven by the heating of troposphere through the release of latent heat. The satellite observations have provided vital inputs to our understanding of various phenomenon like El-Nino/Southern Oscillation (ENSO), Indian Ocean Dipole Mode (IODM) and Madden Julian Oscillation (MJO) that are

associated with monsoon on different time-scales (Sikka and Gadgil, 1980; Gadgil and Srinivasan, 1990; Madden and Julian, 1994).

In the initial period of satellite era imageries provided by polar orbiting satellites were main satellite inputs to monsoon forecasting. The new generation satellites have provided a wide range of products that are used in monsoon predictions on different time-scales. Satellite observations constitute major part of data that is assimilated into numerical weather prediction models.

The application of satellite data in monsoon monitoring and forecasting can be divided into following major categories, viz,

(a) Use in synoptic analysis in real-time forecasting

(b) Real-time inputs to numerical weather prediction models

8(i).2. Synoptic analysis and forecasting

Satellite imageries and products find wide ranging applications in day-to-day monitoring of onset/withdrawal and activity of monsoon over different regions including oceanic areas where very little information is available through other aids.

8(i).2.1. Monitoring of Monsoon onset

The characteristics of monsoon onset, which can be observed in satellite imageries and products, are described below.

8(i).2.1.1. Features observed in INSAT. visible, infrared and water vapour imageries

Just before the onset of monsoon over Kerala, deep convective cloud area starts increasing over southeast Arabian Sea and neighboring areas. Sequence of imageries (IR & Visible) indicates slow but gradual northwards movement of clouds north of the equator (Fig.8(i).1). Water vapour imagery shows moisture influx at the time of monsoon onset (Fig.8(i).1(d)) .The depth of convection ,which is well monitored with the help of cloud top temperature(CTT) increases at the time of

monsoon onset (increase in depth of convection implies decrease in CTT and *vice-versa*).This feature can be seen in Fig 8(i).2.

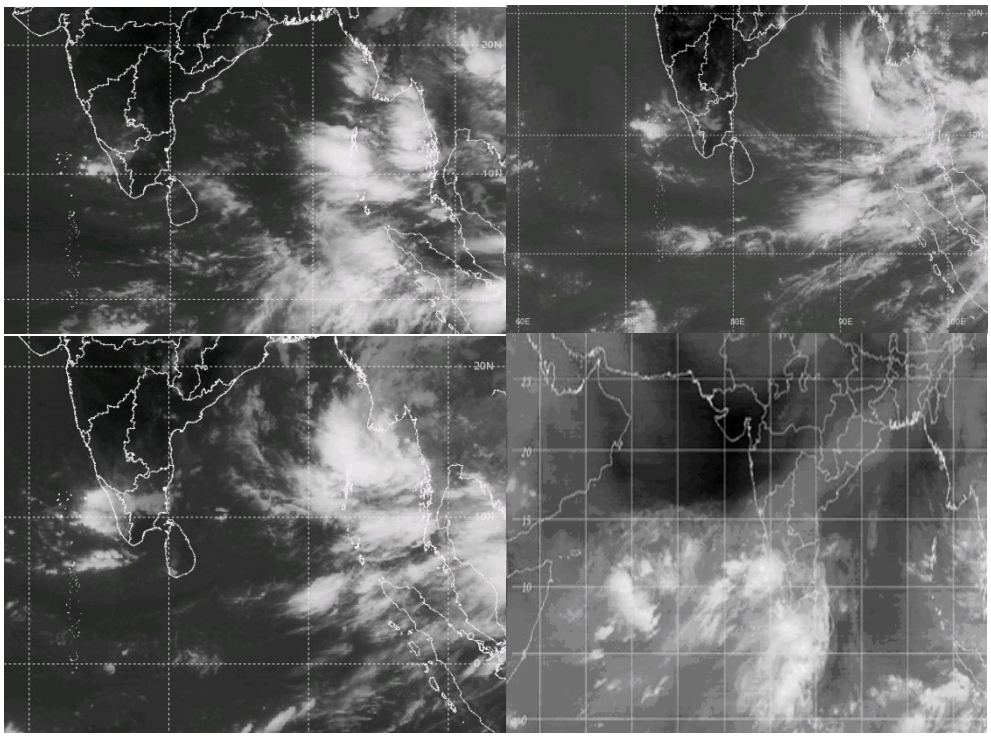


Fig.8(i).1 : Kalpana- IR imageries for (a) 10th (0300UTC) (b) 11th (0900UTC) (c) 12th (0300UTC) indicating onset of SW-Monsoon over SE Bay and Nicobar Islands on10th and by 12th May'2008 over Andaman islands and (d) 28th May'2008 (0300UTC) water vapour image showing moisture inflow on the day of onset of monsoon over Kerala.

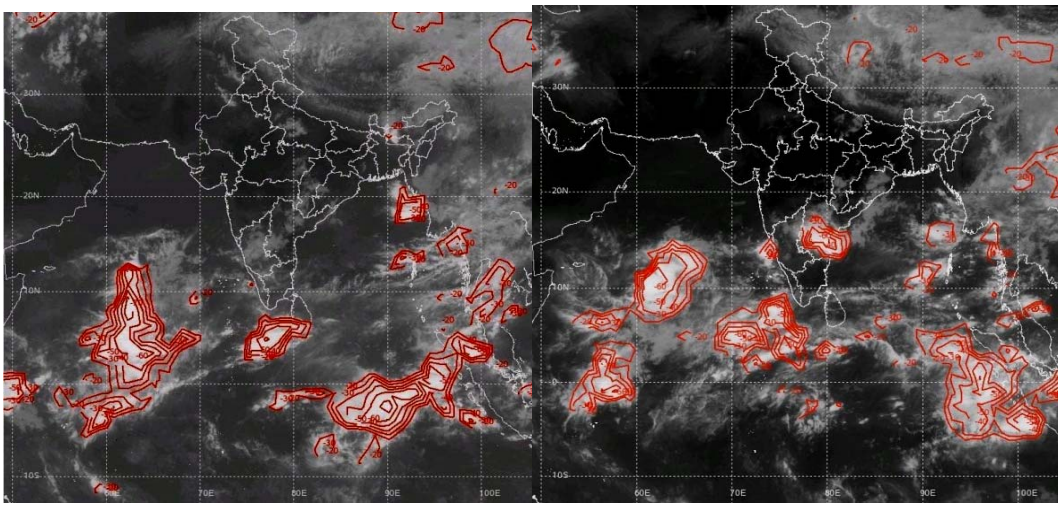


Fig. 8(i).2: Cloud top temperatures during the onset phase of 2009 Monsoon (29 and 30 May 2009, 0300UTC).

8(i).*2.1.2. Application of satellite- derived Outgoing Longwave Radiation (OLR) in monitoring of monsoon onset*

A low value of OLR (below 200 W/m^2) indicates the presence of deep clouds. Before the onset of monsoon over Andaman and Nicobar Islands & over Kerala, the daily mean distribution of OLR shows gradual decrease of OLR value over these areas. One of the criteria, adopted by IMD in 2006, for declaring the date of monsoon onset over Kerala is based on OLR. The INSAT derived OLR value should be below 200 Wm-2 in the box, Lat 5-10°N and Long 70-75° E. On 26th May'2007 (Fig.8(i).3- a) shows daily mean OLR values less than 200 Wm-2 were observed over the south Arabian Sea, but they were more than 200 Wm-2 over the extreme south peninsula and the Kerala coast. On 27th May (Fig.8(i).3 -b), OLR values less than 200 Wm-2 were observed over the Kerala coast which persisted over the area on 28 May (Fig.8(i).3 -c) also, suggesting persistence of enhanced convection off Kerala coast . (Fig.8(i).3 –d) shows one- day mean grid point values for 30th May 2008, suggesting onset of monsoon over Kerala.

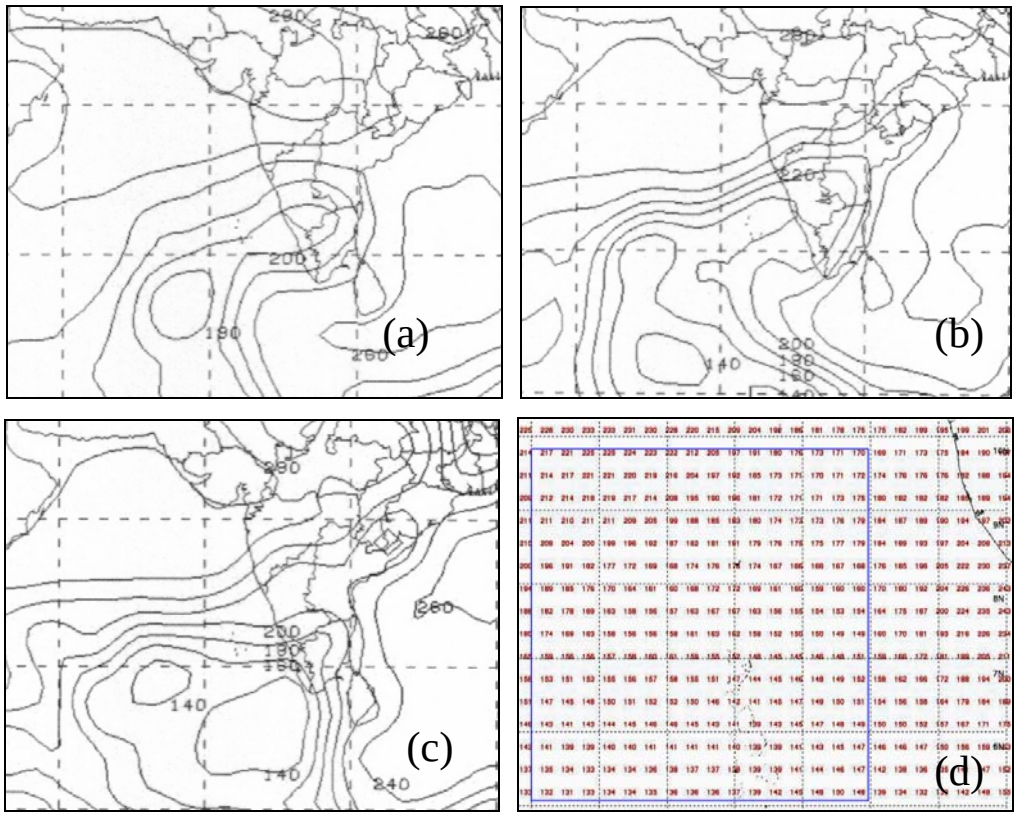
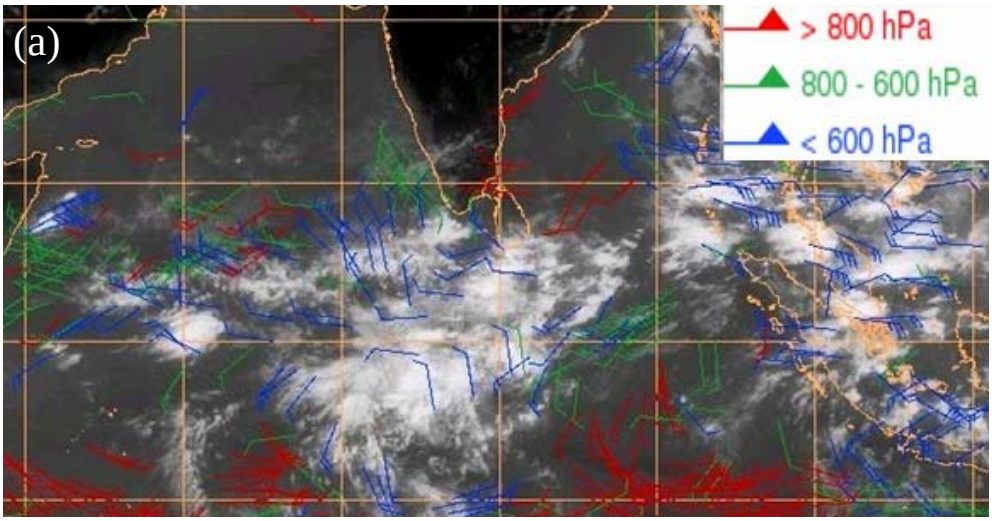


Fig. 8(i).3: (a), (b) & (c) daily mean of OLR distribution for 26th, 27th and 28th May, 2007 And (d) OLR grid point value (daily mean) for 30th May 2008 within the box 5-10º N and 70- 75º E.

8(i)*.2.1.3. Application of satellite- derived winds*

The wind field based criteria adopted by IMD in 2006 for declaring the date of monsoon onset over Kerala states: “Depth of westerlies should be maintained up to 600 hPa, in the box equator to Lat 10°N and Long 55°E to 80°E. The zonal wind speed at 925 hPa over the area bounded by 5°N -10°N, 70°E to 80°E should be at least of the order of 15-20 knots. The source of data can be RSMC wind analysis / Satellite derived winds. Satellite derived Cloud Motion Vectors (CMV), Water Vapour Winds (WVW) and Scatterometer (Quik SCAT and ASCAT) Ocean surface wind vector are very useful in monitoring monsoon features including: (a) cross-equatorial flow; (b) the strength of south-westerly/westerly-Low Level Jet (LLJ) (c) depth of westerly winds, and (d) Southwest Indian Ocean High (Figs.8(i).4 & 8(i).5).



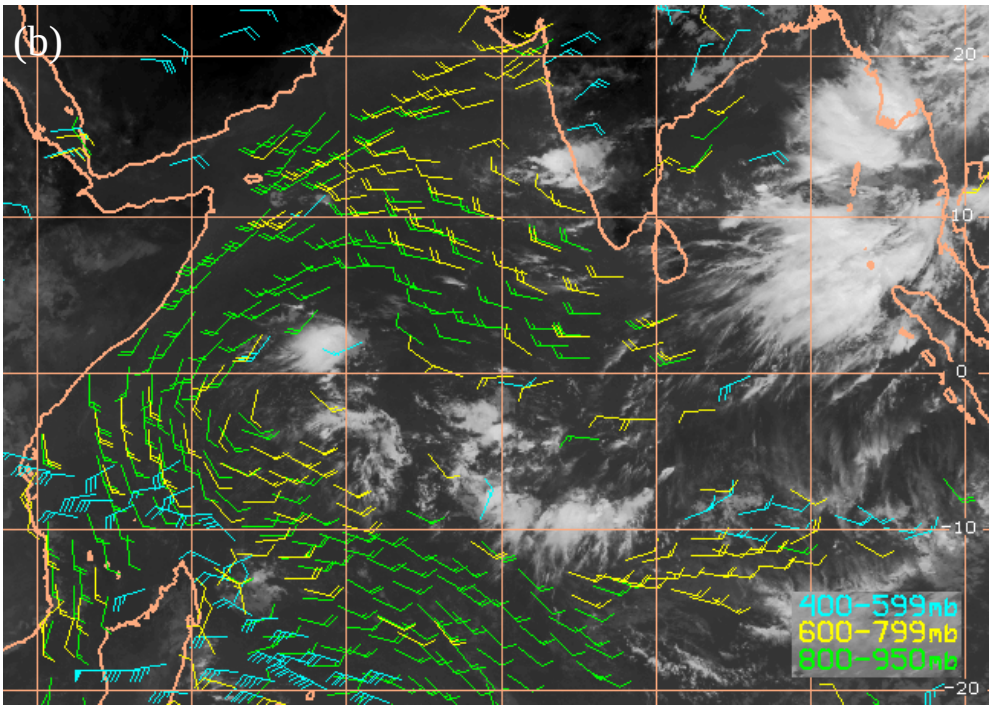


Fig.8(i).4: (a) Kalpana- CMV dated 08.05.2008 (b) METEOSAT-7 CMV dated 29.05.2009 showing setting in of cross-equatorial flow.

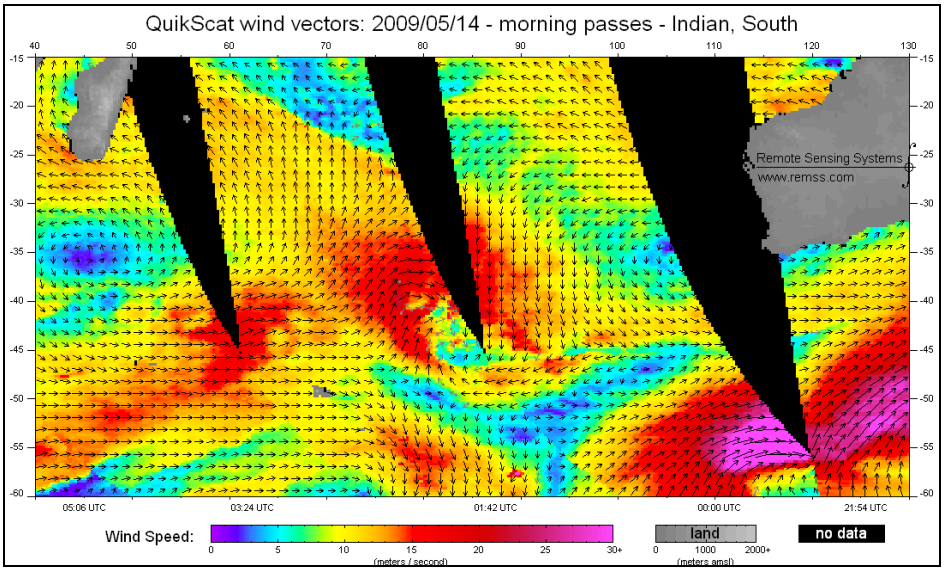


Fig.8(i).5: Southwest Indian Ocean anticyclone as observed in QuikSCAT winds.

8(i)2.1.4. Application of satellite derived integrated water vapour

TRMM/TMI and OCEANSAT-I derived Integrated Water Vapour (IWV) over western Arabian Sea (between lat 0-10°N long 50-65°E) increase by 20-25% about 8-10 days before the onset of monsoon over Kerala (Simon et. al., 2001). Fig.8(i).6

shows the increase in IWV over western Arabian Sea before monsoon onset in 2008.

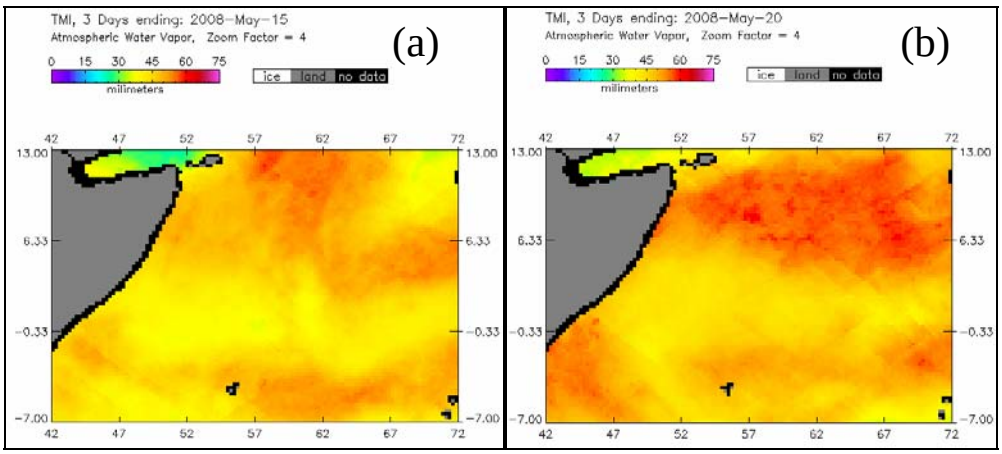


Fig. 8(i).6: TMI derived water vapour for (a) 15th May, 2008 and (b) 20th May, 2008 showing an increase over western Arabian Sea ten days before the onset of monsoon over Kerala.

8(i)*2.1.5. Satellite derived Quantitative Precipitation Estimate (QPE)*:

KALPANA- derived rainfall estimates can be utilized as a tool to observe the rainfall pattern over Indian sub-continent during different phases of monsoon on real-time bases. As shown in fig.8(i).7 QPE is able to capture the rainfall over Kerala and Karnataka coasts during onset of southwest monsoon.

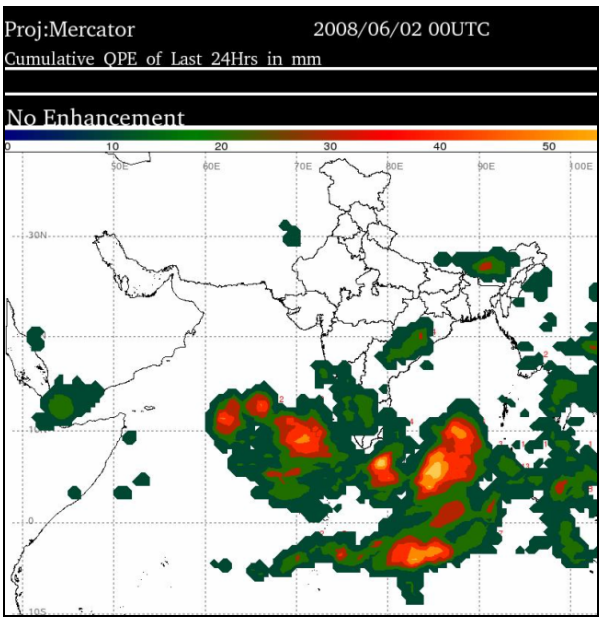


Fig. 8(i).7: Kalpana- derived QPE showing cumulative QPE for 1st June'2008.

8(i).2.2. Monitoring of Progress/advance of monsoon

Outgoing Longwave Radiation (OLR) has been measured by satellites for the past 47 years. A low value of OLR (below 200 W/m2) indicates the presence of deep connective clouds. INSAT (VHRR) and NOAA (AVHRR) derived OLR data has been used extensively to understand cloud organization and propagation in the tropics. Madden and Julian (1994) have brought out possible importance of large-scale convective heating over the Indian Ocean and the western Pacific Ocean. They showed that 40-50 day oscillation exists as a low latitude eastward propagating signal in the zonal wind component from the Indian Ocean to the central Pacific. The Madden Julian Oscillation (MJO) is a dominant signal of cloud movement in the equatorial region. The MJO is identified through deep cloud systems and precipitation that moves slowly eastward along the equator at a speed of about 5m/s (Figs.8(i).8 & 8(i).9). The spatial scale of MJO is in the range of 10,000–20,000 km and hence MJO has an influence on climate in the entire tropical region. The deep clouds in the MJO exist in the Indian and west Pacific oceans only.

Sikka and Gadgil (1980) used the visible cloud imagery from National Oceanic and Atmospheric Administration (NOAA) satellite data to demonstrate that the maximum cloud zone (MCZ) had two preferred regions during the Indian summer monsoon. The primary location of MCZ was around 20°N around the monsoon trough while the secondary location was in the equatorial region. They showed that these cloud bands moved from the equatorial region to the monsoon trough at the rate of around 1° latitude per day. It is apparent from these studies that over the Indian sub-continent region and over the western Pacific Ocean significant cloudiness fluctuations on this time-scale are present.

Meridional propagating motion systems on the 30-50 days time-scale over the monsoon region is only a small part of the total picture, since globally there are many other facts of this phenomenon (Chang and Krishnamurti, 1987)

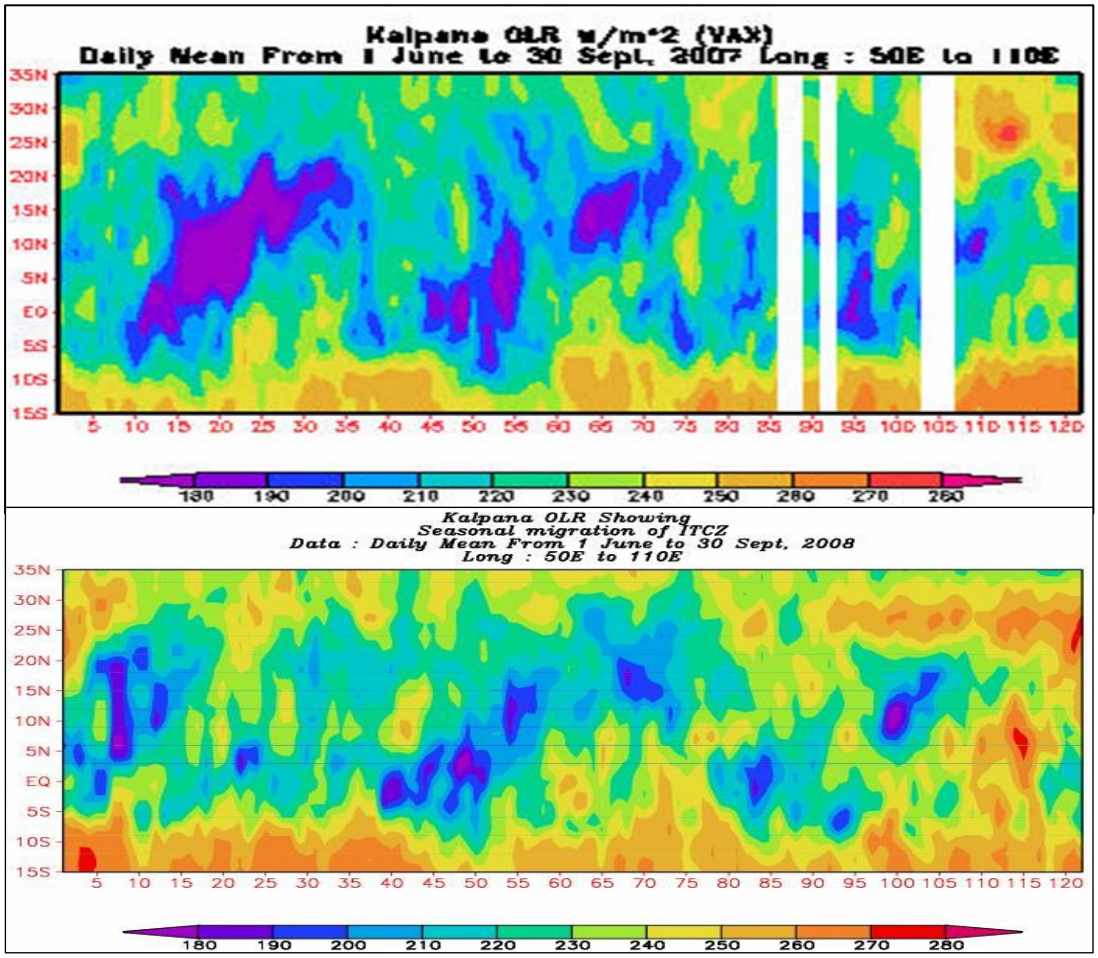


Fig. 8(i).8: Kalpana-I derived OLR for monsoon (a) 2007 and (b) 2008.

Before the advent of satellites it was not possible to estimate the amount of rainfall over the oceans. Satellites can be used to estimate rainfall because deep clouds produce more rain than shallow clouds. Cloud height can be estimated from infrared emission from the top of the clouds and hence one can obtain an estimate of rainfall from infrared sensors in satellites. Arkin *et al*. (Arkin et al., 1989) showed that the brightness temperature data from INSAT-1B can be used to estimate the rainfall over India. Since 1986, India Meteorological Department has been using Arkin's technique operationally for the computation of satellite derived quantitative precipitation estimate (QPE) on daily, weekly and monthly basis. It is an empirical method and is computed from the histogram data by calculating fractional cloud cover colder than a chosen threshold temperature and multiplying the same with a regression coefficient.

$$QPE = K \times Fc \times Nc$$

where Fc is the fractional clouding, K is a constant (in mm/day) related to rain rate ,and Nc is the number of days .The threshold temperature chosen by Arkin was 235K while a value of 71.2 mm/day was used for the constant K. Further studies showed that the convective precipitation as assessed by this technique is fairly good over the oceanic regions off the subcontinent. Over the land areas QPE is able to capture rainfall pattern associated with different synoptic conditions and spatial distribution of mean rainfall of each season. But it has some temporal and spatial biases. Due to threshold value of 71.2 mm/day it cannot capture very heavy amounts of rainfall associated with strong systems like cyclones and depressions and orographically induced motions.

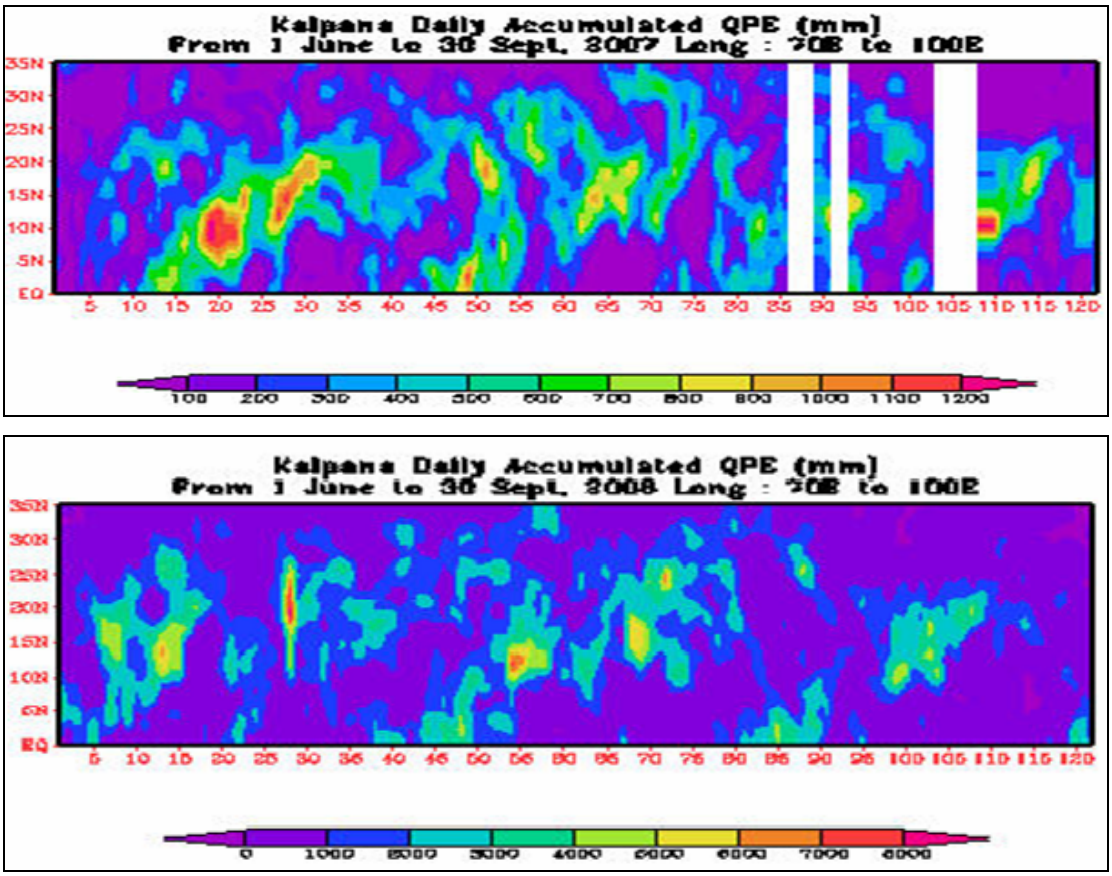


Fig.8(i).9: Kalpana-I derived daily accumulated QPE for the period 1st June to 30thSeptember (a) 2007 and (b) 2008.

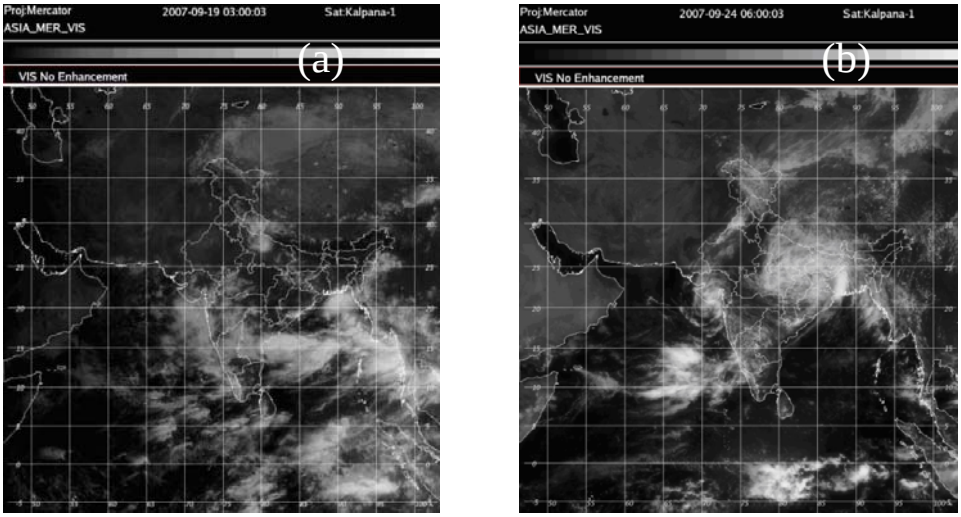
8(i)2.3. Monsoon Trough (MT)

During monsoon season the location of the monsoon trough has biggest influence on the rainfall over Indian sub-continent, particularly over the central parts of the country. The monsoon trough shows periodical movements to the north and south of its normal position (Singh, 2006). In the normal position, it extends from northwest India to the north Bay of Bengal region. The trough, particularly the western half, has tendency to move northward to the Himalayan foothills under the influence of a westerly trough moving eastward across northwest India. On the other hand, monsoon lows/depressions periodically forming over the north Bay of Bengal moving northwestward across the country maintain the normal position and activity of the trough. Sometimes, the low-pressure systems do not form over the Bay, and in this situation a westerly trough affecting northwest India results in the shift of the whole trough to the foothills of the Himalayas. This situation of the monsoon trough over the foothills of the Himalayas is referred to as a 'break' in the monsoon, since except for the sub-Himalayan area and Tamilnadu, the whole country gets very little rain. The sub-Himalayan area receives heavy rainfall during the break period. The strengthening/weakening and fluctuations of MT can be effectively monitored with the help of satellite imageries and products like OLR and QPE. For detailed climatology of 'breaks', one may refer to Ramamurthy (1969) and De, et. al (2002).

8(i).2.4. Off-shore troughs and vortices

A weak off-shore trough is formed along and off Kerala coast in association with the advance of monsoon over rest parts of the country. This type of trough develops quite frequently during monsoon season which is responsible for active to vigorous monsoon conditions over the west coast of India. It can extend from Kerala coast upto Gujarat. In satellite imagery (mainly visible and IR imagery) off-shore trough can be identified by the presence of heavy cloudiness along the west coast. These troughs are generally seen from surface to 0.9 km a.s.l. or sometimes rarely upto 1.5 km a.s.l. Off-shore vortices are formed when relatively weaker southwesterly monsoon winds are returned by the western Ghat mountains and are embedded within the off- shore trough. An off-shore trough in Sep'2007 was

observed during 14th -23rd from Maharashtra to Kerala coast which extended from south Gujarat to Kerala coast during 24th- 30th Sep'2007 as seen in Fig. 8(i).10-a.



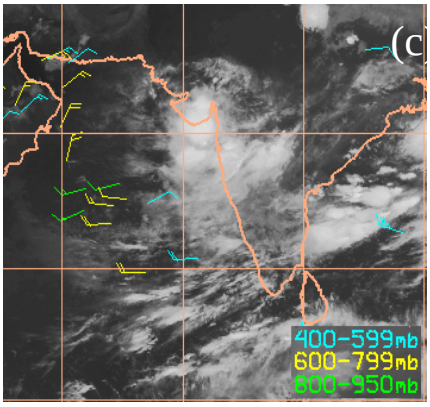
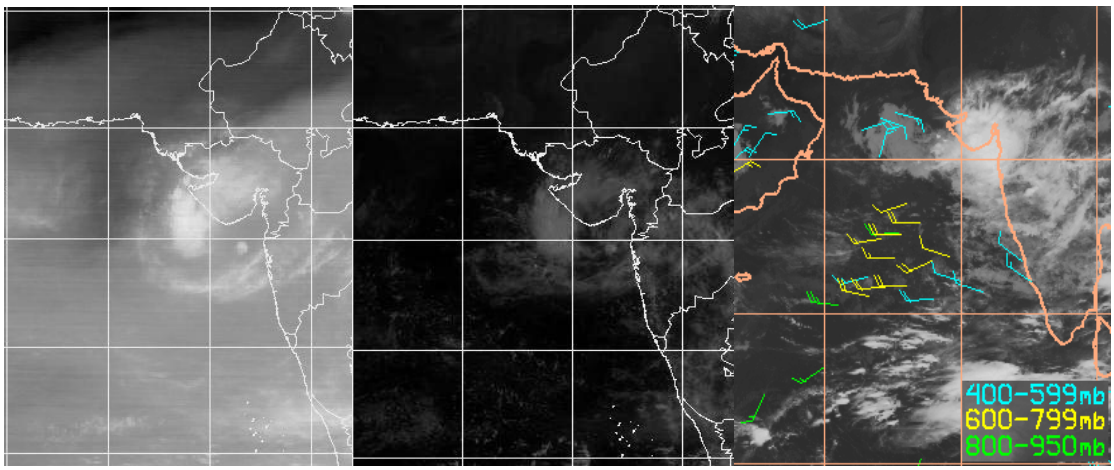


Fig. 8(i).10: (a) Kalpana-I visible imagery dated 19th and 24th Sep'2007 and (b) Meteosat derived CMV for 19th Sep'2007.

8(i).2.5. Mid-Tropospheric Circulation (MTC)

During the International Indian Ocean Expedition in late sixties, it was observed that middle atmospheric cyclonic vortices formed over NE Arabian Sea and adjoining Gujarat and north Maharastra coasts were responsible for very heavy rainfall over the northern sectors of the west coast during southwest monsoon season. These vortices confined to the middle atmosphere between 3 to 6 km with a vorticity maxima, maximum moisture content and maximum convergence between 600 and 500 hPa level. Heavy to very heavy precipitation is concentrated in the southwest sector of a MTC. The most peculiar characteristic of these circulations is that they are only confined to the middle atmosphere and are either not visible on the surface or best seen as a trough. MTC can be identified in visible and infrared satellite imageries as a circulation in the low and medium clouds. These middle level

circulation are observed as well organised circulations in water vapour imagery (6.7µm) from Geostationary satellites(Figs.8(i).10-b and 8(i).11).



(a) (b) (c)

Fig. 8(i).11: 0300UTC 21st Sept'2007 (a) wv (b)vis (c) Meteosat CMV for the same day.

8(i).2.6 Low level Jet (LLJ) or Somali Jet

After crossing the equator southeasterly trade winds become strong due to Corolis force and known as low level Jet. These winds appear to flow from Mauritius across Kenya, Ethiopia, Somalia before reaching southeast Arabian sea near 9-10°N. Their direction of movement keeps on changing from southeasterly to southerly and finally southwesterly in Arabian Sea. Low level Jet can be seen in satellite derived low level cloud motion vector having a core speed of about 40-60 kts. Findlater (1969) showed that this current account for about half of the total cross equatorial transport of air in the lower troposphere in July. Fig. 8(i).12 shows LLJ METEOSAT-7 winds.

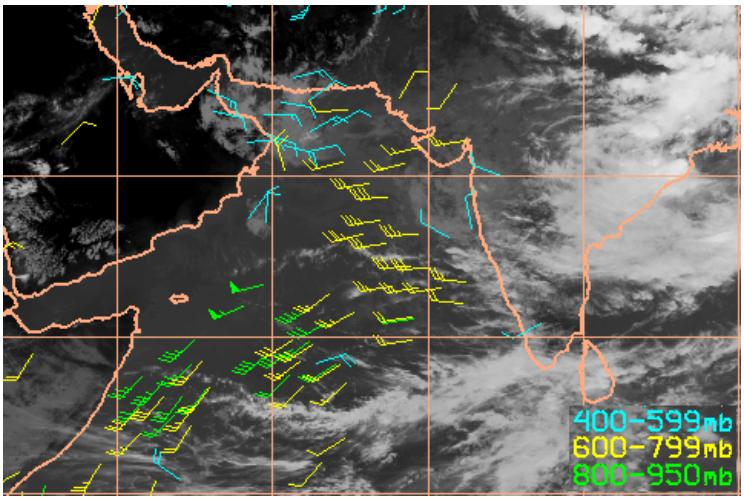
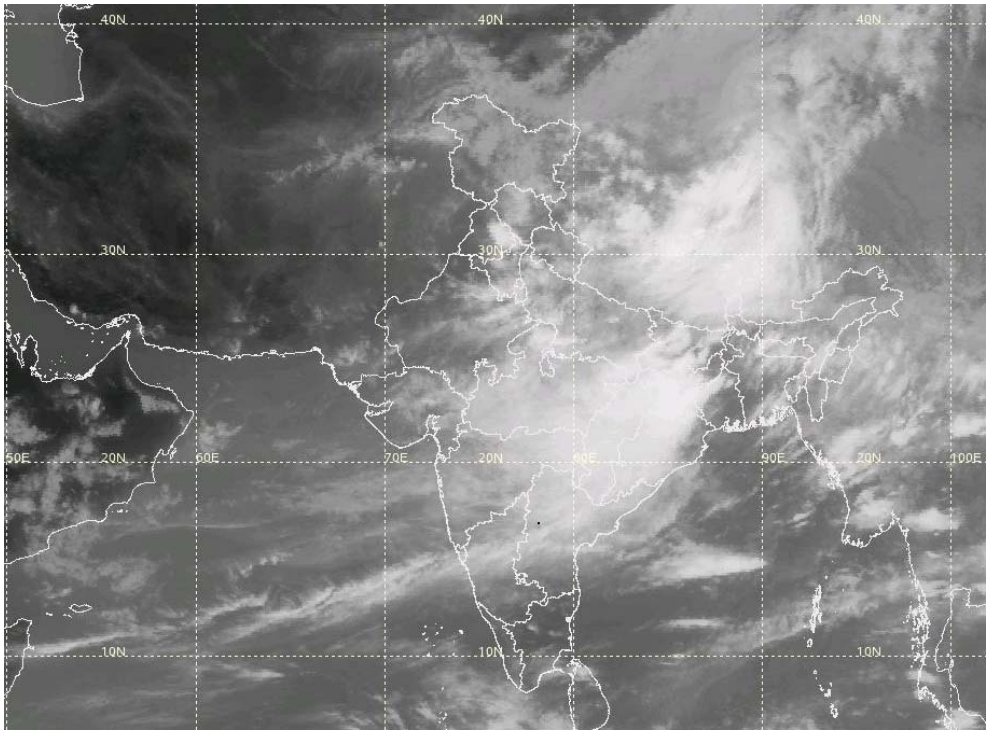


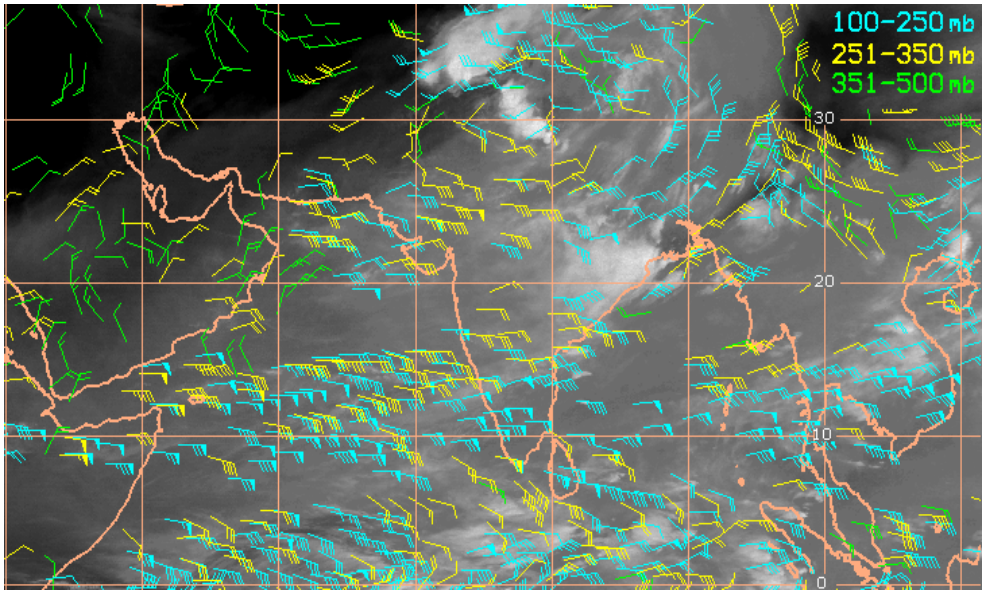
Fig.8(i).12: LLJ stream as observed in METEOSAT-7 derived low level winds of 11th August 2007 (1200 UTC).

8(i)*.2.7 Tropical Easterly Jet(TEJ)*

These are strong easterlies just south of Tibetan anticyclone. The jet is strongest near 100/150hPa with a core speed of 60-80kts. TEJ can be recognized in infrared imagery by the presence of cirrus clouds. But high level water vapour winds are best to see this jet between Vietnam to Africa coast during southwest monsoon season. (Fig.8(i).13).



(a)



(b)

Fig. 8(i).13: 13th August' 2007 (a) Kalpana-I IR imagery and (b) METEOSAT-7 water vapour winds indicating tropical easterly jet flow from Vietnam to Africa coast.

8(i).2.8. Monitoring of activity of monsoon; Active and Break spells

8(i).*2.8.1. Monitoring of monsoonal cyclogenesis*

Mooley (1976) has shown that during large-scale drought years, monsoon depressions, which are generally westward moving systems, either dissipate or recurve north or northeast before reaching longitude 80.0°E. Sikka (1980) has shown that the main features that distinguish years of heavy and deficient monsoon rainfall are the number of monsoon low pressure systems and the number of rainy days in a monsoon. The genesis and movement of monsoon depressions can be monitored with the help of satellite imageries and scatterometer winds (Fig.8(i).14).

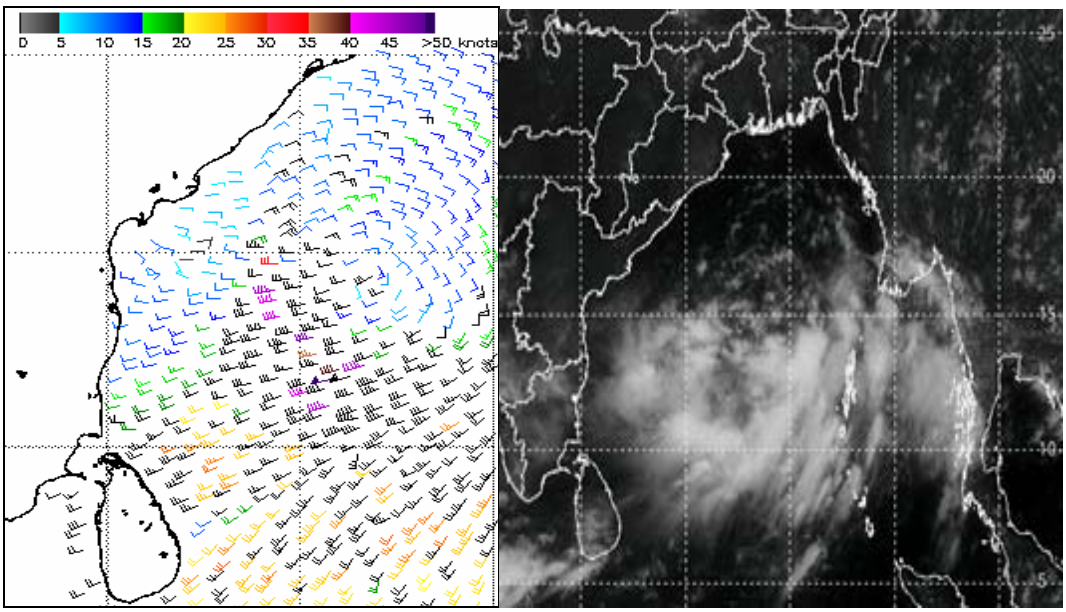


Fig. 8(i).14: A weak vortex can easily be recognized in Quikscat winds as compared to any other source. For example Quikscat winds (ascending pass) and visible imagery (0600UTC) for 22nd May, 2009.

8(i).*2.8.2. Active spells*

Active monsoon spells can be monitored with the help of satellite products. A few examples are shown below (Figs. 8(i).15 - 8(i).18).

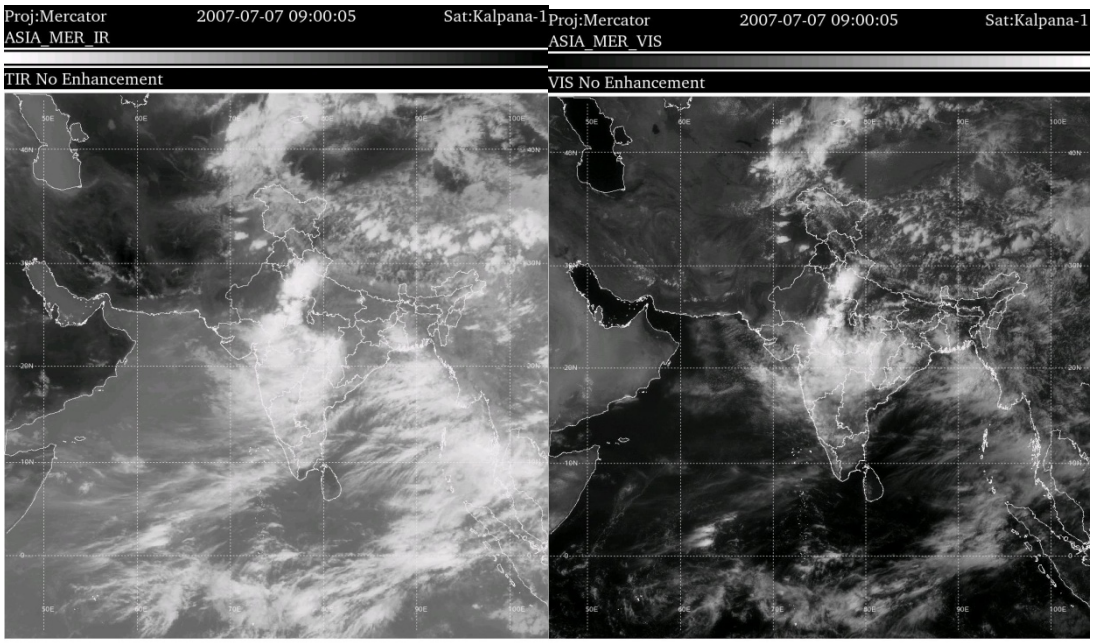


Fig.8(i).15: Kalapana- infrared and visible imagery (7th July2007) showing active spell of monsoon.

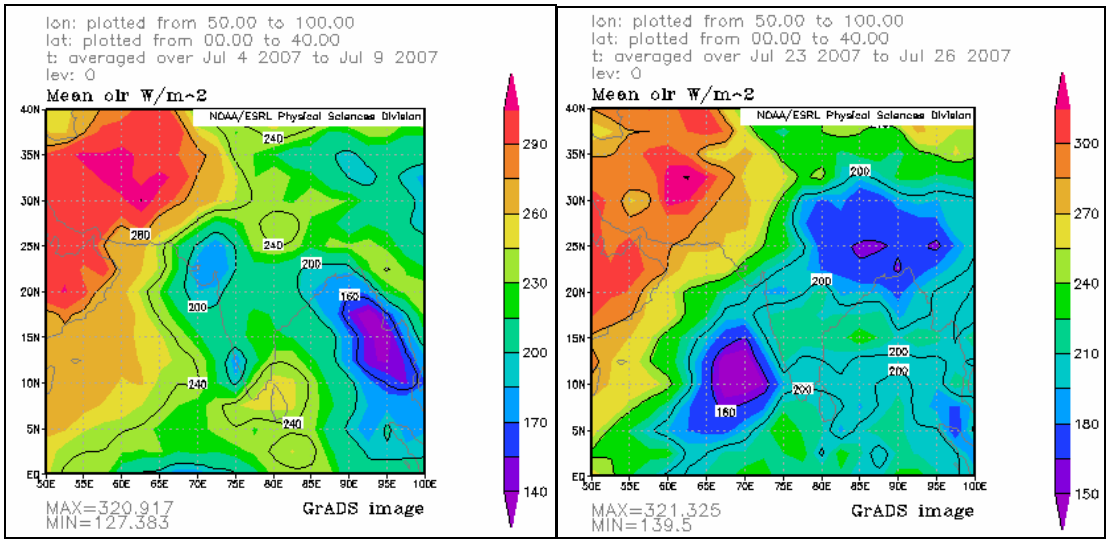


Fig. 8(i).16: NOAA Interpolated Outgoing Longwave Radiation (OLR) plots for two active spells of monsoon (July) 2007 (a) average for 4-9July (b) average for 23-26July.

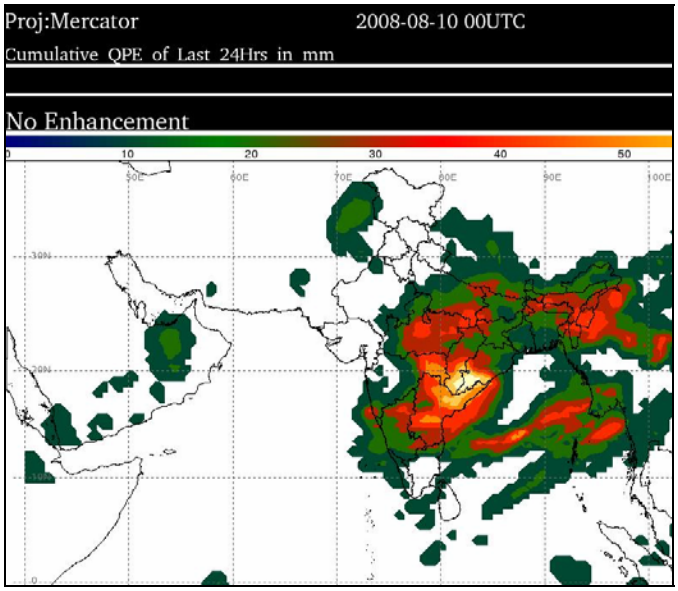


Fig.8(i).17: Kalapana-I derived QPE for 10th Aug. 2008.

The scatterometer aboard the QuikSCAT is a microwave radar designed specifically to measure ocean near-surface (at a 10m height) wind speed and direction. Scatterometer winds are very crucial to examine the strength of monsoon winds during active and break/weak phases of monsoon. As shown in Fig. 8(i).18, on 4th July'2007 (morning and evening passes) Southwesterly winds off Somalia coast, Westcentral Arabia Sea and northeast adjoining Eastcentral Bay are strong.

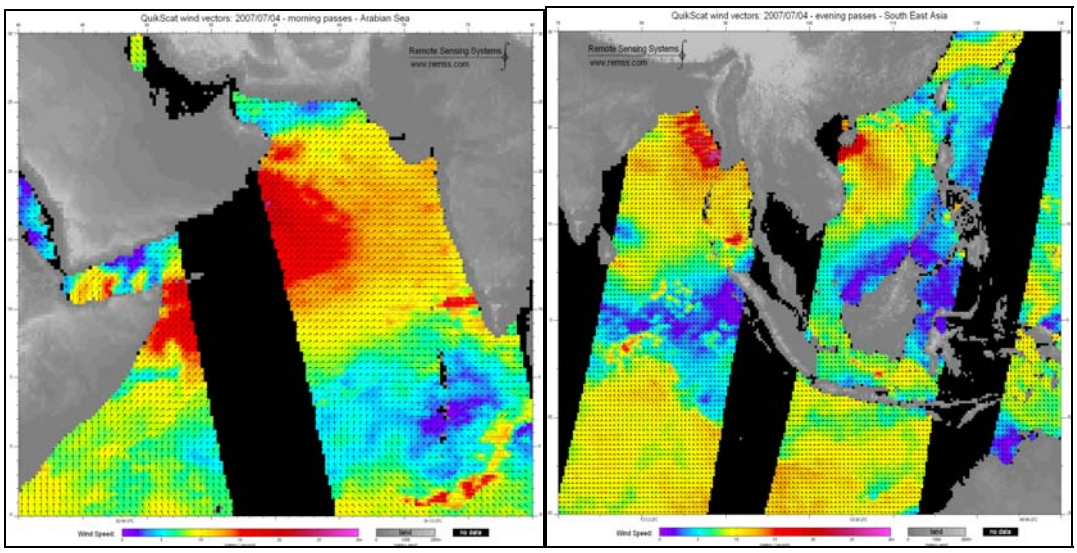


Fig. 8(i).18: Scatterometer winds for 4th July 2007 (morning and evening passes).

8(i).*2.8.3.* *Weak /Break spells*

The criterion used by India Meteorological Department (IMD) for identifying a 'break', is the synoptic situation associated with large rainfall anomaly. In Ramamurthy's (1969) comprehensive study of breaks during 1888 – 1967, a break situation is defined as one in which the surface trough (the "monsoon trough') is located close to the foothills, easterlies disappear from the sea level and 850hpa charts, provided the condition persisted for at least two days. The break composite of Ramamurthy (1969) shows large negative anomalies over a belt around the normal position of the monsoon trough and positive anomalies near the foothills of the Himalayas and Southeastern peninsula. He concluded one or two major breaks in the monsoon area a frequent occurrence.

It is generally recognised that long-lasting breaks in the monsoon are a large-scale phenomena. Sadler et al. (1968) identified the meridonial propagation of an equatorial buffer zone during the occurrence of breaks. They identified the presence of a clockwise circulation during the active monsoon spell prior to the occurrence of a break. The arrival of this anticyclone over the monsoon trough latitudes weakens the latter somewhat as a break in the monsoon rainfall encountered. A few examples of weak/break monsoon conditions as revealed by satellite products are presented in Figs. 8(i).19 -8(i).24.

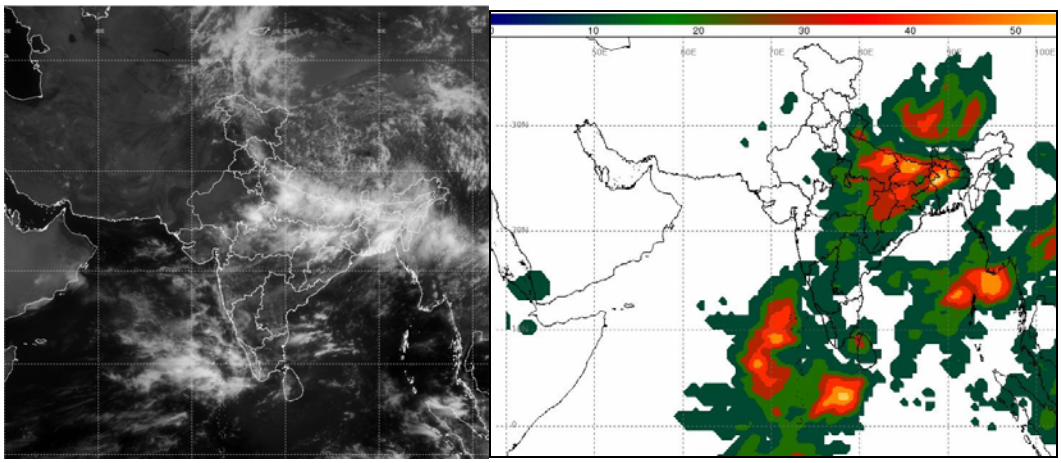


Fig. 8(i).19: Kalpanal (a) Visible imagery for 26th July 2007 0600UTC, and (b) QPE for 22nd July 2007; both products indicates that MT is close to foothills of Himalayas.

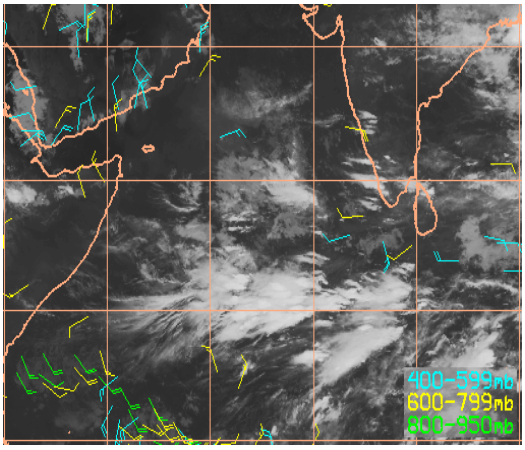


Fig. 8(i).20: Meteosat derived low-level winds for 11th Sep'2007 (0000UTC) showing very weak Southwesterlies current.

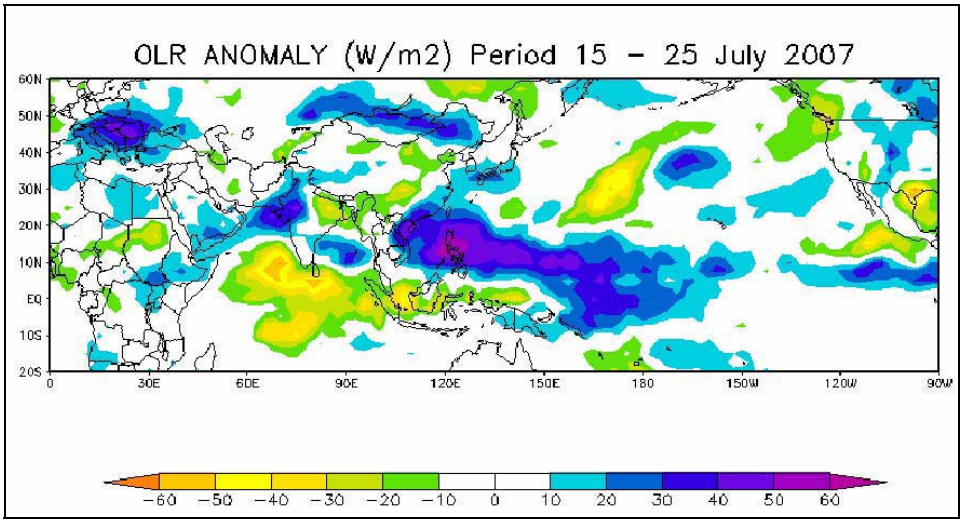


Fig.8(i).21: OLR anomalies during the break event period 15-25 July'2007. The anomaly chart shows large scale suppressed convection (positive OLR anomalies) extending from Northwest India to northwest Pacific across Bay of Bengal. Enhanced convection was observed over the equatorial Indian Ocean and Northeast India. Enhanced convection also was observed over the area adjoining the Caspian Sea. Over the equatorial Central Pacific, suppressed convection was observed, suggesting the effect of cooling over this area.

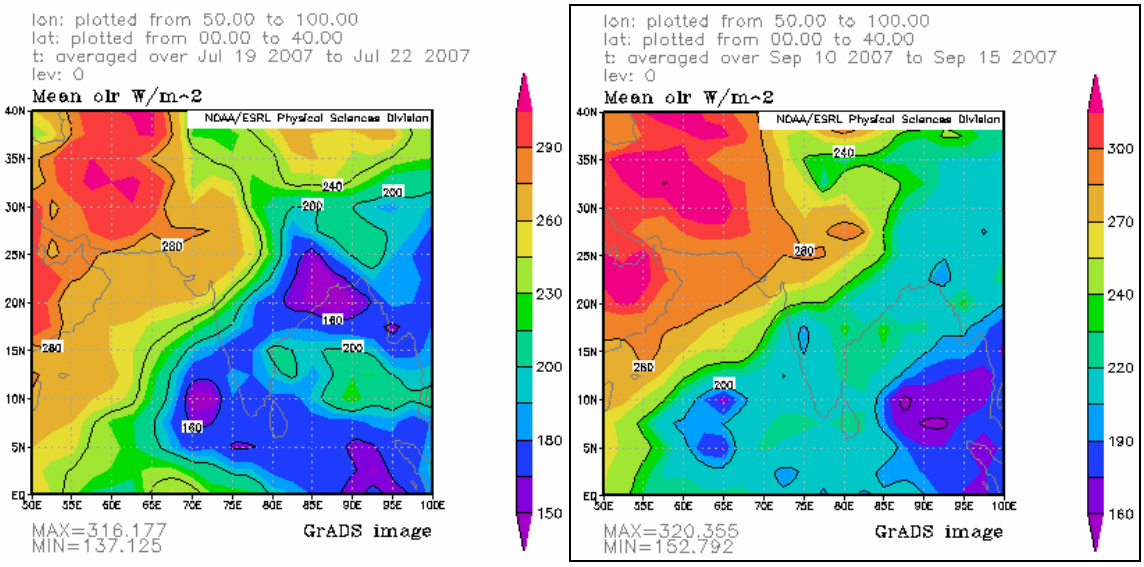


Fig.8(i).22: NOAA OLR average for the period (a) 19-22Jul. 2007, and (b) 10-15Sep. 2007.

By comparing the scatterometer winds data for the active (Fig.8(i).18) and break/weak (Fig.8(i).23) monsoon one can easily make the difference.

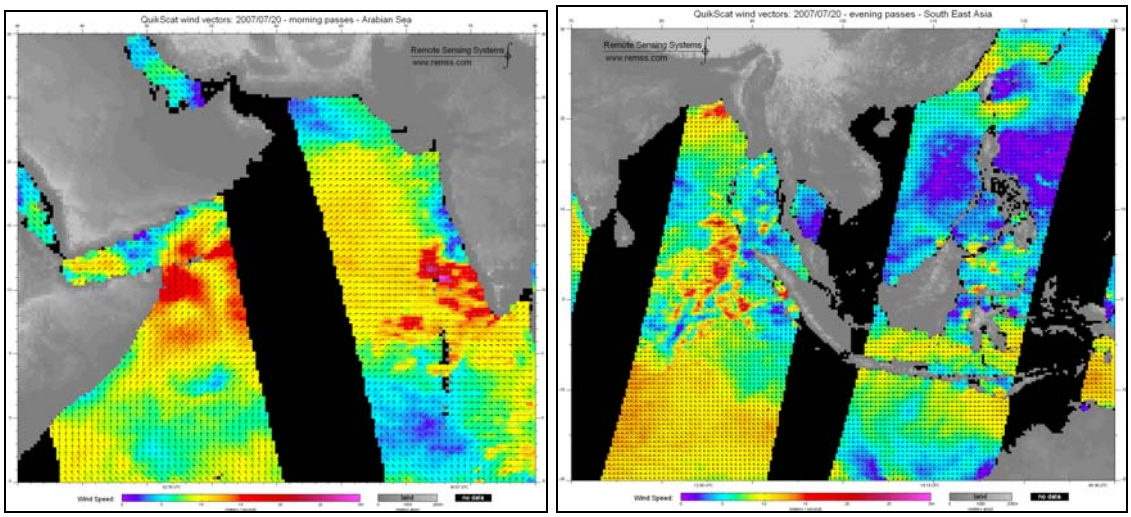


Fig.8(i).23: Scatterometer winds for 20th July 2007 (morning and evening passes).

A strong southern hemispherical equatorial trough (SHET) is an indicative of weak monsoon phase and *vice-versa, De & Prasad et. al (1988).* Fig.8(i).24 depicts strong SHET during weak monsoon phase in July, 2008.

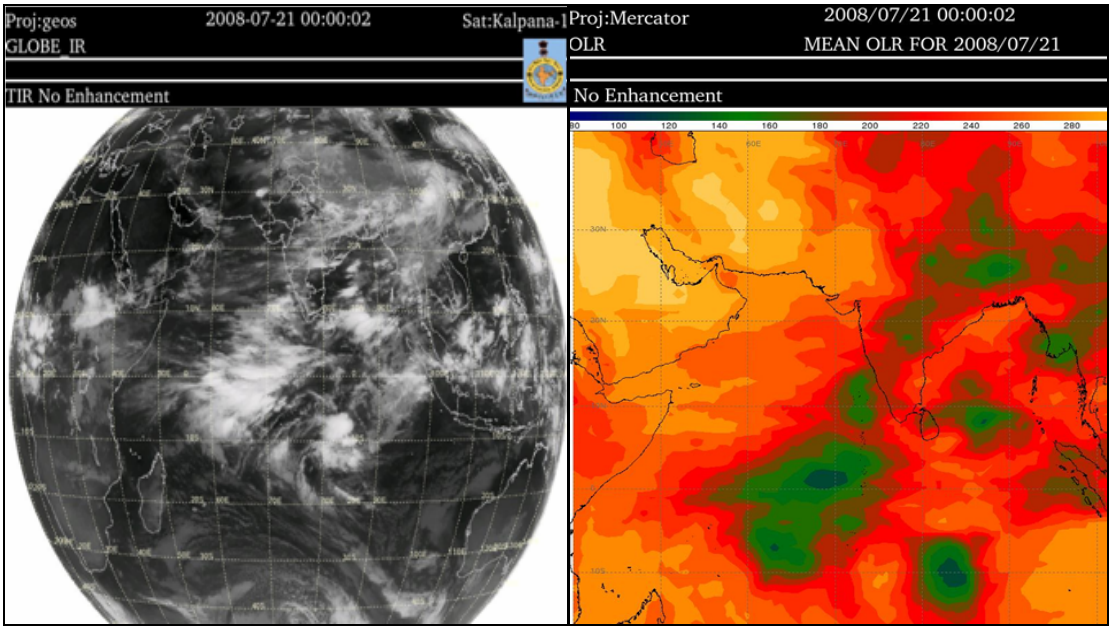


Fig. 8(i).24: Appearance of SHET during break/weak monsoon.

8(i).2.8.4. Relationship between Break and Active Monsoon spells over India and MJO Phases

Pai et al. (2009) used the real-time multivariate MJO indices (RMM1 & RMM2) of Wheeler and Hendon (2004) for defining the various phases of MJO. OLR data of NOAA satellites have been used in the computation of MJO indices. In order to examine the association of the break and active spells with the various phases of MJO, the break and active monsoon spells for the period 1974-2007 were considered from the list of these events provided for the period 1951-2007 by Rajeevan et al. (2008). During the period 1974-2008 (excluding1978), there were 47 break events (265 break days) and 57 active events (237active days).

More critical examination of break and active events revealed that 109 of the 265 break days (41%) and 113 of the 237 active days (48%) were in the weak MJO category. The Fig.8(i).25 depicts the frequency distribution of the break days during the 8 Phases of MJO for both strong and weak categories. It can be seen that when the MJO is strong, maximum number of break days were associated with the phase 1 followed by phase 2. For the weak MJO category also, the maximum number of break days were associated with the Phase 1. However, for the weak MJO category,

the sum of the number of break days during phases 7 & 8 was nearly equal to that during Phases 1 & 2. Thus on considering all the 265 break days irrespective of the strength of the MJO, the maximum number of break days (86%) were associated with the first two (Phases 1 & 2) and the last two (phases 7 & 8) phases.

Fig. 8(i).26 is same as Fig. 8(i).25 but for active days. As seen in the Fig.8(i).26, 124 active monsoon days were associated with the 8 strong MJO Phases with maximum frequency during the Phases 5 & 6. However, within the active days associated with weak MJO category, the maximum frequency was during the Phase 4 followed by Phase 3. Both strong and weak categories taken together, relatively highest number of active days (62%) were associated with the 4 phases from phase 3 to phase 6.

These results suggest that the onset and duration of break and active events were related to the strength and Phase of MJO. However, the association of break monsoon events with MJO was relatively stronger than the association of active events with MJO.

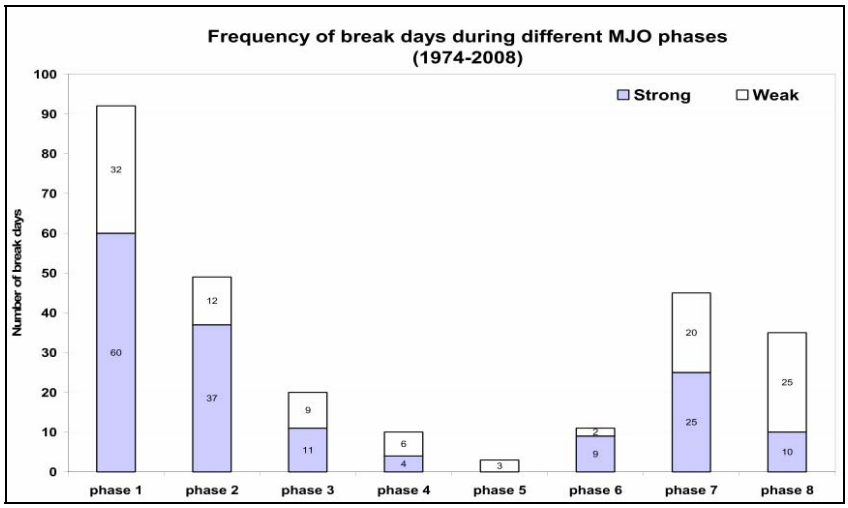


Fig. 8(i).25: Frequency distribution of break days for various Phases of MJO. (Source: Pai et.al. 2009)

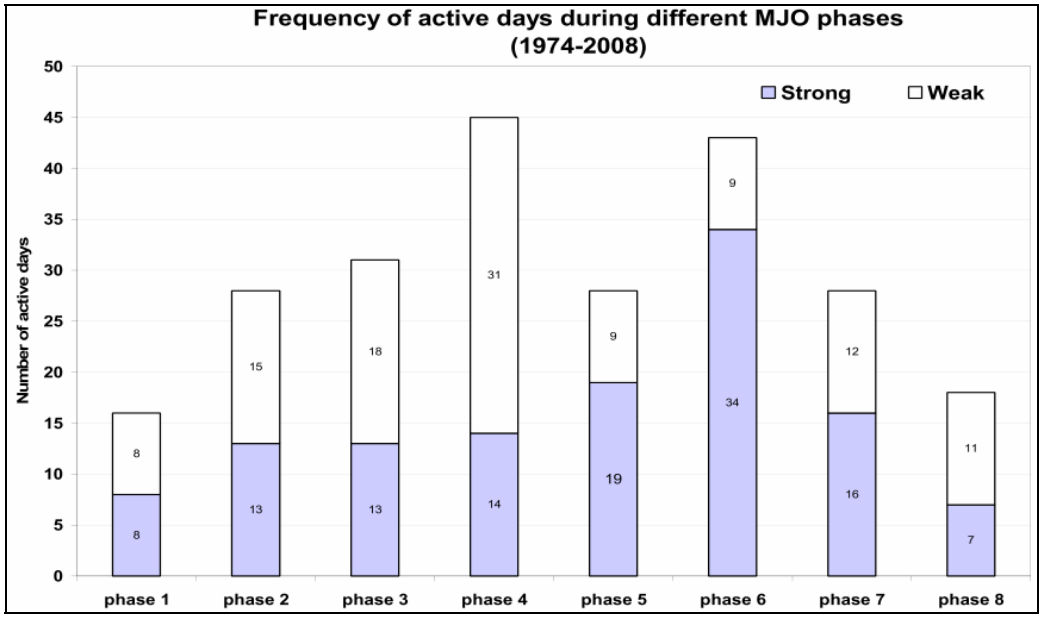


Fig. 8(i).26: Frequency distribution of active days for various Phases of MJO (source: Pai.et.al.,2009)

8(i).2.9. Monitoring of withdrawal of monsoon

As per IMD, following major synoptic features should be considered for the first withdrawal from the western parts of northwest India:

1. Cessation of rainfall activity over area for continuous 5 days.
2. Establishment of anticyclone in the lower troposphere (850hPa and below)
3. Considerable reduction in moisture content as inferred from the satellite water vapour imageries and tephigrams.

Further withdrawal over the country may be declared keeping the spatial continuity, reduction in moisture as seen in the water vapour imageries and prevalence of dry weather for 5 days.

In 2007, there was an unusual delay in the withdrawal of monsoon from extreme west Rajasthan due to the prevalence of cyclonic circulations, availability of moisture and sporadic rainfall activity over the northwest region. The southwest monsoon withdrew from western parts of Rajasthan and some parts of Punjab and Haryana on 30th September 07, with the withdrawal line passing through Amritsar, Hissar, Jaipur, Ajmer, Barmer, Lat.24°N/Lon.65°E, Lat.24°N/ Lon.60°E. It withdrew

from the entire northwest India on 2nd October and some more parts of north and central India by 10th October. During the period 1960-2006, the most delayed date of monsoon withdrawal from extreme west Rajasthan was 28th September, which occurred in 1964 & 1970.

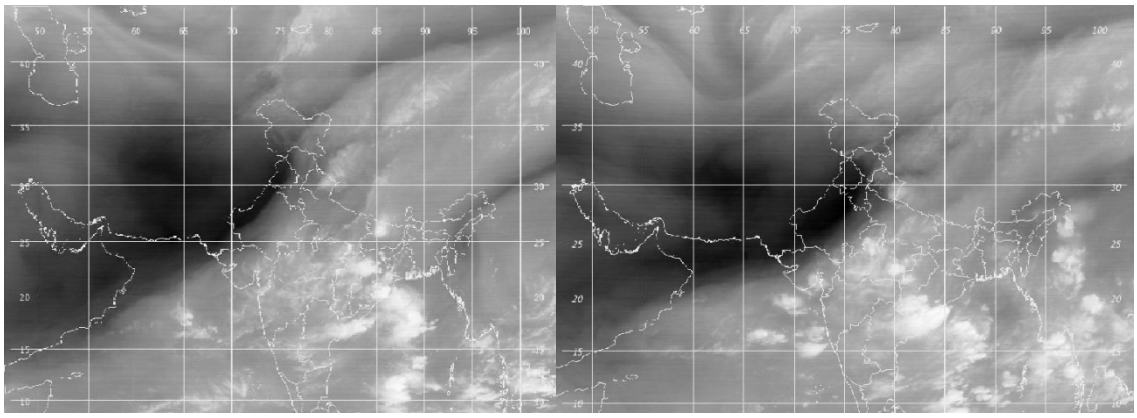


Fig. 8(i).27: Water vapour imagery for (a) 0600UTC of 26-09-2007 (b) 0000UTC of 27-09-2007

8(i).2.10. Satellite data application in understanding the intra-seasonal variability of monsoon

Cloud systems over oceans and land can be readily seen from the daily maps of the outgoing long-wave radiation derived from satellite measurements. Low values of OLR correspond to emission from higher levels and hence higher cloud tops (and more rain when the clouds are deep), and OLR is used as a proxy for rainfall in the tropics. Hence large negative (positive) anomalies in OLR are associated with large positive (negative) anomalies in rainfall over a region.

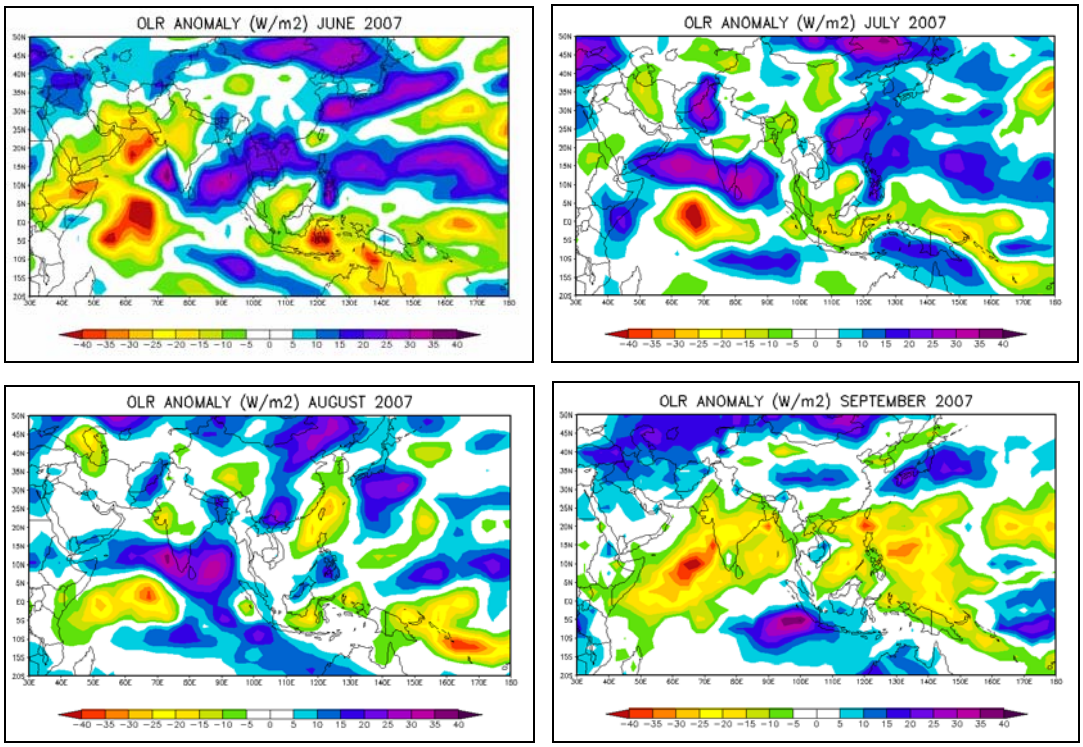


Fig.8(i).28: OLR anomalies during June to September 2007.
(Source: IMD monsoon report 2007)

Outgoing Long wave radiation (OLR) anomalies during June to September 2007 months are shown in Fig.8(i).28. In June, negative OLR anomalies suggesting enhanced convection were observed over a large area from the west equatorial Indian Ocean to northwest India across the Arabian Sea. Over Bay of Bengal and northeast India, OLR anomalies were positive suggesting suppressed convection during the month. In July, OLR anomalies were negative over central parts of India and northeast India. However, over south peninsula, anomalies were positive suggesting suppressed convection. During the month of August, OLR anomalies were negative over Gujarat and adjoining area and the west equatorial Indian Ocean. Positive anomalies were observed over extreme south peninsula and parts of northwest India. During the month of September, negative anomalies were observed over most parts of India and Arabian Sea, suggesting enhanced rainfall activity over India and Arabian sea. Over the equatorial East Indian Ocean, positive OLR anomalies were observed suggesting suppressed convection over this region. The positive (negative) OLR anomalies over the east (west) equatorial Indian Ocean resembles a positive phase of the Indian Ocean Dipole. Pentad wise evaluation of

OLR anomalies have been studied by De & Mukhopadhaya (2002) for shadowing the commencement/cessation of breaks with good success.

The composite rainfall anomaly analysis associated with the various phases of MJO revealed strong intraseasonal variation in the spatial rainfall anomaly distribution over India. During MJO phases of 1 & 2, break monsoon type rainfall distribution was observed over India (Fig. 8(i).29). Subsequently, as the MJO propagated eastwards, a gradual northward shift of the above normal rainfall band from south Peninsula to north India was observed. During phases 5 & 6, the above normal rainfall band was observed along monsoon trough region and active monsoon type rainfall distribution was observed. During the subsequent phases (7 & 8) a general decrease in the rainfall was observed over most parts of the country. The observed intra-seasonal variation in the rainfall can be explained by the changes in the convective and circulation anomalies observed in association with the formation of MJO induced positive convective anomalies over equatorial Indian Ocean during Phases 1 & 2 and its northward propagation.

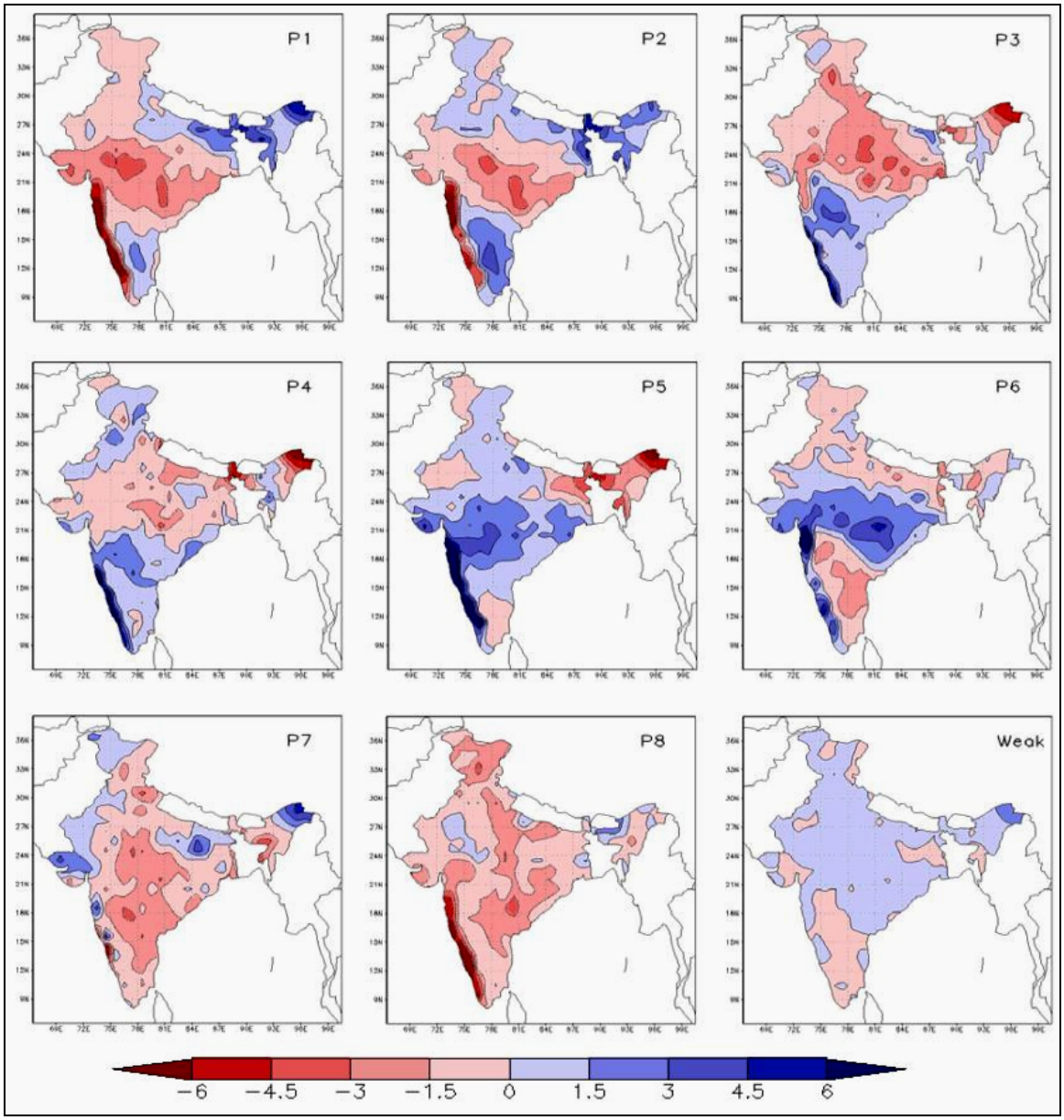


Fig. 8(i).29: Maps of composite rainfall anomaly (mm) in respect of 8 strong phases and the weak category of MJO derived using data for the period 1974-2008 (excluding 1978). Maps for the 8 strong MJO phases are labeled as 'P1', 'P2' etc. and the map for weak category is labeled as 'weak' (Source: Pai et al., 2009).

8(i).3. Satellite data assimilation into numerical weather prediction (NWP) models

Retrievals of temperature and moisture profiles from polar orbiting satellites have been available operationally since 1978. Generally, inclusion of the retrievals improved the forecast skill of numerical weather prediction models inspite of

limitations in the retrieval methods. As the numerical models and data assimilation techniques improved throughout the 1980's, the quality of the retrievals also improved.

The forward radioactive transfer problem, calculating radiances from an initial temperature and moisture profile, is becoming a better-understood problem due to improvements in surface emissivity models and atmospheric absorption models. However, the inverse problem of calculating a temperature profile from the observed spectral radiances is mathematically ill-posed (Rodgers 1976). In order to constrain the problem, prior information must be specified. For most retrieval methods, this prior information is independent of the NWP model and typically contains less information about the current atmospheric state than is provided by a NWP forecast.

In addition, the satellite retrievals have relatively poor vertical resolution due to the broad, overlapping spectral weighting functions, so that the number of independent pieces of information is less than the number of channels. This means that the sounder is fundamentally "blind" to shallow vertical structures and any small-scale vertical structure in the retrieval must come from the prior information and not the instrument (Eyre and Lorenc 1989; Goldberg et al. 1988). Furthermore, the contribution of errors in the prior information to the total retrieval error is significant and difficult to characterize correctly.

Most of the efforts to improve the quality control and use of the satellite retrievals failed due to the inherent limitations discussed above. The solution appeared to be to not use the satellite observations as if they were poor quality radiosondes, but directly as radiances. This realization led to research into methods to assimilate the radiances directly into the analyses.

The simplest version of variational assimilation of radiances to implement is to consider the problem in one (vertical) dimension (known as 1D-VAR). The observed radiances are combined using optimal estimation theory, a forward radiative transfer model and its adjoint, with the co-located temperature and humidity profiles from a short term forecast from the NWP model. The resulting maximum likelihood estimates of temperature and moisture may be assimilated directly into the existing

operational optimum interpolation analysis. Results indicated consistent positive impact in the Northern Hemisphere from the 1DVAR retrievals. Both NCEP and ECMWF have implemented three-dimensional variational assimilation systems for all observations, including TOVS radiances; ECMWF recently implemented a four-dimensional variational system. 1DVAR has a fundamental technical limitation when used for update cycling - namely, the same background is used for both the 1DVAR retrieval and the subsequent full three-dimensional analysis: this leads to 1DVAR retrievals whose errors are correlated with the analysis background errors.

The primary advantage of variational assimilation methods is that nonlinear observation operators can be readily included allowing for the consistent use of indirect observations. The disadvantages are that the development of the observation operator and its adjoint and appropriate error covariances require considerable effort, knowledge and experience. For the radiance problem, this includes knowledge of the satellite sensors, radiative transfer theory, principles of remote sensing, data assimilation, numerical modeling and parameterization schemes and meteorology. Many components of the problem are unique to each individual sensor. These include the forward radiative transfer model, expected error covariances of the observations and forward operator, quality control and bias corrections.

8(i).*3.1.* *Data assimilation*

Data assimilation is a technique for observational data modification process, in which observational data is modified in dynamically consistent fashion in order to obtain a suitable set of data for model initialization. It is well known that a model forecast critically depends on the initial conditions employed for integrating the model. Data assimilation is divided into two processes:

(a) Objective analysis and

(b) Data initialization.

In the objective analysis, all data acquired for a given time from the irregularly spaced observational network of surface and upper air stations are checked for accuracy and interpolated to point on regular latitude-longitude grid at standard pressure levels. A background or first guess estimate derived from a short-range

forecast is used to fill in grid points in data-sparse regions such as ocean. Such objectively analysed data still contain noise that is likely to be interpreted as spuriously large gravity waves when data are used as initial data in the model. So, in initialization process, these objectively analysed data are modified in order to minimize the gravity wave noise, and hence reduce the magnitude of initial velocity and pressure tendencies. In four-dimensional data assimilation (4D-VAR), the two processes: objectively analysis and initialization are combined.

8(i)*.3.1.1. One-dimensional variational assimilation*

One-dimensional variational analysis (1DVAR) minimizes the cost function using Newtonian iteration to arrive at the maximum likelihood. The background is given by a short term forecast from the NWP model interpolated bilinearly to the observation location and pressure levels required by the forward model (40 for RTTOVS 3.0; Eyre 1991). In 1DVAR, the final retrieval step is performed explicitly, and the resulting profiles of temperature and humidity may be readily used in current multivariate optimal interpolation (MVOI) analysis system. 1DVAR is a very useful tool for understanding specific aspects of a problem.

8(i)*.3.1.2. 4 D-VAR Assimilation*

In many oceanic regions there are very few conventional data source available, and it is necessary to rely on asynoptic data (i.e., observations at nonstandard times, such as those from ships, aircraft, and satellites).Fig.8(i).30 depicts schematic for 3D-VAR and 4D-VAR techniques.

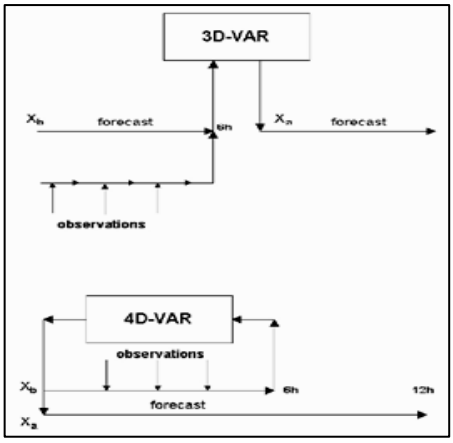


Fig. 8(i).30: Schematic diagram of 3D-VAR and 4D-VAR technique.

8(i).4. Impact of Satellite Observation on Operational Forecasts

The instruments onboard Nimbus 6 (HIRS and SCAMS (latter called MSU) became operational on TIROS N launched in 1978. NESDIS operationally produced 14 layer retrievals from these radiances called SATEMs. In1991 SATEM, were temporary remove from the Northern Hemisphere due to spurious temperature increments caused by NESDIS retrievals.

In June 1992 a 1D-VAR retrieval method (Eyre 1993), replaced the statistical SATEMs. Using this method the observation error was better defined and a positive impact from SATEMs was found in both Northern and Southern Hemispheres. In 1996, 3D-VAR was introduced into operations using the satellite radiances directly by the analysis (Andersson, et al. 1994). The direct use of radiances in the analysis greatly simplifies their use. Fig. 8(i).31 shows the day 3 forecast skill (rms geopotential height error verified using own analysis) for ECMWF (red) MetOffice (blue) and NCEP (green). The introduction in 1998 of 4D-VAR (Klinker 2000), marked another important operational change at ECMWF and has been since followed with further analysis and model improvements (Fig.8(i).32).These examples show that there is a need for effective satellite data assimilation into NWP models for improved monsoon monitoring and forecasting.

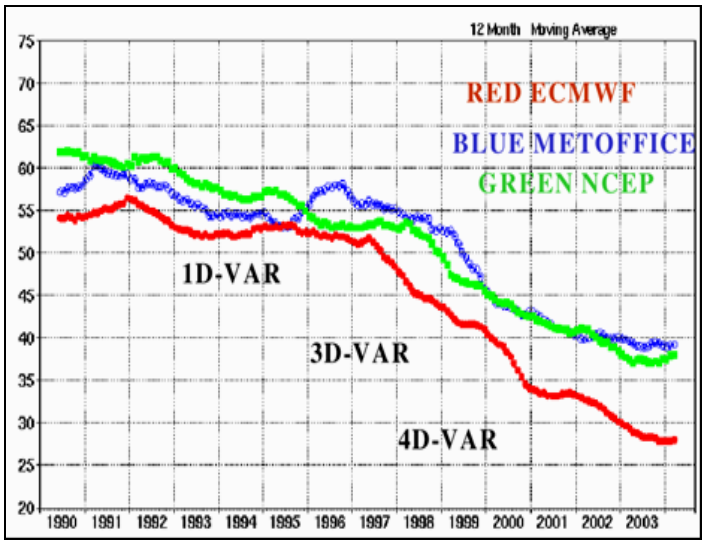


Fig. 8(i).31: Graph shows the 500 hPA rms error for day 3 forecast. (Source: Kelly et al., 2004)

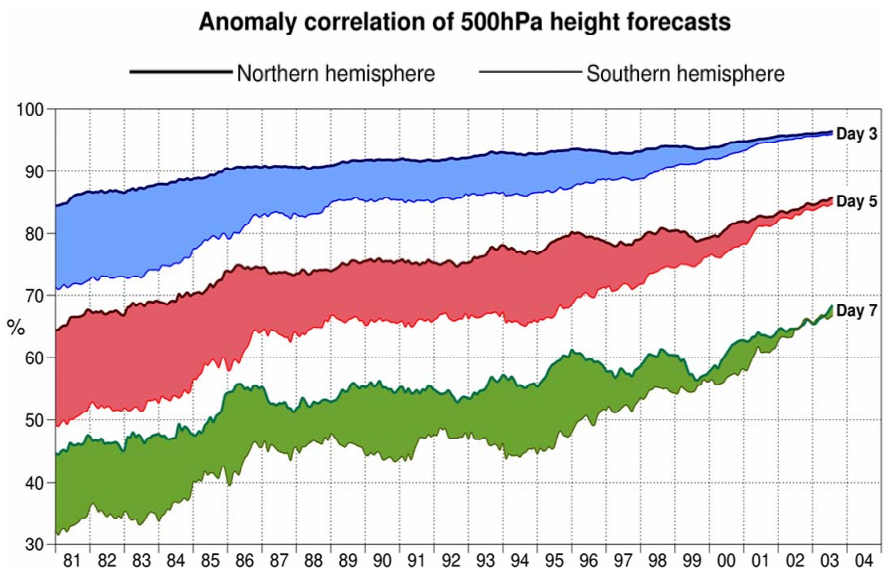


Fig. 8(i).32: Anomaly correction of 500hPa height for 3, 5, and 7-day forecasts for the ECMWF operational model as a function of year. Top and bottom of each band correspond to Northern and Southern Hemispheres, respectively. Note that the difference in skill between the two Hemisphere has nearly disappeared in the past few years, due to the successful assimilation of satellite data.

8(i).5. Future programmes

Several new meteorological satellites including OCEANSAT-2 and MEGHATROPIQUES have been launched recently which provide products like ocean surface winds, rain rate, radiation budget etc which are vital inputs to monsoon monitoring. Future missions like INSAT-3D will carry an advanced imager and a sounder payload. The INSAT-3D will provide many products, besides vertical profiles of temperature and humidity which will find applications in monsoon monitoring and forecasting. There are exciting proposals for satellite measurements of aerosol and precipitation. These products will help in better monitoring of monsoon. Assimilation of more satellite products into numerical models is likely to improve monsoon prediction on different scales. The future of satellite meteorology is indeed promising.

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CHAPTER 8 (ii)

OBSERVATIONAL AIDS AND MONITORING OF MONSOON
(ii) RADAR, AN OBSERVATIONAL TOOL FOR MONITORING MONSOON

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8(ii).1. Introduction

Wind and precipitation are the two main criteria deciding the onset, progress, strength and effectiveness of monsoon. Precipitation can be considered the most crucial link between the atmosphere and the earth-surface in weather and climate processes. Quantitative precipitation estimates at high spatial and temporal resolution are of increasing importance for water resource management, for improving the precipitation prediction scores in numerical weather prediction (NWP) models, and for monitoring seasonal to inter-annual climate variability. A dense and high-temporal resolution ground-based measurement network is required to achieve accurate precipitation observations. However, in several regions, especially over the tropical land areas and over the oceans, the coverage by rain gauges is insufficient. Due to the sparse distribution of rain gauges, monitoring of the monsoon rains is rather difficult.

Local economy, hydrology, and ecology in India are heavily dependent on the availability of monsoon rains. Less rainfall during the monsoon season also results in an increased surface albedo (because of decreased soil moisture content), increased dust generation, and less agricultural yield. Thus it is of utmost importance to have a reliable prediction system in place for planning and management of monsoon bounties. Monsoon predictions specifically using computer models require high spatial and temporal distribution of rainfall data for improving their skill-score.

Therefore a continuous monitoring of monsoon rain with finer details in space and time is of great importance. Rainfall estimates using ground based radar network as well as space borne global precipitation radars do have the potential to provide the degree of resolution and timeliness for improving this monitoring. Apart from these, data from mobile, ship-borne and airborne radars as well as wind profilers do augment the real-time data set available for close and precise monitoring of monsoon activity.

Onset and progress of monsoon is indicated by the horizontal profile of wind over the land and neighboring sea area of the country as well as the vertical extent of the wind system. A network of weather radars with the Doppler measurement capability can provide wind field data with high spatial and temporal resolution. Recognising the potential of radars in capturing and depicting details of wind and precipitation fields of monsoon in near real-time, India Meteorological Department (IMD) has embarked upon augmenting its radar network as part of its modernisation programme. A radar network covering the entire nation with a reasonable overlap over coastal region is being established. Fig.8(ii).1 depicts the radar network coverage of IMD [both non-Doppler radars prior to 2001 and the Doppler Radar network being established].

8(ii).2. Types of radars in use for monitoring monsoon

Based on the operating frequency and purpose of use, Weather radars can be classified into different categories. For precipitation estimates S-band radars are the most preferred ones as they can see through heavy precipitation for long distances, without getting attenuated, and provide reliable estimates. However, owing to their huge size and heavy infrastructural constrains they are relatively expensive and hard to maintain. Recent advancement in technology has devised some workarounds for accounting the attenuation effect and made radars operating in the higher frequency bands like C and X-band also provide reliable precipitation estimates, though limited to shorter ranges. Radars used for study of cloud dynamics operate in still higher frequencies even up to Ka and Ku bands, where as wind profilers are operated in HF and VHF bands. Mobile, airborne, and ship borne radars operate mostly in X-band, as heavy radars can’t fit in their limited structural and space constraints.

Weather radars are installed on fixed platforms as well us moving platforms like, spacecrafts, ships, Aircrafts and even earth movers like Vans. Space borne precipitation measuring radar under the famous TRMM (Tropical rain Measuring Mission) of NASA is a befitting example of use of radars in monsoon rain monitoring.

8(ii).3. Capabilities of typical Doppler weather radar in capturing monsoon parameters

Bhatnagar et.al (2003) provides a fairly detailed description of typical Doppler weather radar and its products. General principles and theory behind radar meteorology are well explained by Atlas (1990). Detailed discussions on capabilities & limitations, choice of frequency, scan strategies, anomalous propagation echoes, electro magnetic theory of atmospheric sounding, calibration and quality control of the data set etc. can be seen in Doviak & Zrnic (1993) and Reinhearts (1999).

8(ii).4. Products of Doppler weather radars (DWR)

A number of products are available from Doppler weather radars. From non-Doppler radars, information on reflectivity factor alone is available, whereas from DWRs, in addition to reflectivity, radial velocity and spectrum width information are also available as base data. These data can be used directly for base-product display and also for deriving further products based on standard algorithms. A few products, which are commonly used in operational meteorology with relevance to monsoon monitoring, are briefly described here.

8(ii).4.1. Base parameters

Reflectivity factor (Z), radial velocity (V) and velocity spectrum width (W) are the three base data directly observed/measured by the DWR. The logarithmic radar reflectivity factor is defined as $Z = 10 \log_{10} [\sum (N_i D_i^6)/(1 \text{ mm}^6/\text{m}^3)]$, where N is the number of droplets of diameter Di to $Di + \delta D$, δD being the diameter interval used in making the measurements, present in unit volume of sample being probed. For the conditions prevailing in most of the weather systems and for the wavelengths used, the scattered power received back is directly proportional to Z (derived by Lord

Rayleigh in 1870s). Hence the weather signal power available at the receiver output is a direct measure of Z. Autocorrelation of time series formed by the received power spectrum is the basis for deriving Z and other Doppler moments. The zeroth lag autocorrelation R_0 of the time series is proportional to weather signal power, and hence Z is computed from it. Mean radial velocity of hydrometeors inside the sample volume is given by the first lag autocorrelation R_1, and the velocity spread inside the sample volume is obtained from the first and the second lag autocorrelations R_1 and R_2 together, assuming a Gaussian distribution. A point to note while interpreting velocity information available from DWRs is that the radial wind is not the actual prevailing wind, but is the *component of the air motion in the direction of the probing radar beam*. The base data available from DWRs are generally displayed in the following formats.

8(ii).4.2. Radar products for visual interpretation

8(ii).4.2.1. Plan Position Indication (PPI)

In this form of radar image, the parameter pertaining to a constant elevation surface (conical surface with radar as the vertex), is projected on a plain surface to form a radar centric two-dimensional image. The parameter generally displayed in this type of image is anyone of Z, V, W, rain-rate (R) etc.(Fig. 8(ii).2 & Fig. 8(ii).3)

8(ii).4.2.2. Constant Altitude PPI (CAPPI)

In this form of radar display, the parameter pertaining to a constant altitude surface (curved surface around the radar and parallel to the earth's surface) is projected on a plain surface to form a radar centric two-dimensional image. Any of the parameters displayed in PPI can be displayed in CAPPI format also. (Fig. 8(ii).10)

8(ii).4.2.3. Surface rainfall Intensity (SRI)

This form of display is similar to the CAPPI but the parameter displayed is exclusively the rain-intensity over the surface above ground at the selected altitude. This is a derived product in the sense that the measured Z is converted to rain rate

using a power regression relation similar to the well known Marshel –Palmer equation $Z = 200 R^{1.6}$, where Z is the linear radar reflectivity factor in mm^6/m^3 and R is in mm. (Fig. 8(ii).4)

8(ii).4.2.4. Precipitation Accumulation (PAC)

This is another form of visualizing radar derived rainfall quantity. This is a third level product derived by integrating the rain-rate over time. The resulting parameter is the rain accumulation during the period of integration (Fig.8(ii).5). To have a reliable product, the time interval between rain-rate sampling must be kept as small as possible.

8(ii).4.2.5. Vertically integrated Liquid water content (VIL)

The parameter for this product is derived by converting the Z values to liquid water content and further integrating the same in vertical (using a relation similar to the Marshal-Palmer relation). VIL is displayed in a radar-centric 2-D display. (Fig. 8(ii).8)

8(ii).4.2.6. Vertical Wind Profile

All available radial velocity data from a volume scan and within a cylindrical volume of limited radius (within which the assumption that the wind field is linear holds good) are processed together using a multidimensional linear regression method to derive the horizontal wind at different vertical levels. This so-called volume velocity processing (VVP) technique introduced by Waldteufel and Corbin (1979) provide one of the very reliable wind products in the form of vertical time section of wind profile (Fig. 8(ii).6). Optionally the vertical velocity and air-vergence also can be computed and displayed as X-Y plot.

8(ii).4.2.7. Horizontal wind display

From the distribution of radial velocity, which is the component of the air motion in the direction of probing radar beam, prevailing wind in the vicinity of the radar

(typically up to a range of about 100 km) is derived using proven algorithms. The derived wind is displayed in a CAPPI like surface using conventional wind barb notation at selected grid points within the selected maximum range (Fig. 8(ii).10).

8(ii).5. Strength and limitations of DWRs

The two major limitations of radar in quantitative estimation of weather parameters are

(i) Inability to see the lower layers of the atmosphere at farther ranges due to the curvature of the earth and

(ii) Poor spatial resolution of data and products at farther ranges due to angular spreading of the sample volume (beam widening).

Owing to the above limitations utility of radar data for quantitative estimates is limited to about 100 to 150 km from the radar. There are many other errors and limitations of varying significance in the radar derived met parameters due to many assumptions and exposure conditions involved in the data acquisition and estimation system.

Despite all these limitations, radars continue to occupy an unique position in remote sensing of weather owing to their unparalleled strength in capturing the minute details of small scale variations of weather both in space and time and making the high resolution data available in near real time and at frequent intervals. Radars are also capable of collecting data over ocean and difficult terrains around them, which would have remained unfilled otherwise (Fig.8(ii).9).

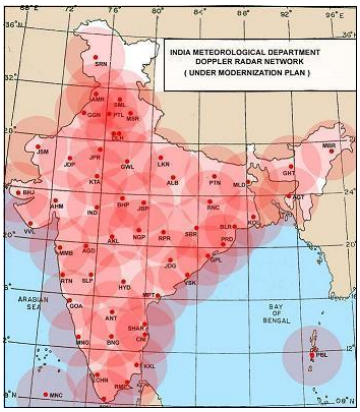
It is seen that despite uncertainties and errors involved in the radar based quantitative precipitation estimates, final product shows a good correlation to gauge recorded rainfall (Fig. 8(ii).7). Adjusting radar recorded rainfall using standard data from gauge / disdrometer render the radar data more reliable for both near real-time use as well as for water resource management and flood forecasting purposes.

Continuous operation of weather radars using scan strategies designed for precipitation measurement for an extended period will enable building a detailed

hydro-climatology of the region around the radar. Once the envisaged radar network-density is achieved, it will be possible to build the hydro-climatological mapping of the country with spatial resolution down to sub-village level. Such a huge and detailed database will serve as a unique and powerful tool in monitoring of onset, advance, coverage and intensity of monsoon and eventually in monsoon management.

IMD X-band radar network (2001)

IMD S-band radar network(2001)



IMD DWR network being established since 2001

Fig.8(ii).1: Radar network of IMD – Both Non-Doppler radars prior to 2001 and the Doppler Radars being established

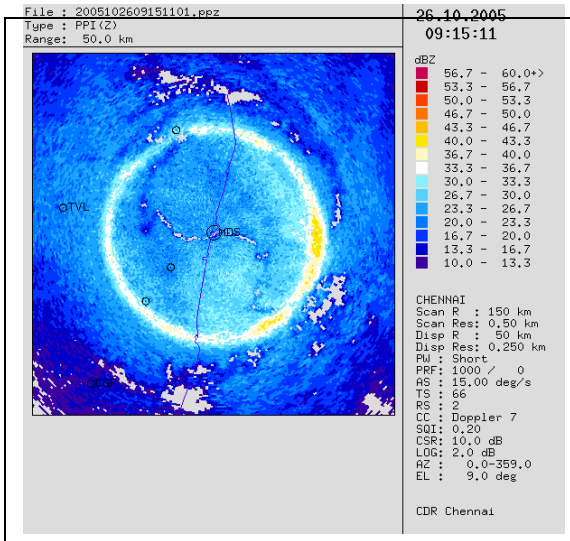


Fig. 8(ii).2: Plan Position Indication of reflectivity Bright band due to reflection from melting layer is seen

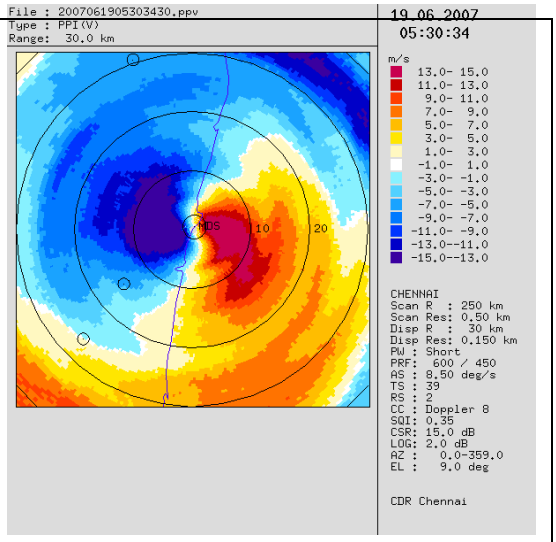


Fig. 8(ii).3: Plan Position Indication of radial Velocity Westerly wind veering with height is seen

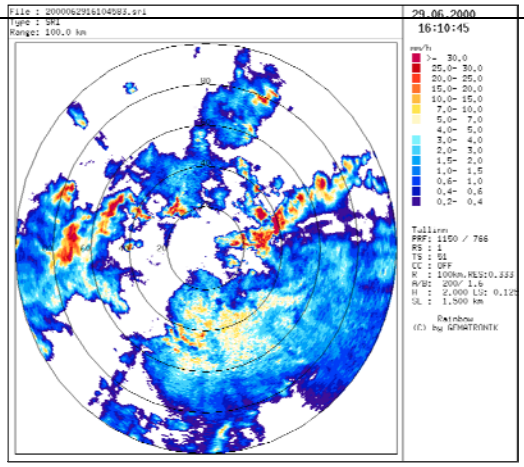


Fig. 8(ii).4: Surface Rainfall Intensity in mm/h

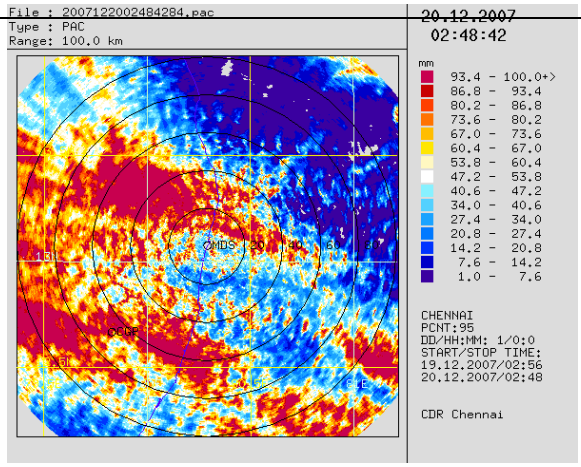


Fig. 8(ii).5: Precipitation Accumulation on a widespread rainy day

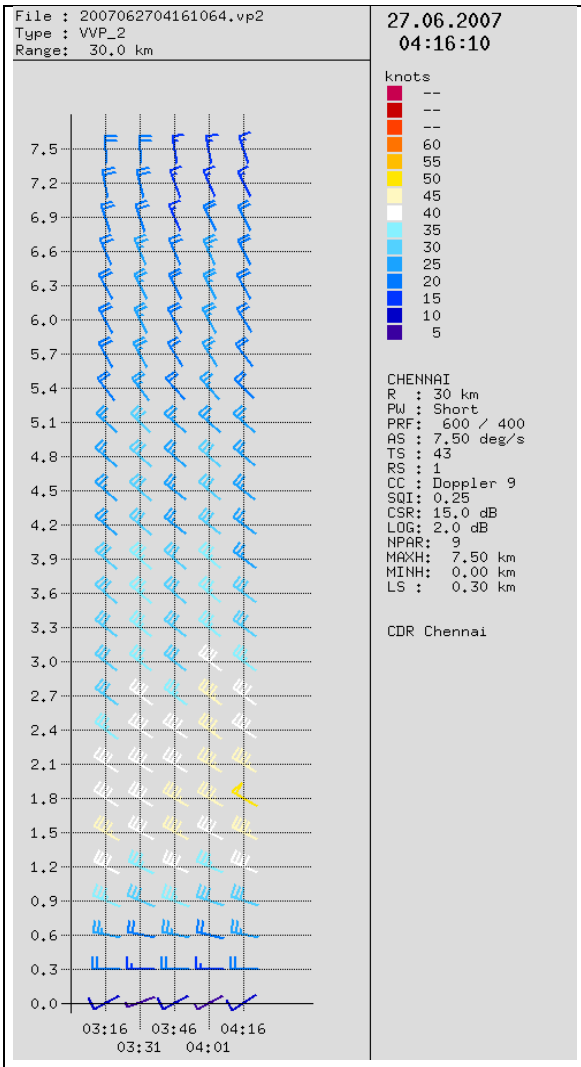


Fig. 8(ii).6: Vertical time section of wind given by VVP2 algorithm

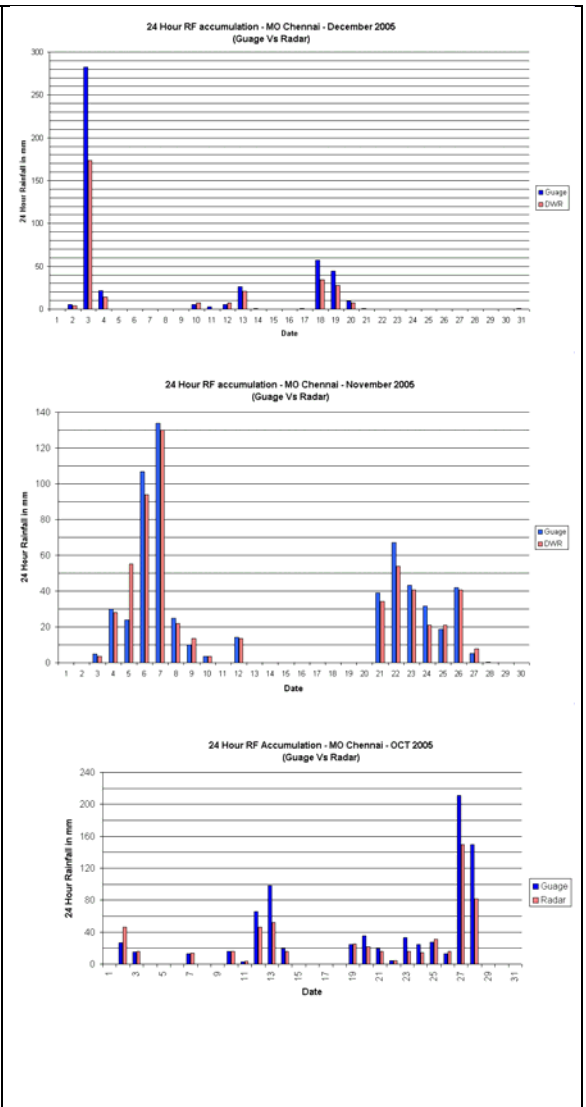


Fig. 8(ii).7: Comparison Radar and Gauge reported rainfall over Meenambakkam during NE monsoon 2005

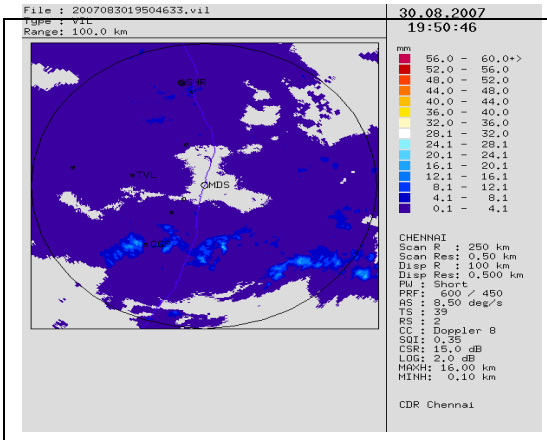


Fig. 8(ii).8: Vertically integrated Liquid (0.1 to 16 km vertically and 100 km horizontally)

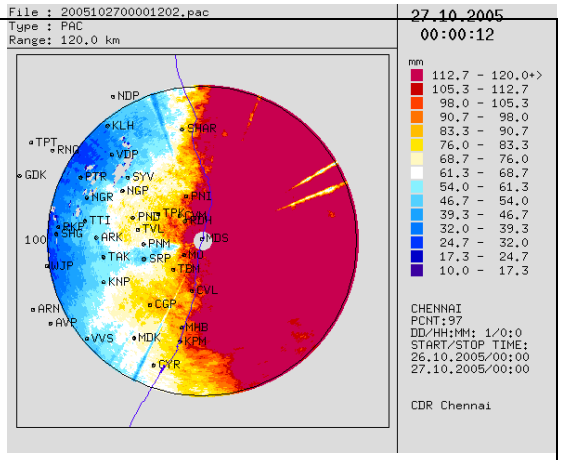


Fig. 8(ii).9: Heavy rain over sea (Gauge coverage is nil)

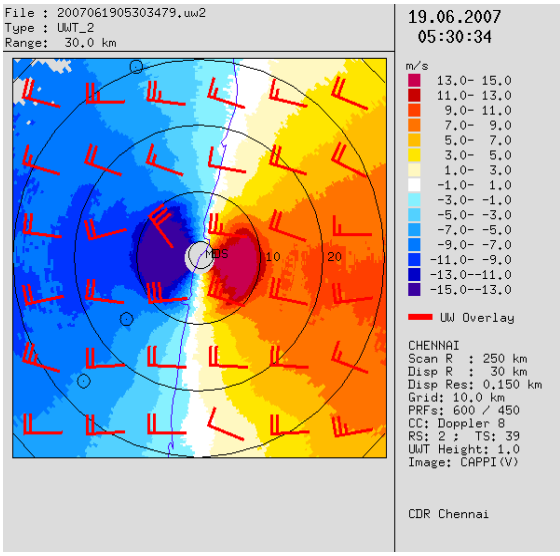


Fig. 8(ii).10: Horizontal wind on grid points – UWT algorithm

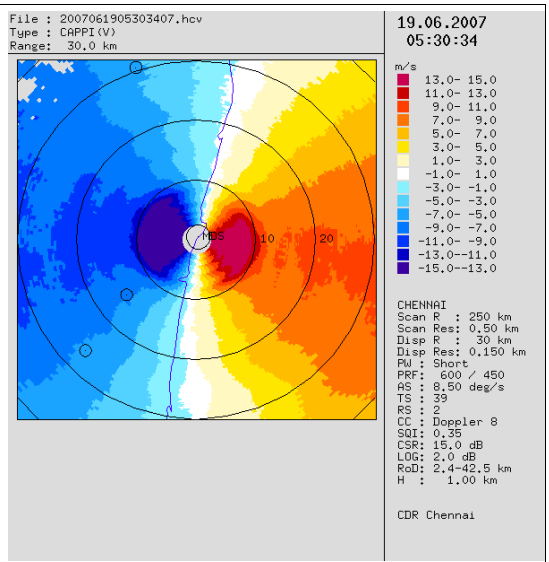


Fig. 8(ii).11: Radial wind on horizontal layer at 1km

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CHAPTER 8 (iii)

OBSERVATIONAL AIDS AND MONITORING OF MONSOON (iii) SYNOPTIC AND ASYNOPTIC

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8(iii).1. Introduction

The variability of monsoon performance has been well documented thanks to the observational aids. Floods in some areas and droughts in some other areas of India during the monsoon season has been witnessed almost every year and observational evidence to document these facts were coming only from synoptic observations recorded by the India Meteorological Department (IMD) in the early years. In view of vast difference in the co-efficient of variation of rainfall during southwest monsoon season between states and to some extent within a particular state as well, economy resulting from agricultural operations is affected. Hence the former Prime Minister of India, Pandit Jawaharlal Nehru had made a statement to the effect that the Indian economy is a gamble of monsoon.

8(iii).2. Synoptic observations

The surface and upper air weather observations are recorded by IMD, the country's apex weather agency, by maintaining a vast network of observatories. Table 8(iii).1 lists the observatories maintained by IMD for taking the main synoptic observations at 0000, 0600, 1200 and 1800 UTC and auxiliary synoptic observations at 0300, 0900, 1500 and 2100 UTC.

Table 8(iii).1: Network of IMD observatories

Type of Observatory	Nos.
Surface Observatories	559
Aviation Current Weather Observatories	71
High Wind Speed Recording Stations	4
Automatic Weather Stations	615
Automatic Rain gauge stations	456
Hydro-meteorological Observatories	701
Agro-meteorological Observatories	219
Evaporation Stations	222
Soil Moisture Recording Stations	49
Dew-fall Recording Stations	80
Evapo-transpiration Stations	39
Radiation Stations … Surface	45
Upper air (radiometer-sonde)	9
Air Pollution Observatories:	
- Background Pollution Observatories	10
- Urban Climatological Units	2
- Urban Climatological Observatories	13

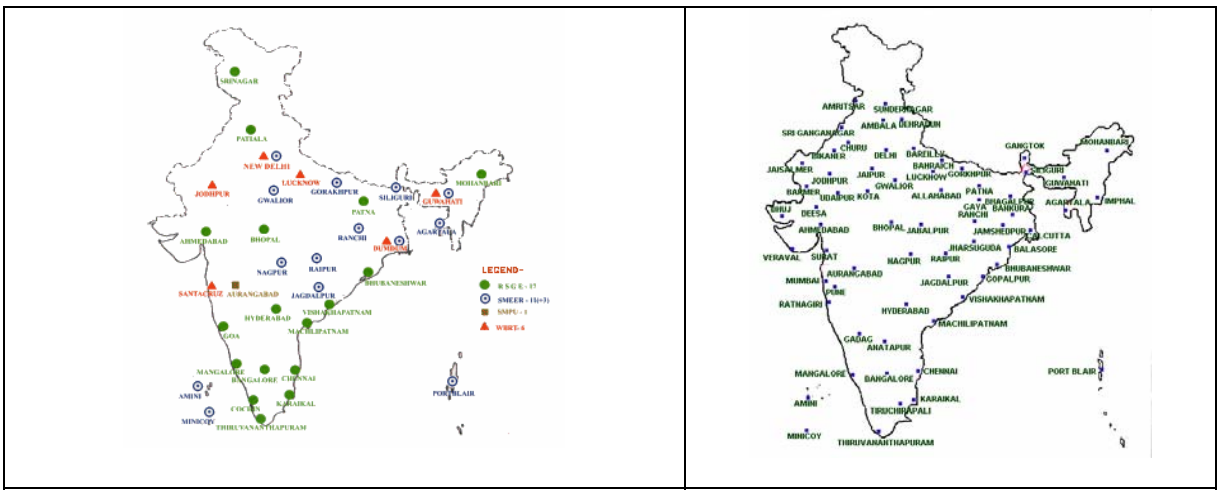
As part of modernization programme of meteorological instrumentations and in replacement of 100 INSAT based data collection platforms (DCP) functioning since late 1980s, 615 automatic weather stations (AWS) and 456 automatic rain gauges (ARG) have been installed as on 8th November 2011. It is expected that under phase I of IMD's modernization programme, there will be, in all 650 AWS by 31st December 2011 and a sizeable number of ARGs, augmenting to the current total of 456 ARGs, will be installed out of 1350 ARGs planned. In addition to the above, a number of non-departmental observatories, especially for measuring the amount of rainfall and snow, have been functioning in India to provide meteorological information (see Table 8(iii).2). The data received from AWS, ARG and radiation stations are uplinked to web site (http://www.imdaws.com) and freely made available in the public domain.

Table 8(iii).2: Details of non-departmental observatories

Non-Departmental Rain gauge Stations	
Reporting	3540
Non-Reporting	5039
Non-Departmental Glaciological Observatories (Non-reporting)	
Snow gauges	21
Ordinary Rain gauges	10
Seasonal Snow Poles	6

8(iii).3. Upper air observations

Upper air temperature, humidity and wind information are obtained at 0000 and 1200 UTC through Radio Sonde / Radio Wind techniques, GPS receivers and by means of radio-theodolites (38 upper air observatories are functioning in India). The upper wind information alone are obtained at main synoptic hours, viz., 0000, 0600, 1200 and 1800 UTC through 62 Pilot balloon observatories. Locations of RS/RW, GPS, Radio theodolite stations and Pilot balloon observatories have been shown in Fig. 8(iii).1.



(a) RS/RW, GPS, Radio theodolite network of India.	(b) Pilot balloon observatories of India.

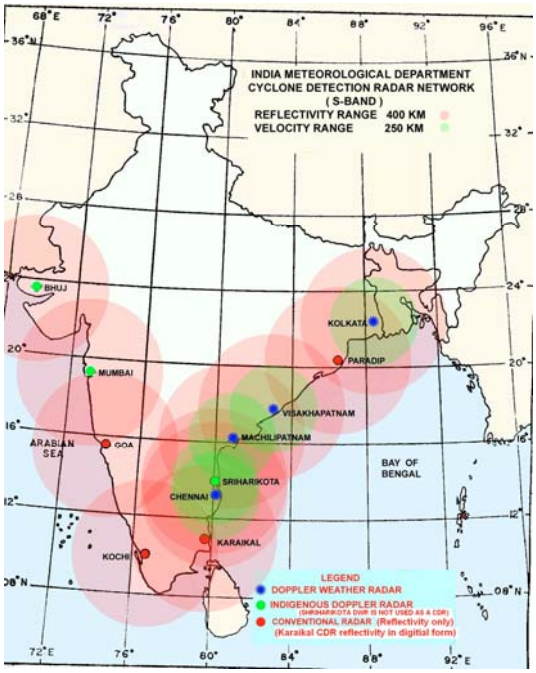
Fig. 8(iii).1: Locations of upper air observatories in India

8(iii).4. Asynoptic observations

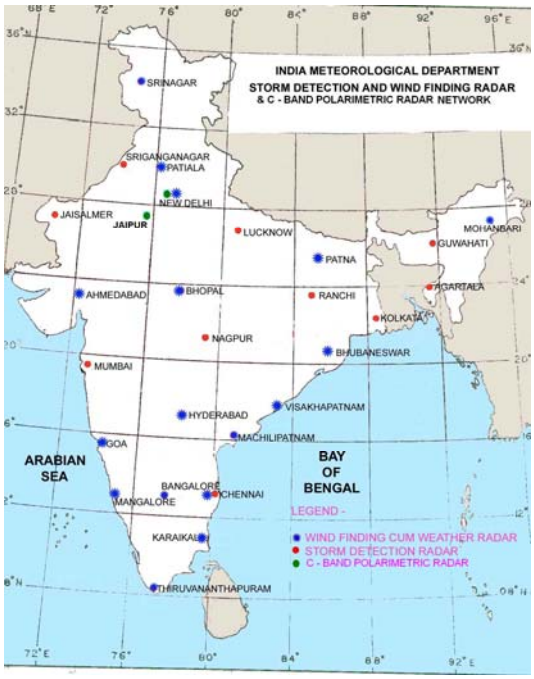
Observations recorded at non-synoptic hours are categorized as asynoptic observations. The asynoptic observations are taken through conventional and non-conventional meteorological instrumentations of both land, ocean (ship) and space based. Regular as well experimental modes of asynoptic observations have increased our understanding of monsoon dynamics. For example, asymmetric thermodynamic structure and the existence of high liquid water content over southwest quadrant of monsoon depressions have been brought out by analyzing the satellite data of asynoptic hours (Wang, 2006). The data recorded by Weather Radars, AWS and ARGs form a solid base for synoptic and asynoptic observations.

8(iii).5. Weather radars

IMD maintains *fourteen S-band Doppler Weather Radars* (DWR) out of which six are used as Cyclone Detection Radars (CDR) at Chennai, Macilipatnam, Visakhapatnam, Kolkata , Mumbai and Bhuj and the remaining eight DWRs are used for weather surveillance (see Fig. 8(iii).2). *Five more DWR*s have already been received at site and the site preparation work is going on. These five DWRs are likely to be commissioned 2012-2013 making the total as 19 S-band DWRs by 2012-2013. In addition, two *C-band Polarimetric DWR*s have been installed at Delhi and Jaipur thus making the total number of DWRs as 21 (16 existing + 5 being installed). For thunderstorm tracking, especially for aviation weather surveillance, 26 X-band radars (10 digital + 16 analogue) have been installed (see Fig. 8(iii).2).



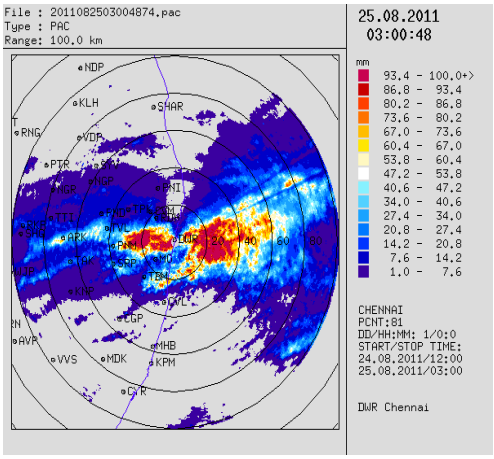
S – band *Cyclone Detection Radar (CDR)* network of IMD



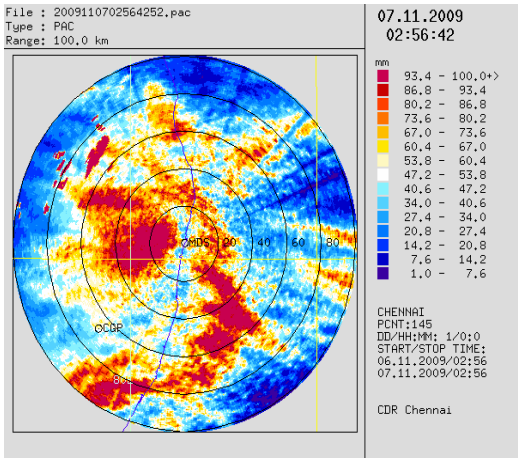
X-band thunderstorm tracking / upper wind finding Radar and C-band Polarimetric Radar network.

Fig. 8(iii).2: Cyclone detection, thunderstorm tracking and upper wind finding network of IMD

The digital X-band radars has both storm tracking and wind finding capabilities. However, these digital X-band radars are being phased out in view of ageing and their obsoleteness (spares not available readily). While analogue radars take observations at synoptic hours, normally, the digital and Doppler Weather Radars (DWR) are operated round the clock and hence provide data continuously even during asynoptic hours. The atmospheric data probed by the DWRs are collected at IMD, New Delhi in near real time basis through virtual private network (VPN) connectivity and these data are used for ingesting into high resolution NWP models. Under IMD's modernization programme Phase II and Phase III, installation of 34 DWRs have been planned to replace all the existing 26 X-band storm detection / Multimet radars and to install at 8 new locations additionally.



Precipitation accumulation from 1200 UTC/ 24.8.2011 to 0300UTC/25.8.2011



Precipitation accumulation from 0256 UTC/7.11.2009 to 0300UTC/8.11.2009.

Fig. 8(iii).3: Precipitation accumulation for a user defined time interval over 100 km radius around Chennai DWR during a typical SW monsoon season (1200UTC/ 24.8.2011 to 0300UTC/ 25.8.2011) and a typical NE monsoon season(0256 UTC / 7.11.2009 to 0256 UTC/ 8.11.2009).

The digital and DWR data, when calibrated with ground truth and after establishing the radar reflectivity (*z*) – rain rate (*R*) relationship, are useful to estimate the instantaneous rain rate and the amount of accrued rainfall over a any period of time. For example, the precipitation rate can be integrated for 24 hrs as well similar to the conventional rain gauge data (from 0300 UTC to 0300 UTC of the next day). Rain rate is estimated in operational set up for all scan volumes using the $z - R$ relationship, $z = 267\ R^{1.345}$ (Suresh et al., 2005) and the time integration for 24 hrs, similar to conventional practice, of 0300 UTC ending the next day is carried out from Chennai DWR and the accumulated precipitation is compared / validated with the available ground truth values. An 85% efficiency has been attained. Precipitation accumulation from 1200 UTC / 24th August 2011 to 0300 UTC / 25th August 2011 over an areal extent of 100 km radius around Chennai DWR and from 0256 UTC / 6th November 2009 to 0256 UTC / 7th November 2009 have been shown in Fig. 8(iii).3 as a sample to give a flavour of the accumulated precipitation over a vast area in which no rain gauge is available.

8(iii).6. Radiation observations

Upper air infra red (IR) radiation flux measurements are taken at fortnightly interval from Bhubaneswar, Delhi, Kolkata, Jodhpur, Nagpur, Pune, Srinagar and Thiruvananthapuram. Upper air radiation flux measurements from these stations and radiation observations recorded at surface from 45 stations help meteorologists to understand the monsoon behaviour and to build up monsoon climatology. IMD's radiation network is shown in Fig. 8(iii).4.

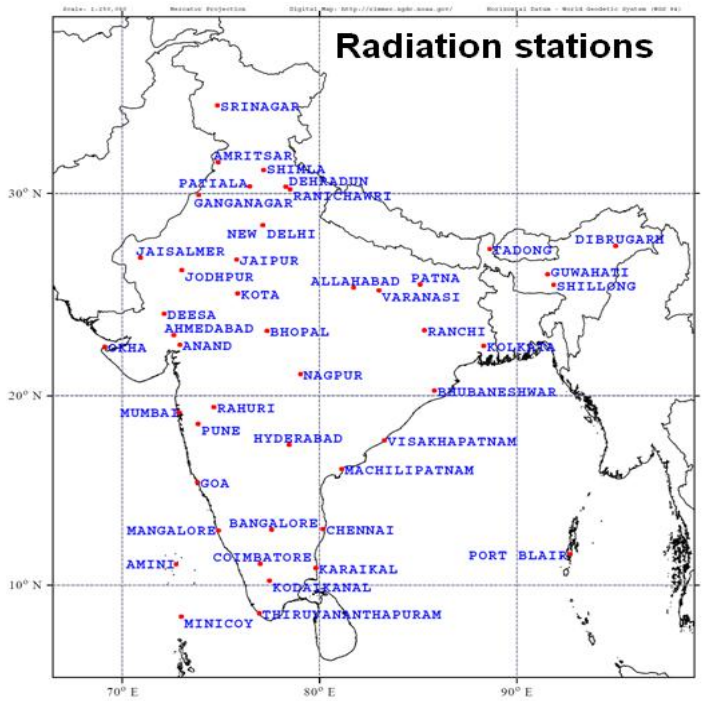


Fig. 8(iii).4: Radiation station network of IMD.

8(iii).7. Ship observations

As many as 203 voluntary observing fleet (VOF) of merchant navy ships have been supplied with meteorological instrumentations by IMD. Voluntary observation ships take both synoptic and asynoptic observations. These valuable observations are passed on through coastal radio to the nearest storm warning centres and the hard copy form of observations are handed over to the port meteorological office of IMD when the ship comes to a port. The ship data are very useful in analyzing monsoon flow over oceanic region.

8(iii).8. Satellite observations

Satellite data from imagers and sounders play a key role in identifying the monsoon characteristics and dynamics (Kelkar, 2007). Satellite imagery data are helpful to identify the onset and advance phases of monsoon system and to reveal the active, weak spells of rainfall. The sea surface temperature as estimated from satellite data has been proved to be a good tool to know about the El-Nino and El-Nino southern oscillation (ENSO) and the atmospheric component of Indian Ocean Dipole, viz., Equatorial Indian Ocean Oscillation (EQUINOO) which are having a bearing on the performance of Indian summer monsoon (Gadgil, 2003; Gadgil et al, 2002; Gadgil et al, 2003; Wang, 2006).

8(iii).9. Experiments

Though the monsoon clouds and rain have been documented in historical literature for centuries, a campaign on intensive, systematic meteorological observations over land and ocean conducted by IMD as early as during 1893-1894 made it possible to scientifically document the onset of monsoon and a clue on cross equatorial flow (Eliot, 1896 and Bhat and Narasimha, 2007). This early scientific finding emphasizes the necessity of continuous observations to understand the monsoon dynamics. However, presumably due to political reasons, no systematic observational campaign had been launched for more than 65 years, till the International Indian Ocean expedition (IIOE) was organized during 1960-1965. Rao (1976) gives an excellent review of results obtained from IIOE.

With the participation of Scientists from twenty countries deploying 40 ships to take oceanic observations and a number of aircrafts to take upper air observations through drop sondes over Indian ocean and adjoining land areas during 1960-1965, interesting results such as cross equatorial flow, low level jet stream, low level inversion over central Arabian sea were discovered (Sikka, 2005; Joseph and Raman, 1966; Findlater, 1969). Based on new impetus from the interesting results on the fascinating monsoon weather phenomenon, the meteorologists throughout the world were actively involved in research and undertaken the following experiments over the monsoon region.

- ISMEX - Indian Summer Monsoon Experiment (1973)
- MONSOON - Indo-Soviet monsoon experiment (1977)
- MONEX - Monsoon experiment (1979).

By deploying a number of ships to have synoptic and asynoptic sea level observations and a number of aircrafts to have drop sonde measurements, a huge data base was created in conjunction with routine synoptic observations recorded by IMD during the campaign period. The combined effort of ISMEX, MONSOON and MONEX experiments had brought out the existence and characteristics of monsoon onset vortex (Krishnamurti et al, 1981) besides the influence of western ghat in augmenting monsoon rainfall over west coast of India (Sarkar, 1966, Grossman and Durran, 1984; Mukherjee et al, 1984).

8(iii).10. MONEX

Two aircrafts (WP-3D of NOAA and Electra L-188 of NCAR) obtained drop sonde data over a path length of 3000 km on 7th July 1979 over a monsoon depression area in the Bay of Bengal and took photographs. 17 drop wind sondes were released. Sounding data from polar orbiters (DMSP F3, TOVS from TIROS-N) were collected for the monsoon depression during the MONEX campaign. Convergence of the order of $-2 * 10^{-4}$ s^{-1} (approx. – 1 Pa s^{-1}) at 945 hPa in south and western sector and divergence of similar magnitude towards northeast of the depression have been observed. Subsidence of magnitude 0.1 Pa s^{-1} was observed in southeast sector of the depression at 694 hPa (Warner, 1984). Predominance of stratiform rain extending 100 to 300 km from a monsoon depression (3-8 July 1979) during MONEX, cloud liquid water content (a maximum 0.1 g m^{-3}), virtual absence of cloud liquid water above 0^0 C, glaciations of clouds above 0^0 C and arc shape convective cloud alignment have been documented using aircraft measurements (Houze and Churchill, 1987).

The need for taking measurements over boundary layer has been emphasized by MONEX. Although in all the aforesaid experiments India had participated as a participating country, based on knowledge gained and active

participation by Indian scientists, the following experiments were conducted indigenously.

- MONTBLEX - Monsoon Trough Boundary Layer Experiment (1990) – [Goel and Srivastava (1990), Sikka and Narasimha (1995)]
- LASPEX - Land Surface Processes Experiment (1997–98) – [Sastry et al (2001)]
- BOBMEX - Bay of Bengal Monsoon Experiment (1999) –[Bhat et al, (2001)]
- ARMEX - Arabian Sea Monsoon Experiment (2002–2005) – [Rao (2005)]

In addition to the above experiments conceived and carried out by Indian Scientists, an international experiment, viz., 'Joint Air–Sea Monsoon Interaction Experiment' (JASMINE) was carried out in the year 1999 to understand the pre-monsoon characteristics in two spells (7–22 April and 1 May–8 June) and the monsoon dynamics during the cessation and withdrawal phase of monsoon (2–28 September) over the Indian ocean (Webster et al, 2002).

8(iii).11. MONTBLEX

An indigenous multi-institutional MONTBLEX experiment had been conducted to get more data in the boundary layer of monsoon trough region during June – September 1990 with 14 intense observation periods. The experiment had been designed to cover both deep moist convective regions with 100 to 200 cm annual rainfall (85 – 95 ^{o}E) and dry convective regions with annual rainfall of 20 to 60 cm (70 – 75 ^{o}E). In addition, the experiment obtained data over a mixed region (deep moist and unsaturated) having an annual rainfall of 60 to 100 cm (75 – 85 ^{o}E). The MONTBLEX experiment was conducted from Jodhpur on western side to Kharagpur on eastern side of India covering a span of about 2000 km with transition zones at Delhi and Varanasi during 1990. Measurements / observations, both synoptic and asynoptic, were taken from conventional surface and upper air observatories of IMD. Two X-band Radars of IMD at Kolkata and Ranchi provided continuous observations. The imageries were obtained from INSAT on three hourly basis routinely. In addition to the above, data were obtained from the following instrumentations specially installed for this experiment.

- Surface micro-meteorological towers (30 m) at Jodhpur, Delhi, Varanasi and Kharagpur
- SODARs at Jodhpur, Delhi, Varanasi and Kharagpur
- Ship mast
- Tether sonde and aircraft measurements.

Applicability of Monin – Obukhov similarity theory and O'Brein scheme, estimation of heat and momentum fluxes and their variability during various phases of monsoon, variation of fluxes during passage of monsoon depression, applicability of various Numerical weather prediction models with different types of processes (thermodynamical, moist), vertical wind shear in the mixed layer and frontal behaviour etc have been documented by utilising both synoptic and asynoptic observations (Mohanty et al, 1992; Vishwanadham and Satyanarayana, 1992; Sivaramakrishnan et al, 1992; Rao et al, 1992; Singal and Gera, 1992, Gera et al, 1996; Vernekar et al, 1992; Rao and Narasimha, 2006).

8(iii).12. Indian Climate Research Programme (ICRP)

In order to understand the monsoon variability on timescales ranging from sub-seasonal to inter-annual / decadal scales, to further understand the coupled ocean – atmosphere – land system processes and to investigate the climate variability links with agriculture, ICRP Science Plan has been mooted by the Department of Science and Technology, Government of India as a multi-institution research project during 1996. Two major experiments were conducted based on ICRP Science Plan, viz., Bay of Bengal Monsoon experiment (BOBMEX) during 1999 and Arabian Sea Monsoon experiment (ARMEX) during 2002-2003 (Bhat et al, 2001; Sikka, 2005; Rao, 2005; Rao and Sikka, 2005).

A number of ocean buoys from the National Institute of Ocean Technology (NIOT), Chennai, a ship named *ORV Sagar Kanya* from the Department of Ocean Development (DOD), and another ship named *INS Sagar Dhwani* of the National Physical Oceanographic Laboratory (NPOL), Kochi and two ships from Indian Navy were deployed for the extensive observations with the support facilities from Indian

Navy, Coast Guard and Indian Air Force. Special observations from coastal and island stations were taken from IMD network.

Results on rapid increase of sea surface temperature (SST) during March – April well above the convection threshold of 28°C and maintaining the SST at more than 30°C during May but without organized convection have been brought out by these experiments. Building up of mini-warm pool over south east Arabian Sea and the maintenance of the same have been excellently brought out by Vinayachandran et al (2007).

8(iii).13. Influence of western ghat in rainfall variability and precipitation enhancement

As the Western Ghats form a major barrier to the south-west monsoonal current on the west coast of India, the influence of western ghat over the SW monsoon rainfall was studied in-depth by many researchers. It has been documented that the Western Ghats enhance rainfall activity substantially under favourable conditions, with increasing trend on the windward and decreasing pattern on the leeward side. Sarkar (1966, 1967) has made extensive studies on dynamical lifting of neutrally stratified and moist air in the westerlies over the crest of Western Ghat and the associated rainfall. The maximum rainfall appears to occur in the high altitudinal zones, prior to the crest of the mountain (Sarkar 1967; Patwardhan and Asnani 2000) and on the windward side ahead of the geographical peak (Venkatesh and Jose, 2007). However, the dynamic model has limitations in explaining the rainfall over the leeside of the mountain, presumably due to non-availability of high resolution ground truth rainfall data for comparison until now. With the introduction of more and more DWRs, it is hoped that the variability of rainfall, the influence of western ghat on rainfall especially on the lee side of the western ghat can be better studied in the coming years.

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CHAPTER 9

EXTREME WEATHER EVENTS - FLOODS, DROUGHTS, HEAVY RAINS

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9.1. Introduction

Floods and droughts are the most fatal natural disaster in developing countries. Due to the high spatial variability of rainfall in spite of good performance of southwest monsoon over the country as a whole in a particular year, smaller regions may suffer drought conditions. Similarly in spite of poor performance of southwest monsoon over the country as a whole in a particular year, smaller regions may get floods. The developing countries fall victim to floods every year causing several deaths and economic losses. Damages from floods take several forms, including the destruction of footbridges that often provide the only link between remote villages, demolition of irrigation diversions, mass wasting by undercutting of steep, stream-adjacent slopes and damage to flood plain agricultural land by erosion and sedimentation. Heavy rain and floods also cause the occurrence of landslides over mountain regions. Most of the people affected by floods in one year do not get time to recover their losses before experiencing another year of flood though the country as a whole may receive normal or even deficient monsoon rainfall in those years.

Different communities use the word flood in different ways. It is a very complex phenomenon and has various effects on common people, scientists, meteorologists, agriculturists, economics and government.

According to the meteorological point of view, a flood is defined as a quantum of rainfall/precipitation received over an area in excess of long period average

precipitation over that area during the period when it is expected to occur. Generally if the precipitation for a particular period exceeds the long period average precipitation (referred as normal) by one standard deviation, meteorological floods occur (Sikka, 2000; WMO, 1994). Bhalme and Mooley (1980,1981) presented a Flood Area Index (FAI) based on the data of monsoon rainfall for the period 1891-1979 as the areal percentage of India for which the mean monsoon rainfall departure exceeded by 2. Choudhury and Mhasawade(1991), using criteria of monsoon rainfall departure 25% or higher from the mean seasonal rainfall, have identified areas of all sub-divisions receiving higher than 25% rainfall and obtained an index defined as

$$MAI = \frac{A - \overline{A}}{S.D.}$$

where A is the sum of the area of meteorological subdivisions receiving greater than 25% of the normal rainfall for the, $\overline{A}$ is the corresponding mean value and S.D. is the standard deviation. Valuable information about the different types of extreme weather events, their occurrence during the last 100 years is available in De et al.(2005). Floods and droughts over India are the two aspects of the weather associated with the abundance or deficit of monsoon rains. A large number of studies are available on various aspects of floods and droughts. A study by Chowdhury et al. (1989) have ranked the year 1918 as the worst drought year of the last century, a year when about 68.7% of the total area of the country was affected by drought. It is of interest to note that the year 1917 had exceptionally high seasonal rainfall. Likewise the severe drought years of 1877 and 1987 were followed by flood years of 1878 and 1988. In the 19th century the droughts of 1877 and 1899 followed by the early droughts of the twentieth century. In the last century the drought of 1987 and 1972 are the next in order of severity. Occurrences in drought of consecutive years have been reported in 1904-05, 1965-66. These pairs of years were associated with moderate droughts, where at least 25% of the country was affected. During 1999, 2000 and 2001 drought conditions prevailed over some parts of India, not affecting the country as a whole significantly. During 2002, twelve out of 36 subdivisions of the country came under the grip of moderate to severe drought when about 29% of the total area of the country was affected by drought. The

seasonal rainfall departure (%) for west Rajasthan and east Rajasthan were -69% and -55% respectively. The seasonal rainfall during the summer monsoon in the country as a whole was 19 percent below normal qualifying 2002 as the first all-India drought since 1987. Rainfall deficits during July were most noteworthy, at a historical low of 51 per cent below normal. Remarkable recovery in rainfall occurred in August, which prevented the situation becoming worse (WMO, 2003). However, the El-Nino episode that developed during 2002 was significantly smaller than 1997 /1998 event. Of all the major natural disasters, droughts account for nearly 22% of significant damages though the number of deaths is only 3% world wide (De & Joshi 1998). Sinha Ray & Shewale (2001) studied the probability of drought on sub-divisional scale. According to their study frequency of droughts was generally high over western and central India and northern peninsula. In spite of normal monsoon years many parts of our country were severely affected by drought. There was a spell of consecutive 14 years from 1988 -2001 when the country experienced normal monsoon. Guhathakurta (2003) has computed probability of occurrence of drought for each of the districts of India during the period 1988-2001 and has shown that the probability was high for Rai Barielly district of east UP and Sangrur of Punjab. The occurrence of droughts during the summer monsoon season in the nineteenth century led to the attempts by Blanford, the then Chief of the Indian Meteorological Service, in seasonal forecasting. The first long range forecast issued by Blanford in 1886 was based on the inverse relationship between the Himalayan snowfall during early spring and the subsequent Indian Summer Monsoon rainfall during June to September. Repeated failure of monsoon during the beginning of the twentieth century led to introduction of multiple regression techniques in 1906, by Sir Gilbert Walker for long range forecasting of monsoon rainfall.

But actually flood occurs not in a seasonal scale but in most cases weekly or further smaller time scale. Considering the day to day and the extensive damages caused by the flood, hydrologists define flood as a high river flow which overtakes the natural runoff, or a high flow which overtakes the natural or artificial embankments of a river or a stream, can excessively high water level in a river or discharge level above an threshold level known as flood level or danger level which are already decided for all the rivers considering several aspects (WMO, 1994). In India, Central Water Commission (CWC) is the nodal agency for water resources

and CWC has fixed Danger Level or all important river catchment or discharge sites. Dhar and Nandargi (2000, 1998,1989) have studied rainfall features of different river basins of India and also different flood characteristics.

Critical studies of floods, their past occurrence, and monthly, annual and spatial variability are therefore required for better planning of flood management and for detecting the flood prone areas. India Meteorological Department brings out annual disaster reports for twelve extreme weather events and these reports which are available from 1967 are very valuable information for disaster mitigation and management planners.

9. 2. Causes of floods

Excessive precipitation is the main cause of flood. Synoptic situations like monsoon depressions, cyclonic storms, low pressure areas, position of monsoon trough over north Indian regions, movement of westerly trough play important roles in causing excessive precipitations. Main components responsible for the flood are as follows:

(i) Flash floods: In a short duration heavy to very heavy rainfall in a smaller catchment can create flash floods. Such floods develop so rapidly that there is little time to be left for the flood occupants to lean the affective area.

(ii) Direct runoff is the major cause of floods. Rain reaching the soil surface can infiltrate into the soil or runoff directly in streams and rivers. Once the absorbed rainfall exceeds the infiltration capacity, the excess water flows as a direct run off. During the course of monsoon season with continuous excessive rain for a few days, the soil becomes fully saturated and additional rain would then form direct run off.

(iii) Deforestation: It is a major influence in developing countries that increases run off and in turn causes flood. The trees hold the soils of the forest flood together and make a deep litter of fallen leaves. Both of these cause increase in infiltration capacity. Due to the deforestation, soils are soon eroded resulting in increase in runoff of sediment – laden water. The resulting erosion of hillsides is very deleterious and this also leads to increase sediment loads in rivers and sitting up of reservoirs

downstream. Overgrazing cause loss of grass covers on hillsides, which in turn increases faster runoff and erosion.

(iv) Sediments from the erosion settle at the river bottom, decrease the depth of river and gradually raise water level.
(v) Due to the increased urbanization, ground absorption of water is prevented and this leads to increase in runoff and contributes to flash floods in urban areas.
(vi) Poor farming techniques also increase soil erosion.
(vii) Flood Plains: Due to less and inexpensive land values and high soil fertility, proximity of water, availability of construction materials such as sand and gravel and flatness of land, flood, plains are attracted by poor urban dwellers and farmers, industrialists, etc. During the flash flood, these areas are affected causing severe damages and loss of life.
(viii) Storm Surges: Storm surges are the major causes of flood in coastal areas and in river estuaries. A cyclonic storm (with intense low pressure in the center) over the sea causes the rise of sea level due to barometric effects and also the strong winds associated with the cyclonic storm if directed to the onshore, drives the sea to the land. Coastal areas like Orissa and Andhra Pradesh are commonly affected by the storm surges. In the Super Cyclone of 29th October, 1999, strong winds of speed 250 kmph lashed most parts of Orissa coast and storm surge of height 12 – 14 meter inundated low lying areas along Orissa coast causing deaths of nearly 10,000 human lives.

Government of India has launched ‘National Flood Control Programme’ in 1954. Rashtriya Barh Ayog (RBA) constituted by the Government of India in 1976 carried out an extensive analysis to estimate the flood-affected area in the country. RBA in its report (1980) has assessed the area liable to floods as 40 million hectares. It was determined by summing up the maximum area affected by floods in any one year in each state during the period from 1953 to 1978 for which data was analyzed by the Ayog. This sum has been corrected for the area that was provided with protection at that time and for the protected area that got affected due to failure of protection works during the period under analysis to arrive at the total area liable to floods in the country as per break-up given below: -

State	Area liable to Floods (million Ha.)
1. Andhra Pradesh	1.39
2. Assam	3.15
3. Bihar	4.26
4. Gujarat	1.39
5. Haryana	2.35
6. Himachal Pradesh	0.23
7. Jammu & Kashmir	0.08
8. Karnataka	0.02
9.Kerala	0.87
10. Madhya Pradesh	0.26
11. Maharashtra	0.23
12. Manipur	0.08
13. Meghalaya	0.02
14. Orissa	1.40
15. Punjab	3.70
16. Rajasthan	3.26
17. Tamil Nadu	0.45
18. Tripura	0.33
19. Uttar Pradesh	7.336
20. West Bengal	2.65
21. Delhi	0.05
22. Pondichery	0.01
Total	**33.516**

Dams in India and any where else are made for different purposes including power generation, irrigation, flood prevention, land reclamation, and water diversion. Given the multiple objectives of dams, dam management and food control involves different stakeholders with different interests and responsibilities, which makes the flood control related decision making a complex process.

Flood forecasting activities in India made a small beginning in November 1958 when the Central Water Commission created a Flood Forecasting Unit for flood forecasting for the river Yamuna at Delhi, the National Capital. The disastrous floods in 1968 in many parts of the country necessitated the setting up of forecasting centers on interstate rivers.

9.3. Types of drought

Several definitions of drought are available in literature. In India, National Commission on Agriculture (1976) has categorized drought into three types, viz., meteorological drought, hydrological drought and agricultural drought based on the concept of its utilization. The National Oceanic and Atmospheric Administration (NOAA) defines agricultural drought as a combination of temperature and precipitation over a period of several months leading to substantial reduction (less than 90%) in yield. National Commission on Agriculture (1976) classified drought as an occasion when the rainfall for a week is half of the normal or less, when the normal weekly rainfall is above 5 mm or more. If four such consecutive weeks occur from middle of May to October, it is considered as agricultural drought. From agriculture perspective, drought is a condition, in which, the amount of water needed for transpiration and direct evaporation exceeds the amount available in the soil.

In meteorological terms, a drought is "a sustained, regionally extensive, deficiency in precipitation". All other definitions are related to the effect or impact of below normal precipitation on water resources, agriculture and social and economic activities; hence the terms hydrological drought and agricultural drought. In quantitative terms, the definitions could vary among countries and regions. In India, the definition for "meteorological drought" adopted by the Indian Meteorological Department (IMD) is a situation when the deficiency of rainfall at a meteorological sub-division level more than 25 per cent of the long-term average (LTA) of that sub-division for a given period. The drought is considered "moderate", if the deficiency is between 26 and 50 per cent, and severe if it is more than 50 per cent. Based on this definition, the National Commission on Agriculture has given the following broad classifications:

Hydrological drought occurs due to prolonged meteorological drought resulting in depletion of surface water from reservoirs, lakes, streams, rivers, cessation of spring flow and fall in groundwater levels causing severe shortage of water for livestock and human needs.

Agricultural drought happens when soil moisture and rainfall are inadequate during the crop growing season to support healthy crop growth to maturity, which causes extreme crop stress and wilting. It is defined as a period of four consecutive weeks (of severe meteorological drought) with a rainfall deficiency of more than 50 per cent of the LTA or with a weekly rainfall of 5 cm or less during the period from mid-May to mid- October (the *Kharif* season) when 80 per cent of the country's total crop is planted, or six such consecutive weeks during the rest of the year.

Drought differs from other natural hazards in many respects -most complex and least understood of all disasters. While it is difficult to demarcate the onset and end of drought but the effects of drought accumulate for a considerable period of time. Prolonged droughts or abnormal weather conditions such as extended winters, cold summers, floods, biological factors like plague of locusts or rodents result in famines. On an average, severe drought occurs once every five years in most of the tropical countries, though often they occur on successive years causing misery to human life and live stock. The crisis brought out by this hazard directly hit poorest and most deprived sections of our society, thus destroy the life, economy, infrastructure, environment and society because all are inter linked.

Period	**Drought years**	**No. of years**
1801-25	1801,04,06,12,19,25	6
1826-50	1832,33,37	3
1851-75	1853,60,62,66,68,73	6
1876-1900	1877,91*, 99	3
1901-25	1901,04,05,07*,11, 13*,15*,18,20,25*	10
1926-50	1939*,41	2
1951-75	1951,65,66,68,72,74	6
1976-2000	1979,82,85*, 87	4
2001-2008	2002,2004	2

The above Table shows the all India drought years from 1801 to 2008. The years with * indicates that the total drought affected area was more than 20% but all India monsoon rainfall departure was not less than -10%. It may be mentioned that here while calculating drought affected area smallest unit is considered as meteorological sub-divisions. The values may change if district is considered as the smallest geographical unit.

Shewale and Kumar (2005) in their studies have prepared drought climatology of India using 130 years of data (1875-2004). The time series of drought areas of the country made by them were based on the sub-division areas only. But the smallest geographical unit i.e. districts were not considered for identifying the drought areas. Realizing the need for district-wise rainfall monitoring a District-wise Rainfall Monitoring Scheme (DRMS) has been introduced by the India Meteorological Department collaborating with different state governments. This helps to monitor rainfall situation (excess, normal or deficient or drought affected) for each of the districts, states, sub-divisions as well as country as a whole in real time basis. A better analysis and thus monitoring of drought situation is made possible after implementation of DRMS and thus helps different agencies for better disaster management. Both floods and drought are mainly localized in nature, not often the country as a whole. Excessive or less rainfall is the major cause of these extreme events, the information of rainfall pattern of India is very much important. The severe rainstorm that is associated with the occurrence of extreme rainfall is also the subject of study.

9.4. Rainfall Features of India.

Fig. 9.1 shows the distribution of the mean south-west monsoon rainfall of the country. South west monsoon rainfall constitutes the major rain producing season as, on an average 76% of annual rainfall is received in this season. However, there is large spatial variation of monsoon rainfall over the country as the percentage of annual rainfall received varied from lower value of 35 % over the sub-division Tamilnadu & Pondicherry to the maximum value of 95% over the sub-division Gujarat Region (Fig.9.2).

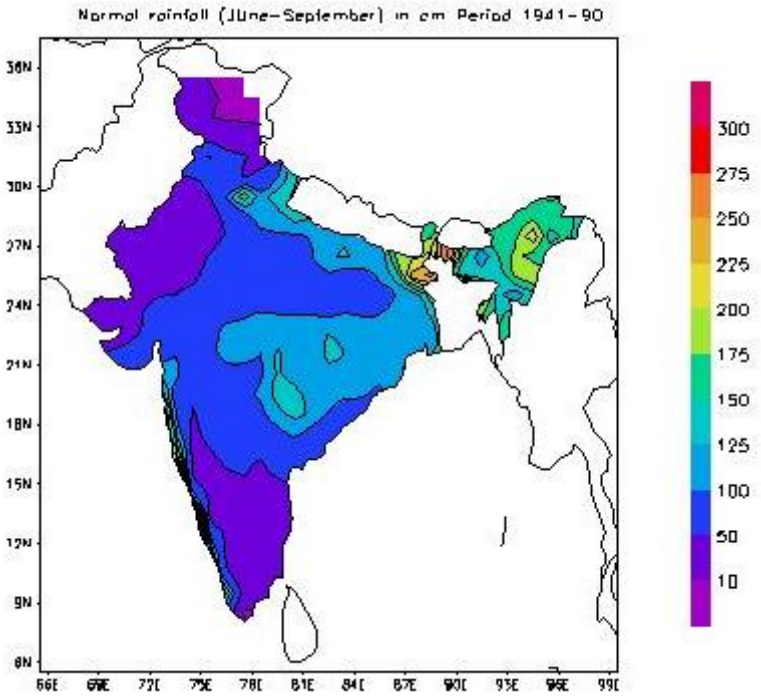


Fig. 9.1: Distribution of the mean south-west monsoon rainfall of the country

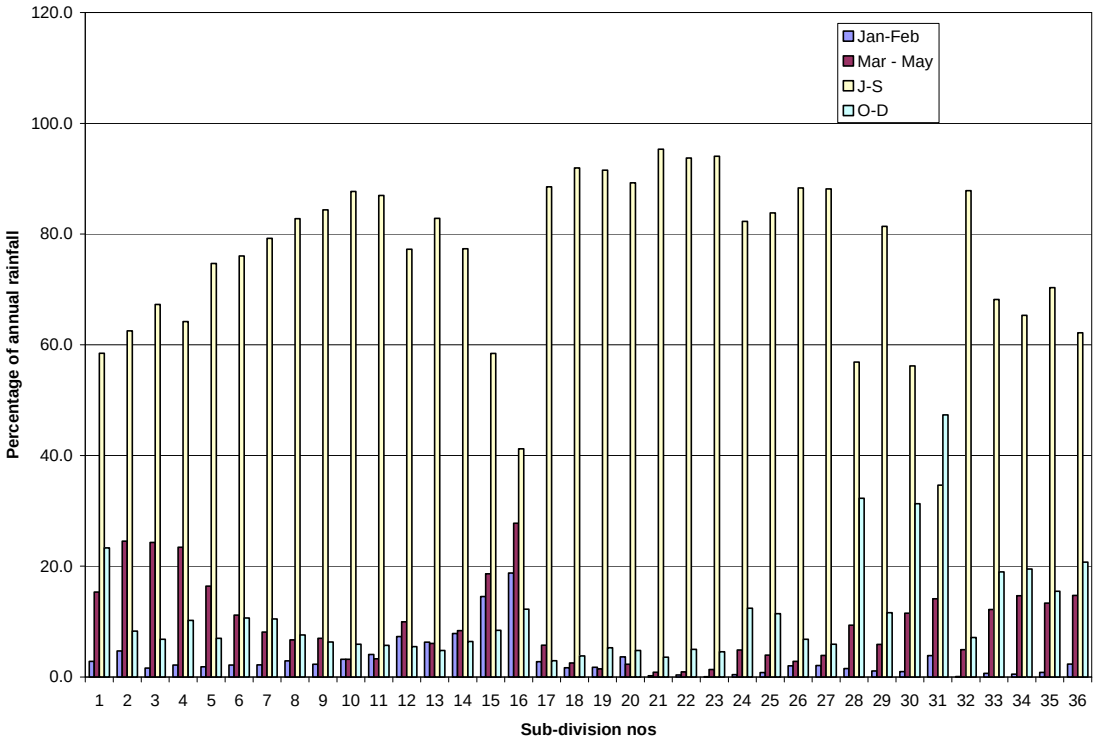


Fig. 9.2: Rainfall as percentage of annual over the 36 met- subdivisions and for four seasons

9.5. Extreme Rainfall Events

In recent years, heavy precipitation events have resulted in several damaging floods in India. The consecutive flash floods over three major metro cities in the same year i.e. Mumbai in July 2005, Chennai in October 2005 and again in December 2005 and Bangalore in October 2005 caused heavy damages to properties, loss of life etc. The information on the changes on extremes weather events is more important than the changes in mean pattern for better disaster management and mitigation. According to the latest report of Intergovernmental Panel on Climate Change (IPCC, 2007) another aspect of these projected changes is that "*wet extremes are projected to become more severe in many areas where mean precipitation is expected to increase, and dry extremes are projected to become more severe in areas where mean precipitation is projected to decrease. In the Asian monsoon region and other tropical areas there will be more flooding".* Karl et al. (1999) assessed the changes in climate extreme over many parts of the world during the past century.

The study of spatial variability of extreme rainfall events helps to identify the zone of high and low value of ever extreme rainfall events. A detailed regionalized study is practically useful for the planners and other users. Earlier Rakhecha and Pisharoty (1996) have studied the heavy rainfall events during the southwest monsoon season for some selected stations over the country. Rakhecha and Soman, (1994) analyzed the annual extreme rainfall series in the time scale of 1 to 3 days duration at 316 stations, well distributed over the Indian region, covering 80-years of rainfall data from 1901 to 1980 were analyzed for trend and persistence using standard statistical tests. They had reported that the annual extreme rainfall records of most stations are free from trend and persistence. However, the extreme rainfall series at stations over the west coast north of 12°N and at some stations to the east of the Western Ghats over the central parts of the Peninsula showed a significant increasing trend at 95% level of confidence. Stations over the southern Peninsula and over the lower Ganga valley have been found to exhibit a decreasing trend at the same level of significance. Stephenson et al (1999) using the data for the period June to September 1986–89 have investigated extreme daily rainfall events and their impact on ensemble forecasts of the Indian Monsoon. Most of the

studies on extreme rainfall over India (Sen Roy & Balling, 2004; Rakhecha and Soman, 1994) used limited number of stations. However, their results are very useful for studies and management of disaster. From the annual frequency of rainy days, rain days and heavy rainfall days, average or normal rainy days, rain days and heavy rainfall days were computed. High spatial variations of climatology of frequency of rainy days, rain days and heavy rainfall days can be seen in Fig. 9.3. Annual normal rainy days vary from the low value of 10 over extreme western parts of Rajasthan to the high frequency of 130 days over northeastern parts of the country. The northeastern parts of the country as well as sub- Himalayan West Bengal and also extreme western coast line of the country received on an average more than 100 rainy days in a year.

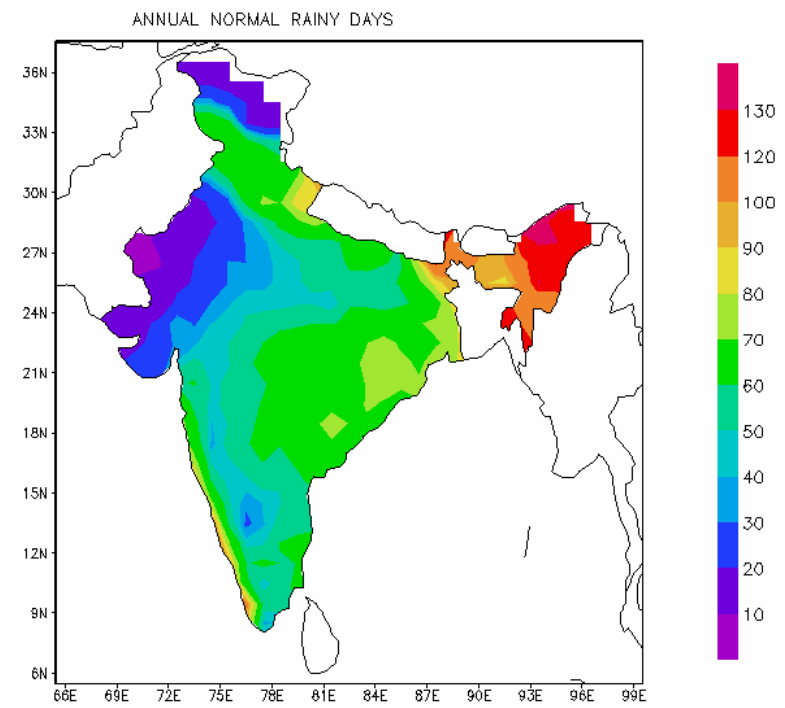
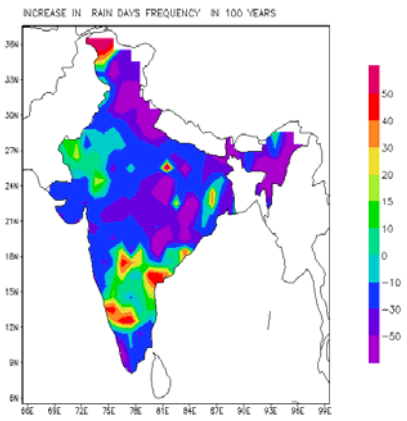


Fig. 9.3: Climatology of frequency of rainy days

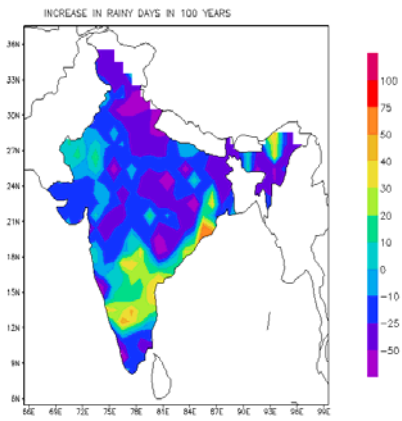
An important feature during the southwest monsoon season of India is the occurrence of heavy rainfall (rainfall amount $\geq$ 6.5cm) associated with certain meteorological situations all over the country. Persistent copious rains through the season are associated with orographic lifting of moisture-laden winds, continuous rain for period of days or short period heavy falls due to intense thunderstorms or cloud burst.

Annual frequency series of raindays (considering the frequency of days for any amount of realized rain), rainy days (rainfall amount ≥ 2.5mm) and heavy rainfall days (rainfall amount ≥ 6.5cm) were analyzed to see the significant changes both temporal and spatial using daily rainfall data of 2599 stations having good data availability for the period 1901-2006. Fig.9.4 a, b and c show the increase/decrease (significant) in frequency of rain days, rainy days and heavy rainfall days in 100 years respectively. Significant increasing trends in rain days are being noticed over most of the parts of Andhra Pradesh and Karnataka, major parts of Rajasthan, some parts of Marathwada and Jammu and Kashmir, Jharkhand. Frequency of rainy days is increasing over most of the parts of Andhra Pradesh, Karnataka, and Orissa, major parts of Rajasthan. Increasing activities of the occurrence of heavy rainfall are noticed over most parts of south peninsula except Kerala, north east India, West Bengal, Sikkim, Bihar, Jharkhand, Orissa, Rajasthan, Gujarat, Madhya Maharashtra, Konkan & Goa and Marathwada.

a)



b)



c)

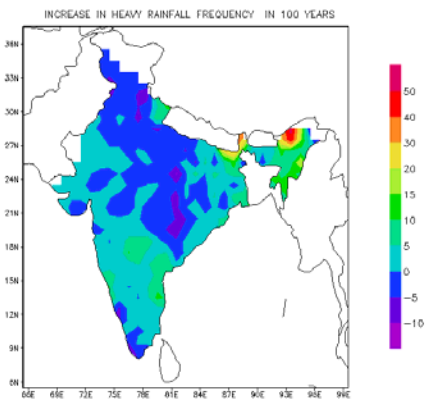


Fig. 9.4: Increase/decrease (Significant) in frequency of (a) rain days, (b) rainy days and (c) heavy rainfall days in 100 years

9.6. Highest One day point Rainfall

Highest rainfall values of one-day duration are computed from the daily rainfall values. Fig 9.5 shows the pattern of highest 1-day point rainfall over India. Isohyets of highest 1-day rainfall range from less than 20cm over large parts of interior peninsula and the arid regions of west Rajasthan to 100 cm over Cherrapunji. Highest one day rainfall of more than 40 cm have occurred over western coast, Rayalseema, some parts of coastal Andhra Pradesh, Orissa and West Bengal, Karnataka, Gujarat and Saurashtra coast, Mount Abu and Meghalaya. From all the daily rainfall data of all the raingauge (more than 9000) available at the archive of India Meteorological Department highest ever recorded one day rainfall values are extracted. Table 9.1 lists the highest observed one day point rainfall of more than 90 cm. It may be noted that the country's highest and second highest one-day rainfall occurred during the recent two decades. Out of nineteen occurrences of more than 90 cm rainfall, twelve cases occurred since 1970 onwards. Occurrence of 156.3cm rainfall in one day at Cherrapunji on 16-Jun-1995 broke the previous long time record of highest one day rainfall 103.6cm which also occurred in Cherrapunji. Consecutive two days rainfall of 15-16 June1995 created a record in world highest 2-day rainfall as reported by Guhathakurta (2007). Table 9.2 gives the world record rainfall for specified periods.

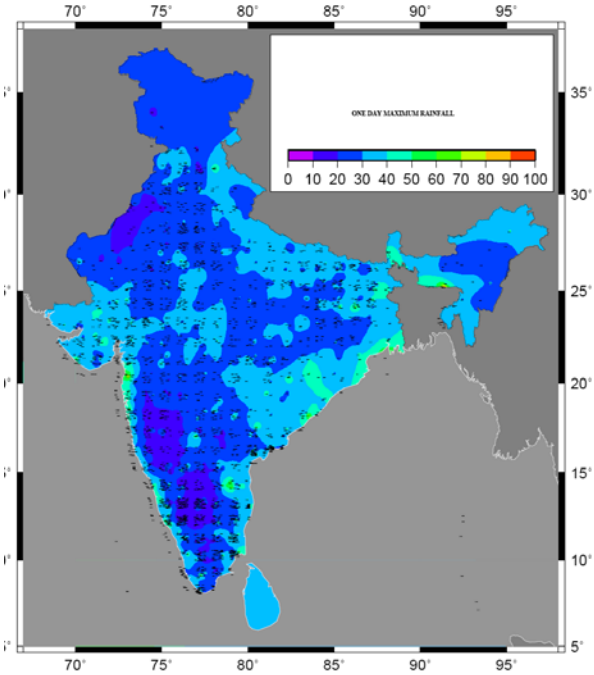


Fig.9.5: Highest One-day Extreme Rainfall in India

Table 9.1: Highest observed one day point rainfall of more than 90 cm in India

	Station	State	1-day rainfall in cm	Date of occurrence
1	Cherrapunji Obsy	Meghalaya	156.3	16-Jun-1995
2	Amini Divi	Lakshadeep	116.8	6-May-2004
3	Cherrapunji	Meghalaya	103.6	14-Jun-1876
4	Ambarnath	Maharashtra	101.0	27-Jul-2005
5	Cherrapunji	Meghalaya	99.8	12-Jul-1910
6	Mausynram	Meghalaya	99.0	10-Jul-1952
7	Dharampur	Gujarat	98.7	2-Jul-1941
8	Cherrapunji	Meghalaya	98.5	13-Sep-1974
9	Mawsynram	Meghalaya	98.0	4-Aug-1982
10	Tamenlong	Manipur	98.0	10-Aug-1970
11	Cherrapunji	Meghalaya	97.4	5-Jun-1956
12	Mawsynram	Meghalaya	94.5	7-Jun-1966
13	Mumbai	Maharashtra	94.4	27-Jul-2005
14	Tamenlong	Manipur	94.0	28-Jul-1970
15	Cherrapunji	Meghalaya	93.0	15-Jun-1995
16	Guna	Madhya Pradesh	92.8	23-Aug-1982
17	Cherrapunji	Meghalaya	92.5	27-Jun-1934
18	Hawraghat	Meghalaya	92.0	11-Nov-1995
19	Cherrapunji	Meghalaya	90.7	25-Jun-1970

Table 9.2: World's record rainfall for specified periods (Source: Guhathakurta, 2007 ; WMO,1994; Blanford , 1886–88.)

Period	Depth (mm)	Location	Date
24 h	1825	Foc Foc, La Réunion The island is located 400 miles east of Madagascar in the Indian Ocean.	8 January 1966
2 days	2493	Cherrapunji Obsy	15-16 June 1995
* 2 days	2467	Aurere, La Réunion	8-9 April 1958
3 days	3130	Commerson, La Réunion	7-9 April 1980
4 days	3721	Cherrapunji Obsy	12-15September 74
31 days	9300	Cherrapunji	1-31 July 1861
2 months	12767	Cherrapunji	June-July 1861
3 months	16369	Cherrapunji	May-July 1861
4 months	18738	Cherrapunji	April-July 1861
5 months	20412	Cherrapunji	April-August 1861
6 months	22456	Cherrapunji	April-September 1861
One Calendar year	22992	Cherrapunji	1861
12 months (from 2 different yrs)	26461	Cherrapunji	August 1860-July 1861
2 years	40770	Cherrapunji	1860 - 1861

* World record of highest consecutive 2 days rainfall before 16 June 1995

9.7. Floods and Drought in the regional scales

The meteorological floods and drought as mentioned earlier are defined as excess or deficient of rainfall departure from normal by more than 25%. India Meteorological Department has long series of sub-divisional rainfall data since 1875. Sub-division rainfall data available in the Hydrology section of the office of

ADGM(R), India Meteorological. Department is used for construction of this table. Table 9.3 shows the years of meteorological flood for each of the thirty-six meteorological sub-divisions during the southwest monsoon season during the period 1875-2009. Frequency of flood years is highest in the mainland Saurashtra and Kutch where southwest monsoon rainfall was more than normal by 25% over 40 years out of 135 years (1875-2009). Not only that in the recent years i.e. since 2001, it has experienced 5 flood years namely 2003,2005,2006,2007 and 2009. The next highest is Punjab (37 years) and then Haryana, Delhi, Chandigarh (36 years). However none of these two sub-divisions have experienced flood year since the year 2000. The sub-divisions Assam & Meghalaya (4) and Chattisgarh (7) in mainland are having lowest number of occurrence of flood years. There are19 sub-divisions which have never experienced a flood year since the year 2000.

Table 9.4 provides the list of the meteorological drought years for each of the 36 meteorological sub-divisions for the period 1875-2009. Maximum number of drought occurs over west Rajasthan (31 years) followed by Gujarat (30 years) and Saurashtra and Kutch (29 years). The sub-divisions Orissa and Coastal Karnataka had received lowest number (4 years) of occurrence of monsoon rainfall below 75 % of normal. There are eight sub-divisions, which have never experienced drought years since the year 2000. Thus the sub-division Saurashtra and Kutch has experienced both the extreme flood and drought conditions in higher frequency during the last 135 years than any other sub-divisions during the southwest monsoon season.

9.8. Floods and Drought during northeast monsoon

The period October to December is referred to as Northeast Monsoon season over peninsular India. This period is also referred to as "Post-Monsoon Season" or "Retreating southwest Monsoon Season". Northeast Monsoon season is the major period of rainfall activity over south peninsula, particularly in the eastern half comprising of the meteorological subdivisions of Coastal Andhra Pradesh, Rayalaseema, Tamilnadu & Pondicherry, South Interior Karnataka and Kerala. Contribution of northeast monsoon rainfall of these five sub-divisions to annual rainfall is about 32%, 31%, 47%, 20% and 16% respectively. Table 10.5 lists the

years in which northeast monsoon rainfall departure from normal for each of these five sub-divisions were more than 25% and less than -25%.

Both Coastal Andhra Pradesh and Rayalaseema have highest number of flood years (35 years) during the northeast monsoon. Kerala has experienced only 16 occasions of flood years in northeast monsoon. Among the recent years since 2000, the year 2005 was the best where four met sub-divisions Coastal Andhra Pradesh, Rayalaseema, Tamilnadu & Pondicherry, South Interior Karnataka out of five met sub-divisions have experienced flood year. Number of deficient rainfall (<-25%) years was highest for the sub-division South Interior Karnataka (38 years). In the recent years since 2000, the year 2000 has three sub-divisions viz. Coastal Andhra Pradesh Tamilnadu & Pondicherry and Kerala and the year 2004 has three sub-divisions viz. Coastal Andhra Pradesh, Rayalaseema and South Interior Karnataka of drought years with northeast monsoon rainfall being less than -25%.

Table 9.3: List of the meteorological flood years for each of the 36 meteorological sub-divisions for the period 1875-2009

Sub. Div. name	Year	Total Year
Andaman & Nicobar Isl	1934,1949	2
Arunachal Pradesh	1918,1921,1938,1948,1974,1987,1988,1990,1998	9
Assam & Meghalaya	1878,1918,1974,1987	4
N. M. M. T.	1897,1918,1947,1960,1969,1970	6
SHWB & Sikkim	1878,1890,1916,1965,1987,1998	6
Gangetic West Bengal	1880,1886,1900,1913,1922,1933,1971,1978,1981,1988, 1991,1999,2007	13
Orissa	1896,1900,1925,1933,1943,1961,1994,2006	8
Jharkhand	1893,1971,1977,1984,1990,1999	6
Bihar	1886,1890,1899,1922,1936,1938,1984,1987,2007	9
East U.P.	1888,1890,1894,1898,1915,1916,1922,1930,1936,1938, 1948,1953,1955,1980	14

West U.P.	1879,1881,1884,1885,1888,1891,1894,1897,1916,1917, 1921,1922,1924,1925,1933,1936,1942,1948,1958,1961, 1963,1964,1967,1975,1978,1988,2003	27
Uttaranchal	1879,1889,1890,1894,1898,1909,1910,1916,1917,1921, 1922,1924,1936,2007	14
Haryana, Delhi, Chandigarh	1875,1880,1881,1884,1886,1887,1890,1892,1893,1894, 1900,1908,1909,1914,1916,1917,1924,1933,1942,1945, 1958,1960,1961,1964,1967,1971,1975,1976,1977,1978, 1983,1988,1994,1995,1996,1998	36
Punjab	1875,1876,1878,1880,1881,1882,1884,1886,1887,1892, 1893,1894,1900,1908,1909,1914,1917,1925,1933,1942, 1945,1950,1956,1958,1959,1961,1962,1975,1976,1977, 1978,1980,1984,1988,1990,1995,1998	37
Himachal Pradesh	1875,1876,1879,1880,1881,1894,1906,1914,1917,1922, 1942,1943,1950,1971,1988	15
Jammu & Kashmir	1878,1881,1882,1890,1892,1893,1894,1903,1908,1909, 1914,1916,1917,1942,1943,1950,1953,1956,1957,1958, 1959,1961,1975,1988,1994,1996,2006	27
West Rajasthan	1876,1878,1879,1881,1884,1892,1893,1894,1908,1909, 1916,1917,1926,1931,1933,1944,1945,1955,1956,1961, 1970,1973,1975,1976,1977,1978,1983,1990,1992,1994, 1995,1996,1997,2006	34
East Rajasthan	1875,1879,1882,1884,1887,1889,1892,1908,1916,1917, 1924,1933,1942,1945,1946,1961,1973,1975,1977,1994, 1996	21
West M.P.	1878,1916,1917,1919,1933,1934,1942,1944,1948,1961, 1973,1994	12
East M.P.	1884,1891,1919,1934,1961,1994,2003	7
Gujarat	1878,1884,1892,1894,1913,1917,1921,1926,1927,1937, 1942,1944,1945,1946,1950,1954,1956,1959,1970,1976, 1977,1994,1997,2005,2006	25
Saurashtra & Kutch	1878,1881,1884,1886,1894,1896,1897,1900,1909,1910, 1912,1913,1914,1917,1921,1926,1927,1933,1937,1944, 1945,1953,1954,1956,1959,1961,1964,1967,1970,1976, 1977,1979,1983,1988,1994,2003,2005,2006,2007,2009	40
Konkan & Goa	1878,1890,1892,1914,1931,1954,1955,1958,1959,1961, 1963,1983	12
Madhya Maharashtra	1878,1882,1896,1900,1914,1933,1959,1969,1976,1988, 1991,2005,2006,2007	14

Marathwada	1875,1878,1879,1882,1883,1889,1892,1893,1894,1903, 1909,1910,1914,1916,1917,1933,1938,1942,1954,1955, 1958,1959,1963,1970,1973,1975,1980,1983,1988,1989, 1990,1998	32
Vidarabha	1883,1884,1887,1933,1938,1959,1961,1990,1994	9
Chattisgarh	1876,1877,1900,1961,1994	5
Coastal A.P.	1878, 1886, 1889, 1892,1903, 1910, 1916, 1917, 1938, 1949, 1951, 1954, 1956, 1958, 1959, 1964, 1975, 1978, 1983, 1988, 1989, 1991, 1996, 1998, 2007	25
Telangana	1878, 1892, 1893,1903, 1908, 1914, 1917, 1938, 1955, 1956, 1958, 1959, 1970, 1975, 1978, 1983, 1988, 1989	18
Rayalaseema	1878, 1889, 1890, 1892, 1897,1909, 1910, 1916, 1921, 1938, 1949, 1964, 1981, 1983, 1988, 1989, 1995, 1996, 1998, 2000, 2007	21
Tamilnadu & Pondicherry	1878, 1879, 1886, 1889, 1892, 1895, 1897,1903, 1924, 1961, 1966, 1975, 1981, 1983, 1985, 1988, 1996	17
Coastal Karanataka	1878, 1889,1914, 1959, 1961, 1970, 1975, 1994	8
North Int. Karnataka	1878, 1892,1907, 1910, 1914, 1938, 1953, 1956, 1959, 1964, 1975, 1981, 1983, 1988, 1998,2007, 2009	17
South Int. Karnataka	1882, 1886, 1889, 1896, 1897,1912,1923, 1924, 1953, 1959, 1961, 1975, 1979, 1983,2005, 2007,2009	17
Kerala	1882, 1897,1920, 1923, 1924, 1959, 1961, 1968, 2007	9
Lakshadweep	1875,1876,1877,1878,1879,1880,1881,1882,1883,1884, 1885,1886,1887,1888,1889,1890,1891,1892,1893,1894, 1895,1896,1897,1898,1899,1900,1902,1917,1919,1924, 1933, 1947, 1949, 1961, 1964, 1970, 1975, 1989, 1992, 1995, 1998, 2007	42

Table 9.4: List of the meteorological drought years for each of the 36 meteorological sub-divisions for the period 1875-2009

Sub Div name	Year	Total Year
Andaman & Nicobar Islands	1886,1892,1893,1899,1900,1907,1915,1919,1924,1925, 1927,1979,1984,1990,1999,2000	16
Arunachal Pradesh	1937,1942,1961,1992,1994,2001,2006	7
Assam & Meghalaya	1884,1896,2005,2006,2009	5
N. M. M. T.	1884,1887,1888,1889,1895,1896,1899,1972,1980,1981, 1986,1994	12
West Bengal & Sikkim	1891,1896,1904,1908,1972,1975,1976,1994	8
Gangetic West Bengal	1892,1895,1966,1975,1976	5
Orissa	1901,1924,1974,1987	4
Jharkhand	1903,1966,1979,1982,2005	5
Bihar	1877,1882,1884,1891,1901,1908,1932,1951,1959,1966, 1972,1982,1992,2009	14
East U.P.	1877,1880,1883,1896,1907,1913,1918,1928,1932,1959, 1965,1966,1972,1979,1987,2009	16
West U.P.	1877,1883,1905,1907,1913,1918,1928,1929,1941,1965, 1972,1979,1987,2004,2006,2007,2009	17
Uttaranchal	1877,1883,1905,1907,1913,1918,1941,1951,1965,1972, 1976,1979,1982,1987,1991,1996,2009	17
Haryana, Delhi, Chandigarh	1877,1896,1898,1899,1905,1907,1913,1915,1918,1920, 1928,1929,1938,1939,1941,1951,1965,1968,1974,1979, 1986,1987,1989,2002,2006,2007,2009	27
Punjab	1877,1891,1899,1904,1905,1911,1915,1918,1920,1921, 1928,1929,1938,1939,1951,1964,1965,1969,1974,1979, 1987,2002,2004,2007,2009	25
Himachal Pradesh	1877,1899,1902,1905,1907,1911,1918,1928,1965,1968, 1972,1974,1979,1981,1982,1983,1984,1986,1987,1991, 2004,2007,2009	22
Jammu & Kashmir	1876,1877,1879,1883,1884,1885,1886,1887,1889,1891, 1895,1896,1898,1900,1902,1911,1918,1920,1937,1949, 1951,1965,1971,1977,1979,1982,1991,1993,2009	29
West Rajasthan	1876,1877,1879,1883,1884,1885,1886,1887,1889,1891, 1895,1896,1898,1900,1902,1911,1918,1920,1937,1949,	31

	1951,1965,1971,1977,1979,,1982,1991,1999,2002,2004, 2009	
East Rajasthan	1877,1898,1819,1901,1905,1907,1911,1913,1915,1918, 1925,1928,1939,1941,1951,1965,1966,1979,1987,1991, 2000,2002,2009	23
West M.P.	1899,1907,1911,1915,1918,1920,1925,1951,1965,1966, 1979,2000,,2009	13
East M.P.	1877,1899,1905,1913,1941,1965,1966,1979,2000,2007, 2009	11
Gujarat	1877,1885,1888,1895,1899,1901,1904,1911,1915,1918, 1920,1923,1925,1936,1939,1948,1951,1957,1960,1965, 1972,1974,1982,1985,1986,1987,1991,2000,2002,2009	30
Saurashtra & Kutch	1875,1877,1890,1899,1901,1904,1905,1911,1915,1918, 1923,1925,1931,1939,1948,1951,1968,1969,1972,1974, 1982,1985,1986,1987,1991,1993,1995,1999,2000	29
Konkan & Goa	1877,1899,1905,1918,1920,1941,1968,1972,1986	9
Madhya Maharashtra	1877,1899,1904,1905,1911,1918,1972,1985,1987	9
Marathwada	1877,1885,1899,1905,1907,1912,1918,1920,1925,1929, 1939,1941,1971,1972,1974,1984,1985,1994,1997	19
Vidarabha	1877,1899,1902,1904,1918,1920,1950,1952,1965,1971, 1972,1974,1982,1984,1985,1987,2004,2009	18
Chattisgarh	1886,1902,1941,1965,1974,1979,1987,1998,2000,2009	10
Coastal A.P.	1883,1888,1899,1904,1920,1952,1968,1984,1987	9
Telangana	1876,1877,1888,1899,1918,1920,1939,1941,1952,1968, 1971,1972,1985,1997,2004,2009	16
Rayalaseema	1876,1891,1896,1899,1901,1904,1907,1911,1913,1918, 1920,1922,1923,1934,1941,1948,1952,1985,1994,2002	20
Tamilnadu & Pondicherry	1884,1891,1899,1904,1918,1934,1952,1969,1980,1982, 1999,2002	12
Coastal Karanataka	1881,1899,1918,2002	4
North Interior Karnataka	1876,1899,1905,1918,1921,1922,1937,1995,2002,2003	10
South Interior Karnataka	1875,1876,1881,1884,1891,1894,1905,1918,1976,1985, 1990,2002	12
Kerala	1880,1881,1899,1918,1944,1951,1952,1965,1966,1976, 1990,2002,2003	13
Lakshadweep	1901,1918,1927,1928,1934,1941,1948,1952,1956,1957, 1969,1980,2002	13

Table 9.5: List of the meteorological flood and drought years for each of the five meteorological sub-divisions during northeast monsoon for the period 1875-2009

Sub. Div. name	Year of excess rainfall (>25%)	Total excess (>25%) Year	Year of deficient rainfall (<-25%)	Total deficient (<-25%) Year
Coastal A.P.	1878,1880,1882,1883, 1886,1892,1893,1902, 1903,1915,1916,1917, 1919,1923,1925,1930, 1931,1939,1941,1944, 1946,1956,1958,1962, 1969,1972,1976,1977, 1987,1991,1994,1996, 1998, 2003,2005	35	1876,1877,1881,1891, 1896,1897,1899,1900, 1904,1905,1907,1908, 1909,1914,1956,1934, 1935,1938,1951,1957, 1959,1960,1964,1965, 1967,1970,1971,1978, 1981,1983,1984,1988, 1989,2000,2002,2004	36
Rayalaseema	1877,1880,1882,1883, 1884,1885,1886,1887, 1889,1893,1903,1912, 1916,1922,1930,1940, 1943,1946,1953,1956, 1966,1969,1972,1975, 1977,1987,1990,1991, 1993,1994,1996,1997, 1998,2001,2005	35	1875,1876,1881,1896, 1897,1899,1900,1904, 1907,1908,1909,1911, 1914,1920,1923,1924, 1926,1927,1938,1942, 1945,1947,1949,1950, 1951,1957,1959,1965, 1970,1980,1984,1986, 1988,1989,1995,1999, 2004	37
Tamilnadu & Pondicherry	1877,1880,1884,1887, 1896,1898,1902,1920, 1922,1925,1930,1931, 1932,1940,1944,1946, 1966,1969,1972,1977, 1978,1979,1993,1997, 1998,2005,2008	27	1875,1876,1878,1879, 1886,1889,1892,1894, 1897,1904,1908,1909, 1924,1926,1927,1938, 1947,1949,1950,1951, 1952,1958,1961,1966, 1968,1974,1984,1986, 1988,1995, 2000	31
South Interior Karnataka	1877,1880,1883,1885, 1887,1893,1898,1902, 1903,1910,1916,1922, 1930,1932,1933,1939, 1940,1943,1946,1953,	32	1875,1876,1879,1881, 1889,1891,1892,1896, 1897,1899,1904,1907, 1908,1913,1920,1923, 1924,1926,1927,1938,	38

	1956,1957,1962,1966, 1987,1991,1992,1993, 1994,1997, 1999,2005		1945,1951,1959,1965, 1967,1968,1970,1971, 1974,1982,1984,1985, 1988,1989,1995,2004, 2006,2008	
Kerala	1877,1885,1887,1920, 1930,1932,1943,1946, 1966,1977,1987,1992, 1993,1997,1998,2002	16	1875,1876,1890,1894, 1897,1899,1900,1904, 1908,1923,1926,1927, 1938,1947,1949,1959, 1961,1967,1968,1970, 1971,1974,1982,1984, 1985,1986,1988,2000	28

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CHAPTER 10

OROGRAPHIC INFLUENCE

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10.1. Introduction

Weather and climate of a place is substantially influenced by the presence of orographic barriers at that place. It is known that in India the location and orientation of orographic barriers play a major role in modulating the distribution of seasonal rainfall during southwest monsoon season. The problem of airflow over an orographic barrier, and, in particular, the formation of stationary atmospheric lee waves has received considerable theoretical attention in the past. It is also quite well known, at least in qualitative terms, that orography plays a dominant role in determining the precipitation of a place.

In a stably stratified atmosphere a fluid parcel, displaced vertically, undergoes buoyancy oscillation which gives rise to gravity waves. Now these gravity waves can propagate vertically to a great distance carrying energy and momentum to higher levels in the atmosphere. Sometimes, they are associated with the formation of clear air turbulence (CAT). The information about standing waves, which under favourable meteorological conditions form on the lee side of the mountain barrier, is very important for the safety of aviation. Many aircraft accidents reported in mountainous areas are often attributed to the vertical velocities of large magnitude associated with the lee waves.

The studies of the influence of orography on airflow and rainfall can be classified into two main groups:

(i) Observational

(ii) Theoretical.

10.2. Observational work

The early observational investigations on the airflow over orographic barriers were made from the experience of pilots, gliders and powered aircraft. Later, instrumented field projects and use of satellite imageries gave a comprehensive picture of the phenomenon. These observations have established the existence of the lee waves under favourable conditions of wind and thermal stability. The deformations of air-stream into wave motion on the lee of mountains are visibly revealed in a variety of orographic clouds.

Kuettner (1939), through field investigations in the European Alps, recognized the existence of powerful gravity waves over and to the lee of mountains, and noted their quasi-stationary character. Förchtgott (1949, 1957) made observational study of mountain waves at several different localities in the rugged mountainous country of Czechoslovakia. He observed lee wave clouds and the associated rotors to move slowly downstream for some appreciable fraction of a wavelength. Larsson (1954) made observational study of mountain waves using orographic cloud picture taken in the region of the central Swedish mountain. Observed wavelength was between 5km and 25km. Workers like Ludlam (1952, 1967), Reiter and Nania (1964) made exhaustive observational study regarding the different types of orographic clouds. Holmboe and Klieforth (1957) studied physical and dynamical structure of airflow across “Sierra-Nevada” in United States of America.

Although the observational evidence relating to mountain waves has come from different observational tool and from several different countries there appears to be a remarkable degree of agreement and consistency. Compiling the results of observational studies on mountain wave, their general structure and conditions relevant for their formation are discussed below:

The general structure of mountain waves:

The gravity waves frequently occur to the lee of hills and mountains in all parts of the world. They take the form of more or less vertical oscillations or undulations in the air stream to the lee. These waves may be regarded as oscillations about the dynamically stable state of the undisturbed air stream, with the mountain providing the source of disturbance and gravity providing the restoring force, necessary for oscillations by virtue of static stability of the atmosphere. The horizontal wavelength is generally in the range of 5 to 25km. It appears that wavelength increases with the wind speed and decreases with static stability.

In most cases the vertical amplitude increases from the ground upward to achieve some maximum in middle troposphere. The vertical double amplitude may reach 2000m or more, whilst maximum vertical currents maybe around 25m/s. The waves may propagate up to stratospheric level.

A succession or train of several lee waves may be formed, especially if the obstacle is a mountain ridge of considerable length lying across the air stream. In contrast, the waves to the lee of an isolated hill or mountain quickly die away downstream and there may be one or two observable lee waves.

The meteorological conditions relevant for lee waves:

(a). Static stability

In an air-stream conducive for the formation of lee waves, static stability should be larger in the lower layer as compared to the overlying upper layer. For strong waves there should be marked stability at levels where the air is disturbed by the mountain. Maximum amplitude in lee waves is usually attained somewhere in or near the layers of maximum static stability.

(b). Wind

Lee waves occur, when the air stream blows more or less perpendicularly across the ridge or has a substantial component across the ridge, and this must be maintained through a considerable depth of the troposphere. Several authors suggest a rough limiting wind speed, ranging from 7 to 15m/s, which must be attained at crest of the barriers for lee waves to occur.

Kuettner and Jenkins (1953), Harrison (1965) reported presence of severe turbulence while flying to the downwind side of an orographic barrier. Starr and Browning (1972) described several radar observations. Kuettner and Lilly (1968), Vergenier and Lilly (1970) have reported observations regarding lee waves with the help of super pressure balloons, and instrumented sailplanes/gliders during their comprehensive field project in the Rockies.

10.3. Satellite cloud observations

The orographically generated clouds are seen in satellite pictures as gray to white bands of clouds parallel to the orographic barrier. Satellite observations of orographic cloud were reported by Döös (1961, 1962), Conover (1964), Fritz (1965), Cohen et.al (1966, 1967), Wooltridge and Lester (1969), De (1971), Sarker and Calheiros (1974), Cruette (1976). Waves to the lee of the barrier were identified in these satellite pictures and compared with lee wavelengths, computed theoretically.

Lee waves due to an isolated obstacle were also studied with the help of satellite imagery. Fujita and Tecson (1977) and Pitts et.al (1977) presented photograph from Skylab mission showing waves in the lee of islands in the Aleutian group, Gough and Bouvet islands in the southern Atlantic Ocean and also Auckland, Chambell and Antipode islands south and southeast of New Zealand. The waves were found within a wedge shaped region behind the islands, with diverging and transverse wave system, similar to the wave system observed behind ships. Fujita and Tecson (1977) noted that when the wind speed exceeds a critical value the transverse waves tend to disappear leaving the diverging waves only. Gjevic and Marthinsen (1978) analyzed satellite pictures to study the lee wave patterns generated by isolated islands in the Norweigian sea and Barents sea. They found that trapped lee waves were located within a wedge shaped wake behind the islands. They found divergent type as well as transverse type lee wave pattern. In the former case the crests were observed to be oriented outwards from the centre of the wake where as in later case the crests were nearly perpendicular to the wave direction. Lee waves of diverging wave type were rather frequently observed at Jan Mayen in a situation with similar stratification as in the case when vortex shedding is observed. The wind velocity was, however, usually higher on the days with lee

waves. The waves were observed in situations with a strong stable layer inversion extending from 0.5 to 3.0 km above sea level. The wind speed in the stable layer was always more than 15m/s.

Peltier and Clark (1983) made a theoretical study for cases when ship waves were observed in NOAA-5 satellite imageries in the lee of islands in Norwegian and Barents sea.

Hoinka (1984) studied a typical mountain wave phenomenon with the help of data collected during ALPEX above the Pyrenees on 23rd March1982. Mountain waves were also detected in the satellite picture of that day. He found that the character of flow was fairly three-dimensional (3-D) below 7km and above that it was more or less two-dimensional (2-D) in structure.

In India, De (1970) reported lee waves as evidenced by satellite cloud pictures over northeast India. Observed wavelength of lee waves across the Assam-Burma hills was between 17km and 34km.

Sinha Ray (1988) reported mountain waves over the western Himalayas as evidenced by satellite cloud imageries. Observed wavelength of lee waves across western Himalayas was between 7km and 23km.

The wavelengths observed in the two studies given above were verified with the computed values using a 2-D model.

Tyagi and Madan (1989) reported mountain waves over the Western and Central Himalayas, using satellite imageries for the period January 1981 to July 1984. They documented several cases of mountain waves to the lee of Western and Central Himalayas over Tibetan plateau. The observed wavelengths, reported by them, vary between 13 to 22 km.

De (1994) discussed the importance of mountain waves and their impact on aviation with particular reference to the Indian scenario. Recently, Kumar (2000) made a climatological study of lee waves in the different mountainous areas of India. He reported lee wavelengths between 15 and 18km in the eastern Himalayas up to

90^0E, between 15 and 22km over Assam, Burma and Yunun regions (China), between 9 and 14km over central India, between 8 and 11km over peninsular India.

These studies show that range of wavelength of lee waves occurred in different mountainous region of India lies between 10 and 50km.

From the above observational evidence we get a fair idea about the influence of orography on airflow. The influence depends on the physical and dynamical structure of the air stream and on the scale of the phenomenon.

10.4. Theoretical studies

The theoretical study of airflow across an orographic barrier is a complex problem. The complexity of the problem may be attributed to the following:

(a). Apart from the shallow adiabatic layers near the ground on a hot day, the atmosphere has a continuous thermal stratification.

(b). Atmosphere is compressible and has no definite upper boundary. The formulation of definite upper boundary conditions in the problem of airflow across an orographic barrier is difficult and it has given rise to considerable controversy.

(c). Complexity of the problem is also due to different scales of the barrier. For example the nature of disturbance caused by a narrow hill is quite different from that caused by a broad plateau. Also the disturbance caused by a long ridge differs from that caused by an isolated 3-D barrier.

The theoretical studies can broadly be categorized into three following types:

(i). Linerized steady state perturbation theory

(ii). Time dependent theories

(iii). Large-scale flow across a large barrier.

10.4.1. Linearized steady state perturbation theory

The studies on the effect of orographic barrier on airflow may again broadly be divided into two categories: in one category the orographic barrier is assumed to have a semi-infinite extension in the direction normal to the prevailing basic flow, so that the flow can be considered to be essentially two dimensional (2-D). In other

category the orographic barrier is assumed to have finite extension in the direction of prevailing wind as well as in the direction normal to it, so that the flow essentially becomes three dimensional (3-D). We shall briefly discuss these two different types of formulations.

10.4.1.1. Two dimensional (2-D) linearized steady state perturbation theory

Lyra (1943) considered the problem of airflow over a rectangular obstacle. He considered a 2-D model with uniform air-stream of constant static stability and obtained solutions using Green's functions. He obtained lee waves, which decreased downstream and increased upward. But this upward increase of wave amplitude was contrary to the observation.

Queney (1947, 1948) proposed a complete theory of adiabatic perturbations in a stratified and rotating atmosphere, and applied this theory to the flow of air-stream over a 2-D bell shaped mountain with half width 'a'. Like Lyra (1943) Queney also took a uniform basic flow and constant static stability.

From physical consideration he found that, there could be three different horizontal scales of motion, (i) when 'a'<10km, the motion is non-hydrostatic and non-geostrophic, (ii) for 'a'>1000km, the motion is hydrostatic and quasi-geostrophic. But Queney did not discuss about the motion when the value of 'a' lies between 10km to 100km.

Queney (1948), Scorer (1953) and Corby (1954) showed that Lyra's (1943) results were due to the fact that he used a steep rectangular obstacle.

Scorer (1949) first introduced the variation of wind and stability with height. He assumed a frictionless, steady, non-rotating and incompressible flow. In his study Scorer divided the atmosphere into two layers. Introducing the concept of layered atmosphere, he proved that the necessary condition for the occurrence of 2-D mountain wave in a two layer atmospheric model is $l_1^2 - l_2^2 > \frac{\pi^2}{4h^2}$, where l_1^2 and l_2^2 are the values of l^2 in the lower and upper layer respectively and h is the depth of

the stable lower layer and $l^2 = \frac{g}{\theta U^2}\frac{d\theta}{dz} - \frac{1}{U}\frac{d^2U}{dz^2}$ is the Scorer's parameter, U is the undisturbed prevailing wind normal to the ridge, k is the horizontal wave number in the direction of U and θ is the potential temperature of the undisturbed prevailing wind. He also showed that no lee wave is possible for a constant l^2 with height.

Sawyer (1960) studied the 2-D mountain wave problem numerically. For that a steady, non-rotating, non-viscous, adiabatic 2-D flow was considered. Wind velocity and static stability were assumed to be constants or some simple variation with height in the basic air stream. He showed that the lee wave system consists of more than one wave superimposed on each other. He also showed that if two or more lee waves are present, the shortest dominates the lower troposphere and longer waves are conspicuous at higher levels and the crest of streamlines tilted upstream.

Eliassen and Palm (1961) studied the flow of wave energy for stationary, two-dimensional gravity waves in a basic current where velocity and stability varies with height. They showed that the vertical flux of wave energy varies with height in proportion to the wind speed. They established that the vertical flux of horizontal momentum does not change with height except possibly at levels where basic current is zero.

In India, Sarker (1965) first time made theoretical studies of mountain waves over the Western ghats in India. It was shown by him that the air-stream of winter season has the favourable stable stratification to produce mountain waves on the lee of Western Ghats, provided the wind is westerly. He showed that 3 to 4 waves generally superimpose on one another. He showed that during winter season the lee waves with wavelength between 25-80km only give rise to significant amplitude across the Western Ghats. Sarker (1966, 1967) also investigated mountain waves over the Western gnats during the southwest monsoon season. During this season the air-stream does not have much stable stratification and it is more or less neutral and the lapse rate is close to moist-adiabatic. His theory showed that lee waves are possible in a statically neutral atmosphere also, if the vertical shear of wind varies

favorably in the vertical. Lee wavelength as computed by him was between 20-30 km.

De (1970, 1971) computed the wavelength of the lee waves over the Assam-Burma hills using a similar approach with necessary modification for the mountain profile and wind direction in that region. Computed wavelength varies between 17 to 34 km and agreed well with those observed from satellite pictures.

Farooqui and De (1974) used a two dimensional model to calculate the flow over a small obstacle (half width 2 km), a large obstacle (half width of 20 km) and across the Assam Hills (200-300 km). Their results in the later experiment show long waves of length (20-40 km) and other large perturbations mainly between heights 1 and 9 km. From 9 to 15 km perturbations are very small.

Smith and Lin (1982) investigated theoretically the response of a stratified air stream to combined thermal and orographic forcing. They computed the magnitude of heating aloft from observed rainfall rates. They had shown that for typical wind speeds and rainfall rates the thermally generated waves would equal or exceed the orographically generated waves. They showed that the heat released associated with observed rates of orographic precipitation were found to have a significant influence on the airflow. In a hydrostatic atmosphere the typical response to localized heating was seen to be a down ward displacement in the vicinity of heating. They showed that the momentum flux and mountain drag are strongly influenced by moist processes.

Kumar et al (1995) considered a single layer two-dimensional mountain wave model with exponential scorer parameter profile to study the effect of latent heat, released associated with precipitation, on the lee waves across western Himalayas. They showed that for non-hydrostatic non-precipitation case, when balanced heating/cooling takes place on the windward/lee ward side of the mountain, the effect of heating is negligibly small. They showed that amplitude of the wave is more in precipitation case as compared to non- precipitation case. They also showed that for a hydrostatic case it is observed that the “Shear effect” is opposite to that due to thermal forcing.

10.4.1.2. Three dimensional (3-D) linearized steady state perturbation theory

One fundamental difference between the 2-D and 3-D approach is the direction of propagation of wave energy away from the mountain. In two dimensions as the mountain becomes wider and the flow more nearly hydrostatic, the group velocity (relative to the mountain) becomes directed vertically with the result that the wave energy is found directly above the mountain. This result does not carry over to three dimensions. Some of the hydrostatically waves generated by 3-D mountain lie down stream of the mountain and to the side tending to form trailing wedges of vertical motion. Thus for practical and theoretical reasons, it is necessary to understand the three-dimensional mountain flow problem.

Wurtele (1957) represented the 3-D orographic barrier in the form of semi-infinite plateau of height 'h' with narrow width '2b' in the crosswind direction. He considered the incoming wind (U) and buoyancy frequency (N) to be independent of height. His theory predicted the region of updraft, which had a horseshoe shape, and was located some distance downstream of the barrier.

Crapper (1959) presented a 3-D small perturbation approach of waves produced in a stably stratified air stream flowing over a mountain. He obtained the fundamental solution for a doublet disturbance in an air stream in which Scorers parameter remains constant and then it was extended to that for a disturbance caused by a circular mountain in the same air stream. He showed that circular mountain can give rise to waves which have greater amplitude than those produced by an infinite ridge in the same air stream. Crapper (1962) considered the airflow across a 3-D barrier with elliptical contour for two types of air stream. In one case the Scorer parameter l was constant with height, in other case it was assumed to fall off exponentially with height. In each of the above cases $\frac{1}{U}\frac{d^2U}{dz^2} = q^2$ was kept constant. The result showed that when l is constant, and then the form of the waves was determined by the value of q. He also showed that when l falls off exponentially, the waves closely resembled ship waves for any value of q.

Sawyar (1962) studied gravity waves in the atmosphere as a 3-D problem. He derived an equation, for the vertical variation of the amplitude of the standing waves, when the wind varied with height and the wave was periodic in the horizontal. He solved the equation numerically for specified two or three layer atmosphere to determine possible wavelengths in the horizontal directions for lee waves. He obtained results for the cases when wind direction changed with height as well as for the cases when wind direction remained same in the vertical. He showed interestingly that Scorer's (1949) condition for the occurrence of lee wave was no longer applicable for wave motion in 3-D. He showed that in 3-D, lee waves are always possible in a two-layer atmosphere.

Onishi (1969) solved 3-D linearized equations for arbitrary upstream conditions by including friction in the governing equations. Pekelis (1971) extended his 2-D work to solve linearized 3-D problem. Vertical velocity fields obtained by him compare well with those of Sawyar (1962).

Smith (1980) examined the stratified hydrostatic flow over a bell shaped 3-D isolated mountain using linear theory. Solutions for various parts of the flow field were obtained using analytical method and numerical Fourier analysis. The flow aloft was found to be composed of vertically propagating mountain waves. The maximum amplitude of these waves occurred directly over the mountain, but there was considerable wave energy, trailing downstream along the parabolas $y^2 = \frac{Nzax}{U}$; where U, N are respectively the constant basic zonal wind and buoyancy frequency.

Bluemen and Dietze (1981) considered a 3-D linear hydrostatic model of stationary mountain wave in a stably stratified air-stream. They took both the incoming flow and Burnt-Vaisalla frequency to be independent of height, but lateral variation of incoming flow was incorporated by assuming a hyperbolic secant profile ($U = \sec h\, y$). The results in their solution for different shape of hill showed that the pressure pattern and the velocity at the ground level were similar in many respects to the field obtained by Smith (1980) for constant basic flow. The incoming air-stream tends to circumvent the hill resulting in a permanent streamline deflection.

Somieski (1981) studied the stratified hydrostatic flow over a three dimensional circular mountain. He derived a 2nd order wave equation from the primitive equation including constant rotation and vertical wind shear of the mean flow. He solved the equation numerically. He showed that in case of no shear and constant static stability, the nodal lines are parabolic for a circular mountain of diameter 50km.

Bluemen and Dietze (1982) extended their earlier model by including the vertical variation of the basic flow and static stability. To take into account the vertical structure of basic state, they introduced stretched vertical co-ordinate. The energy flux computed by them was comparable with the results of Elliassen and Palm (1961).

Olafssen and Bougeault (1996) explored the hydrostatic flow over an elliptical mountain barrier of aspect ratio 5. They took upstream profiles of wind (U), stability (N) constant and ignored the effect of Coriolis force. Under such conditions their result showed the flow characteristics to be dependent mainly on the non-dimensional mountain height $\frac{Nh}{U}$. They found that for all values of $\frac{Nh}{U}$, a substantial part of the flow was diverted vertically above the mountain. They found generation of potential vorticity in the wake of the mountain, leading to the creation of lee vortices.

Dutta et.al. (2002) developed a 3-D lee wave model for an idealized air stream, where, both stability and the basic flow was solely normal to the major ridge of the barrier, remain invariant with height. The vertical structure equations neglecting rotation of the earth from scale considerations are:

$$\frac{\partial^2 \hat{w}_1}{\partial z^2} + m^2 \hat{w}_1 = 0$$

$\frac{\partial^2 \hat{\eta}_1}{\partial z^2} + m^2 \hat{\eta}_1 = 0$ for perturbation vertical wind $w'(x, y, z)$ and streamline displacement $\eta'(x, y, z)$.

Here, $\hat{w}_1(k, l, z) = \hat{w}(k, l, z)\sqrt{\frac{\rho_0(z)}{\rho_0(0)}}$,

$\hat{\eta}_1(k,l,z) = \hat{\eta}(k,l,z)\sqrt{\frac{\rho_0(z)}{\rho_0(0)}}$ and $\hat{w}(k,l,z), \hat{\eta}(k,l,z)$ are double Fourier transforms of $w'(x,y,z)$ and $\eta'(x,y,z)$ respectively.

Here, $m^2 = \frac{k^2+l^2}{k^2}\left(\frac{N^2}{U^2} - k^2\right)$, N is the vertically averaged Brunt-Vaisalla frequency, U is the vertically average of the component of prevailing wind and k,l are horizontal wave numbers in the direction normal and parallel to the major ridge respectively. The orographic barrier has been represented by an analytical function $z = \frac{H}{1+\frac{x^2}{a^2}+\frac{y^2}{b^2}}$.

$w'(x,y,z)$ and $\eta'(x,y,z)$ are expressed in terms of double integrals. These double integrals are difficult to evaluate exactly. Hence their asymptotic values have been evaluated and side-by-side these integrals have been also numerically approximated by a quadrature scheme. Results obtained from the above study are:

- For both the wave parameters, w' and η', the asymptotic solution as well as the numerical solution show upwind tilting and lateral spreading of waves with height.
- In the asymptotic solution as well as in numerical solution the regions of updrafts are approximately crescent shaped.
- Asymptotic solution as well as Numerical solution show that both w' and η' falls off downwind of the barrier, more over in asymptotic solution they vary as x^{-1} in the central plane.

Dutta (2003) had considered the effect of both the components of basic flow, viz., and the component 'U' normal to the major ridge as well as the component 'V' parallel to the major ridge. Similar to Dutta et al., (2001), in this study also, both the components of basic flow and the Burnt-Väisala frequency (N) assumed to be invariant with height. The vertical structure equations neglecting rotation of the earth for scale considerations:

$$\frac{\partial^2 \hat{w}_1}{\partial z^2} + (k^2+l^2)\left[\frac{N^2}{(Uk+Vl)^2} - 1\right]\hat{w}_1 = 0$$

$\frac{\partial^2 \hat{\eta}_1}{\partial z^2} + (k^2 + l^2)\left[\frac{N^2}{(Uk + Vl)^2} - 1\right]\hat{\eta}_1 = 0$ for perturbation vertical wind $w'(x, y, z)$ and streamline displacement $\eta'(x, y, z)$.

Here, $\hat{w}_1(k,l,z) = \hat{w}(k,l,z)\sqrt{\frac{\rho_0(z)}{\rho_0(0)}}$,

$\hat{\eta}_1(k,l,z) = \hat{\eta}(k,l,z)\sqrt{\frac{\rho_0(z)}{\rho_0(0)}}$ and $\hat{w}(k,l,z), \hat{\eta}(k,l,z)$ are double Fourier transforms of $w'(x, y, z)$ and $\eta'(x, y, z)$ respectively.

Here, N is the vertically averaged Burnt-Väisala frequency and U, V are the vertically averaged components of prevailing horizontal wind and k, l are horizontal wave numbers in the direction normal and parallel to the major ridge respectively. The orographic barrier has been represented by the same analytical function as in the first part of this chapter.

$w'(x, y, z)$ and $\eta'(x, y, z)$ are expressed in terms of double integrals. These double integrals are difficult to evaluate exactly. Hence their asymptotic values have been evaluated and side-by-side these integrals have been also numerically approximated by a quadrature scheme. We obtain following results from the study:

- The asymptotic solution as well as numerical solution show that, in the central plane along the line a $Uy - Vx = 0$ both of w' and η' decrease downwind of the barrier. Moreover, in the asymptotic solution they vary at a rate inversely proportional to the distance along this line.
- Incorporation of 'V' seems to rotate the axes of crescent shaped nodal lines through an angle $\tan^{-1}\left(\frac{V}{U}\right)$.
- In the horizontal plane contours of w', as obtained by asymptotic method, show that the regions of updraft are crescent shaped which are inclined at an angle $\tan^{-1}\left(\frac{V}{U}\right)$ with the central plane.

- Asymptotic solutions for both w' and η' show upwind tilting along the line $Uy - Vx = 0$ and lateral spreading about the same line with height.
- The numerical solution for w' also shows approximately crescent shaped updraft region in the horizontal plane.

Dutta (2005) developed a 3-D lee wave model for an air stream, where, both stability and wind velocity of the basic flow had realistic vertical variation. For simplicity, the basic flow was assumed to have only one component 'U' normal to the major ridge of the barrier.The vertical structure equations neglecting rotation of the earth for scale considerations was:

$$\frac{\partial^2 \hat{w}_1}{\partial z^2} + \left(\frac{N^2(k^2+l^2)}{(Uk)^2} - \frac{1}{U}\frac{d^2U}{dz^2} - \frac{1}{\rho_0}\frac{d\rho_0}{dz}\frac{1}{U}\frac{dU}{dz} - (k^2+l^2) + \frac{1}{4\rho_0^{\,2}}\left(\frac{d\rho_0}{dz}\right)^2 - \frac{1}{2\rho_0}\frac{d^2\rho_0}{dz^2} \right)\hat{w}_1 = 0$$

Where, $\hat{w}(k,l,z) = \sqrt{\frac{\rho_0(0)}{\rho_0(z)}}\,\hat{w}_1(k,l,z)$, $\hat{w}(k,l,z)$ is the double Fourier transforms of $w'(x,y,z)$

$N^2 = N^2(z) = \frac{\frac{g}{\theta_0}\frac{d\theta_0}{dz}}{(U(z))^2}$ is the square of the Brunt-Vaisalla frequency of the prevailing basic flow, $U(z)$ is the basic state wind, $\theta_0(z)$ is the basic state potential temperature and $\rho_0(z)$ is the basic state density at level z.

The above equation is integrated downward from the top of the model domain located at certain height z=z_1, at and above which $\hat{w}_1(k,l,z)$ decays exponentially. Finally $w'(x,y,z)$ is expressed as,

$$w'(x,y,z) = \sqrt{\frac{\rho_0(0)}{\rho_0(z)}}\,\mathrm{Re}\; i\int_{-\infty}^{\infty}\int_{-\infty}^{\infty} kU(0)\,\hat{h}(k,l)\,\frac{\psi(k,l,z)}{\psi(k,l,0)}\,e^{i(kx+ly)}\,dk\,dl$$

$$\eta'(x,y,z) = \sqrt{\frac{\rho_0(0)}{\rho_0(z)}}\,\mathrm{Re}\int_{-\infty}^{\infty}\int_{-\infty}^{\infty} \frac{U(0)}{U(z)}\,\hat{h}(k,l)\,\frac{\psi(k,l,z)}{\psi(k,l,0)}\,e^{i(kx+ly)}\,dk\,dl\,,$$ where,

$\hat{h}(k,l)$ is the double Fourier transform of the ground profile $z = h(x,y)$ of the barrier at surface and $\psi(k,l,z)$ satisfies the equation

$$\frac{\partial^2\psi}{\partial z^2}+\left(\frac{N^2(k^2+l^2)}{(Uk)^2}-\frac{1}{U}\frac{d^2U}{dz^2}-\frac{1}{\rho_0}\frac{d\rho_0}{dz}\frac{1}{U}\frac{dU}{dz}-(k^2+l^2)+\frac{1}{4\rho_0{}^2}\left(\frac{d\rho_0}{dz}\right)^2-\frac{1}{2\rho_0}\frac{d^2\rho_0}{dz^2}\right)\psi=0$$ and

the boundary condition that, at and above z=z_1, $\psi(k,l,z)$ decays exponentially and also $\psi(k,l,z)$ =1 at z=z_1.

For a given divergent wave number, $l=l_0$ (say), the transverse lee wave number was found out by the following formula:

$k_r' = k_r - \frac{(k_{r+1}-k_r)\psi(k_r,l_0,0)}{\psi(k_{r+1},l_0,0)-\psi(k_r,l_0,0)}$, where, k_r and k_{r+1} are any two successive values of 'k' in the range of integration such that $\psi(k,l_0,0)$ differs in sign between them.

The total contribution of all possible transverse lee waves for all divergent lee waves towards the $w'(x,y,z)$ and $\eta'(x,y,z)$ can be obtained (following Sawyar, 1962) as:

$$w'_{lee}(x,y,z) = -2\pi\sqrt{\frac{\rho_0(0)}{\rho_0(z)}}\sum_l\sum_{k'_r} k'_r U(0)\hat{h}(k'_r,l)\frac{\psi(k'_r,l,0)\cos(k'_r x+ly)}{\left[\frac{\partial\psi(k,l,0)}{\partial k}\right]_{k=k'_r}}\delta l$$, and

$$\eta'_{lee}(x,y,z) = 2\pi\sqrt{\frac{\rho_0(0)}{\rho_0(z)}}\sum_l\sum_{k'_r}\frac{U(0)}{U(z)}\hat{h}(k'_r,l)\frac{\psi(k'_r,l,0)\sin(k'_r x+ly)}{\left[\frac{\partial\psi(k,l,0)}{\partial k}\right]_{k=k'_r}}\delta l,$$

Where $w'_{lee}(x,y,z)$ the lee is wave part of $w'(x,y,z)$ and $\eta'_{lee}(x,y,z)$ is the lee wave part of $\eta'(x,y,z)$. The variation of $w'_{lee}(x,y,z)$ and $\eta'_{lee}(x,y,z)$ in the central plane (y = 0) are given in fig () and their contour in the horizontal plane are given in fig () for different seasons as mentioned above.

From the study following results have been obtained:

- The air stream characteristics during winter, Post monsoon and south-west monsoon (SWM) seasons in India are favourable to give rise both divergent and transverse lee waves across the Western gnats (WG) and Khasi-Jayanti (KJ) hills.
- During SWM season, in the statically neutral air-stream, a given divergent lee wave gives rise only one transverse lee wave across both the barrier under study.

- But during winter-Post monsoon period, in the statically stable air-stream, a given divergent lee wave may give rise to a number of transverse lee waves across WG.
- During SWM period longer transverse lee waves corresponds to shorter divergent lee waves and vice-versa for both the barrier.
- During the period from December to March, it is seen that in general the wavelength of transverse lee wave increases with that off divergent lee wave.
- In the central plane at any level, w'_{lee} decreases downwind of the barrier in association with lee wave across both barriers during SWM season. In winter season also, w'_{lee} decreases downwind of the barrier in association with lee wave across WG in a comparatively less regularly.
- During SWM season for KJ hills, the transverse lee wave length varies from 30km to 90km corresponding to all possible divergent lee wave length ranging from 30km to 150km and that for WG varies from 90km to 135km. During December-March for WG, the transverse lee wavelength varies from 30km to 147km corresponding to all possible divergent lee wavelengths ranging from 30km to 150km.
- In both the season, for both the barriers, the updraft regions are approximately crescent shaped. During December-March, crescent shaped updraft regions are concave downwind, where as during June-September they are convex down wind. Furthermore, during December-March, crescent shaped updraft regions tilts upwind and spreads laterally with height. Such tilting or lateral spreading has not been found during June-September.
- The maximum value of w'_{lee} during December-March across WG has been found to be 6-15mtr/sec, attained at 9.5km and that during June-September it ranges between 30mtr/sec-50mtr/sec.

In the absence of static stability, only wind shear can give rise to convergent-trapped 3-D lee wave. To obtain divergent lee wave a threshold minimum value of static stability (dry or moist) is always required.

10.5. Time dependent solution

A first attempt concerning time dependent mountain wave problem was made by Hoiland (1951). He considered a 2-D, incompressible uniform flow over a corrugated bed with a free surface and no vertical shear. Using perturbation technique, the equations were linearized and results showed the existence of unique solution (the steady state solution) as time tends to infinity.

Foldvic and Wurtele (1967) numerically investigated the transient nature of lee waves by numerical computations for idealized and realistic air-stream at various time intervals. They have shown, by retaining non-linear terms in the governing equation, that intensity of down-slope wind is intensified and that of up-slope winds weakened, thus producing hydraulic jumps over the lee slope in some areas.

Furukawa (1973, 1977) studied the airflow over mountain as an initial value problem. He considered the following four cases

(i) Linear equation with linear lower boundary condition.

(ii) Linear equation with non-linear lower boundary condition.

(iii) Non-linear equation with linear lower boundary condition.

(iv) Non-linear equation with non-linear lower boundary condition.

His main results were:

(i) Non-linear effects are generally dominant near the mountain. They reduce the approaching flow upstream in the lowest layer and generate a strong wind zone near the level of mountain crest. Further down-slope winds intensify and extend far down-stream in the lowest layer.

(ii) Closed circulation pattern is produced in full non-linear case, a feature not found in linear approach.

(iii) The non-linear effect enhance when the mountain height is increased beyond the critical value for overturning instability.

(iv) In the case of relatively high mountain, the flow does not necessarily reach a steady state, if full non-linear model is adopted; where as the corresponding linear model reaches a quasi-steady state.

Furukawa concluded that the inclusion of the finiteness of the mountain is an essentially important factor in simulating the airflow near the mountain.

Klemp and Lilly (1978) developed a numerical model to simulate the stably stratified airflow over a mountain assuming that the flow is 2-D, hydrostatic, non-rotating but time dependent. The model had a viscous layer beneath the upper boundary, which acts as a sponge layer and eliminates the possible reflection of upward propagating wave energy from the upper boundary. The model showed good agreement when compared with two observed cases of strong mountain waves.

Hatwar (1982) considered a 2-D frictionless, incompressible stratified model. He found out time dependent solution for 2-D flow over a triangular mountain. He performed few numerical experiments with different boundary conditions.

Peltier and Clark (1983) considered a 3-D time dependent incompressible model to simulate the evolution of non-linear internal waves which are forced by stratified flow over isolated topography. Detailed comparison of the output of his non-linear model with the results of 3-D linear steady state theory showed that the linear theory might considerably underestimate the forced wave amplitude for symmetric topographic excitation.

10.6. Large-scale flow across a large barrier.

Rossby (1939), following his investigations on the propagation of planetary waves in the atmosphere, hinted that the topographic features could give rise to a system of stationary waves, which coincided with semi-permanent features of circulation pattern around the globe. He drew it's analogy with the system of standing waves which are sometimes observed in the clouds on the lee side of the mountain ridge.

Charney and Elliassen (1949) established that the stationary perturbations in the westerlies are forced perturbations created by the forced ascend of the westerly current over topography or continental elevation and modified by friction. Considering latitudinal belt of 45^0-50^0N they obtained two stationary troughs around

longitude 70°-150°W, which were due to forced ascend of the westerlies over the Rocky mountains and the Himalayas respectively.

Murakami (1956) studied the influence of topography by evaluating the stationary flow pattern at the equivalent barotropic level by using vorticity equation along with the divergence field caused by the mountain. He showed that the splitting of the jet stream in the far east and the difference of intensity of wind maxima, in the three regions around (150°E, 30°N), (70°E, 40°N) and (40°E, 25°N) could be explained due to differences in topographic features of the Himalayas and the Rockies. Smagorinsky (1953) considered the variation of zonal wind as a linear function of height and solved the problem of quasi-stationary perturbations produced by heat source and sinks.

Saltzman (1963) considered the problem of meridional circulation produced by heat sources and orographic barriers. The solutions were obtained as influence functions for the forced geostrophic response of the atmosphere to the combined zonal asymmetries in topography and heat sources. The computations were made using idealized representation of the forcing function.

Das (1964) studied the influence of the Himalayas using a linear baroclinic model which included the variation of ‘ f ’ with latitude. The solutions were obtained for the downstream waves in an asymptotic form. It was shown that the wave distortion depends on a non-dimensional parameter, which is a combination of Froude number $\left(U^2/gd\right)$, Rossby number $\left(U/fL\right)$ and the static stability $\left(l = \frac{1}{\theta}\frac{d\theta}{dz}\right)$. Considering the center of the circular mountain at 85°E and 30°-40°N, the major trough line was found between 105°-110°E, which is in reasonable agreement with the observations over eastern Tibet in winter.

Krishnamurty et.al (1973) has studied the effect of Tibetan plateau as a mechanical barrier in the generation of warm anticyclone in the upper troposphere over Asian highlands. Once the anticyclone is formed it influences the Climatology of the tropical zonal flow pattern. By performing long-term integration of the non-

divergent barotropic vorticity equation with β-plane approximation for a finite domain 45°N to 25°S, they have successfully generated the stationary feature of 200mb northern hemisphere flow pattern.

Hahn and Manabe (1975) had studied, in detail, the role of mountains in south Asian monsoon circulation. They made a detail numerical experiment with 11sigma $\left(\sigma = \frac{p}{p_s}\right)$ layers extending from surface to 31km, which has prescribed seasonal variation of insolation and sea surface temperature (SST). The model was integrated for three model years. The model was global in domain and incorporated smoothed mountain topography. In order to show that the mountains play a very important role in the south Asian monsoon circulations, they also conducted a second experiment where mountains were removed, other things being same. The second experiment was integrated with respect to time from 25th March to 31st July. The model with mountain was designated as M-model and that without mountain was designated as NM-model.

The comparison of the simulation with mountain to that of without mountain reveals that the presence of mountain is instrumental in maintaining the south Asian low-pressure system. In the M-model with much higher temperatures are maintained in the middle and upper troposphere over Tibetan plateau, a region, where upward motion and latent heating dominate. Without mountain, downward motion and sensible heating by the earth's surface would have dominated in this region.

Das et.al (1976, 1978) made a detailed study of the impact of orographic barriers on the monsoon circulation. Both sigma and pressure co-ordinate systems were used with three layers in the vertical. They took 200hpa as the upper bound and the surface of earth as lower bound. The horizontal region of their study was the rectangle bounded by the latitudes 0°N and 60°N and the longitudes 0°E and 140°E. In their model Himalaya was represented by an elliptic block. The salient features of their study were the formation of a deep trough to the east of Himalaya, anti cyclonic curvature of the stream lines induced by the Western Ghats and a low pressure

area, resembling the monsoon trough, if the initial zonal profile had meridional and vertical shear.

Tokioka and Noda (1986) studied the effects of large-scale orography on January atmospheric circulation with the help of a five-layer general circulation model (GCM). The results of the numerical experiment carried by them showed that a large scale orographic barrier tend to divert flows around it and also showed that the ascending centre shifts pole ward on the upwind side whereas the descending centre shifts equator ward to the downwind of the barrier.

Chakraborty, Nanjundiah and Srinivasan (2002) investigated the Role of Asian and African orography in the Indian summer monsoon using a general circulation model. The study revealed that removal of the African orography increases the seasonal precipitation over the Indian subcontinent by 28%, whereas removal of orography over the entire globe reduces it by 25%. They also found a substantial delay in all-India monsoon onset in the experiment in which mountains were removed globally; mainly due to the intrusion of mid latitude dry air west of 80°E. The increase in precipitation in which orography over Africa was removed was attributed to the positive feedback between the wind over the East African coast/Arabian Sea and precipitation over Bay of Bengal, with the latter leading the former by about 2 days.

10.7. Orographically induced gravity wave drag or orographic drag

When air flows across an orographic barrier, then stationary wave disturbances are set up in the air current. The dynamical property of these waves depends on the size of the barrier. A meso-scale mountain can excite only gravity waves. Because of these gravity waves, pressure is systematically higher on the upwind slope of the barrier than on the lee slope of it. Due to this difference of pressure there is a net force exerted on the ground. This is known as pressure drag or mountain drag. Now in response to this net pressure drop between the windward and leeward slope of the barrier, these gravity waves transport momentum from a stably stratified air stream to the earth's surface. This is known as wave momentum flux or orographically induced gravity wave drag or orographic drag.

Sawyar (1959) first pointed out the relative importance of this momentum loss of stably stratified air-stream by the above mentioned momentum transport. He examined a case of 2-D flow over a bell shaped obstacle and determined that the typical surface stress due to wave momentum flux was of the order of 1-10 dyne/cm². Eliassen and Palm (1961) had shown that vertical flux of this wave momentum does not change with height, except possibly at those levels where basic flow becomes zero. Bluemen (1965) confirmed the findings of Sawyar (1959). He had shown that the magnitude of the wave drag is sensitive to the vertical wavelength. He also showed that the maximum value of the drag is attained when the vertical wavelength is twice the maximum height of the mountain.

Bretherton (1969) reviewed the theories concerning the propagation of internal gravity waves (IGW) in a horizontally uniform shear flow. He noted that an upward transport of horizontal momentum inevitably accompanies the generation of such waves in the atmosphere, the mean flow being affected only precisely at those levels, where the waves are dissipated. He had also shown that if the mean wind depends on horizontal position, there might be a continuous transfer of momentum from the waves to the mean flow during propagation. In the absence of intense clear air turbulence or a critical level, where the intrinsic frequency vanishes, many waves propagate vertically upward to a great height. His computation showed that for a 19m/s gradient wind over hilly terrain in north Welsh, the wave drag amounted to 4 dyne/cm², of which 3 dyne/cm² probably acted on the atmosphere above 20km.

A conclusive verification of the importance of the wave drag, at least over the Front Range of the Colorado Rockies, was obtained from the data collected by instrumented aircraft and reported by Lilly (1972).

Bluemen and McGregor (1976) studied the effect of both crosswind and vertical shear of the basic flow in a linear, hydrostatic model of stationary mountain lee waves in a stably stratified air-stream. Using a constant lapse rate basic flow, analytical solutions were determined first. They compared the solution for the basic flow $\bar{U}(y)$ =Sech y, with that for the constant basic flow $\bar{U}$=constant. In the former case they found the wave energy to be trapped by the background shear. They also found that wave drag was higher in the former case. It was also found that wave

drag exerted by an isolated 3-D mountain was always less than that exerted by a 2-D semi-infinite ridge. In their study they had also considered a two-layer model in order to examine the effect of stable stratosphere. In their study the wave drag was shown to be sensitive to the phase difference between the transmitted and reflected waves in the lower layer. They found that this sensitiveness becomes more pronounced in the presence of crosswind shear.

Smith (1978) had determined the pressure drag on the Blue-ridge Mountain in the central Appalachians. During the first two weeks of January 1974, several periods with significant wave drag were observed by him with pressure differences typically 50N/m ^{2} across the ridge. Emies (1990) has discussed the different types of wave drag. According to him wave drag can be splitted into three main components, viz., form drag, wave drag and hydrostatic drag.

Satomura and Bougeault (1994) simulated the airflow over Pyrenees in two lee wave events during PYREX (Pyrenees experiment) programme, using a 2-D, non-hydrostatic, compressible model. In both the cases the mean flow was initialized by a single upstream sounding. The bottom topography used in the model was taken from the real mountain profile after smoothing out the wavelengths shorter than 10km. In both the cases the simulated downward momentum fluxes agreed well with the observed fluxes around 4km height. Simulated flux was constant almost throughout the troposphere, whereas the observed flux decreased with height above 4km. This overestimation of simulated flux above 4km was attributed to the overestimation of the amplitude of long mountain waves at these heights. Again this overestimation of the amplitude of long mountain waves was attributed to the time evolution of the mean wind and the lateral momentum flux divergence found in the real atmosphere.

Broad (1995) have argued that since in general, the environmental wind velocity turns with height, the critical levels exist continuously in the vertical direction. At the critical level corresponding to a wave vector, the component of the momentum flux vector projected on this wave is absorbed, as a result of which momentum flux vector is azimuthally filtered continuously in the vertical.

Shutts (1995) demonstrated analytically that, except for the azimuthal filtering, the momentum flux vector for an environmental flow with a constant shear is almost the same as that for a uniform environmental flow.

Vosper and Mobbs (1998) derived a 3-D wave action equation for linear IGW. Using this equation, in the absence of the turning of the basic flow with height, they showed that for steady flows in the absence of dissipation, the vertical flux of both horizontal components momentum is independent of height. They showed that when basic flow turns with height then the above result does not hold good.

Satomura and Sato (1999) studied the simulation of the generation of new secondary gravity waves associated with the breaking of primary mountain waves in the lower stratosphere using a 2-D non-hydrostatic model. The wave parameters of secondary waves found by them were 3-8km and 2-20km for horizontal and vertical wavelengths respectively, and phase velocities were –1.5m/s and 4.0m/s in the horizontal and vertical direction respectively. In their study they have suggested three mechanisms for the secondary wave generation:

(a) An obstacle effect due to convective motions in the shear flow near the breaking zone,

(b) An unstable normal mode of slowly growing breaking mountain wave,

(c) A mode growing non-linearly at the edge of the breaking mountain wave.

Kanehisa (2000) derived an approximate formula for the momentum flux. He considered a steady, non-rotating, hydrostatic flow across a 3-d mountain. The basic flow taken by him had vertical shears of both magnitude and direction. His result confirmed the earlier findings of Broad (1995), Shutts (1995).

In India, Dutta (2001), Dutta and Naresh Kumar (2005) and Dutta (2007) addressed the issue of momentum flux and energy flux associated with internal gravity wave resulted from airflow across mountainous region along west coast of India and in the north east region. In the former two studies, vertically averaged wind and stability was used, where as in the last study realistic vertical variation of them was considered. Both Dutta (2001) and Dutta and Naresh Kumar (2005) addressed

this problem following 2-D approach. Dutta (2007) addressed this problem following 3-D approach.

Dutta (2001) studied momentum/energy flux associated with mountain wave across Mumbai-Pune section of the Western Ghats in an idealized air stream. He showed that both the fluxes were independent of height and the half width of the bell shaped part of the barrier. Further it was shown that the plateau portion of the section does not contribute to the above fluxes.

Dutta and Naresh Kumar (2005) offered a simple model for parameterizing momentum flux and energy flux associated with internal gravity wave across the Assam-Burma hills (ABH). They considered vertically averaged basic flow normal to the ridges of ABH with vertically averaged static stability. Salient features from the study of Dutta and Naresh Kumar (2005) are given below:

- Wave momentum flux is vertically downward and wave energy flux is vertically upward for a vertically propagating mountain wave across the Assam-Burma hills (ABH).
- It is found that the directions of vertical fluxes across the valley between the two ridges of the ABH are opposite to those across the two ridges and across the entire ABH.
- A long valley (length of the valley exceeds the sum of the half widths of individual ridges, separated by the valley) acts as a source in the atmospheric momentum budget and as a sink in the atmospheric energy budget.
- The downward momentum flux (averaged over a length of 100 km along the flow) across the entire ABH varied between 4.5 – 10.0 Nm^{-2} and that across the two ridges varied between 2.0 - 7.0 Nm^{-2} and upward momentum flux across the valley varied between 0.5 - 1.2 Nm^{-2} and upward energy flux (averaged over a length of 100 km along the flow) across the entire Assam- Burma hill varied between 45.0 - 180.0 Wm^{-2} and that across the two ridges varied between 20.0 - 125.0 Wm^{-2} and downward energy flux across the valley varied between 5.0 - 21.0 Wm^{-2}.
- For the ABH, magnitude of both the fluxes increases with height but decreases with half with of the individual ridges.

Dutta (2007) offered a sound dynamical model to parameterize the vertical energy flux (E_z) and the horizontal components of momentum flux (τ_{zx} and τ_{zy}) associated with orographically excited internal gravity waves (IGW) in a baroclinic background flow. This model was based on a linear dynamical model Dutta (2005) for airflow across a three-dimensional meso-scale elliptical barrier. This model can be used for an arbitrary orographic barrier, although in the present study an elliptical barrier has been considered. Realistic vertical variations in the stability have been incorporated in the model. The solution of model has been obtained quasi numerically for some selected cases near the Indian mountains [The Western Ghats (WG) and the Khasi-Jayanti (KJ) hills]. The vertical profile of energy flux and momentum fluxes for typical cases over the Western ghats and that over the Khasi-Jayanti hills are shown in fig ().The salient results from this study are:

(i) The proposed model can compute the energy flux (E_z) and the horizontal components of momentum flux (τ_{zx} and τ_{zy}) associated with orographically excited gravity waves at each level for any barrier.

(ii) Computations for the selected cases show that the fluxes vary in the vertical. The vertical variation is not uniform with height. It depends on the vertical profiles of the wind and temperature of basic flow for respective cases. However, the influence of the '*V*' component (parallel to major ridge axis of elliptical barrier) of basic wind has been observed in all cases.

(iii) For the WG, τ_{zx} is invariant in the vertical in absence of '*V*' component. Other fluxes are invariant in the vertical in some layers. Whether this component is present or not, an upper layer is found where all the vertical fluxes are non divergent.

(iv) For the KJ hills, in absence of '*V*' component τ_{zx} and τ_{zy} both are invariant in the vertical; presence of '*V*' component makes τ_{zx} and τ_{zy} divergent/convergent.

10.8. Influence of orography on rainfall

Orographic precipitation processes strongly shape the climate in and around mountainous regions. Orographic influences can be pronounced on spatial scales ranging from the size of individual hills to the scale of major mountain ranges, and on

temporal scales from the duration of a brief snow squall to the long-term climatology. Almost all orographic influences are fundamentally caused by topographically driven ascending and descending atmospheric motions that force condensation and evaporation. However, these basic forcing combine with a wide range of dynamical and microphysical processes to shape the precipitation distribution. Since different physical processes can be important for different storms and for different mountain ranges, orographic precipitation influences may take many forms. Characterizing and understanding the effects of topography on precipitation remains an active field of research.

Banerjee (1929, 1930) emphasized the effect of mountain ranges on the distribution of rainfall. Assuming that all the important mountain ranges stand as walls of equivalent height normal to its surface, he tried to explain the influence of Indian mountain ranges on the southwest monsoon current. However, the precipitation in mountainous regions may also be due to several other causes, viz., horizontal convergence and convective instability. Qualitative evaluation of rainfall in mountainous regions is a very complex problem as it involves different physical and dynamical aspects of the processes of different scales. First, there are synoptic scale factors, which determine the characteristics of the air stream, which crosses the orographic barrier, its speed, direction, thermal stability and moisture content. Second aspect is the microphysics of the rain and cloud forming process, Third aspect is the dynamics of the airflow over and around the barrier, which determines the distribution of vertical velocity and displacement of the air stream at each level and along the entire path.

Douglas and Glauspoole (1947) investigated the role air mass characteristics and large scale synoptic factors in orographic precipitation. Ludlam (1955, 56) studied the microphysics of the cloud and the rain, which determine whether the water vapour, which is condensed as cloud, will reach the ground as rain or snow or it will be re-evaporated on the leeward side.

Murray (1948) gave a very simplified model, based on the assumption that the saturated air is lifted enmass over the Khasi and Jayanti hills in India. The order of rainfall intensity computed by him agreed with observed value.

Sawyar (1956) considered the rainfall over British Isles using simplified assumptions that the air due to orography is lifted at all levels and to the same extent. On these assumptions he computed rainfall and compared with observed rainfall over the Welsh mountain. However realistic distributions of vertical motion were not made in their studies.

Sarker (1966, 1967) for the first time proposed a sound 2-D dynamical model of orographic rainfall based on theories of airflow over mountains. To compute rainfall using vertical velocity, the conservation of mass and moisture were considered. The model gives the amount of rainfall due to forced orographic lifting and also accounts for the variation of rainfall along the slope. The agreement between the observed rainfall and that computed by him over the Western Ghats during southwest monsoon season was remarkable.

De (1973) applied this model over Khasi and Jayanti hills in Northeast India. The orographic rainfall, computed by him was in good agreement with the observation. Sarker et.al (1978) had modified the earlier model of Sarker (1966, 67) by incorporating the modified vertical velocity for different southwest monsoon conditions.

Smith (1979) suggested three independent mechanism of orographic rainfall, viz. large-scale upslope precipitation, enhancement of rainfall over small hills and orographic control of the formation of Cumulonimbus (CB) clouds in a conditionally unstable air mass.

Browning (1980) discussed the structure, mechanism and prediction of orographic rainfall in detail with special reference to south Wales. According to him, rain over hills of small or medium size can be very large for a strong and also saturated low-level airflow.

Grossman and Duran (1984) studied the influence of the Western Ghats, in India, in the formation of deep convective cells away from the mountain crest on the windward side of the Western Ghats. He suggested that principal areas of

convection occur off the western coast of the mountains of India, Burma, Thailand and Philippines.

Alpert (1986) computed orographic precipitation assuming that orographic precipitation approximately equals the convergence of moisture in the mountainous boundary layer. He also found that rainfall rate distribution was sensitive to the horizontal wind strength, surface temperature, and temperature lapse rate and specific humidity lapse rate.

Watnabe and Ogura (1987) made a case study of heavy precipitation in the western parts of Japan. He found that during the spells of heavy rainfall, convective cells form in succession over the sea about 50 km off the coast. As they moved eastward and approached the coastline, they developed rapidly and organized into a band structure. He applied the one layer model developed by Danard (1977) to investigate the topographic effect. The model result indicates that the surface flow over land is deflected mainly by the effect of topographic barrier and partly by increased surface friction over land. His result shows that even a mountain range of modest height could enhance precipitation significantly.

Sinha Ray (1988) developed a dynamical model for the perturbation vertical velocity, induced by an orographic barrier, including friction. He solved the model numerically and used this perturbation vertical velocity induced by an orographic barrier to compute the orographic rainfall. The orographic rainfall computed by him was in good agreement with the computed rainfall from earlier models and also with observed rainfall.

Doer and Choularten (1992) used a three dimensional model of airflow developed by Carruthers et.al (1989) and the orographic rain enhancement model of Caruthers and Choularton (1983) was incorporated into the airflow model. His result shows that in conditions of either stably stratified air or with inversion layer aloft, three-dimensional airflow effects can lead to much reduced rainfall over a bell shaped conical hill. He also found that three-dimensional effects were particularly prominent in the presence of an upper layer inversion. His result shows that maxima in the rainfall rate across the central axis of the hill occur just to the lee of the peak of

the hill, which is in contrary to the observation and the results obtained from 2-D study that maxima in the rainfall rate across the central axis of the hill occur just to the windward side of the hill.

In investigating heavy orographic rainfall events over the Southern Alps, Buzzi and Foschini (2000) found that the following synoptic and mesoscale environments are conducive to heavy orographic rainfall over Alps: (1) strong low-level confluence over the western Alps between the post-frontal south-westerly flow and the pre-frontal south-easterly flow, (2) an 850 hPa pre-frontal low-level jet (LLJ) which serves as a warm conveyor belt, and (3) a deep trough approaching the Alps with an upper tropospheric, quasi-stationary pressure ridge located to the east. In addition, the Atlas Mountain in North Africa, mountains in Sardinia and Corsica, the Apennines in Italy and the coastal range to the east of the Adriatic Sea have also played important roles in forming LLJ's toward the Alps (Tripoli et al. 2000). Lin et al. (2001) proposed that these features belong to essential ingredients for heavy orographic rainfall, which have also been observed in other parts of the world, such as the U.S. and East Asia.

Kitoh (2004) made a series of coupled general circulation model (CGCM) experiments to study the effects of progressive mountain uplift on East Asian summer climate. Eight different mountain heights were used: 0% (no mountain), 20%, 40%, 60%, 80%, 100% (control run), 120%, and 140%. The land–sea distribution was kept same for all experiments and mountain heights are varied uniformly over the entire globe. Systematic changes in precipitation pattern and circulation fields as well as sea surface temperature (SST) appeared with progressive mountain uplift. In summertime, precipitation area moves inland on the Asian continent with mountain uplift, while the Pacific subtropical anticyclone and associated trade winds become stronger. The mountain uplift also resulted in an SST increase over the western tropical Pacific and the Maritime Continent and an SST decrease over the western Indian Ocean and the central subtropical Pacific. There is a drastic change in the East Asian circulations with the threshold value at the 60% mountain height. With the mountain height below 60%, the southwesterly monsoon flow from the Indian Ocean becomes strong by uplift and transports moisture toward East Asia, forming the baiu rainband. With higher mountain heights,

intensified subtropical trade winds transport moisture from the Pacific into the Asian continent.

Chen and Lin (2004), using the Weather and Research Forecast (WRF) model, made idealized numerical simulations. Three flow regimes, based on the moist Froude number, were identified for a conditionally unstable, rotational, horizontally homogeneous, uniformly stratified flow over an idealized, three-dimensional, mesoscale mountain stretched spanwise to the impinging flow, viz. (I) a quasi-stationary upslope convective system and an upstream-propagating convective system, (II) a quasi stationary upslope convective system, and (III) a stationary upslope convective system and a quasi- stationary downstream convective system. One important finding of this study is that relatively strong mean flow produces a quasi-stationary mesoscale convective system (MCS) and maximum rainfall on the windward slope (upslope rain), instead of on the mountain peak or over the lee side. They found that the Coriolis force helps produce heavy upslope rainfall by making transition from ‘‘flow-around’’ the eastern part of the upslope to ‘‘flow-over’’ the western part of the upslope (transits to a higher flow regime) by deflecting the incident southerly flow to become east–southeasterly barrier winds. A lower-Froude number flow tends to produce a rainfall maximum near the concave region. Several other important facts can also be found in this study. The ratio of the maximum grid scale rainfall to the sub-grid scale rainfall increases when the moist Froude number increases. This study suggests that, besides Froude number, CAPE might play an important role in determining moist flow regimes.

In India, Dutta (2007) first time proposed a 3-D meso scale dynamical model for orographic rainfall. This model has two parts, namely, a dynamical part and a thermodynamical part. In the dynamical part the vertical velocity induced by a mesoscale elliptical orographic barrier has been computed using the perturbation technique. In the thermodynamical, part rainfall intensity (RFI) has been computed using the computed vertical velocities, with the help of continuity of moisture and mass. To study the sensitivity of orographic rainfall enhancement to atmospheric static stability and vertical shear of horizontal wind, a number of numerical experiments were carried on with different atmospheric static stability and vertical shear of horizontal wind. Salient results from this study are:

- In the longitudinal vertical plane, the maximum rainfall intensity takes place about 5-10 km behind the orographic peak.
- The computed RFI has been compared with observed RFI as well as with that computed by 2-D model.
- The spatial distribution of RFI across the barrier shows that there are four regions of maximum rainfall, one primary on the windward side behind the peak of the barrier and three secondary on the leeward side.
- The symmetry in the locations of these secondary rainfall maxima appears to be critically dependent on the component of basic flow parallel to the major ridge axis of the barrier.
- During the southwest monsoon season (SWMS), orographic rainfall enhancement in the WG area appears to be solely due to the vertical shear of the basic flow and its variation with height. Stability appears to have very little influence on it.

In a recent study by Dutta et al. (2007), dynamical role of the Western Ghats on the rainfall enhancement over Pune, in presence of a cyclonic vortex over Bay of Bengal, has been examined. To examine it, an already developed dynamical model of airflow over a meso-scale barrier has been used. Six cases have been studied, in which there is an enhancement of rainfall over Pune in presence of vortex over Head bay during southwest monsoon months. The study shows that:

- In many occasions the WG plays some dynamical role in the 'Distance effect' of vortex over bay on rainfall enhancement over Pune.
- In four, out of six cases, studied, the model has captured, at least qualitatively, the fluctuations in the observed rainfall over Pune, during such period.
- In these four cases westerly along west coast was strong with considerable depth in vertical and the vertical profile had two maxima.
- In one case model has failed to capture observed fall in rainfall intensity, although it has captured the observed rise in rainfall intensity. In another case model has failed completely, even qualitatively, to capture the observed fluctuation in rainfall intensity.

- It appears from the study that the strength and depth of westerly at Santacruz along with double maxima in the vertical profile of westerly at Santacruz are favourable for the dynamical role of the WG in the 'Distance effect' of vortex over the Bay of Bengal on rainfall enhancement over Pune.

10.9. Effect of mountain in releasing the convective instability

It has been long known that mountains play an important role in the development of deep convection. First, mountains act as an elevated heat sources that makes the air at mountain summits unstable, favouring the formation of convective cells. Moreover, in response to the sustained heating at high elevations, horizontal pressure gradients are developed with lower pressure over the heated mountains than air at the same level, some distance away from the peak. This induces air to flow up the heated slopes and converge at the tops in a heated chimney, which provides moisture convergence as well.

Bonacina (1945) surveyed the regions of intense orographic rain around the world and the possible mechanisms. He concluded that orographic rain does not occur every time an air-stream impinges on a mountain, but rather that air-stream must have been conditioned by the prevailing synoptic situation. In particular he emphasized the importance of convective instability for the generation of intense orographic rain.

Mukherjee and Ghosh (1965) suggested a possible role played by Khasi-Jayanti hills in triggering pre-monsoon convective instability over Brahmaputra valley at night. They argued that during night only, when KJ hills cool down more rapidly than the plane, strong Katabatic wind blows down the slope towards the valley, which is being obstructed at least up to 1.5km at daytime. This in turn brings down southerly/south westerly moist air on the valley, where already easterly/east northeasterly dry air originating from Tibetan plateau prevails. So, a front like structure developed over the valley at night, which causes convective activity to take place at night.

In India, Sarker (1966,1967), De (1973), have indicated that in such occasions in addition to forced orographic ascent, synoptic scale convergence, convective instability are also very important factors for heavy to very heavy precipitation rate on the wind ward slope.

From the study of Browning et.al (1974) it appears that convective cells and meso-scale precipitation areas may be accentuated by orography. They are persistent over six hours or more and account for much of heavy precipitation.

Mukherjee and Ramana murthy (1978) studied the contrasting rainfall features and associated thermodynamic behaviour over Mumbai, lying on the windward side of the coastal mountain Western Ghats along west coast of India. They found that higher rainfall days are associated with higher static stability, less convective instability and higher precipitable water content (PWC). They also found that differences in static stability and PWC between consecutive days of contrasting rainfall were not significant.

Smith (1979) suggested that convection triggered by smooth orographic ascent brings the air to saturation and after some delay, raindrops form and fall to the ground. He also suggested that heating of the mountain slopes by insolation causes upslope winds leading to thermals above the mountain peak. Smith (1979) suggested the two important factors which can determine whether the orographic precipitation is convective or non-convective and whether orographic precipitation maxima is at the summit or on the windward slope:

i. The weather type (or season), which can determine the relative importance of stable, verses convective rain.

ii. The size of the mountain which determines whether the orographic precipitation will occur on the upwind slope with a rain shadow in the lee (i.e., larger mountains L>100km), or with the maxima more nearly at the summit (i.e., smaller mountains L<20km).

Now whether the precipitation is convective or stable and whether precipitation maxima is on the windward slope or on the summit, there must be a mechanism to lift a moist air parcel to the lifting condensation level (LCL). Lifting may be due to:

(a). Low-level synoptic scale convergence and/or upper level divergence.

(b). Convective instability as suggested by Smith (1979)

(c). Frontal up glide

(d). Forced orographic lifting

Grossman and Duran (1984) studied the influence of the Western Ghats, in India, in the formation of deep convective cells away from the mountain crest on the windward side of the Western ghat. They attributed the discrepancies between observed coastal rainfall and that computed by Sarker (1967), using dynamical model, to the frequent shallow and occasional deep convection near the coast. Watnabe and Ogura (1987) made a case study of heavy precipitation in the western parts of Japan. He had shown that during the heavy rainfall period convective cells form in succession over the sea about 50 km off the coast. As they moved eastward and approached the coastline, they developed rapidly and organized into a band structure. His results show that a mountain range of modest height could trigger the convective instability to enhance rainfall significantly along windward slope.

Often the above factors act together. Frontal up gliding is not important at tropics. In 10.4, we have discussed the different studies on forced orographic lifting. Quite often convection is triggered by smooth orographic ascent, which brings the air at LCL to saturation and after some delay, raindrops form and fall to the ground. It is manifested by heavy to very heavy rainfall on the windward slope of mountain.

In section 10.8, it has been mentioned that a dynamical model for computing orographic rainfall has been developed by Sinha Ray (1988). But the observed rainfall rate in many cases was significantly more than the computed rainfall. He also attributed this discrepancy in the rainfall rate observed and computed to the other factors, like synoptic scale convergence, convective instability.

Cotton et.al (1989) studied the effect of the Colorado Rocky on the formation of meso scale convective systems (MCS) to the lee of it. They showed that heating of mountain slopes induces mountain-plane solenoidal circulations responsible for organizing convection on a broad range of scales. These shallow solenoidal

circulations are strengthened and deepened by the systematic latent heating by the clusters of Cumulonimbus.

Quite often it is observed that places along the windward slope of orographic barrier receive heavy rainfall on an isolated day, preceding and succeeding days with comparatively very less rainfall or dry weather. The dynamical model for orographic rainfall, discussed in previous section, cannot explain a significant part of such observed heavy rainfall on isolated days. So, in such cases forced orographic lifting alone is not the only mechanism responsible for such observed heavy rainfall. Consequently, a complex relationship exists which operates through convective instability in the mountainous regions.

De and Dutta (2005) attempted to study the role of convective instability in enhancing orographic rainfall rate. They developed a convective rainfall model. The model had two parts, viz., a dynamical and a thermodynamical part. In the dynamical part, vertical velocity, solely due to buoyancy, was computed and then in the thermodynamic part, using the computed vertical velocity and conservation of moisture/mass, the convective rain rate was computed. Initially this model was applied to a coastal orographic station, viz., Santacruz in Mumbai, India then this model was applied to another orographic place, viz., syllhete in Bangladesh (Dutta 2007). For both places, qualitative agreement of the model with observed day to day fluctuation in daily rain rate was fairly well.

Fuhrer and Schär (2007) investigated the underlying dynamical mechanisms for small-amplitude topographic variations in triggering and organizing banded convection. For that three-dimensional simulations of moist flows past a two-dimensional mountain ridge using a cloud-resolving numerical model were conducted systematically. Most simulations address a sheared environment to account for the observed wind profiles. Results confirm that small-amplitude topographic variations can enhance the development of embedded convection and anchor quasi-stationary convective bands to a fixed location in space. The resulting precipitation patterns exhibit tremendous spatial variability, since regions receiving heavy rainfall can be only kilometers away from regions receiving little or no rain. In addition, the presence of banded convection has important repercussions on the

area-mean precipitation amounts. For the experimental setup in their study, the gravity wave response to small-amplitude topographic variations close to the upstream edge of the cap cloud (which is forced by the larger-scale topography) is found to be the dominant triggering mechanism. Small-scale variations in the underlying topography are found to force the location and spacing of convective bands over a wide range of scales. In their study the small-amplitude topographic roughness was found to trigger banded convection and to control the spacing and location of the resulting bands.

Park and Lee (2007) studied the Synoptic features associated with a record-breaking heavy rainfall event, with a 24-hr accumulated rainfall of 870.5 mm, occurred in a coastal area at the foot of a mountain range in the central-eastern part of the Korean Peninsula during the passage of Typhoon Rusa (2002). High-resolution numerical analysis and forecast fields obtained from the PSU/NCAR MM5 were used for this study. The suggested main causes of this localized heavy rainfall include: 1) strong low-level convergence of moist air from the sea into the coastal/ mountainous area; 2) consequent orographic lifting; 3) low levels of lifting condensation and free convection; and 4) release of potential instability by orographic lifting to trigger deep convection.

Tanaka et al., (2008) performed numerical experiments using a mesoscale meteorological model (MM5) to evaluate the mountainous orographical effects on heavy rainfalls brought by Typhoon 0514 (NABI), which caused a flooding disaster in the southeast Kyushu area of Japan. Three terrain conditions were used, viz., a flat terrain with altitude 1 m above mean sea level; an idealized, line-shaped mountain terrain; and a complex terrain using topography data from the U.S. Geological Survey. Although the total observed rainfall due to Typhoon 0514 was greater than 1,000 mm, the rainfall value calculated using the flat terrain conditions was 250–300 mm; and the value calculated using the complex terrain conditions was 500–900 mm. This discrepancy was attributed to the evolution of convective cells, generated by water vapor lifted along the mountain slope in the windward areas.

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CHAPTER 11

INDIAN SUMMER MONSOON EXPERIMENTS

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11.1. Introduction

Surface and upper air data are regularly collected at meteorological observatories round the globe. Such data provide synoptic view of weather systems, and are useful for understanding climatology on different spatial and temporal scales. However, routine observational data are not adequate to address several processes and interactions that occur in the atmosphere and ocean because the information available through them is not adequate to understand complex physical processes, interactions and feedbacks taking place in nature. For example, turbulent fluxes are parameterized in terms of synoptic data and to develop these (bulk) flux algorithms, measurements are required on micrometeorological towers with multi-level instrumentation that include fast response sensors for wind, temperature and water vapor. Most of the monsoon lows and depressions that give widespread rainfall over the Indian subcontinent originate over the Bay of Bengal. It is expected that formation of the monsoon systems is influenced by the sea surface conditions, that themselves get modified by the physical processes associated with cloud systems. Similarly, atmospheric boundary layer feeds the convective clouds with energy rich moist air during their growing stage, however, precipitating clouds cool and dry the boundary layer through downdrafts and thereby reduce convective instability. Such physical process cannot be addressed fully using routine synoptic observations. Specially designed field expeditions/experiments have to be carried out to fill gap areas in spatial coverage (e.g., open oceans where continuous *in situ* measurements are difficult) and knowledge.

Indian summer monsoon involves complex interactions between land, atmosphere and oceans. Most of the water vapour needed for the monsoon cloud systems come from the surrounding as well as far away oceanic regions. Therefore, to understand the monsoon rainfall on the Indian subcontinent, it is necessary to understand the planetary scale circulation associated with monsoon, cloud systems over the oceans, their link to ocean surface conditions, hydrological feedbacks over land, role of aerosols in cloud formation and precipitation, etc. Understanding such issues calls for specially designed field experiments. Several observational experiments have been carried out in the last 5 decades targeting the Indian summer monsoon (Table 11.1). The motivation came from a desire to understand the planetary scale monsoon system as well as to identify and address outstanding issues that could lead to better prediction of monsoon.

The spatial scale of the monsoon flow was not scientifically explored till India Meteorological Department (IMD, established in 1875) carried out a two-year intensive programme to collect surface observations from land and ocean (ship based) in the so called monsoon area during 1893-94 (Eliot 1986). These observations led to the discovery that at the time of monsoon onset, strong cross equatorial flow develops off the east coast of Africa bringing in moisture from the south Indian Ocean. Then there was a gap of more than 60 years with no major monsoon experiment (two world wars contributed to this). In the last 5 decades, 9 major monsoon observational experiments have been carried out and another one in progress (Table 11.1).

Each of these field campaigns was carefully planned addressing scientific issues and questions considered important at respective time periods. These experiments brought together people from different countries/groups/disciplines to address important issues and problems related to monsoon. Each resulted in a data set that is unique. Basically the field programme provides new and special observations targeting the objectives. Related findings are published within the first few years of the field phase, however, studies continue and many interesting papers have been published several years later. When taken collectively, the published results addressed most of the objectives of the programme. Data collected also

motivated studies some of which were probably not anticipated during the planning stage.

Detailed description of individual monsoon experiments is not attempted here, nor a comprehensive reporting of results. Instead the focus in the following pages is to give a flavor of how some international and national monsoon experiments were planned and their contributions towards the advancement in our understanding of the monsoon phenomenon and processes.

11.2. The International Indian Ocean Expedition (IIOE, 1960-1965)

Indian monsoon is part of the planetary scale circulation, and phenomena in remote places affect rainfall over the subcontinent through teleconnections. Early monsoon experiments are no exception, and they were influenced by the interactions among scientists working in places far away from India. (*Many materials given below are taken from UNESCO reports and other articles found on web; in some cases even authorship is not clear. See section B in bibliography for the details.)* IIOE was conceived and initially planned in the USA, and started as an ocean observational programme in late 1950s. Indian Ocean (IO) covers over 14% of earth's surface and more than a quarter of the world's people lived in countries surrounding the Indian Ocean. Population growth called for expanding fisheries. There was fragmentary evidence then for unusually high productivity in north IO. Seasonal reversal of winds was known but not its effect on currents and organisms in water. No systematic study had been attempted, reported observations did not give more than a preliminary picture of behavior and characteristics, and IO was an unexplored frontier. Initially IIOE started with three major objectives. First, to know the IO potential for fishery resources, since most of the countries bordering IO were deficient in proteins in their diet; second, to assess the role of the northern IO in effecting the monsoonal changes, which are vital for agricultural operations in the Indian sub-continent, but which also influence the current patterns, upwelling systems, productivity and the carbon-dioxide cycle; and third, to determine the limits to the use of the oceans for dumping human wastes, including spent nuclear fuels, etc. It was also suggested that, during the first two years of the expedition, participating countries should encourage standardization of equipment and methods

of analysis and data logging so that the results obtained by different ships would be comparable.

IIOE evolved receiving comments, suggestions and inputs from many scientists. For example, R. B. Montgomery wrote: '…this programme can be so designed as to aid directly the development of one or more oceanographic centers in the countries bordering the Indian Ocean. Active oceanographic centers on the Indian Ocean are essential for continuing intensive studies.' in *Indian Ocean Bubble* (Issue 2; 27 Feb 1959), a magazine that was started as a forum to discuss IIOE. The birth of National Institute of Oceanography, Goa was a result of such a thinking. Emphasis was also placed on involving people who are genuinely interested in the problem. For example, Gene LaFond wrote 'To spread the gospel and attain any lasting results, the work has to be carried on by the scientists of Indian Ocean area. This does not mean just coming along for a ride but actually given a major share in planning, analysis and reporting. As a final remark, I would like to suggest that the overall programme be planned by the people who will take the ship to the sea and then use the data.'

A large number of ship cruises were planned covering the entire IO north of 40°S (Fig. 11.1), and ship cruises started collecting data from late 1960. Peak activity was planned in 1962-63 and expeditions continued in 1964 as well. The international programme called for the study of the entire system from below the bottom, through the water itself (biological, chemical and physical characteristics), through the boundary between sea and the atmosphere and on upwards to the upper atmosphere. Since nothing much was known about IO, more emphasis was given in the initial stages to collecting the basic data. Continuous revision and reexamination of the plans was envisaged as new data became available, and IIOE programme evolved with time. IIOE Working Group initially targeted to study (a) Physical oceanography, (b) Chemical Oceanography, (c) Meteorology to obtain understanding of the energy exchange between sea and atmosphere, (d) Marine biology, and (e) Marine geology and geophysics. Initially there were some discordant voices too, doubting whether such a large programme was feasible. Next five years proved the pessimists wrong and IIOE became a pioneering example of planning and execution of international scientific programmes.

The IIOE officially ended in 1965. More than 40 oceanographic research vessels belonging to 13 countries surveyed the Indian Ocean and collected useful data in almost all disciplines in the marine sciences. Since the completion of the expedition, hundreds of papers have been published and some of them reprinted and included in the 8 volumes of collected reprints of the International Indian Ocean Expedition published by UNESCO. With a large number of cruises and detailed observations, IO would become the ocean with maximum information on ocean physical, chemical and biological characteristics at the end of IIOE. Based on data and findings a set of atlases was published for the Indian Ocean which is unique for a field programme (Table 11.2).

India played a significant role in the overall operations and co-ordination of the IIOE, as well as being an active participant. With the inauguration of the 1st Scientific Cruise of INS *Kistna on 9 October 1962,* the Indian Programme of Work during the IIOE was officially launched. Besides INS *Kistna, the Indian Programme* included scientific cruises by RV *Varuna, of the* Indo-Norwegian project, RV *Conch, of the* University of Kerala, and FV *Bangada, an exploratory* fishing vessel of the Ministry of Food and Agriculture, Government of India. All the cruise tracks and programme of work were coordinated so that a better coverage of important coastal areas in the Bay of Bengal and the Arabian Sea could be achieved.

IIOE and the Indian Monsoon

IIOE was initially an ocean programme and monsoon was included mainly for its effects on ocean currents and energy budget. Realizing the potential of gathering meteorological data over the ocean by becoming a partner, meteorological community joined IIOE and evolved a detailed observational component subsequently. Participating nations and the International Association of Meteorology and Atmospheric Physics established working groups to design meteorological programmes to be integrated with the oceanographic expeditions. Detailed plans were worked out in a meeting at Bombay in July 1961 that included the following.

(a) Establishment of International Met Centre in the Indian region

(b) Make immediate and practical use of vast data from ship expeditions, island and shore stations

(c) Data quality control, analyse, process and do research

(d) Train people.

(e) Study of the monsoon addressing (i) how does the reversal of the monsoons affect the oceanic circulation in the northern Indian Ocean, and (ii) how may the onset and intensity of the southwest monsoon be predicted.

It was hoped that IIOE would help in finding answers to the following issues/ problems in physical oceanography and meteorology, many of which are relevant even 5 decades later.

a. How can the large scale atmospheric circulation of Indian Ocean be described?

b. How can evaporation and vertical heat flux be determined over large areas?

c. What is the areal distribution of vertical flux of water vapour, heat and momentum?

d. What are the local interface fluxes at the air sea boundary?

e. What is the average energy transfer or heat budget for the atmosphere and the ocean?

f. How do we interpret satellite cloud photographs and infra-red measurements in terms of such meteorological observations?

IIOE benefitted the meteorological activities in India in a major way with the establishments of an Indian Ocean Expedition Directorate, the International Meteorological Centre at Bombay and the Indian Ocean Biological Centre at Cochin, all under the Council of Scientific and Industrial Research. IMC started functioning from January 1962 with Prof. C. S. Ramage of the University of Hawaii as Director and with the help of IMD personnel. An IBM 1620 computer for data processing was contributed by international agencies. Meteorological and satellite (*Tiros VII and Nimbus)* data reception facilities were established at IMC. On a typical day the total coverage was like: Surface reports - 1155; Ships - 384; Upper air - 429. Aircraft reported from long-distance international flights on three or four air routes. At the IMC, synoptic charts were prepared for two principal times - 00 and 12 hours GMT - for surface and standard isobaric levels; namely, 50, 100, 200, 300, 500 and 700mb.

Scientific contributions of IIOE

IIOE was a landmark experiment and revolutionized oceanographic and monsoon research in India in several ways. First, it accumulated a vast quantity of information whose analysis improved our knowledge of the circulation pattern of the monsoon winds and its influence on the ocean. In particular, great influence of conditions in the sea on the weather pattern of the Indian sub-continent along with the see-saw game played by the monsoon winds with the surface currents became evident. During the southwest monsoon season, the Somali current near 8°- 10°N up to 7 knots. Its inner edge was close to the Somali shore, where the temperature of the surface waters was sometimes I6°C or less while the rest of the Arabian Sea was at nearly 30°C. It was also noticed that a strong set of the Somali current preceded the onset of the southwest monsoon along the west coast of India by almost a month, thereby indicating a possible relationship between the two events.

IIOE led to new discoveries including the low level jet during monsoon (Joseph and Raman 1966, Findlater 1969), strong atmospheric inversions over the western and central Arabian Sea (Colon 1964). Wind speeds as high as 60-70 knots were observed near the equatorial region off the coast of Africa at height between 1 and 2 km (Findlater 1969). Width of the core is about 100 nautical miles and depth less than 1 nautical mile. The low level jet induces strong upwelling along the east coast of Africa and also has a strong influence on the air-sea fluxes over the Arabian Sea. The IIOE data also enabled the estimation of transport of moisture across the equator and into the Indian landmass and evaporation over the Arabian Sea (Sikka 2005). Dropsonde data showed strong low level inversions in the atmosphere with the height of the inversion about 500 m off east coast of Africa gradually increased eastward of 65°E and vanished as the flow approached the Western Ghats (Colon 1964). The Arabian Sea inversion is basically caused by the flow of hot and dry air (originating from the surrounding deserts) over a relatively cooler Arabian Sea water (Fig. 11.3).

Another important finding from IIOE data is the mid tropospheric cyclone (MTC) in the region near the west coast of India (Miller and Keshavamurty 1968). MTC is a copious rain producing transient system whose signature is not prominent

in the surface pressure chart. Maximum convergence and maximum vorticity is observed in 700-500 hPa layer. It has been argued that MTC is an instability phenomenon of a special class of baroclinic basic state created by factors such as westward moving depressions/low pressure systems from Bay of Bengal, heat low over Pakistan and Tibetan low play a role (Mak 1975).

The heavy rainfall associated with MTC suggests that vertical velocities are strong. Examination of the vertical section in north-south direction of the temperature field associated with MTC (e.g., Fig. 11.4) shows slightly warmer temperatures in the mid troposphere but not the kind of temperature gradient in 700-500 hPa layer (e.g., those associated with mid-latitude cyclones) that is likely to cause strong convergence and thereby updrafts. Therefore, there must be other driving mechanisms as well. An instability in the tropical atmosphere that is known to cause strong updrafts is moist convective instability, one measure of which is convective available potential energy (CAPE). CAPE should be positive for convective instability since, (theoretically) vertical velocity is proportional to the square root of CAPE. For air parcels originating above 800 hPa, values of CAPE are either very small or absent in the tropics (e.g., Emanuel 1994). As an example, Fig. 11.5a shows values of CAPE for parcels of air lifted from different levels in the atmosphere for Singapore for July 2010 and 2011. It is observed that CAPE is high for air below 900 hPa and decreases rapidly as the height of parcel origin increases (i.e., pressure decreases) and CAPE is either absent or negligible above 700 hPa. CAPE for the corresponding period based on Mumbai soundings are shown in Fig. 11.5b. For Mumbai, CAPE decreases from 1000 hPa to 700 hPa, but then picks up again and is positive between 600 hPa and 400 hPa with a secondary peak around 500 hPa. (It may be noted that Mumbai too follows Singapore trend on many days with no CAPE above 700 hPa, and CAPE seems to be positive in the mid troposphere mainly on days when convection is active.) Goa also shows a trend that is similar to Mumbai but CAPE values are somewhat smaller (Fig. 11.5c). This means that air converging in the mid-troposphere in the region where MTC is observed can rise as individual convective cells rather than being restricted to baroclinically driven stratiform clouds (sloping convection) where vertical velocities are typically a m/s or less (Ludlam 1981). In contrast, vertical velocity can reach several m/s if not more through

convective instability. Perhaps this is the reason why MTC can produce very heavy precipitation.

11.3. Monsoon Experiment of 1979 (MONEX)

MONEX was planned under the First GARP (Global Atmospheric Research Programme) Global Experiment (FGGE) programme with the main objective of understanding planetary scale monsoon circulation (e.g., Murakami 1979). It evolved over a period of six years through four international planning meetings (Yerevan, USSR, 1973; Singapore, 1974; New Delhi, 1977; Kuala Lumpur, 1978). It addressed winter monsoon system in the South China Sea region and adjacent island and coastal areas (December 1978 to February 1979, came to be known as winter MONEX or WMONEX) and the summer monsoon system over the Indian subcontinent (May-August 1979, summer MONEX or SMONEX). MONEX observations extended from 30°N to 20°S and from 35° to 150°E (Fig. 11.6a). In the following MONEX and SMONEX mean the same.

As precursors to MONEX, India and Russia carried out monsoon experiments in 1973 and 1977. The Indo-Soviet Monsoon Experiment (ISMEX-73) was conducted during the period 15 May to 10 July 1973, by the USSR and India. Six vessels collected meteorological and oceanographic measurements over the Arabian Sea and the equatorial and southern Indian Ocean. The results of ISMEX-73 provided useful background for the development of plans for Monsoon-77 and MONEX. The Monsoon-77 experiment had ships taking time series observations with a polygon formation in the western and eastern sectors of the Arabian Sea during the summer of 1977. Later part of MONSOON-77 involved four USSR ships forming a polygon over Bay of Bengal centered at 17°N, 89°E during 11–19 August 1977. A trough formed in the North Bay on 16 August and developed into a depression on 19 August. ISMEX-73 and MONSOON-77 observations provided important insights into the onset of the monsoon, active and break monsoon periods, and oceanographic phenomena.

The major objectives of SMONEX were as follows (Murakami, 1979, Fein and Kuettner 1980).

1. Planetary scale aspects: Investigation of heat sources (determination of the components of heat sources and sinks at the earth's surface, within the atmosphere, and at the top of the atmosphere, to evaluate their role in the annual cycle of the planetary-scale monsoonal circulations), monsoon onset and breaks in monsoon.
2. Synoptic scale regional aspects:
   Arabian Sea Component – low level jet and boundary layer inversion, MTC and monsoon onset mechanism.
   Bay of Bengal component - development and structure of monsoon depressions, structure and energetic of monsoon troughs and prediction of monsoon depressions.
3. Interactions with atmospheric circulation in Pacific, southern hemisphere, northern latitudes, and stratosphere (i.e., monsoon teleconnections)
4. Numerical simulation and prediction - global modeling and regional forecast problems.

It may be remarked here that numerical modeling approach to the understanding and prediction of monsoon became an objective of MONEX, an activity that was not envisaged during IIOE. Proposed numerical experiments included four categories: predictability, controlled experiments to investigate individual physical processes, observing system simulation experiments, and theoretical model experiments (Murakami, 1979).

There was an ocean component as well with focus on the Somali and equatorial currents and their associated upwelling which play an important role in the ocean atmosphere feedbacks during monsoon. There was an ongoing oceanographic programme, namely Indian Ocean Experiment (INDEX) with several ship cruises planned in the equatorial Indian Ocean and Arabian Sea. MONEX took advantage of INDEX in addressing the ocean component.

SMONEX held field experiments in 3 phases with bases at Dhahran (Saudi Arabia experiment, early May), Bombay (Arabian Sea Experiment, mid May to late June) and Calcutta (Bay of Bengal Experiment, July-August). The upper air network was strengthened in the MONEX region that included the addition of eight

radiosonde/ radiowind stations in India. Special observational systems consisted of research aircrafts, ships, mini-sonde upper air stations, radiometer sondes, weather radars, constant level low altitude balloons, boundary layer measurements, and rocket soundings supplemented by geostationary and polar orbiting satellite data from USA, Russia and Japan. USA provided Electra (NCAR), P-3 (NOAA) and CV-990 (NASA) research aircraft, and India its AVRO (NRSA). Cloud microphysics, aerosols and trace gases, multispectral radiation were measured and about 600 dropwindsondes were released. Sixteen ships belonging to 7 nations were utilized. Figure 6b schematically shows some of the observational systems deployed during SMONEX. INDEX cruises helped in increasing the areal coverage of surface observations over the Indian Ocean and Arabian Sea.

Some important scientific contributions of SMONEX.

a. Voluminous data collected provided a four dimensional description of the planetary scale 1979 summer monsoon (Krishnamurti 1985).
b. Description and evolution of monsoon heat sources including the role of Tibetan Plateau, and land sea contrast.
c. Capturing the development of monsoon onset vortex over the Arabian Sea, understanding the roles of differential heating and wind shear (Krishnamurti et al 1981).
d. Better estimation of water vapour flux showing importance of cross equatorial transport and evaporation over Arabian Sea (Greco 1984).
e. Extensive use of surface and satellite data to understand air-sea interactions during monsoon, including the rapid cooling of SST over the Arabian Sea following the monsoon onset.
f. Application of GCMs for monsoon studies and prediction, and model sensitivities to parameterizations.
g. Energetics of monsoon circulation (Krishnamurti and Ramanathan 1982).
h. MONEX marked the beginning of atmospheric boundary layer (ABL) measurements in India (Holt and Raman 1987).
i. It motivated studies leading to better understanding of west coast rainfall and effect of orography on rainfall (Grossman and Durran 1984, Mukherjee et al. 1984).

j. Combining ISMEX-73, MONSOON-77 and MONEX-79 observations, the response of the upper ocean to atmospheric forcing during monsoon could be examined in detail (Rao 1990).
k. Atmospheric heat source due to deep convection (Mohanty and Das 1986).

11.4. Post-1980 Monsoon Experiments

Following the global trends, numerical models gained popularity in India during 1980s as research and weather forecasting tools. Accurate representation of the near surface processes and ABL in the models became important. Over the vast areas of the Indian landmass, there had been no organized efforts to study ABL. Surface fluxes and ABL parameterizations used in the numerical models were based on measurements carried out in mid and high latitudes and not in the monsoon region. It is not a priori clear that these relations are applicable to the Indian conditions, particularly over the monsoon trough and needed to be validated (Sikka and Narasimha 1995). MONTBLEX and LASPEX were carried out to fill this gap. (In the following, many materials have been taken from the review article by Bhat and Narasimha (2007) more or less addressing the same topic.)

11.4.1. MONTBLEX

In his classic memoir of 1886 on *The Rainfall of India*, Blanford (1886) noted the existence of a 'barometric trough which runs obliquely across Northern India, and is the chief seat of the convective ascent…'. The trough extends from Rajasthan and Pakistan in the west to the head Bay of Bengal in the east (Fig. 11.7); its position is closely associated with rainfall patterns in India (Rao 1976). The eddy fluxes in the trough region can play a crucial dynamical role, and their proper parameterization has the potential to improve model simulations of the monsoons. A study of ABL in the trough region therefore seemed highly worthwhile. A project proposal for the field experiment MONTBLEX, involving 20 Indian institutions, was approved by DST in 1988.

Main objectives of MONTBLEX were,

(a) the description of the structure of the ABL across the entire extent of the trough
(b) the study of eddy fluxes and energetics
(c) the formulation of better parameterization schemes for the boundary layer for use in atmospheric general circulation models
(d) to understand the interactions between the moist, ascending eastern end of the monsoon trough with the dry subsidence regime at its western end.

A pilot experiment was conducted at Kharagpur in July 1989, and was followed by the full field experiment in 1990. The main phase of MONTBLEX included observations from four 30 m surface layer masts, respectively at Jodhpur, Delhi, Varanasi and Kharagpur (Fig.11.7); ocean cruises by *ORV Sagarkanya*; observations over IMD network; sodar and tethersonde measurements; and extensive aircraft flights carried out by the Indian Air Force. A comprehensive account of the results from the various investigations carried out was published in 1997 (Narasimha et al 1997).

The major outcome of the experiment has been the acquisition of good data on momentum and heat flux by sonic anemometers, and the effort that followed at devising new parameterizations. The bulk aerodynamic coefficients showed an appreciable increase as wind speed decreased (Mohanty et al (1997) and Rao et al (1997)). With the data segregated on a stability parameter like the flux Richardson number, showed that Monin-Obukhov (M-O) theory could be in considerable error at sufficiently low speeds (Rao et al 1996). Rao and Narasimha (2006) provided a new framework for parameterizing eddy fluxes in the low-wind convective regime that prevailed in Jodhpur (which is characteristic of much of the tropics). There is a flow regime that may be called ‘weakly forced convection’ (WFC) where the sensible heat flux is given by the classical free convection formula even in the presence of mild (cross) wind. Within the WFC regime the drag varies linearly with wind speed (and not as its square, as assumed in the formulation of bulk aerodynamic coefficients). The evidence for these conclusions is shown in Fig.11.8. What these results suggest is that the conventional scaling argument of M-O theory, centered around friction

velocity u_* and temperature scale θ_*, needs to be replaced by one scaled by heat flux in the WFC regime. MONTBLEX data also proved valuable in viewing eddy flux processes as a series of events, demanding an episodic rather than a harmonic description (Narasimha et al 2007).

11.4.2. LASPEX

LASPEX was a sequel to MONTBLEX whose planning began in 1992 and specifically carried out to understand land surface process with Indian Institute of Tropical Meteorology (IITM) as a nodal agency. The area chosen for the study was the Sabarmati river basin in Gujarat. The experimental area was spread over 100 km x 100 km area with different land surface characteristics. The main objectives of the programme were the following (Shekh et al 2001).

1. To collect a complete surface and sub-surface atmospheric hydrological data base against which parameterized models for land surface processes, i.e., energy exchange, radiative, sensible and latent heat fluxes can be tested for improvement and further development.
2. To foster the development of observing techniques, data management and assimilation systems.
3. The generated data shall be used in addressing the problem of determining evaporation flux and parameterizing land-surface processes at the scale of GCM-grid square, i.e., approximately 100 km x 100 km.
4. To provide in-situ measurements as ground truths for the data to be obtained through ERS-I, for evaluating soil moisture over the region.

The Pilot phase was carried out during April-July 1995, and the main phase during January 1997 to March 1998. Five 10 m high micrometeorological towers with Anand as center were installed (Aanad also had a 30 m tower). Data on soils, crops and weather, energy fluxes, upper air data were collected. Focus of LASPEX was on processes near the surface (i.e., the boundary layer). Some results of LASPEX can be found in the special issue of Journal of Agrometeorology (Vol. 3, June-December 2001). An important outcome of the analysis of the data collected is the large differences between observed and model produced fluxes.

11.5. Programmes under Indian Climate Research Programme (ICRP)

The emphasis in pre-1980 field experiments was on understanding the large-scale aspects of monsoon circulation. Since 1980, it became clear that the underlying surface (particularly the ocean) is not a mere supplier of energy and moisture to the atmosphere, but influences and is also influenced by the events in the atmosphere, and this interaction can be an important mechanism of climate variability. As the El Nino and Southern Oscillation (ENSO) phenomenon was being unraveled in the 1980s, ocean-atmosphere variations and their coupling on intraseasonal time scales and accurate estimation of surface fluxes emerged as important scientific issues (Webster and Lucas 1992). New relationships were discovered between the occurrence of large cloud systems (organized convection) and the sea surface temperature (SST) (Gadgil et al 1984, Graham and Barnett 1987). A variety of satellite derived data became available and numerical models were becoming powerful tools of research. Monsoon research also benefited from these developments and a number of issues were being addressed by different investigators. In early 1990s, the need to document what we know about monsoon based on all the work done till then including the Indian, and what needs to be done that will enhance our ability to understand and predict monsoon variability better, was increasingly felt within the Indian Scientific community. In order to have maximum impact from the limited resources available in the country, a road map for monsoon research in the country for the coming decade with well focused programmes was required. After several meetings within the country among people working in the areas of meteorology and oceanography, a document titled '***Indian Climate Research Programme Science Plan***' was brought out in 1996 (DST 1996). The major thrust of ICRP is on monsoon variability on timescales ranging from subseasonal to interannual and decadal, and its impact on critical resources. The ICRP science plan lists the following 4 major objectives.

1) Understanding the physical processes responsible for variability of the monsoon, the oceans (specifically the Indian seas and the equatorial Indian Ocean) and the coupled atmosphere-ocean-land system on various time-scales (sub-seasonal, seasonal, interannual, and decadal).

2) Study of the space time variation of the monsoons from sub-seasonal, interannual to decadal scales for assessing the feasibility for climate prediction and development of methods for prediction.
3) Study of change in climate and its variability (on centennial and longer time scales) generated by natural and anthropogenic factors.
4) Investigation of the links between climate variability and critical resources such as agricultural productivity, and for realistic assessment of the impact of the climate change.

Meeting the objectives of ICRP required well focussed programmes which study not only the individual components but also the interactions / feedbacks between different components of climate. Suggested action plans to address the ICRP objectives include the analysis of available data, numerical modeling, and carrying out several special field experiments with focus on process studies. Among the various actions planned in the ICRP Implementation Plan (DST 1998), field programmes have been successfully executed. ICRP field experiments are inter-agency national programmes with support from DST, Department of Ocean Development (DOD, now Ministry of Earth Sciences, MoES), Department of Space, and Ministry of Defence, with the participation of major national institutes, research organizations and universities working in meteorology and oceanography. Atmospheric and oceanographic communities work together in ICRP experiments. The field experiments are so designed that resulting data would enable the testing of some of the hypotheses that were prevailing then regarding the monsoon onset, propagation of monsoon cloud systems, air-sea interactions, oceanic processes, etc. Available resources in the country were utilized. DOD/MoES provided its ship *ORV Sagar Kanya* and National Institute of Ocean Technology (NIOT) Chennai deployed additional buoys and also provided its ship ORV Sagar Nidhi. National Physical Oceanographic Laboratory (NPOL) Kochi participated with its ship *INS Sagar Dhwani*. Indian Navy, Coast Guard and Indian Air Force actively participated and provided their facilities for observations. IMD organized special observations from coastal and island stations in addition to hosting the Scientific Advisory Committee (that monitored the progress and advised observational strategies based on short term weather forecasts) during the field phase. The scientific rationale, objectives and important findings from these experiments are described briefly in the following.

Buoy measurements

Indian monsoon is strongly coupled to the warm oceans surrounding the subcontinent. Most of the monsoon rainfall occurs in association with synoptic scale systems (the monsoon disturbances such as lows and depressions) which are generated over these waters and move onto the Indian landmass. In particular, the Bay of Bengal (bay henceforth) is exceptionally fertile, with a very high frequency of genesis of these systems. However, in situ data, especially in the areas where monsoon systems form and/or intensify, have been lacking. Monsoon observations received a boost with the installation of moored buoys (buoys henceforth) in Arabian Sea and bay by NIOT since 1997.

Time series of SST and wind speed measured by buoys over the bay and Arabian Sea during 1998 are shown in Fig. 11.9. Both Arabian Sea and bay show similar patterns of SST warming between March and the monsoon onset time. During the monsoon onset, SST collapsed suddenly over Arabian Sea but gradually over the bay. The dramatic decrease in SST over Arabian Sea coincided with a sudden increase in the wind speed following the formation of a monsoon onset vortex and its subsequent development into a cyclonic storm (this is consistent with MONEX observations). In some years Arabian Sea SST does not collapse as dramatically and shows gradual cooling over a few weeks after the monsoon onset. Buoy data showed for the first time that ocean and atmosphere undergo coherent variations on intraseasonal time scales in the north Indian Ocean during monsoon (Premkumar et al 2000). The behaviour of SST during July-September differs between Arabian Sea and head bay with intraseasonal variability in the former dull while very strong in the latter. Wind speed also showed a larger variation over the bay with values more than 10 m s^{-1} when organized convection formed there while wind speed decreased below 3 m s^{-1} during the weak phase of convection. On the other hand, winds over Arabian Sea do not drop this low during the monsoon season even when convection is in its weak phase. As a result, the latent heat loss always remains high over Arabian Sea.

Fig. 11.9 demonstrates how continuous *in situ* data can alter our thinking on air-sea coupling over the North Indian Ocean during the summer monsoon. In the

absence of other observations, research studies and conclusions derived are often based on Reynolds SST (Reynolds et al 2002, also called optimally interpolated SST, OISST) which merges satellite derived SST with ground truth. Reynolds SST missed the strong intraseasonal signal over the North Bay, and the conclusions one would reach from buoy and Reynolds SST time series are entirely different (Reynolds SST shown here was downloaded in January 2007, and the earlier comparison was no better (e.g. Premkumar et al 2000)). Buoy data showed that bay is not an infinite reservoir of heat unaffected by the monsoon drama unfolding over it in the atmosphere during summer, but has a top layer that quickly responds to atmospheric forcing even on very short time scale of a few days. Since SST and deep convection relationship is highly nonlinear (Waliser et al 1993), this has strong implications for atmospheric convection. Bay SST undergoes large intraseasonal fluctuations (while remaining above the convection threshold), whereas that of Arabian Sea remains nearly steady just above the convection threshold, but nevertheless supports intense high rainfall events on the west coast of India (Rao 1976, Francis and Gadgil 2006). Thus, Arabian Sea and bay have their unique characteristics. Buoy data raised new questions and some of the objectives of BOBMEX and ARMEX were an attempt to answer them.

11.5.1. BOBMEX

Bay is the breeding ground for the monsoon systems. One of the outstanding problems has been, how does the Bay manage to sustain high SSTs conducive for convection for a period of more than 4 months despite strong winds and frequent clouding? There were several other questions regarding the bay with no clear answer but many possibilities, and it was decided to organize the first field experiment under ICRP over the bay. BOBMEX is the first experiment to collect observations during a peak monsoon period in the north Bay using modern surface flux sensors and high resolution radiosondes (Bhat et al 2001). The emphasis in BOBMEX was on collecting high quality data over the bay and the surrounding coastal areas during different phases of monsoon. Within the bay, there are marked variations in the freshwater flux between the northern and southern parts of the bay. The northern bay receives a large quantity of fresh water through river discharge in addition to local precipitation. This results in very low values of surface salinity in the

northern bay which is not the case in the southern bay (Murthy et al 1992). The low saline water makes the top layer of the ocean very stable for vertical mixing. Thus, we expected the nature of the response of the ocean to atmospheric forcing to be different in the northern and southern parts of the bay. The upper ocean processes in the presence of strong monsoonal winds and low surface salinity needed to be understood. Therefore, it was decided to measure energy fluxes over the bay with sufficient accuracy, along with upper ocean temperature and salinity profiles.

BOBMEX was carried out during July-August 1999 with a Pilot during October-November 1998. [Some initial results from the BOBMEX-Pilot experiment are published in the June 2000 (special) issue of the Proceedings of the Indian Academy of Sciences (Earth and Planetary Sciences)]. Long time series observations over the open sea was given high priority in BOBMEX as previous observations in the bay were less than 2 weeks in duration and could not capture active and break monsoon conditions adequately. Indian research vessels *INS Sagardhwani* and *ORV Sagar Kanya* were deployed at TS1 (13°N, 87°E) and TS2 (17.5^{o}N, 89°E), respectively (Fig. 11.7a). Measurements of all components of surface fluxes, the vertical profiles of atmospheric temperature, humidity and winds, and ocean temperature, salinity and current profiles were planned from both the ships. All the planned measurements could be accomplished only on *ORV Sagar Kanya* and upper air data could not be collected on *INS Sagardhwani*. Synoptic and upper air observations over the coastal and island stations belonging to IMD were also documented. Observations covered active and break monsoon conditions.

11.5.2. ARMEX

After the successful execution of BOBMEX, ARMEX was carried out during 2002-2003 with some measurements completed in 2005. ARMEX was executed in two phases addressing different scientific questions related to the atmosphere and ocean. It is observed from Fig. 11.9 that SST rapidly increases during March-April, and then remains above 30°C (i.e., well above the convection threshold value of 28°C for the Indian Ocean) for over a period of more than a month but organized convection rarely develops. In fact, a mini-warm pool builds up in the south eastern Arabian Sea and there have been theories about its evolution, maintenance

(Vinayachandran et al 2007), and its importance to the Indian monsoon rainfall. One of the objectives of ARMEX was to understand the mini-warm pool dynamics and test the hypotheses. The monsoon onset processes over Kerala and dramatic collapse of SST during the monsoon onset was another important issue. This part of ARMEX involving the study of the evolution, maintenance and the collapse of the Arabian Sea mini warm pool and pre-onset and onset phases of the monsoon, formed phase II of ARMEX and was carried out during March-June 2003. Oceanographic component dominated ARMEX-II objectives and has been described in Vinayachandran et al (2007).

When monsoon is active, many places along the Indian west coast receive more than 200 mm rain in 24 hours, and such cases were designated as intense rainfall events (IREs) in ARMEX. One recent example is the Mumbai rain event of 25 July 2005 where 94 cm rainfall occurred in about 12 hours (Jenamani et al 2006). Documenting the structure of off-shore vortices (whose existence was first suggested by George in 1956) that produce IREs, the off shore trough and mechanism of IREs, and the monsoon rainfall of the west coast was another objective of ARMEX. The experiment to address these issues was carried out during June-August 2002 (ARMEX-I). Previous monsoon experiments over Arabian Sea covered mid May to early July period, whereas, ARMEX-I covered monsoon onset (mid June) to late August period with emphasis on IREs, and the revival and maintenance of monsoon on the west coast after the monsoon onset phase is over.

The focus in ARMEX was the region within 250 km from the west coast of India (Fig. 11.10). Land based observations were enhanced by installing 10 automatic weather stations along the west coast, and ships were deployed for monitoring conditions over the Arabian Sea. Surface and upper meteorological observatories belonging to IMD, Defense establishments and other agricultural and research organizations falling within the ARMEX study area were operated and data made available. Indian Navy deployed two ships and DOD provided *ORV Sagar Kanya*. Some preliminary results can be found in the January 2005 special issue of *Mausam* brought out on ARMEX.

11.5.3. CTCZ

The non-orographic rainfall over the Indian subcontinent during an active monsoon is characterized by large-scale low level convergence in the boundary layer, cyclonic vorticity above the boundary layer and organized deep convection. Having the characteristics of tropical convergence zone (TCZ) observed over the oceans, this low pressure belt over the plains of north India is called the continental TCZ (CTCZ). A southeast-northwest oriented low pressure (heat low) axis exists over north India before the monsoon onset also, however, it is shallow. CTCZ gets established over the subcontinent at the end of the onset phase of the summer monsoon. The major phases of the CTCZ are (i) the spring to summer transition which occurs in the onset phase, and (ii) the peak monsoon months of July and August, when the CTCZ fluctuates primarily in the core monsoon zone (Fig. 11.7b). An important feature of the intraseasonal variation of the CTCZ is the fluctuation between active and weak spells, and CTCZ determines the active and weak phases of rainfall activity. Thus, understanding the monsoon rainfall over India is closely linked to understanding the space time variation of CTCZ.

The third monsoon observational programme under ICRP, i.e., is the sequel to BOBMEX and ARMEX, is called the CTCZ Programme which address CTCZ. The main objective of the CTCZ programme is to understand the mechanisms leading to the space-time variation of rainfall and the embedded monsoon disturbances during the summer monsoon over the core region shown in Fig.11.7b. Land-atmosphere interactions, hydrological feedbacks and aerosols influence the seasonal transitions and intraseasonal variation during the summer monsoon. Most of the cloud systems in the CTCZ (synoptic scale in particular) are generated over the warm oceans around the subcontinent and propagate onto the Indian region. Therefore, it is important to investigate the links of monsoon variability with the oceanic conditions and convection over the surrounding oceans. Accordingly, the science foci of CTCZ comprise of both important phenomena and process studies as listed below.

Phenomena

(a) Seasonal transition and intraseasonal variation of the monsoon.

(b) Links of monsoon variability with the convection over the surrounding oceans.

Process studies

(i) Land-atmosphere interactions and hydrological feedbacks.

(ii) Role of aerosols in monsoon variability.

(iii) Scale interactions in convection ranging from the cumulus to the large scale.

A multi-pronged approach involving field experiments, analysis of existing data from conventional platforms as well as satellites, buoys, and ARGO floats, and theoretical studies with process models, complex atmospheric general circulation models, as well as coupled ocean-atmosphere models has been adopted for attaining this challenging objective. Under the CTCZ programme, special efforts are being made to elucidate the nature of the cloud systems over land and measure critical components of water and heat balance in few river basins/watersheds. The impact of land surface processes in the genesis of cloud systems and their propagations will be studied. The observational systems would pool the resources available within India including Doppler Weather Radars, flux towers to measure components of the surface energy balance (over the vegetated, forested and croplands), high resolution GP Sondes, micro-pulse lidars, instrumented aircraft, moored buoys in the bay and the Indian Ocean, research ships with complete set of ocean-atmosphere instrumentation. Few river basins/sub-basins will be intensively observed and modeled to understand the hydrological cycle.

Recent studies have shown that aerosols can cause substantial alteration in the energy balance of the lower atmosphere and at the earth's surface, thus modulating the hydrological cycle. Observations of the space-time variation of aerosols particularly over regions which are considered to be critical for impact on the monsoon, aerosol life cycles in clouds and impact of aerosols on atmospheric radiation, are planned. The CTCZ programme offers a unique opportunity to address several challenging scientific issues related to the interactions of the monsoon with the atmospheric boundary layer, ocean-atmosphere coupling and land-surface/cloud/aerosol processes.

A pilot phase of CTCZ was carried out during 01 July to 31 August 2009 utilizing most of the existing observational monitoring networks including Radars, aerosols, agro meteorological stations, met-ocean data buoys, Argo floats and

drifters, two ships (ORV Sagar Kanya and ORV Sagar Nidhi), two aircrafts with state-of-the art instrumentation, additional radiosonde systems at Kharagpur and over northern bay, three micrometeorological towers (Kharagpur, Ranchi, and Anand), stand-alone atmospheric observing systems (e.g. micropulse lidars and disdrometers) at few locations north of 18°N, up to foothills of Himalayas. Often weather does not behave the way we want it to study a given phenomenon in one year, but over a period of a few years the chances are much better. Hence, CTCZ is a multi-year programme, and the main phase is planned during the monsoon seasons of 2011 and 2012.

11.5.4. Some results from ICRP Experiments

The questions 'did the experiment lead to new findings?', naturally arise at the end of each experiment. BOBMEX and ARMEX did provide data to study and understand certain physical processes and test some hypotheses. For example, it was known that when organized convection occurs, boundary layer cools and convective instability of the atmosphere (which drives the formation of clouds in tropics) is destroyed. However, how long it would take for the atmosphere to recover its instability to support another active spell of rains over the Bay was not known. BOBMEX observations revealed that the recovery time is about 2 days (Bhat 2001). The changes in the vertical temperature structure of the atmosphere between the active and weak convective conditions could be clearly seen with the BOBMEX upper air data (Bhat et al 2001, Bhat et al 2002). The largest variation in the vertical between active and weak convective conditions is observed in the relative humidity (Fig. 11.11) and wind fields. Another interesting observation is the differences in the dependence of the latent heat flux (LHF) on wind speed over bay and Arabian Sea. At a given wind speed, the LHF is much lower (30 to 40%) over the bay as compared to that over Arabian Sea and the western Pacific warm pool (Fig. 11.12).

During BOBMEX, the differences in the upper ocean structure between northern and central Bay could be clearly established. The decrease of the oceanic mixed layer in the North Bay could be captured (Vinayachandran et al 2002, Rao and Sikka 2005). Fresh water arrived in the last week of July at the ship location (TS2, Fig. 11.7) during BOBMEX, and following this, the mixed layer depth (MLD)

decreased from around 30 m to less than 15 m (Fig. 11.13). While salinity is responsible for mixed layer decreasing in the head bay, it destabilizes the upper layer of the ocean over the Arabian Sea (Fig. 11.13). The intra-seasonal behaviour of SST over the bay and Arabian Sea are very different (Fig. 11.9). The data collected during BOBMEX and ARMEX helped in addressing this issue, and the SST evolution is briefly discussed next.

Surface fluxes and SST evolution

Temporal evolution of the temperature T of the mixed layer is given by,

$$T(t) = T_0 + \int_0^t (Q(t') - Q_{pen}(t'))\, dt' / (\rho_w C_w h) \qquad (1)$$

where Q is the net heat flux into the ocean at the surface, and Q_{pen} is the amount of shortwave radiation penetrating below the mixed layer, ρ_w and C_w are sea water density and specific heat, respectively, and h is often taken as the mixed layer depth (MLD). Q is given by

$$Q = NSW - (NLW + SH + LH) \qquad (2)$$

where, NSW and NLW are net shortwave and net longwave radiation, respectively. SH and LH are the sensible and latent heat fluxes. When the mixed layer is shallow, part of the solar radiation escapes from its bottom and the actual heat flux available for the mixed layer is $Q-Q_{pen}$. During BOBMEX, all terms in equation (2) could be obtained from measured data, while all terms including Q_{pen} were measured during ARMEX. Here we look at the short time SST variations at the ship location driven by the surface heat flux neglecting the diffusion and advection terms and assuming a constant value of h.

In equation (1), T depends on 3 factors, namely Q, h (normally taken to be MLD) and Q_{pen}. MLD could be less than 15 m in the head bay whereas that in the Arabian Sea is more than 60 m (Fig. 11.13). At the time of planning BOBMEX, it was thought that the large difference in MLD is primarily responsible for the rapid warming of SST in the head bay (Fig.11.9). Over the head bay, latent heat fluxes are 30-40% smaller compared to that over other warm tropical basins at a given wind speed (Fig. 11.12). During the weak phase of convection (when ocean tends to warm) wind speed is very low over the bay while over Arabian Sea it remains high

(Fig. 11.9). Further, the atmosphere is very humid over the bay and the net long-wave radiation is around 30-35 Wm^{-2} compared to 40-50 Wm^{-2} over other oceans. The sensible heat flux is negligible in both the cases. The net result is that when all terms are added up during the weak phase of convection, the daily average value of Q is in the range of 140-180 Wm^{-2} in the head bay, whereas over the Arabian Sea, it can vary from +50 Wm^{-2} to –80 Wm^{-2} with a mean value (averaged over several days) very close to zero or slightly negative even during the weak phase of convection (Fig. 11.14). Q_{pen} is not negligible over the head bay, however Q is sufficiently large to provide enough heat to warm the mixed layer by more than 1^oC in just 4 days after organized convection decays. Calculations show that both the shallow mixed layer and significantly larger value of surface heat flux (compared to other warm tropical oceans) contribute almost equally to increase the SST so rapidly over the North bay. Thus, the combination of shallow mixed layer and large net heat flux into the ocean result in the rapid increase in SST over the bay. Therefore, in the head Bay, ocean and atmosphere cooperate to maintain high SST. The situation over the Arabian Sea is just the opposite, and even during clear sky conditions, Q could be negative.

Field experiments need large lead time for planning and execution. At times, ground situation may not favour the original objectives as the weather conditions are not under experimenter's control. For example, monsoon onset was delayed by about 2 weeks in 1979 when MONEX was conducted. Understanding IREs was one of the main objectives during ARMEX-I. However, 2002 was a major drought year against all expectations of a normal monsoon, and the all India seasonal rainfall was about 20% below the normal. The offshore vortex did not form and IRE did not occur in the area known for propensity of IREs where ARMEX-I intense observations were planned. Nevertheless, ARMEX-I data proved very valuable as July 2002 rainfall was the lowest in the recorded history and the data collected over the AS and on the west coast helped in understanding the conditions that prevailed over the eastern AS during one of the worst monsoon years. In particular, strong and persistent inversions were present in the atmosphere over the Arabian Sea and west coast (Bhat 2006). A sample from the radiosonde data collected during ARMEX-I is shown in Fig. 11.3. Two inversions are seen in the Arabian Sea temperature profile (around 2 km and 6 km). Such strong inversions (especially the low level one) suppress the

vertical development of clouds and rain cannot occur. It is observed that relative humidity shows several minima (especially in the inversion layers), suggesting a highly stratified and laminated atmosphere where air parcels from different sources were moving in thin layers with little vertical mixing. The large scale circulation was very different during July 2002 compared to a normal monsoon year and mid latitude air had penetrated southward of 15°N over Arabian Sea during this period (Bhat 2006).

ARMEX-I data has been used for modeling, especially using regional models. A couple of IRE events occurred in the last week of June over Gujarat (Hatwar et al 2005), i.e., slightly to the north of intense observations area during ARMEX. Numerical experiments showed that incorporating ARMEX-I observations (ship radiosonde data in particular) did have an impact in improving the simulation of systems associated with these IRE events (Rao and Prasad 2005).

Aerosols were not measured during BOBMEX. Following the Indian Ocean Experiment (INDOEX) carried out during 1996-1999, its impact on monsoon became a highly debated topic. Measuring the aerosols and their radiative forcing was taken up as an objective of ARMEX and aerosol spectral optical depths were measured over Arabian Sea as a part of ARMEX-I. These are the first measurements of aerosols over northern and central Arabian Sea during Indian summer monsoon season. Estimates show that sea-salt contributes about 60% to the composite aerosol optical depth (Vinoj and Satheesh 2003). The presence of aerosols over Arabian Sea during summer monsoon season decreases the short wave radiation arriving at the surface by as much as 21 W m^{-2} and increases top of the atmosphere reflected radiation by 18 W m^{-2}. Thus, the atmosphere absorbs 3 W m^{-2}. ARMEX data also helped is quantifying the direct and indirect effects of aerosols and it was found that the magnitude of indirect effect is several-fold larger than the direct effect of sea-salt aerosols (Vinoj and Satheesh 2004).

During the CTCZ Pilot cruise, ORV Sagar Nidhi surveyed the southern bay. Temperature and salinity profiles measured by CTD brought out the complex nature of salinity variation (Fig. 11.15). Salinity has a large temporal variation with the

presence of low salinity water on some days at the top and a tongue of high salinity water in the thermocline.

11.6. Concluding Remarks

Monsoon experiments have contributed to monsoon studies in several ways. Each experiment enhanced the infrastructure facilities in the country, brought together scientists from different organizations within and outside the country to a common platform and also attracted new people to this field. A large amount of data has been generated and their analysis has led to new understanding and discovery of new phenomena. However, we believe that these data have a lot more potential, and there is scope for further studies using numerical models and modern data analysis techniques.

A monsoon experiment with even modest objectives requires enormous resources, and several burning issues cannot be addressed owing to the lack of required facilities or man power. For example, monsoon rain comes from clouds, but study of clouds (microphysics and dynamics) could not be taken up in Indian monsoon experiments as India does not have an instrumented aircraft needed for this purpose. This lacunae is being corrected in CTCZ by collaborating with IITM's CAIPEEX programme. Similarly, upper air data could be collected only from *ORV Sagar Kanya* as other ships didn't have the radiosonde facility. One important physical process in the tropics is the interaction between organized convection and the large scale atmospheric circulation. To calculate their interaction, at least 3 simultaneous radiosonde profiles with temperature, humidity and wind measurements are needed over the open ocean. This means simultaneous deployment of three ships. Hopefully research aircraft and additional ships will not be a constraint in future programmes.

Weather knows no national and political barriers and monsoon is a planetary scale phenomenon. Monsoon trough over the Indian sub-continent is a part of the planetary scale system stretching eastward from the Indian longitudes to the central Pacific. The cloud systems that give rain over India are linked Tibetan high, east China monsoon, South China Sea and Pacific Ocean. A major international

monsoon research programme named MAHASRI (Monsoon Asian Hydro-Atmosphere Scientific Research and Prediction Initiative) led by Japan is being planned by some Asian countries during 2008-2015. Another programme, the Asiam Monsoon Year (AMY), which aims to study the Asian monsoon system on different spatial and time scales with coordinated simultaneous measurements in many Asian countries, is being observed during 2007-2012. Clearly, a co-ordination of the CTCZ programme with AMY and MAHASRI would help in collecting observations in critical regions of Asian monsoon spread across nations. Now the time is becoming ripe for India to collaborate with neighboring and other countries in monsoon research with her own well defined programmes. CTCZ could be the Indian contribution to MAHASRI and AMY.

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The Indian Ocean Bubble

Table 11.1: Major experiments that addressed Indian monsoon.

No.	Experiments	Year/Period	Broad area of investigation	Comments
1	IIOE - International Indian Ocean Expedition,	1960-65	Indian Ocean north of 40°S	multinational,20 countries, 40 ships, aircraft (drop-sonde), etc.
2	ISMEX - Indian Summer Monsoon Experiment	1973	Arabian Sea, the equatorial region and southern Indian Ocean	Indo-Soviet, 6 research ships (4 – USSR, 2 Indian)
3	MONSOON-77	1977		Indo-Soviet
4	MONEX-79, Monsoon Experiment (under FGGE)	1979 May-August (summer MONEX)	Arabian Sea, Bay of Bengal, equatorial Indain Ocean	Multinational , about 20 ships, research air craft, etc.
5	MONTBLEX-Monsoon Trough Boundary Layer Experiment	1990	Northern India, Bay of Bengal	Indian,
6	LASPEX – Land Surface Processes Experiment	1997-98	North-Western India	Indian, land surface processes studies in a semi-arid area
7	BOBMEX - Bay of Bengal Monsoon Experiment	1999 July-August	Bay of Bengal	Indian
8	ARMEX - Arabian Sea Monsoon Experiment	2002-05	Arabian Sea, west coast of India	Indian
9	JASMINE – Joint Air-Sea Monsoon Interaction Experiment	1999 Apr.7-22, May 1-Jun 8, and Sept 2-28	International Equatorial IO & Central Bay of Bengal	International

Table 11.2: Some important Atlases based on IIOE observation.

No	Atlas
1	*International Indian Ocean Expedition Plankton Atlas. Panikkar, N.K., editor (1968-1973).*
2	*Oceanographic Atlas of the International Indian Ocean Expedition. Wyrtki, K. (1971). National Science* Foundation, Washington DC, USA.
3	*Meteorological Atlas of the International Indian Ocean Expedition. National Science Foundation,* Washington DC, USA.
4	Ramage, C.S., F.R. Miller and C. Jcfferics (1972). *The surface climate of 1963 and 1964.*
5	Ramagc, C.S. and C.V.R. Raman (1972). *Upper Air.*
6	*Geological-geophysical Atlas of the Indian Ocean. Udintsev, G. B., editor (1975). Academy of Sciences,*Moscow.
7	*Phytoplankton Production Atlas of the International Indian Ocean Expedition. Krey, J. and* B. Babenard (1976). Institut fur Meereskunde, Kiel.

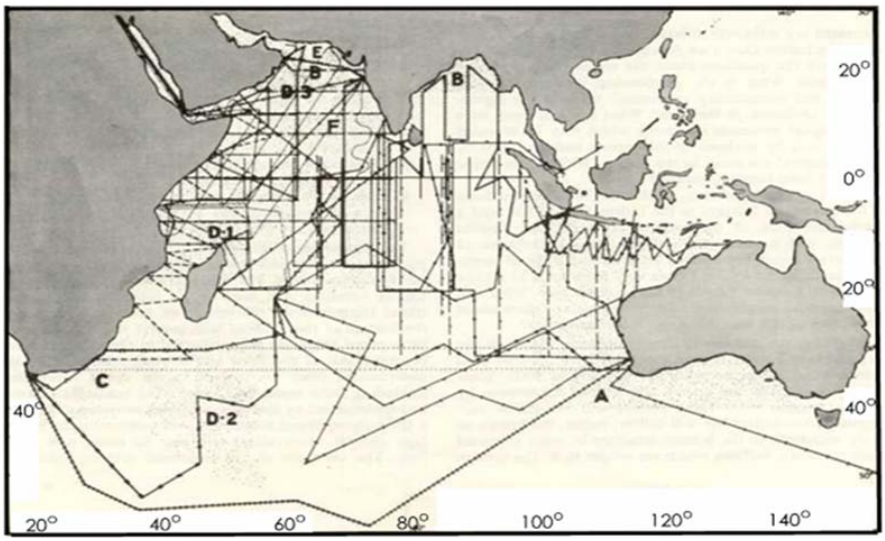

Fig. 11.1: IIOE Cruise tracks as in 1961. More than 40 oceanographic research vessels belonging to 13 countries surveyed the Indian Ocean from 1960 to 1965.
(Based on R. G. Snider in Discovery, 1961)

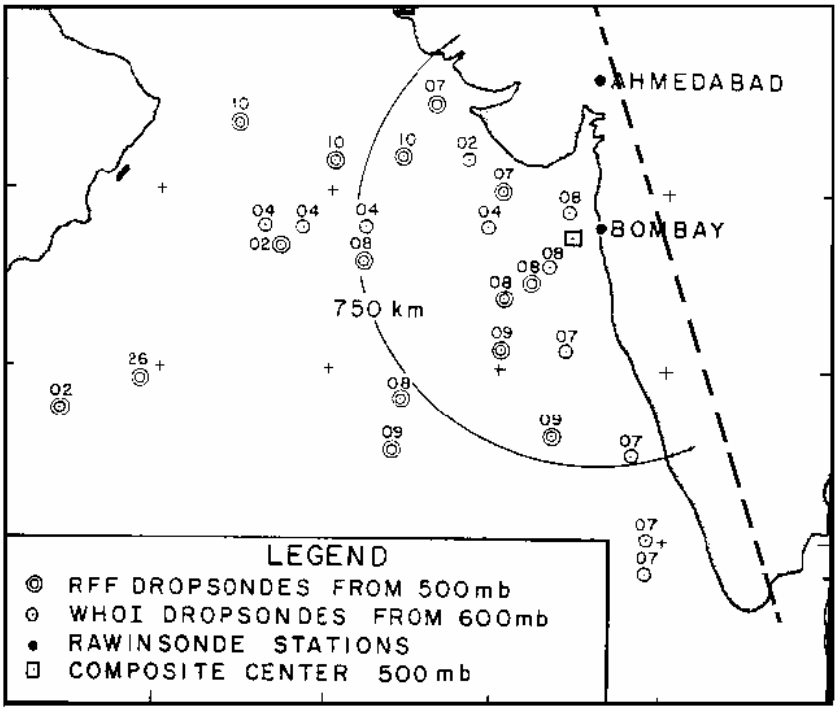

Fig. 11.2: Locations of dropsonde observations made during IIOE. Period: 26 June-10 July 1963. The numbers on symbols refer to the date.Vertical section shown in figure 4 is taken along the dashed line. ***From Miller and Keshavamurty (1968)*****.**

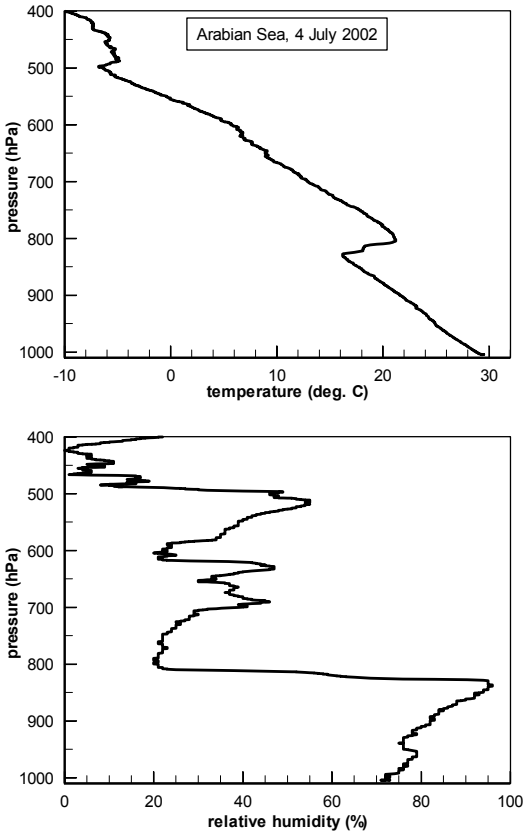

Fig. 11.3: Variation of temperature (top) and relative humidity (bottom) with height over the Arabian Sea. Location 16.94 °N 71.21°E. Upper air data collected during ARMEX.

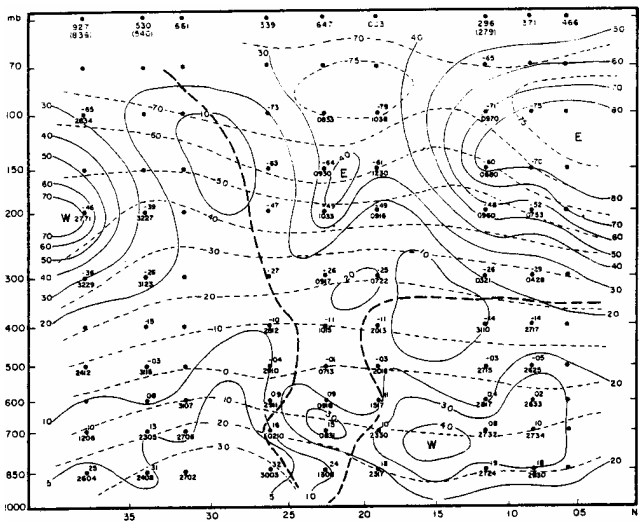

Fig. 11.4: Vertical section of wind speed and temperature along the dashed line shown in figure 2. Dashed line is temperature. Solid lines are isotachs of total wind speed. Date - 4 July 1963. (based on fig 28 Miller and Keshavamurti 1968).

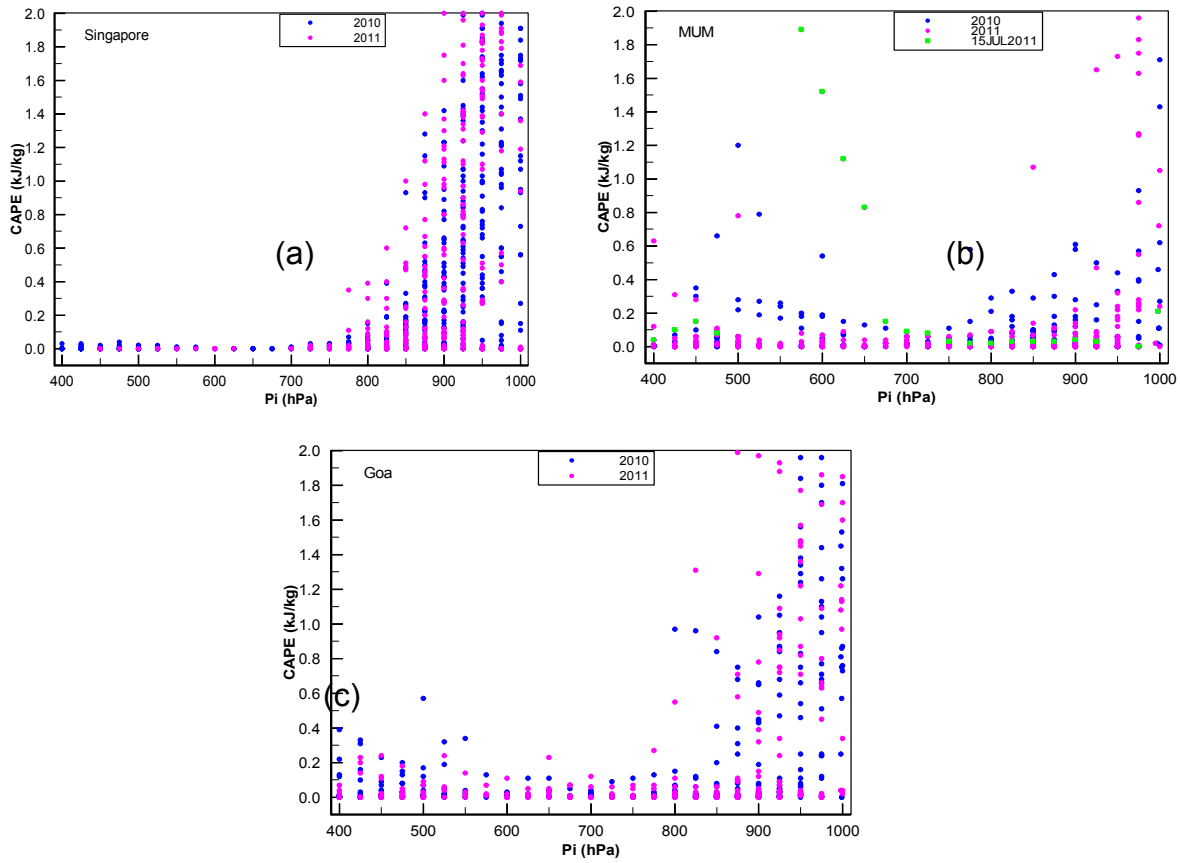

Fig. 11.5: Values of CAPE for parcels lifted from different heights for July soundings. From the sounding data available from the University of Wyoming website, data are interpolated to 25 hPa interval, and then CAPE calculated.

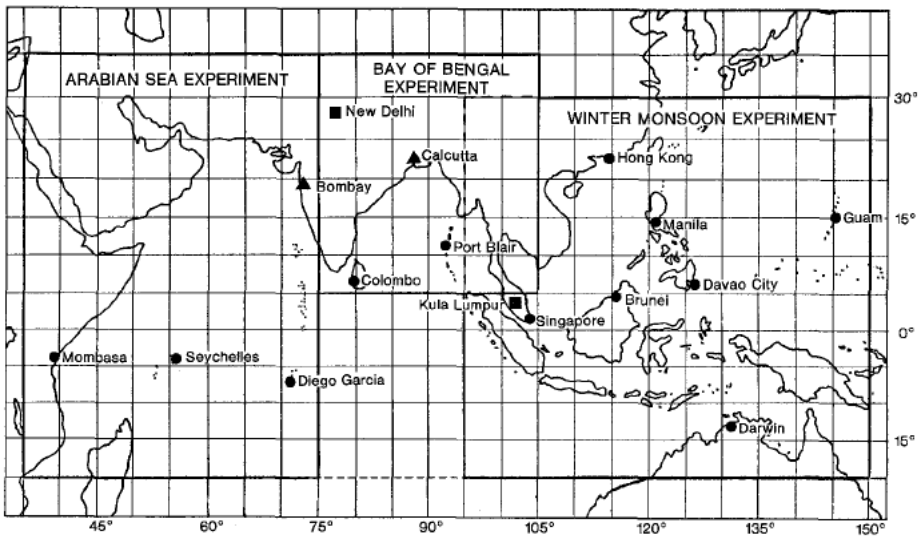

Fig. 11.6a: Areal extent of MONEX observations. (from Murakami, 1979)

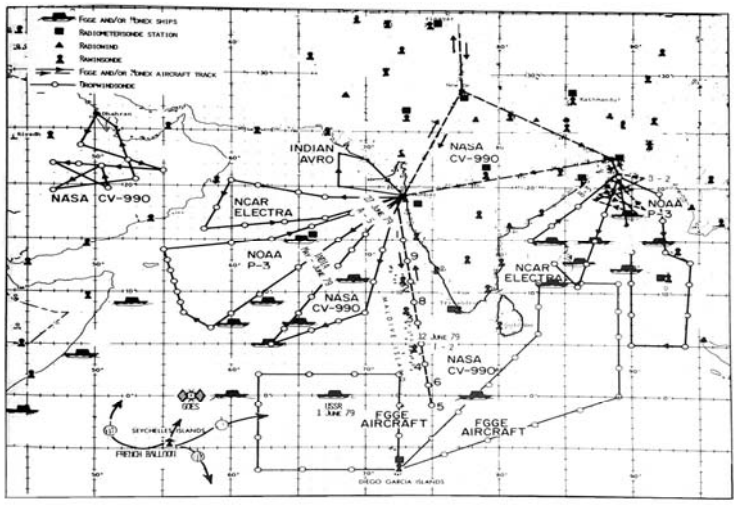

Fig. 11.6b: Areas covered by ship, upper air and aircraft observation s during SMONEX. (based on Fein and Kuettner 1981).

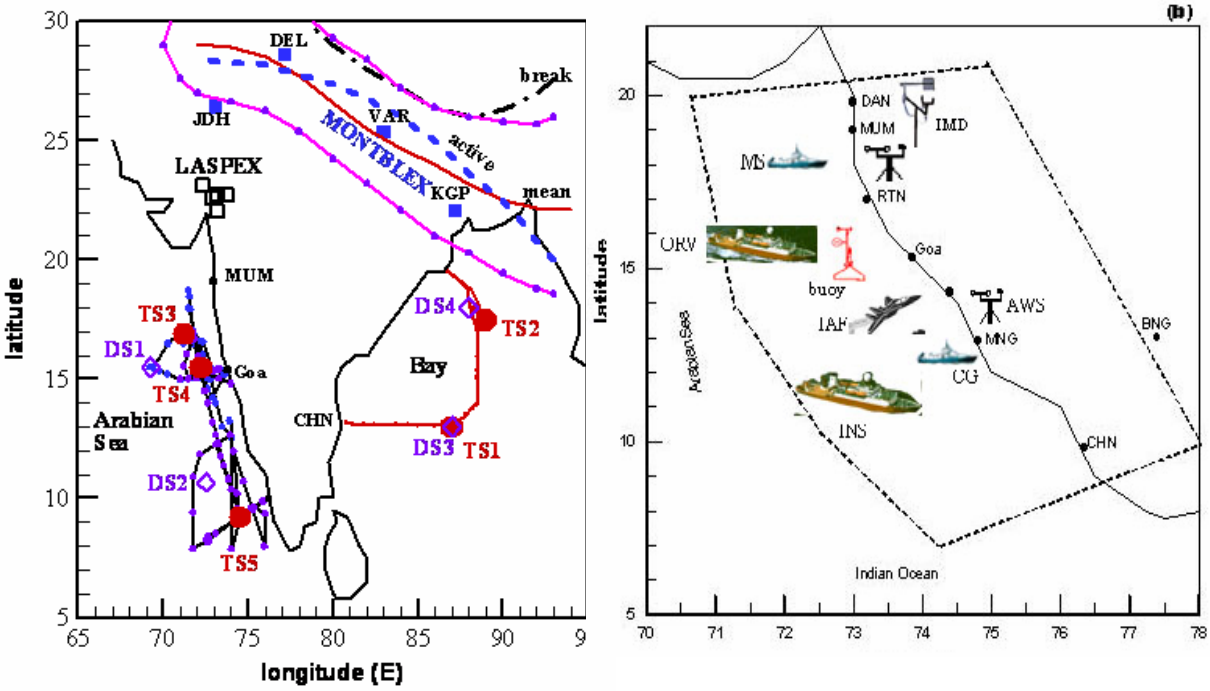

Fig. 11.7a: Study areas of recent monsoon experiments carried out by Indian scientists (left). The fluctuating location of the monsoon trough is shown in northern India (pink line) where MONTBLEX observations were carried out. The filled boxes show the locations of 30 m towers operated during MONTBLEX. The open boxes in western India show the observation sites of LASPEX. The lines over the bay and Arabian Sea show cruise tracks of *ORV Sagar Kanya*, and the corresponding time series observation stations (TS2 to TS5) by filled circles. *INS Sagardhwani* was deployed at TS1 (13°N, 87°E). DS1, DS2, DS3 and DS4 (diamond) are moored buoys deployed by NIOT. ARMEX Phase-I observation facilities shown on right.

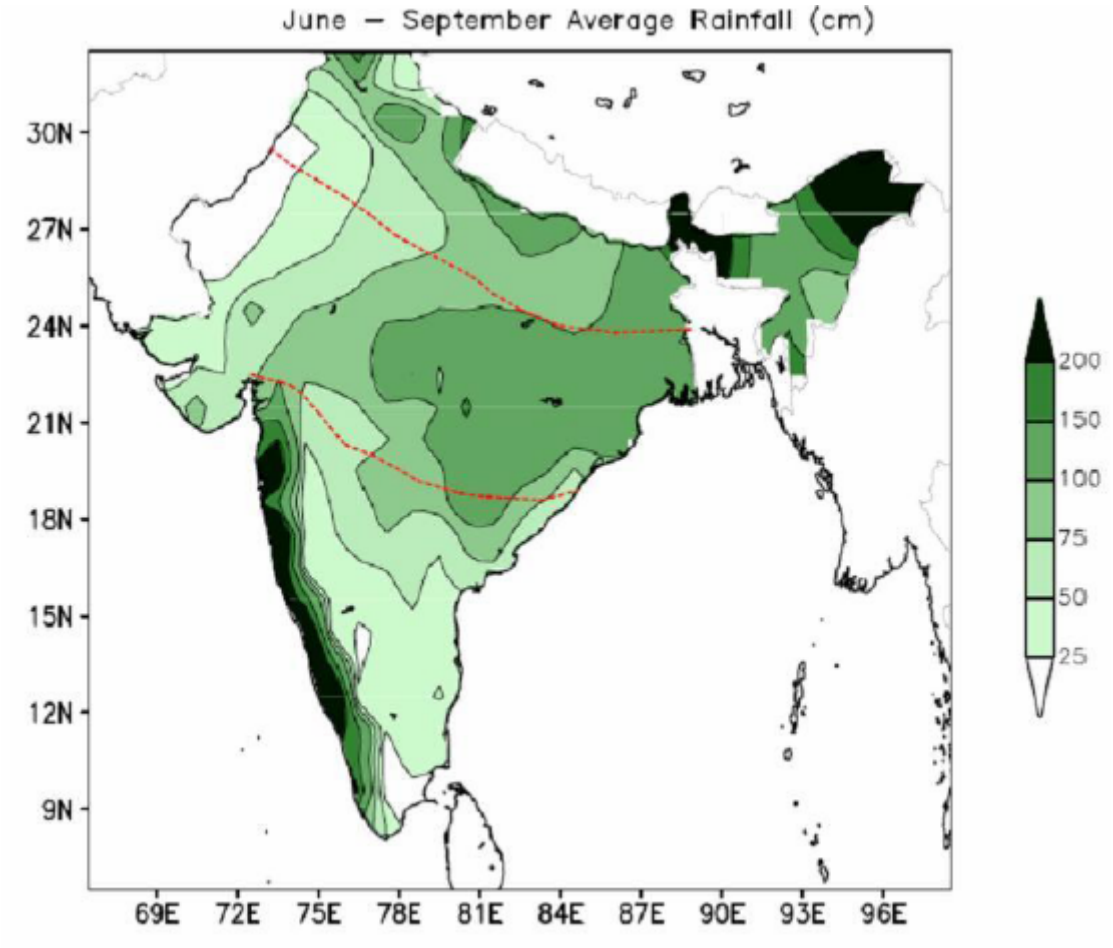

Fig. 11.7b: June-September mean rainfall over the Indian region (cm); approximate limits of the core monsoon zone indicated by red dashed lines. CTCZ area is north of 18°N comprising of the main monsoon zone over the sub-continent. (Courtesy: IMD)

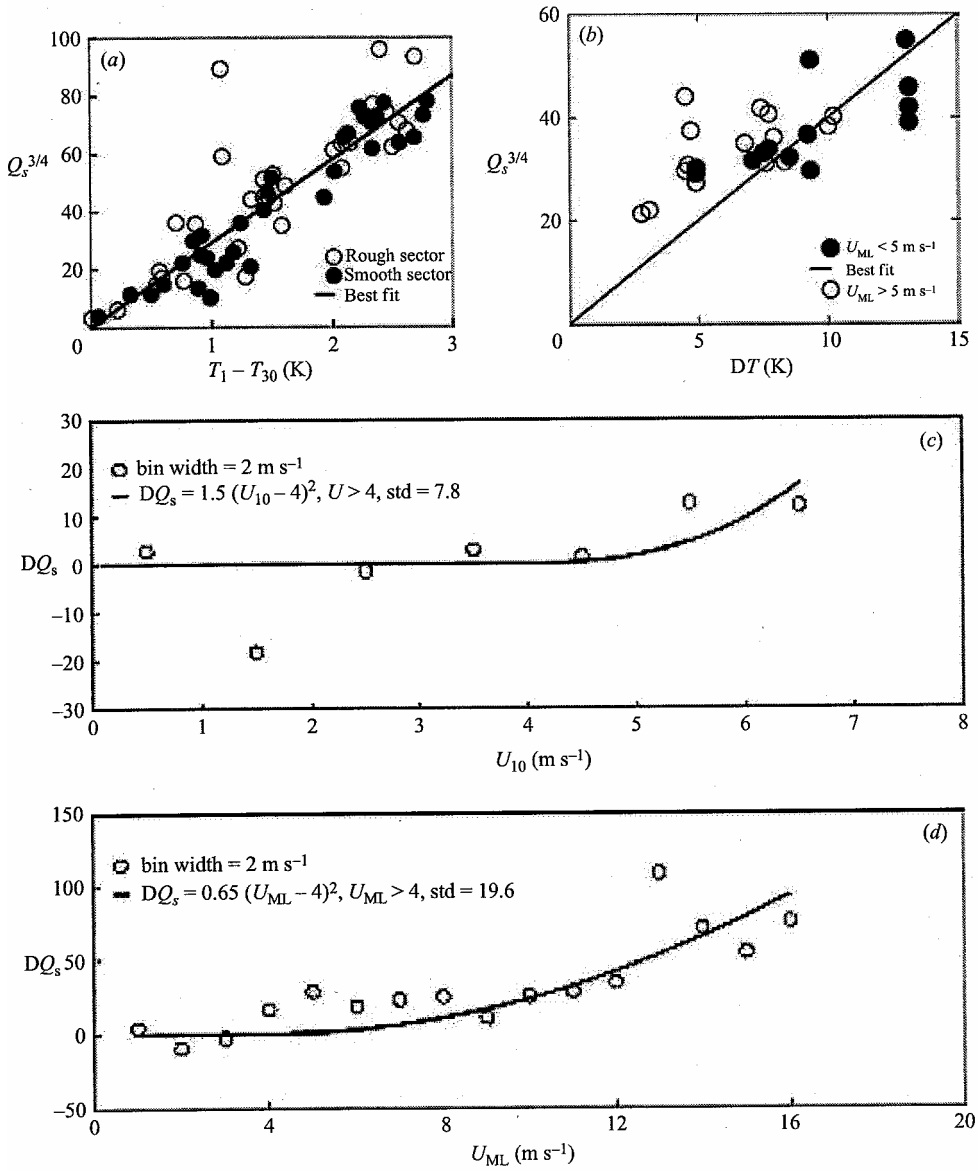

Fig. 11.8: Observed sensible heat flux as a function of a characteristic temperature differential, to show the range of a 4/3 power dependence. (a) Data from MONTBLEX. (b) Data from BLX 83, indicating separately low- and high-wind observations. (c,d) Deviations of Q from best fit as function of wind speed; data from MONTBLEX (c) and BLX83 (d). (from Bhat and Narasimha 2007)

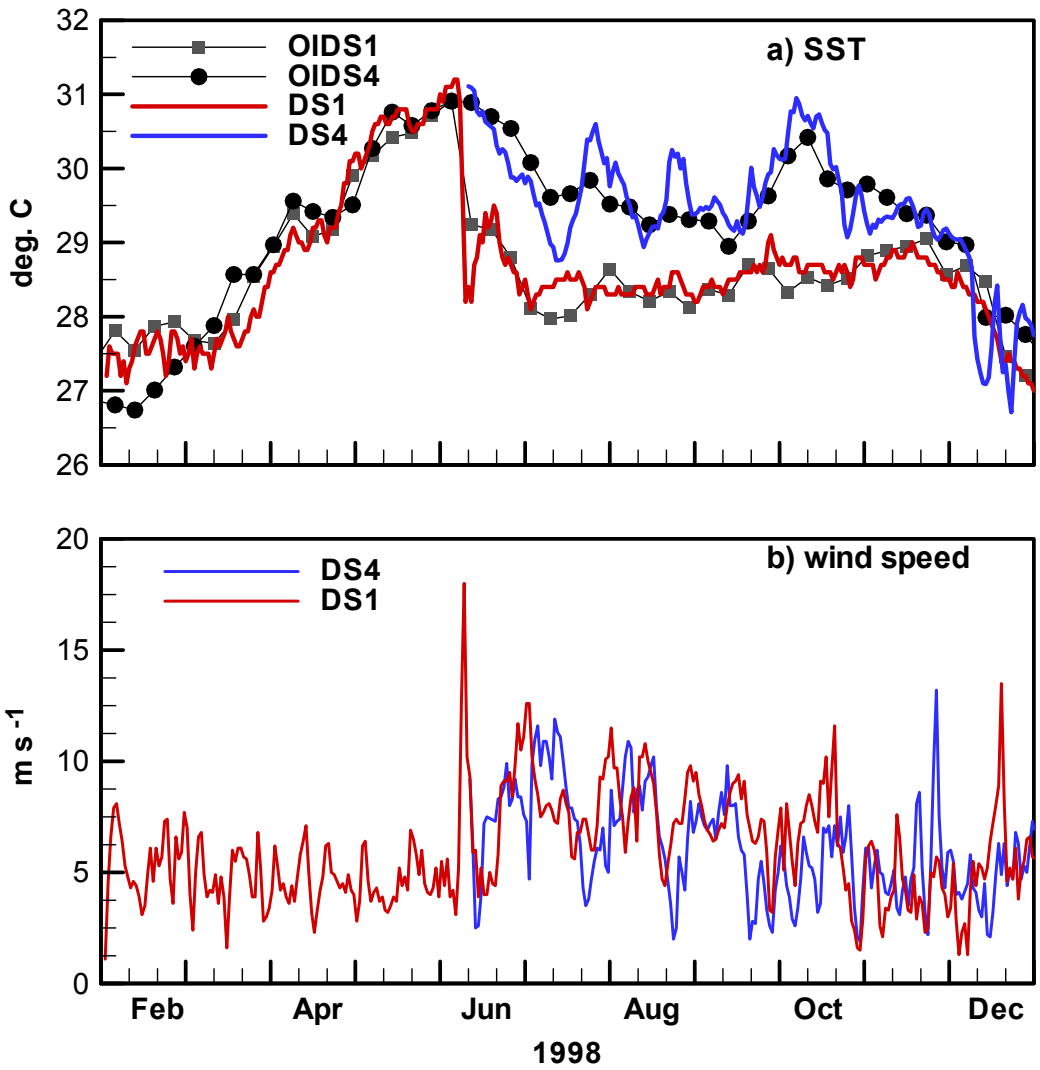

Fig. 11.9: a) The variation of daily average SST measured by NIOT buoys DS1 & DS2 in the year 1998. Corresponding optimally interpolated pentad SST (earlier known as Reynolds SST) downloaded from NOAA website in February 2006 at DS1 (OIDS1) and DS4 (OIDS4) are also shown. (b) Daily average wind speed measured by buoys.

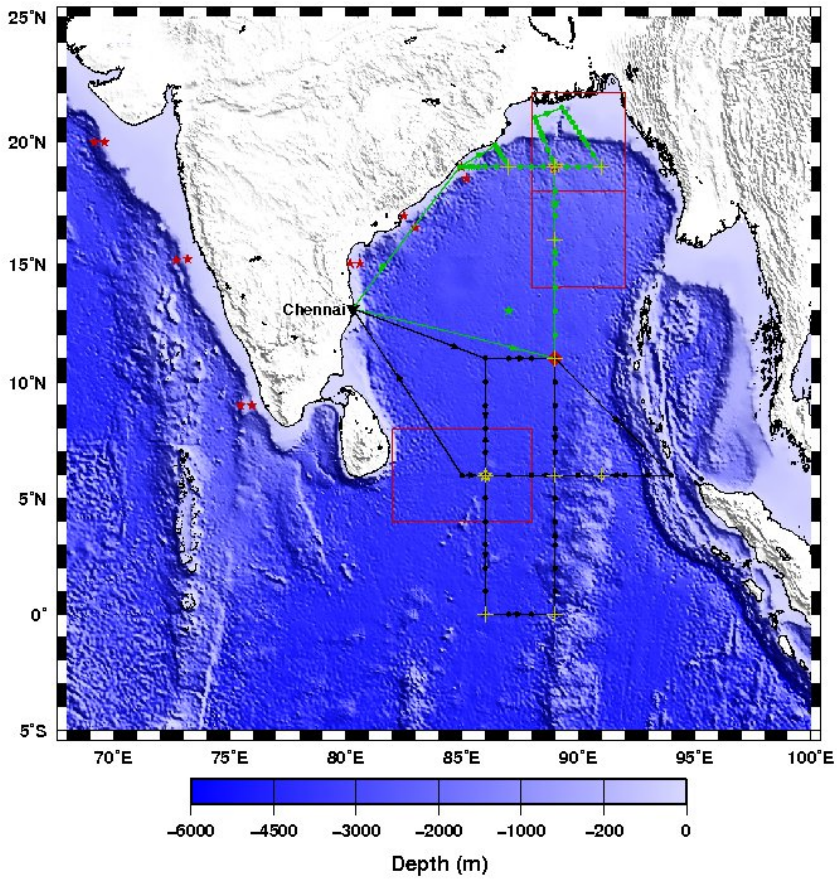

Fig. 11.10: CTCZ-Pilot 2009 Cruise Tracks. Black line – RV Saga Nidhi (16 July-18 August 2009) , Green line – ORV Sagar Kanya (star indicates time series observation locations). (Courtesy – D. Shankar)

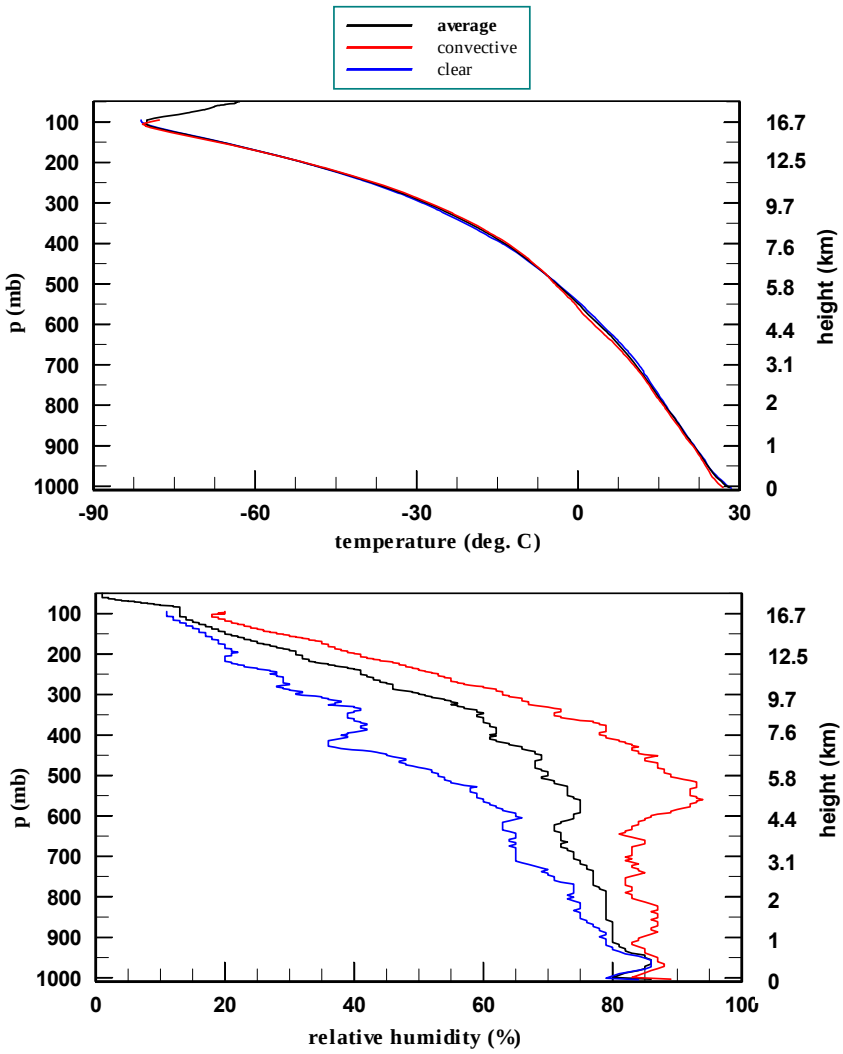

Fig.11.11: Vertical profiles of temperature and relative humidity measured at TS1 during BOBMEX. Convective and clear refer to average for the periods with and without organized convection.

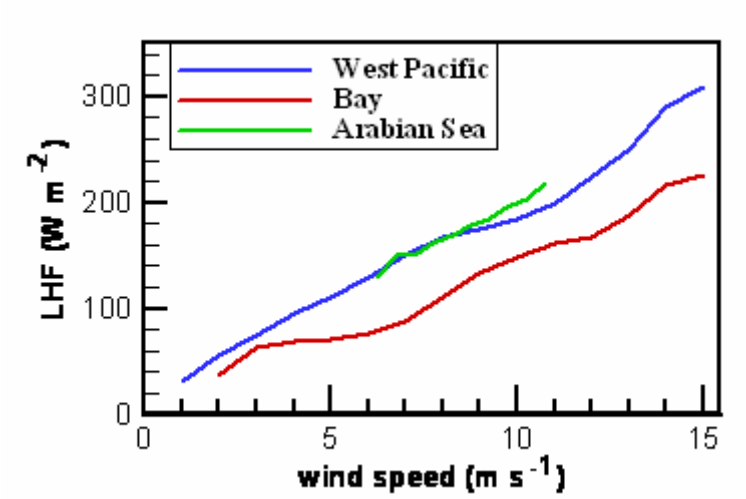

Fig. 11.12: Variation of latent heat flux with wind speed over three warm tropical water bodies.

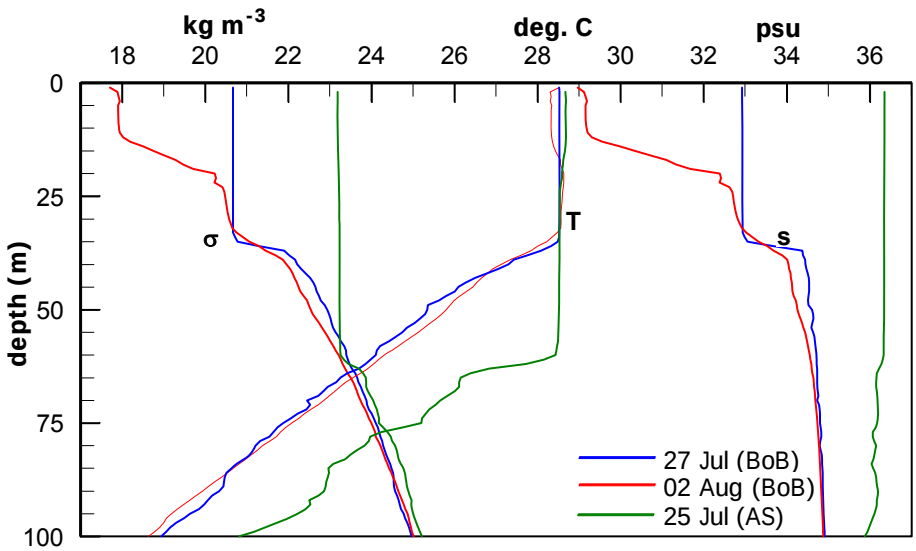

Fig. 11.13: Vertical variation of water temperature (T), salinity (s) and water density (□) in the head Bay (measured during BOBMEX) and AS (measured during ARMEX-I). The constant density layer is the mixed layer, and MLD approximately corresponds to the thickness of this layer.

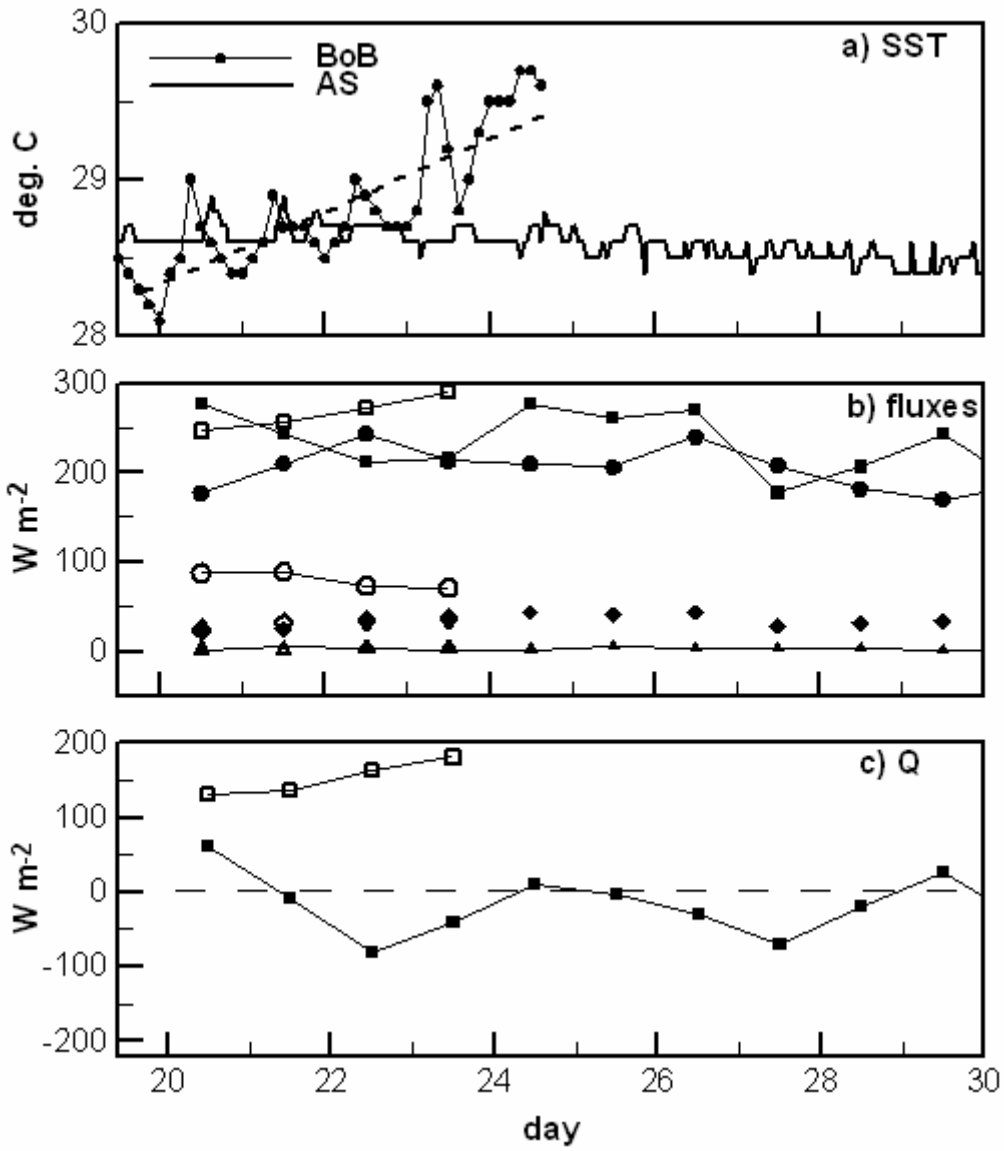

Fig. 11.14: Variation of SST, components of surface energy flux and net heat flux into water at the air-sea interface over the bay and Arabian Sea during the weak phase of convection. The fluxes shown are daily average values. The period is 19-24 August 1999 for BOBMEX and 20-30 July 2002 for ARMEX. A dashed line is drawn in panel (a) to indicate bay SST warming trend. In panels (b) and (c), open and closed symbols refer to bay and Arabian Sea data. In panel (b), square – NSW, circle – LHF, triangle – SHF, diamond – NLW. (from and Narasimha, 2007)

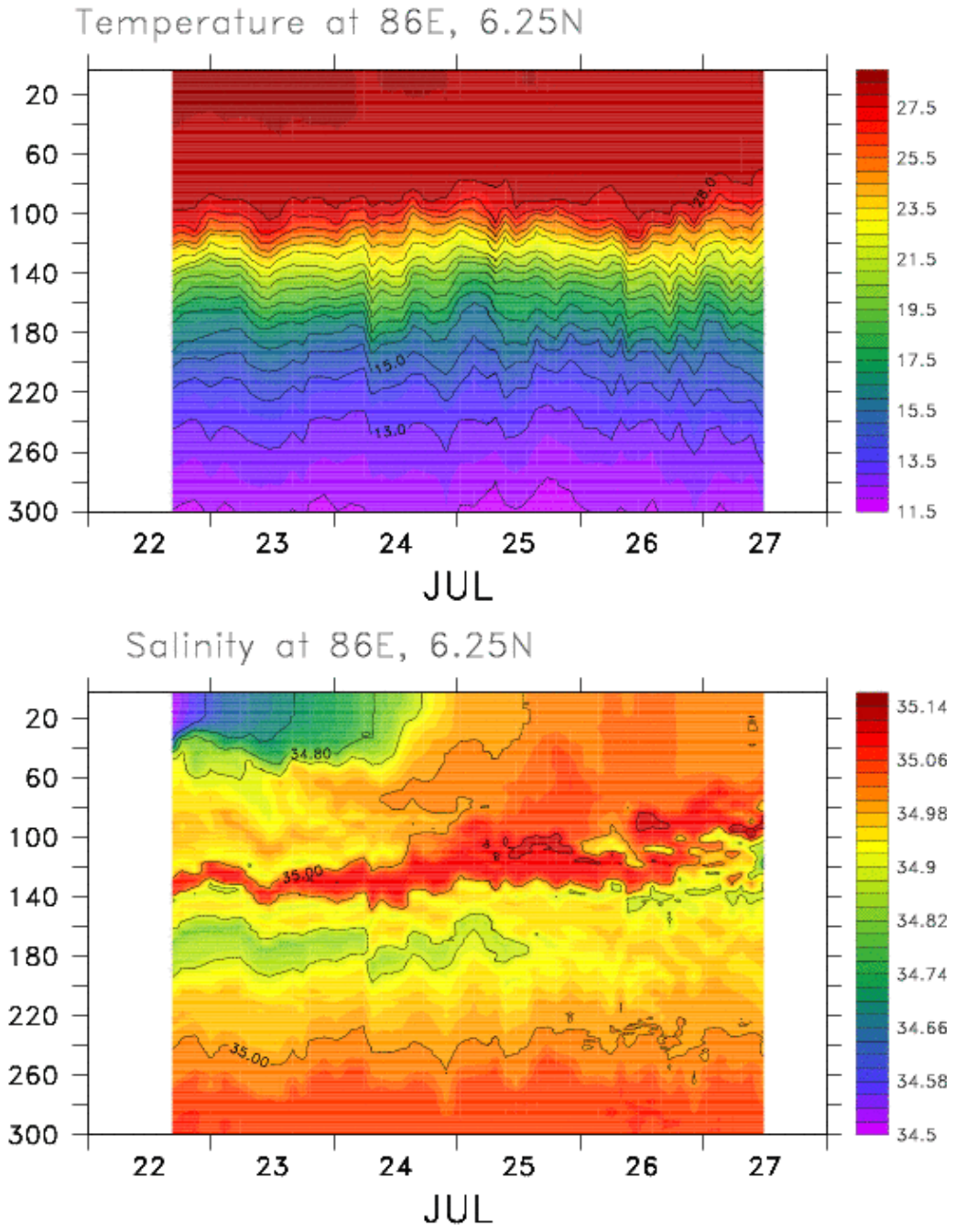

Fig.11.15: Time-height variation of ocean temperature (top panel) and salinity (bottom panel) measured in the southern Bay of Bengal during the CTCZ-Pilot. (Courtesy: P. N. Vinayachandran)

CHAPTER 12

INDIAN MONSOON AND AGRICULTURE

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12.1. Introduction

India is a land of different climates and varieties of soils affording scope for much diversity in agriculture. The geographic location, physical features and climate of the place largely determine the cropping patterns and agricultural production of the country.

The lofty Himalayas run along its entire length in the north. To the south of this barrier are the alluvial plains watered by great rivers. In the farther south, there lies the plateau of peninsular India skirted by narrow coastal strips, the Arabian Sea to the west, the Bay of Bengal to the east and the Indian Ocean to the south. The unique physical features determine the flow of monsoon current and rainfall pattern and in turn, dictate the temporal as well as spatial variation in monsoon rain as well as agricultural operations and cropping pattern.

The climatic variations in the different parts of the country from per-humid in north-east India to arid in Rajasthan with a belt of arid and semi-arid climates which extend from north to south, divide the humid climates of the west coast and the central and eastern parts of the country, where the annual rainfall is generally less than 1,000 mm. This belt is popularly known as the dry farming tract of India where innovative agricultural practices need to be introduced through technology transfer and new crops could be grown to augment food production in the country.

The areas of very heavy rainfall exist on the windward side of the Western Ghats, the Khasi Hills and the Sub-Himalayas. The north-western parts of India are the driest, Rajasthan receiving less than 500 mm rain annually. On an average a large part of the country receives rainfall less than 1000 mm per annum. Moreover, the area weightage average annual rainfall of 890 mm is the highest in any part of the world. Water for the irrigation in the semi-arid or arid belt is not sufficient and the water balance is precarious.

Indian Monsoon has two distinct phases, first one is South-West Summer and another one is North-East Winter Monsoons. Indian summer monsoon results from the differential heating of land and sea. Thus southwest monsoon gets established and moisture laden air begins to flow from the sea to the landmass that is from a region of high pressure to that of a lower one causing heavy rain over a large area of landmass. During winter, the landmass is colder than the sea and so airflow is in the opposite direction that is from land to sea. Direction of flow is from north to ocean in the south, which is called the North-East Monsoon. Since this air current originates from landmass, it carries sparse moisture and so do not cause much rainfall. North-East Monsoon flowing over the Bay of Bengal picks up moisture and results in rains in parts of Andhra Pradesh, Tamil Nadu and Kerala.

12.2. South-west summer monsoon

The South West monsoon enters India in two currents: the Bay of Bengal branch and the other one the Arabian Sea and these two currents are the source for the rains in most part of the country. The southwest monsoon after striking the Western Ghats (Fig.12.1), causes heavy rainfall over Kerala, but Tamil Nadu, which lies on the leeward side remain dry. When this monsoon reaches the Thar Desert area, the wind blows parallel to the Aravallis, failing to produce much rain over Rajasthan, the driest desert region. By the time the monsoon current reaches western parts of the country, it has lost almost all its moisture content and the rainfall decreases accordingly. Physiography is a major controlling factor of rain distribution over the country.

With the exception of Jammu and Kashmir in the extreme north and in Tamil Nadu in the south, 80-90% of the annual rainfall over the country occurs mostly during the south-west monsoon season. SW monsoon season being the major contributor of rainfall, the success of agriculture in India, depends mainly on the timely onset, the amount, intensity and the distribution of rains in the season. The date of the onset of the monsoon in different parts of the country and the intensity and the distribution of rain display large variations in time and space and also from year to year. Normally, the southwest monsoon reaches the Kerala coast by the end of May, advances along the Konkan coast in early June and extends over the entire country by the middle of July. The rains continue up to the end of September, when the south-west monsoon withdraws gradually from north-west to Southern India. In November and December, the north-east monsoon current is the main contributor to the amount of rainfall over the south-eastern portion of the Peninsular India.

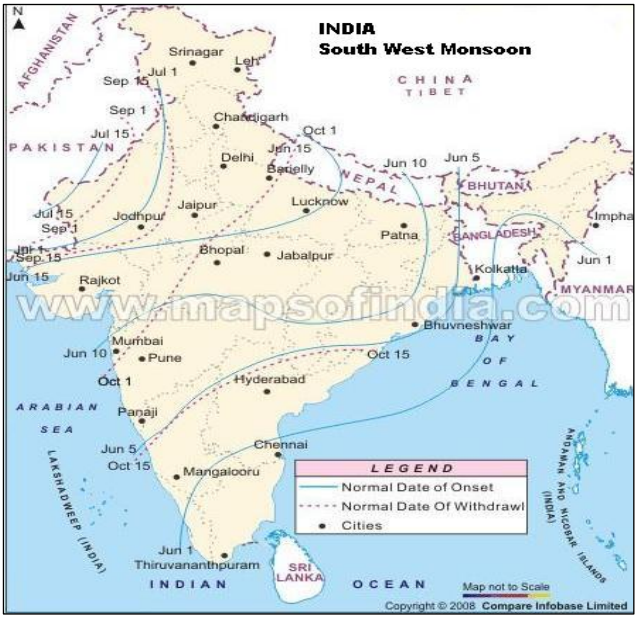


Fig. 12.1: Onset and withdrawal of monsoon

12.2.1. Monsoon and it's forecasting in India

The farmers anxiously await the onset of the monsoon and a well distributed rainfall during the season. As there is a considerable variation from year to year in the advancement of the monsoon over different regions, the predicting of the dates

of its onset and giving advance indications to the probable distribution of rainfall are of great economic significance.

Since 1886, the India Meteorological Department has been issuing long-range or seasonal forecasts for the monsoon rainfall during June-September every year. The method consists in making use of certain weather factors over different parts of the world that have significant relationship with the subsequent monsoon rainfall in India. These forecasts are, however, of limited use, since they cover large areas, such as northern India and the peninsula as a whole. Since the monsoon is a major atmospheric circulation over the globe, a dynamic approach by taking into consideration the global factors is necessary. It is also essential to understand the role of the oceanic picking up of moisture and the underlying oceanic currents in order to enable long term predictions being made with some degree of confidence.

Once the season has begun, forecasts of daily rainfall are attempted by observing and predicting the lengths of "active" and "break" periods. These are naturally occurring phases in the monsoon, lasting from 5 to 7 days, identified by fluctuations in the typical pattern. Several features associated with the active phase, brings rain to the northern Indian Plains and its west coast. They include tropical depressions in the Bay of Bengal, a low-level jet stream along the east African coast, and the variations in the monsoon trough, the area of low pressure that develops over India during the summer monsoon season.

The current monsoon forecast methods are generally either statistical or numerical. Statistical forecasts look at correlation or relationships between known phenomena and the event being analyzed, such as the earlier example of monsoon performance based on the Tibetan Plateau snow pack. However, their strength lies in steering one towards a logical result rather than in absolutes. For instance, in the case of the Asian drought of 1987, the monsoon was weak, resulting in one of the worst droughts of the century. But the El Niño which caused the disruption in world weather patterns was not as strong as the 1982-83 events. In contrast, a numerical model is a mathematical simulation of the atmosphere, represented by known physical relationships such as the equations of motion and thermo-dynamics etc. For example, the various models used by meteorologists to provide temperature and

precipitation forecasts for 5 days are numerical models, run on supercomputers utilizing the large amounts of data being processed.

Improved techniques of weather forecasting could provide timely information for agricultural operations so that desired and appropriate steps may be taken well in advance. A “16 parameter model” subsequently modified to “5 parameter ensembled model” developed by a forecast team of Indian Meteorologists has enabled near successful prediction of monsoon rains in a given time and area.

12.2.2. Drought

Drought is said to have occurred when the principal monsoon fails or is deficient. It leads to crop failure due to insufficient soil moisture, shortage of drinking water as well as undue hardship due to non availability of surface water for day to day household consumption to the rural and urban community.

Drought prone areas are located mostly in arid and semi-arid regions of India (Fig.12.2). Frequency of occurrence of drought in arid region is the highest followed by in semi-arid region. However, droughts though not frequently occurring in humid and sub-humid region yet cause devastation when it occurs in moderate to severe intensity.

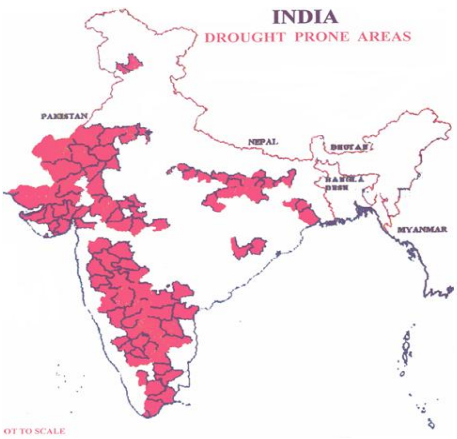


Fig.12.2: Drought Prone Areas of India

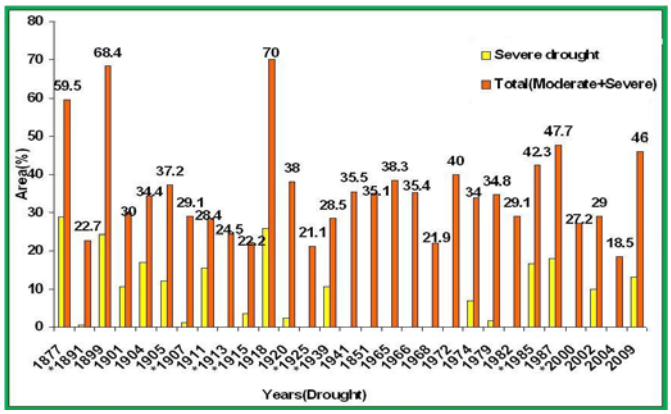


Fig. 12.3: Country’s Total area (%) Under Drought and Severe Drought

Drought prone areas are located mostly in arid and semi-arid regions of India (Fig. 12.2). Frequency of occurrence of drought in arid region is the highest followed by in semi-arid region. However, droughts though not frequently occurring in humid and sub-humid region yet cause devastation when it occurs in moderate to severe intensity.

Fig. 12.3 shows the area affected under moderate and severe drought. It is interesting to note that on an average about 30% the country's total area was affected under drought with exceptionally largest (about 60-70%) area only in 3 out of 29 drought years experienced so far from 1877 to 2009.

12.2.2.1. Drought in monsoon season

The southwest monsoon has a stranglehold on agriculture, the Indian economy and consequently, the livelihoods of a vast majority of the rural populace. An overwhelming majority of cropped area in India (around 68%) falls within the medium and low rainfall ranges regions. Large areas are therefore affected if the southwest monsoon plays truant. Most parts of peninsular, central and northwest India regions are most prone to periodic drought. These regions receive less than 1,000 mm of rainfall. The drought of 1965-67 and 1979-80 affected relatively high-rainfall regions, while the drought of 1972, 1987, 2002, 2004 and 2009 affected low-rainfall regions, mostly semi-arid and sub-humid regions. The percentage departure of rainfall during the major drought years are presented in Table 12.1.

Table 12.1: Percentage departure of rainfall during major drought years

	1918	1965	1972	1979	1987	2002	2004	2009
All India	-25	-18	-24	-18	-19	-19	-14	-22
Northwest	-47	-35	-31	-42	-44	-26	-21	-35
Central	-30	-21	-23	-17	-30	-17	-17	-20
East	22	3	-19	-1	31	-9	-9	-23
Peninsular	-39	-8	-19	-5	-19	-33	-33	-06

Drought poses many problems. Irrigation facilities available in country are limited and therefore, when drought occurs, they cause partial or complete crop failure. If failures occur in consecutive years, it becomes a national calamity, putting great strain on the economy of the country.

According to the practice followed by the India Meteorological Department, drought occurs over an area where the annual rainfall is less than 75 percent of the normal. When the annual rain is less than 50 percent of the normal, it is called severe drought. Meteorological drought happens when the actual rainfall in an area is significantly less than the climatological mean of that area. Attempts to quantify drought in the form of an index have been made by using various techniques. Pilot studies have been conducted in India, with special reference to Bihar State, in order to arrive at a rational definition. The Palmer drought index, which takes into account rainfall, evapo-transpiration and soil moisture, is considered a comprehensive approach to this problem. The computation of the Palmer drought index has been carried out for the different sub-divisions of the country. The computations show that, on the average, drought is experienced on 20-25 percent of the days in each of the months of the kharif season over large areas of the country.

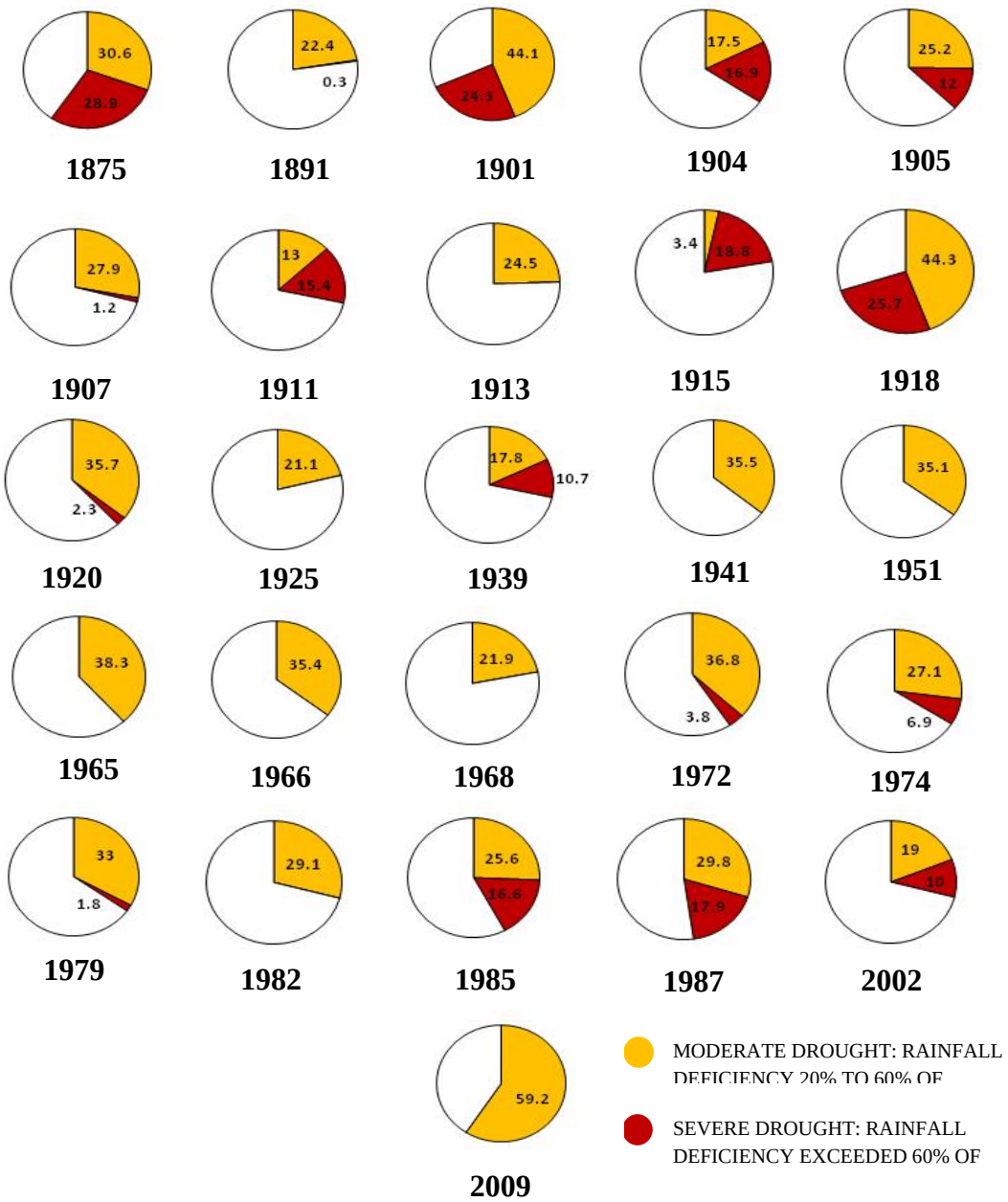
The power-spectrum analysis of the rainfall series and that of the Palmer drought index series show some relation to the quasi-biennial and the eleven year sunspot cycles in some areas. The amplitude of the cycle is, however, too small to exert a significant influence.

It has been found that the rainfall series is of a random nature with no significant trend. Hence it would not be feasible statistically to predict the amounts of rainfall during any particular year from a knowledge of the past rainfall occurrences alone. The probabilities of the rainfall occurrences can, however, be used to estimate the risk of poor rainfall and the attendant drought.

12.2.2.2. Impact of drought

Drought has both direct and indirect impacts on the economic, social and environmental fabric of the country. Depending on its reach and scale it could bring about social unrest. The immediate visible impact of monsoon failure leading to drought is felt by the agricultural sector.

Fig.12. 4: Drought years in India with percentage of area affected since 1875 based on June-September rainfall



The impact passes on to other sectors, including industry, through one or more of the following routes:

- A shortage of raw material supplies to agro-based industries.
- Reduced rural demand for industrial/consumer products due to reduced agricultural incomes.
- Potential shift in public sector resource allocation from investment expenditure to financing of drought relief measures.

Rare droughts of severe intensity occurred once in 32 years, with almost every third year being a drought year. The 1987 drought was one of the worst droughts of the century. Recently drought of the year 2002 ranked fifth in terms of magnitude and caused reduction in food grain production to the tune of 13% in India. In the year 2009 the rainfall deficiency was -21.8% and percentage area affected by moderate drought was 59.2% (Fig. 12.4).

12.2.2.2.1. Impact of drought on agriculture

Drought has an immediate effect on the recharge of soil moisture resulting in reductions of stream flow, reservoir levels and irrigation potential and even the availability of drinking water from wells. The acreage under food crops is also affected by land quality. The cultivation of lands subject to a high degree of rainfall variability makes them extremely susceptible to wind erosion (and desertification) during prolonged drought episodes, as the bare soil lacks the dense vegetative cover necessary to minimize the effects of Aeolian processes.

12.2.2.2.2. Impact of drought on kharif crops during 2002, 2007, 2008 and 2009

Monsoon 2002 was one of the worst drought affected year when only 15 out of 36 meteorological sub-divisions had received normal to excess rainfall during June to September. Rainfall deficiency during July 2002 was 54%, which adversely affected the performance of kharif crops. The crops suffered moderate to severe during 2002 in Karnataka, Tamil Nadu (TN); in parts of Andhra Pradesh (A.P.) and partially in Kerala. In contrast, monsoon 2003 was marked by well distributed normal rainfall throughout the country with 33 out of 36 meteorological sub-divisions

received normal or excess rainfall. Though, the overall performance of kharif crops was better during 2003 in southern India, crops suffered in interior Karnataka due to intra-seasonal moderate to severe aridity in some months. Major Kharif crops suffered marginally in Kerala also during the same year. During 2002, in southern India, crops were affected by moderate drought, except in Kerala and Coastal Karnataka (CK) where kharif crops were moderately satisfactory due to adequate rainfall. Moderate to severe aridity prevailed in the region (excluding Kerala and Coastal Karnataka) during July and by the end of September 2002. However, during kharif 2002, contingent crops, viz., maize, sunflower, soybean in Andhra Pradesh and pulses in Karnataka performed well. The overall crop performance in southern India was partially satisfactory during 2002 which improved considerably during 2003 when well distributed normal rainfall was received in most parts of these states.

In 2007, with the almost normal onset (28th May) of monsoon over Kerala, land preparation/sowing/transplanting of kharif crops started as usual in the southern part of the country. After a hiatus of 9 days, further advance of monsoon took place on 8th June. For the period from 1st June to 19th September, 30 out of 36 meteorological subdivisions in the country received excess or normal rains; only 6 subdivisions received deficient rain – the deficiency being mild to moderate Delay in the progress of monsoon in several parts of the country in June considerably delayed the sowing operations in rainfed areas. Such delay in advance of monsoon affected the sowing operation in central India and adjoining areas in June. Sowing of crops was made up due to good rainfall in the subsequent weeks.

During the year 2008, the South-West monsoon arrived over Kerala on 31st May, one day ahead of the normal date. Further advance took place quite rapidly mainly due to depression over the East Central Arabian Sea and a well marked low pressure area over Saurashtra & Kutch and neighbourhood. By June 16, South-West monsoon had covered most parts of the country except for some parts of Rajasthan. Subsequently, there was a hiatus in the further advance due to the weakening of the monsoon. The monsoon covered the entire country on July 10, nearly 5 days ahead of the normal schedule. The rainfall during the monsoon months was, however uneven during June – September 2008. Most of the districts of Marathwada, Madhya Maharashtra and some parts of Karnataka and Gujarat experienced delayed and

deficient monsoon rainfall which caused reduction in net sown area in those regions. Rainfall was deficient in most of the weeks of the sowing months i.e. June, July 2008. The sowing of kharif crops was reported to decline in some pockets of Marathwada, Madhya Maharashtra, Karnataka and Gujarat during 2008.

The year 2009 was the third highest deficient all India monsoon rainfall year during the period 1901-2009. The season (JJAS) rainfall deficiency for the country as a whole was -22% of Long Period Average (LPA). In 2009 monsoon arrived earlier, hitting the Kerala coast on 23rd May, ahead of the usual date on 1st June and it covered most parts of Karnataka, coastal AP, Rayalaseema, most parts of West Bengal and north eastern States by the first week of June. But the progress of the monsoon was slowed down afterwards. The monsoon was stagnated over southern peninsular region for more than two weeks and delayed for 7-10 days in central India and parts of north India. Further delay in the monsoon affected the sowing of kharif crops in most parts of the country (Maharashtra, Gujarat, Andhra Pradesh, Orissa, Bihar, Chhattisgarh, Jharkhand and UP). Long dry spell during July and 1st fortnight of August in north Interior Karnataka & South interior Karnataka hampered timely sowing to a great extent and normal growth of early sown crops. This has delayed transplantation of paddy also by 15 to 30 days resulting in less numbers of tillers in eastern Vidarbha, Konkan and NE region. Delayed onset and deficient rainfall in June and July has caused reduction in net sown area by about 15%.

Thus 2009, was drought year due to shortfall in rainfall. There was a fall in GDP from 6% to 5.5%. The acreage under “total food grains” was 563 lakh hectares as against 635 lakh hectares in 2008. Paddy was badly affected. The total acreage was 289 lakh hectares as against 358 lakh hectares in 2008. But there was a pick-up in sowing due to late rain in states like Orissa, Chhattisgarh and West Bengal. Oilseeds showed a big shortfall of 7 lakh hectares as against 177.71 lakh hectares in 2008 with groundnut showing the biggest fall. In Maharashtra, the average sowing area for kharif fell from 140.74 lakh hectare to 134.17 lakh hectares. Rice production declined by about 15%. Cyclone Phyan which passed over rice-producing areas was responsible for the damage to the crop. The unseasonal rains also affected the grape production. However, the rains did help the Rabi crops like

tur, jowar, whole Bengal gram, sweet sorghum and wheat. The oilseed production increased from 22 lakh tonne to 25 lakh tonne.

12.2.3. Drought assessment

Regional drought assessment is conventionally based on drought indices for the identification of drought intensity, duration and areal extent. Criteria for drought assessment include weather data including soil moisture, information on crop state and stages, area sown under major crops in each districts and drought indices. All this information helps to declare drought and relief assessment as well as management to be undertaken in a region. Fig.12.5 shows the flowchart for drought assessment and declaration.

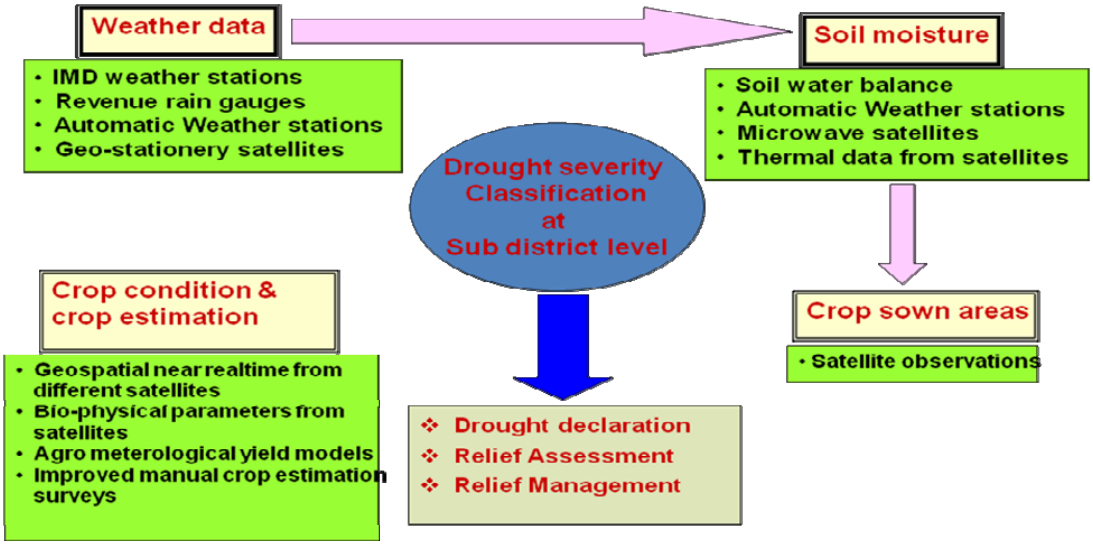


Fig.12.5: Flow chart for drought assessment & declaration

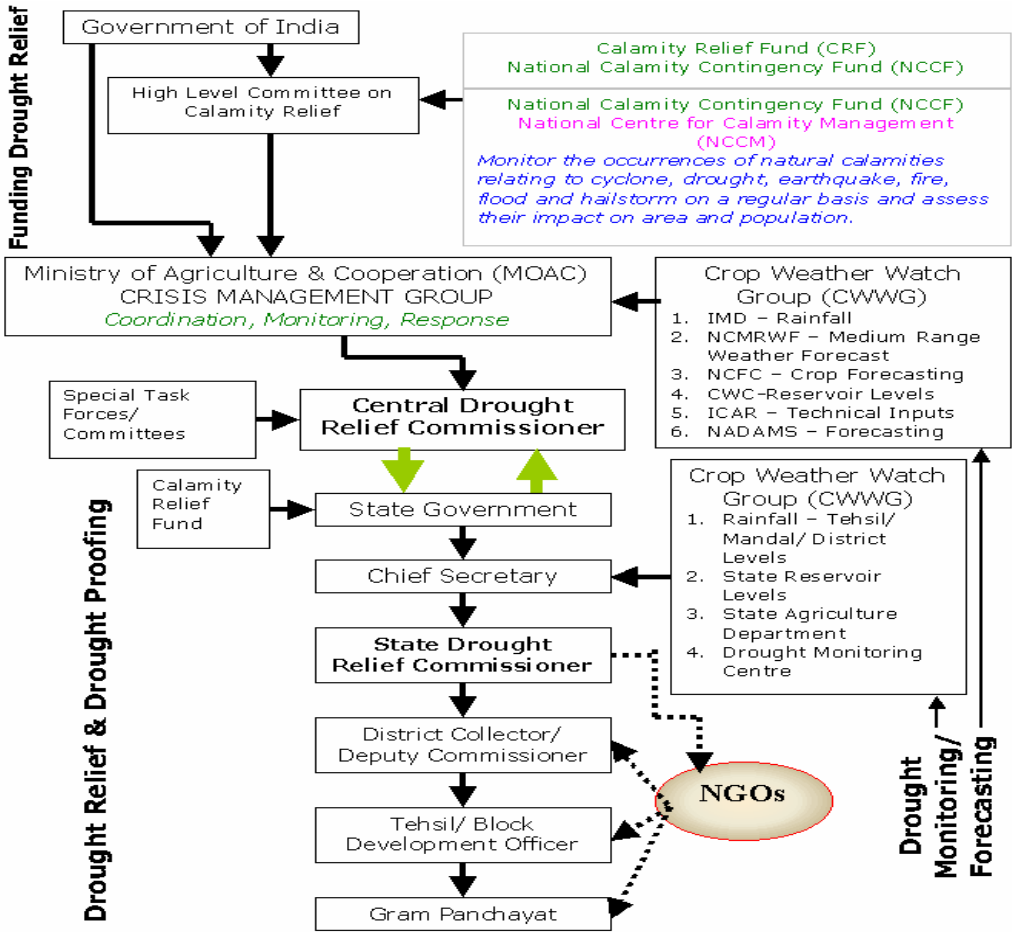


Fig.12.6: Organization of EWS for Drought Management

Fig.12.6 indicates the role of IMD and other organization in drought assessment and providing Early Warning through CWWG.

12.2.3.1. Role of IMD in drought assessment

IMD plays a vital role in Drought warning and monitoring by Early Warning through LRF, providing probabilities of dry/wet spells during crop growing season, monitoring through Aridity Anomaly Index (AAI) and mitigation through Agromet Advisory Services (AAS). IMD has also Desert Locust Monitoring (DLM) in Northwest India and network of AWS for real time data. It has also undertaken the project of 'Forecasting Agricultural output using Space, Agro-Meteorology and Land based observations' (FASAL) for forecasting of Agricultural yield. IMD identifies meteorological drought for subdivisions every year based on rainfall analysis. IMD monitors agricultural drought once every two weeks on a real-time basis during the

main crop seasons (kharif and rabi). Information on rainfall from IMD is taken as most important input for drought monitoring in Crop Weather Watch Group (CWWG) meeting.

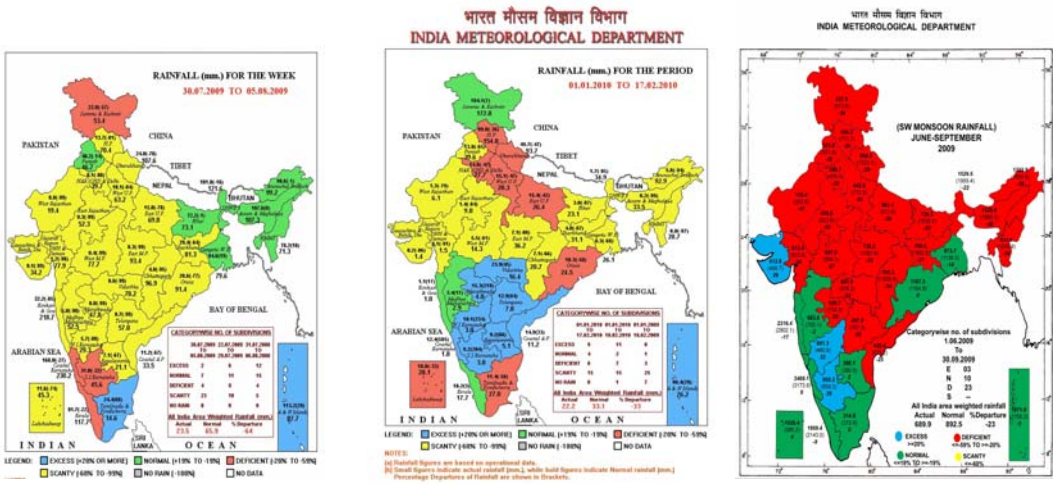


Fig. 12.7: Rainfall maps prepared by IMD

IMD prepares rainfall maps (Fig.12.7) on sub-divisional basis every week throughout the year. These maps show the rainfall received during a week and corresponding departures from normal. During monsoon season, these maps are indicative of progressive occurrence of drought. Cumulative rainfall and its departure during the monsoon season also indicate the severity of drought in the region.

12.2.3.2. Aridity Anomaly Index (AI)

Aridity Anomaly Index (AI) is a useful index for monitoring agricultural drought. Thornthwaite's water balance concept is used to monitor the incidence, spread, intensification, and recession of drought. AI is given as:

$$AI = \frac{PE - AE}{PE} \times 100$$

Where PE is potential evapotranspiration calculated with the help of Penman's formula, which takes into account mean temperature, incoming solar

radiation, relative humidity and wind speed. AE is actual evapotranspiration calculated according to Thornthwaite's water balance technique taking into account PE, actual rainfall and field capacity of the soil.

The arid areas are demarcated as follows:

Aridity Anomaly	**Areas**
0 or negative	Non-arid
1-25	Mild arid
26-50	Moderate arid
> 50	Severe arid

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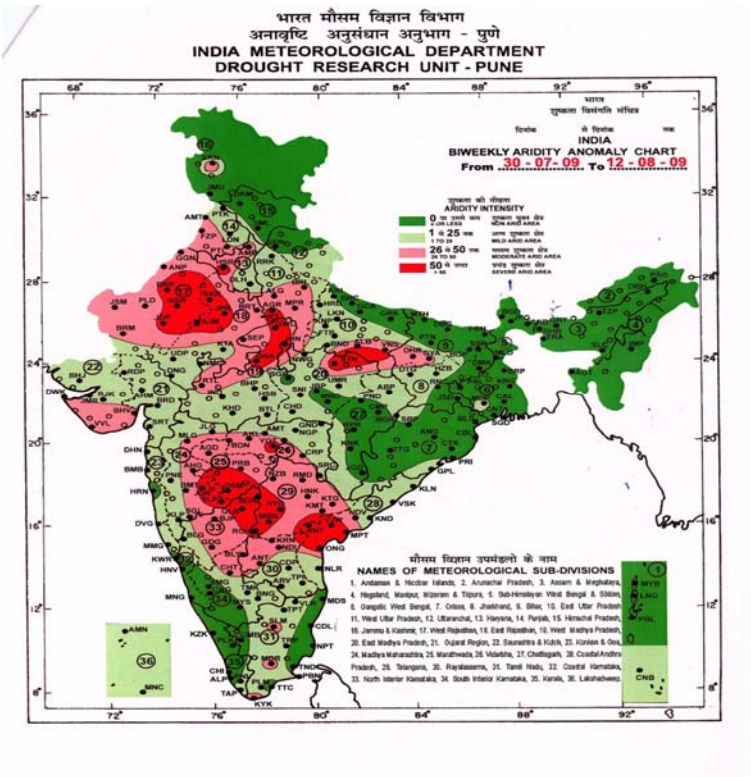

Fig.12.8: Aridity Anomaly Map

Fig. 12.8 shows the aridity anomaly map of India for the fortnight 30-7-09 to 12-8-09. With the help of aridity anomalies, crop stress conditions in various parts of the country are monitored during the monsoon seasons. These anomalies are also used for crop planning and in the early warning system during drought.

12.2.4. Drought management

Drought in India is a recurring phenomenon and occurs almost every year in some parts with varying spread and severity. Amongst various droughts, viz., meteorological, hydrological and agricultural, the later one plays crucial role as it is directly linked with the food production and food security in the country. Thus drought monitoring, assessment and management are important and need greater emphasis for contingency planning. Drought management in India amounted to 'crisis management', a reactive approach to tackle disasters when they are already upon the nation. Studies made have recommended for a policy with 'preparedness' as its cornerstone. The guiding principle is– preparedness over insurance, insurance over relief and incentives over regulation (in respect of crop failure, crop management).

Role of different institutes, agencies and stake holders in this endeavour though different but needs interdisciplinary and collaborative approach to tackle drought, especially agricultural drought. District level multimodel ensemble forecast and Agromet Advisories issued through 130 Agromet Field Units for tackling drought particularly early and mid-season droughts through contingency planning plays vital role in Indian Agriculture. These Agromet advisories include drought management practices which are:

- community nurseries at points where water is available
- transplantation
- sowing of alternate crops/varieties
- ratooning or thinning of crops
- soil mulching if the break in the monsoon is very brief
- weed control
- in situ water harvesting and/or run-off recycling
- broad beds and furrows
- graded border strips
- inter-row and inter-plot water harvesting systems
- intercropping systems for areas where the growing season is generally 20 to 30 weeks

- alternate land use systems
- development of agriculture on the basis of the watershed approach
- alley cropping
- agro-horticultural systems
- watershed approaches for resource improvement and use
- water resources development
- treatment of lands with soil conservation measures
- alternate land use systems
- forage production.

12.2.5. Floods

Occasionally, many parts of the country, particularly the areas drained by large river systems, suffer from devastating floods. Widespread floods occurred in 1878, 1892 and 1917 in the many provinces of India and again in many states of the country in 1954. The National Commission on Floods has assessed the flood prone area in India to be around 12 per cent of the total area. Type of floods in India includes Snow-melt floods, Flash floods / cloudburst floods; Monsoon floods of Single & multiple events, Cyclone floods and Floods due to dam bursts / failure.
Flooding is caused by several factors. Some of the main sources are:

- Excessive rainfall in river catchments or concentration of runoff from the tributaries and river carrying flows in excess of their capacities.
- Backing of water in tributaries at their confluence with the main river.
- Synchronization of flood peaks of the main rivers and tributaries.
- Intense rainfall when river is flowing full.
- Poor natural drainage.
- Landslides leading to obstruction of flow and change in the river course.
- Cyclone and very intense rainfall when the El Nino effect is on a decline.

12.2.5.1. Impacts of flood on agriculture

Flood causes widespread losses of life (human and animal), property, structures, agricultural crop loss and economic losses to the tune of crores of

rupees. Table 12.2 depicts some of the losses based on the information received from ministry of agriculture of respective states. Impact of flood on agriculture can be felt in many ways. Some of them are:

- Total loss of crop due to erosion or rotting due to long stagnation of water.
- Depletion of oxygen available to the plant root zones.
- Creates anaerobic soil conditions that can have significant impacts on vegetation.
- Chemical reactions in anaerobic soils lead to a reduction in nitrate and the formation of nitrogen gas.
- The denitrification can be a significant cause of loss of plant vigour and growth following flooding.
- Causes several physical, chemical and biological changes, some of which are not reversible

Table 12.2: Impact of flood in different States

State	Year	Loss of life		Homes damaged		Structure damage	Area affected Land (in ha)	Population affected in Lakh	Crop Area affected (in ha)	Loss of crops (in Rs. Crores)
		Human	Cattle	No. of Homes	Loss in Rs. (Crores)					
Karnataka	1993	49	-	1466	-	12.35	-	3000	40325	-
	1994	60	2.399	17486	5.32	65736	128000	-	-	25.800
	1995	11	15	734	-	-	-	-	5000	-
	1996	74	101	3348	1.890	20.93	-	-	69 000	2.620
	1997	92	120	6333	-	-	19000	-	19000	-
Kerala	1993	209	824	87000	12.825	105.000	233000	-	-	144.5
	1994	74	-	4318	-	-	78000	-	12000	-
	1995	88	-	18729	6.655	483.262	55000	4.76	40000	68.908
	1996	149	183	16109	-	-	385000	9.5	90000	-
	1997	42	104	11690	2.12	0.588	7000	-	-	5.090
Gujarat	1993	119	662	4895	-	10.480	43000	-	-	12.120
	1994	108	-	435	-	-	-	-	-	-
	1995	41	390	400	-	-	-	-	-	-
	1996	32	1962	24644	3.770	62.97	1822000	164.8	3000	60.50
	1997	55	795	32877	-	-	-	21.18	-	-
Bihar	1993	97	418	228150	122.050	31.800	1521000	-	-	150.920
	1994	83	34	2807	2.178	1.313	601000	-	-	48.068

	1995	528	4122	297765	-	-	926000	87.09	424000	-
	1996	218	110	113974	14.005	8.462	1189000	69.65	734000	70.314
	1997	163	151	174379	-	-	1471000	-	655000	-
Uttar Pradesh	1981	352	1195	-	163.3	-	1502819	-	-	-
	1982	382	-	-	1.5	-	1386000	-	-	-
	1983	376	-	-	3.995	-	300001	-	-	-
	1984	149	-	-	1.509	-	21000	-	-	-
	1985	244	467	-	24	-	1200000	-	-	-
	1986	201	104	-	-	-	532000	-	-	-
	1987	163	-	-	055	-		-	-	-
	1988	611	-	-	-	-	228000	-	-	-
	1989	160	516	-	3.0	-	594000	-	-	-
	1990	210	2000	-	84.507	-		-	-	-
	1991	106	207	-	49.07	-	153260	-	-	-
	1992	107	1167	-	340	-	245100	-	-	-
	1993	206	2347	-	500	-	481000	-	-	-
	1994	31	-	-	3	-		-	-	-
	1995	166	-	-	1.6	-	799000	-	-	-
Maharashtra	1970	-	-	-	-	-	-	-	-	2
	1971	-	-	-	-	-	-	-	-	-
	1972	-	-	-	-	-	-	-	-	-
	1973	-	-	-	-	-	-	-	-	-
	1974	69	-	-	-	6.0	-	-	-	-
	1975	-	-	-	-	-	-	-	-	-
	1976	-	-	381	-	-	-	-	10000	0.9
	1977	4	-	-	-	-	-	-	-	-
	1978	49	-	6817	-	-	-	0.31	-	-
	1979	72	-	14440	-	-	-	-	-	68
	1980	102	-	251	-	-	-	-	11300	.32
	1981	77	13300	92	-	-	-	-	1962	.045
	1982	10	140	100	0.04	-	-	-	-	.0005
	1983	381	163	2128	0.47	4.7	-	-	10669	4.47
	1984	375	15000	132	0.005	-	-	-	21500	10
	1985	56	-	-	-	-	-	-	-	-
	1986	82	-	16348	19.65	-	-	-	3000	-
	1987	16	-	-	-	-	-	-	-	-
	1988	156	-	569	-	4.15	-	-	>100000	-
	1989	921	14629	110000	100	85.68	-	-	-	50.28
	1990	121	199	9225	-	-	55000	-	55000	-
	1991	614	4466	12996	-	300	-	-	70000	-
	1992	128	1225	41000	6.22	17.81	-	-	-	7.797
	1993	75	33	1156	0.04	5.22	-	-	12000	1.17

	1994	218	693	15007	6.123	20.61	-	-	41000	-
	1995	40	84	3534	-	-	9000	-	3000	-
	1996	96	13	2957	5.53	-	-	-	184600	3.72
	1997	199	318	35034	-	-	49000	-	49000	-

12.2.5.2. Early warning system

Flood forecasting is a non-structural measure to control/minimizing loss of lives and damage due to flood. There are 10 Flood Meteorological Offices of IMD (Fig. 12.9) actively involved in flood forecasting. The Central Water Commission is maintaining a network of 157 flood-forecasting stations on various river basins in the country.

Fig. 12.9: Flood Meteorological Offices in India

Table 12.3: Flood Frequency, Casualties and Economic Damages for the period 1981-05

West Coast Region						
Decade	Konkan / Goa (flood freq)	Coastal Karnataka (Flood freq)	Kerala (flood freq)	Total (flood Freq)	Causalities	Economic losses (Rupees in Crores)
1981-1990	49	10	41	100	827	319.21
1991-2000	60	23	82	165	2486	675.7
2000-2005	15	7	23	45	1468	601

Table 12.3 shows the Flood frequency, casualties and economic damages for the period 1981-05. It can be seen from the table that maximum flood frequency is observed in Kerala. The maximum economic loss was during the decade 1991-2000.

Table 12.4 indicates the occurrence of major flood events during 1992-2005. Mumbai & neighborhood (Konkan/Goa) experienced highest rainfall of 944 mm and suffered property loss of 450 crores in 2005.

NA-Information not available (Source: De et al., 2005)

12.2.5.3. Adaptation strategies for flood

- Raft (bamboo and banana plant) for people/livestock/poultry
- Afforestation of the upper catchments areas of the rivers
- Construction of river embankments and the execution of multi-purpose river valley projects.
- Construction of storage dams, reservoirs, embankments, drainage structures as required at suitable locations

Table 12.4: Occurrence of Major Flood Events (1992-2005)

Sub-divisions	Year	Rainfall in mm	Deaths	Amount of properties destroyed
Aizawl (Nagaland, Manipur, Mizoram, Tripura)	9-10th Aug 1992	690	106	NA
Allapuzha, Kollam & Trivandrum (Kerala)	10-11th Oct 1992	240	250	NA
Dakshin Kannad (Coastal Karnataka)	1st week Oct, 1993	460	34	11.35 Crores
East Siang, upper Siang, and Lohit (Arunachal Pradesh	11-16th June 2000	234	2,024	1.5 Crores
Assam (Assam/Meghalaya)	1st week Oct 2004	722	500	NA
Mumbai & neighborhood (Konkan/Goa)	25-27th July 2005	944	927	450 Crores

- Application of fertilizers like urea, muriate of potash for standing crops after draining out excess water.
- Skip, if deemed appropriate, entire cropping season
- Sowing of short duration non-rice minor crop
- If the immediate cropping season is lost, start early for the next season
- Provide weather based agricultural insurance.

12.3. Importance of monsoon in agriculture

Indian economy is largely dependent on agriculture as more than 70% of the total population depends on agriculture for their livelihood. In spite of significant advances since independence to bring more area under sustainable irrigation and the development of new high yielding crop varieties, Indian agriculture is highly dependent on monsoon rainfall. Thus India has taken enormous strides toward freeing herself from dependence of the vagaries of the monsoons. The monsoon affects the Indian agriculture in a substantial manner. The massive impact of the monsoon on Indian economy is indeed very apparent. Therefore the country's budget is referred to as "a gamble in monsoon rains". It is a well-known fact that success of Indian agriculture depends largely on the quantum and distribution of SW monsoon rainfall both in time and space. The monsoon being the epitome of the Indian agriculture, each good monsoon year brought a full harvest, created life from desolation and brought hope for the poor peasantry. Its failure means despair and fear of famine conditions (Sikka, 1999).

Characterized by either being too wet or too dry, an abnormal monsoon can result in the loss of seasonal employment, shortages of food and income, and the spread of disease. That abnormal and violent weather in monsoon season can also be felt in subsequent years. India has made concerted efforts to improve both short and long-term forecasts in order to alert farmers and manage resources. Nevertheless, even today it is not far wrong to say that "In India scarcity is only a missed monsoon away".

It is felt and realized that an abnormally wet monsoon is better than a dry one. Crop production is directly proportional to the seasonal rains. For example in 1965-

66, rainfall was about twenty percent below normal in 14 meteorological subdivisions of India causing rice production to decrease substantially and famine to ensue. Whereas a wet monsoon, despite causing much human misery and suffering, improves groundwater availability and allows good crops to grow after the recession of the excess waters.

The population of India is expected to increase to about 1.5 billion by 2030. Food production must increase by 5 million tons per year to keep pace with this increase and ensure food security. Much of this extra production will need to come from rain-fed agriculture that comprises 70% of the farmed land. But these rain-fed farming systems are acutely vulnerable to climate variability and change. Current ability to forecast crop yields for a season ahead is very limited and improved predictions of rainfall and its space-time characteristics are vital for making progress.

In India projections of food production in future are limited by our knowledge of how climate change will impact on the complex biophysical, social and economic processes that interact to influence food chains, and on specific science questions concerned with how crops will respond to elevated CO_2, extremes of temperature and imbalanced water supply. More basic research on crop responses and better assessment of the potential to transgress key thresholds associated with weather events are needed. The evidence to date suggests that the changes in water and temperature described above will have serious consequences for agriculture. The special IPCC report 2007 has suggested that the global mean surface temperature will rise by 1.0- 3.5^{0}C by 2100 due to increase in carbon dioxide concentration and other Green House Gasses in atmosphere. This will indirectly affect the food production. The possible effect of Climate change on the wheat production is illustrated in Fig. 12.10 (Aggarwal, 2000).

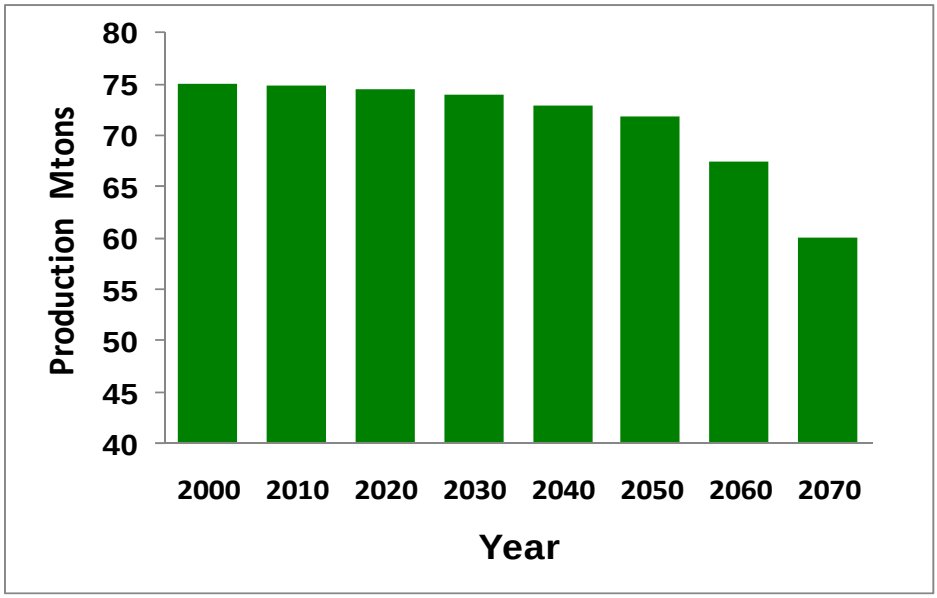

Fig.12.10: Wheat production under climate change

The impact of less rainy days and increased intensity of rainfall events is to reduce the amount of water available for crop growth, since more water is likely to be lost to runoff and drainage. This in turn leads to a reduction in crop yield. Changes in the active/break cycles of the monsoon will also lead to reduced yield if a break occurs at a time when water availability is critical for the crop. The climate change scenario for India is illustrated in Table. 12.5

Table 12.5: Increase (decrease) in Temperature and percentage change in Rainfall under climate change (Aggarwal, 2000)

Year	Season	Increase in Temperature, °C		Change in Rainfall, %	
		Lowest	*Highest*	*Lowest*	*Highest*
2020s	Rabi	1.08	1.54	-1.95	4.36
	Kharif	0.87	1.12	1.81	5.10
2050s	Rabi	2.54	3.18	-9.22	3.82
	Kharif	1.81	2.37	7.18	10.52
2080s	Rabi	4.14	6.31	-24.83	-4.50
	Kharif	2.91	4.62	10.10	15.18

Adaptation to these changes is a possible solution to such problem. A change of crop variety can mitigate to some extent the impact of extreme weather events. Mean changes in planting date may also provide some adaptation. Weather forecasts can aid the shorter-term scheduling of planting and irrigation. The potential to forecast a season ahead exists in some regions and benefit can be derived by dissemination of information in a timely fashion through Integrated Agromet Advisory Services (IAAS) recently launched in all Agroclimatic zones through dedicated 130 Agromet Field Units (AMFU).

India's economy and societal infrastructures are finely tuned to the remarkable stability of the monsoon, with the consequence that vulnerability to small changes in monsoon rainfall is very high. In future, the pressures of an increasing population will bring additional stresses on society and the environment with implications for water resources, health and food security. Consequently,
climate change and the potential for the monsoon to become more unpredictable have major implications for India, itself, and for economies, worldwide.

Monsoon which impact agriculture is an important segment of the economy because:

- It generates about a quarter of the GDP.
- Provides raw materials for a fifth of the industry.
- Generate demand for industrial goods.
- Nearly 40 percent of the manufactured consumer goods are sold in rural areas.
- 70% of the population depends on agriculture for livelihood.

India has made lot of progress in agriculture. It has gone through a green revolution, a white revolution, a yellow revolution and a blue revolution. Today, India is the largest producer of milk, fruits, cashew nuts, coconuts and tea in the world, the second largest producer of wheat, vegetables, sugar and fish and the third largest producer of tobacco and rice. The per capita availability of food grains has risen in the country from 350 gm in 1951 to near about 400 gm per day now, of milk from

less than 125 gm to 226 gm per day and of eggs from 5 to 30 per annum despite the increase in population from 35 crores to more than 100 crores. At present only about 30 percent of the farmers are able to derive any benefits of extension services provided by various government agencies and every year about 20 percent of the crop is lost due to mishandling, spillage, floods, droughts and pests and diseases. In fruits and vegetables the loss is estimated around 30 percent.

Agriculture accounts for about 20 percent of the GDP. Over the years, the agriculture sector has not received as much attention as other sectors in services and manufacturing. The emerging areas in agriculture like horticulture, floriculture, organic farming, genetic engineering, food processing, branding and packaging have high potentials of growth. Development of rural infrastructure, rural extension services, agro-based and food processing industries are essential for generating employment and reducing poverty. However, all these are linked to agricultural production which is directly related to good monsoon rainfall.

12.3.1. Food grain production

Food grains include rice, wheat, corn (maize), coarse grains (sorghum and millet) and pulses (beans, dried peas and lentils). Sorghum and millet, the principal coarse grains, are dry land crops most frequently grown as staples in central and western India.

In FY 1990, approximately 127.5 million hectares were sown with food grains, about 75 percent of the total planted area. The total number of hectares increased by 31 percent over the forty-year period from FY 1950 to FY 1990. CMIE expects food grain production to reach 234 million tones as compared with 230 million tones during FY 08, an increase of 2.2%. Kharif acreage was down by 2.4% in 2008 and about 10% in 2009 due to less and erratic rain in some states in the crucial sowing month of June and July. But increased acreage and favourable weather conditions are expected to boost Rabi production. Growth in total crop production during FY 09 stands revised downwards to about 10% from our earlier estimate. The final estimate for FY 08 soybean production was 10.97 million tonne against 9.4 million tonne as

per the fourth advance estimate. Sugarcane production was raised to 348.2 million tonne from 340.6 million tonne.

For the past few years India has been experiencing fluctuating food grains production but it had never witnessed such a steep fall as in 2002-03 and 2009-10 when the decline was estimated to be anywhere between 10-15 percent. The month of July that normally records highest rainfall in monsoon season in India, registered the lowest rainfall in 2002 in the past 100 years. July normally receives about 30 percent of the monsoon rainfall and the shortfall in 2002 was as high as 49 percent. In 2002 the monsoon rains failed during July, resulting in a seasonal rainfall deficit of 19% and causing profound loss of agricultural production with a drop of over 3% in India's GDP. Against 75.05 percent of the total Full Reservoir Level (FRL) which is the average of last 10 years, country's reservoir storage at the end of the monsoon 2002 stood at 50.49 percent. Rainfall during the 2002 monsoon season (June-September) was 19 percent below the normal rainfall having disastrous impact on different segments of the Indian economy, in irrigation and agriculture in particular. Drought condition reigned over 29 percent of the country and it was severe in 10 percent area.

Similarly for June 2009 rainfall over the country as a whole was 47% of its Long Period Average (LPA) value. During the month, out of 36 meteorological subdivisions, 2 received excess rainfall, 6 received normal rainfalls, 19 received deficient rainfalls and remaining 9 subdivisions received scanty rainfall. Area weighted rainfall over the country as a whole was only 83.9 mm (Normal 162.4 mm), which is the second lowest since 1901 after 1926 (83.2 mm). For July 2009, rainfall for the country as a whole was 96 % of its Long Period Average (LPA) value. During the month, out of 36 meteorological subdivisions, 7 received excess rainfalls, 14 normal rainfalls, 14 received deficient rainfalls and remaining 1 subdivision received scanty rainfall. For August 2009, rainfall for the country as a whole was 73 % of its Long Period Average (LPA) value. During the month, out of 36 meteorological subdivisions, 2 received excess rainfall, 15 received normal rainfalls, 17 received deficient rainfalls and remaining 2 subdivisions received scanty rainfall. For September 2009, rainfall over the country as a whole was 80 % of its Long Period Average (LPA) value. During the month, out of 36 meteorological subdivisions, 7

received excess rainfall, 13 received normal rainfalls, 12 received deficient rainfalls and remaining 4 subdivisions received scanty rainfall.

For the country as a whole, seasonal rainfall at the end of the southwest monsoon season (June to September) was 78% of its Long Period Average (LPA) value. The LPA value of southwest monsoon rainfall over the country as a whole, calculated with the data of 1941-1990 is 89 cm.

The gravity of the crisis was reflected in the fact that country's food grains production dropped by 13.18 percent in 2002-03. According to the Agriculture ministry, country's food grains production in 2002-03 was 184.06 million tonnes - a significant fall from 212.02 million tonnes recorded in 2001-02. This is lowest ever since 1996-97 when food grains production was 199.4 million tonnes. While the kharif and rabi productions in 2002-03 were estimated at 89.45 million and 94.61 million tonnes respectively.

Rice production in 2002-03 was 17.40 percent lower at 76.91 million tonnes than the previous year's 93.08 million tonnes and wheat production declined by 2.15 percent at 70.26 million tonnes compared with 71.81 million tonnes in 2001-02.The maximum impact of drought was felt in the production of coarse cereals where the fall was as steep as about 26 percent at 25.1 million tonnes compared with 33.9 million tonnes in 2001-02. Production of pulses decrease by 10 percent at 11.8 million tonnes in 2002-03.

The severest effect of drought was felt in the oilseeds sector. Production nosedived to about 24 percent - from 20.46 million tonnes to 15.57 million tonnes in 2001-02. Groundnut and soyabean production decreased by 0.2 million tonnes and 1.6 million tonnes, respectively. Rapeseed/Mustard oilseeds production fell by around 1 million tonnes. Sugarcane production too witnessed lower production-by about 5 percent- at 285.4 million tonnes. While cotton production declined by 15.61 percent, in respect of jute and mesta the decline was between 5-6 percent.

A fall in agricultural production therefore leads to a fall in industrial output after a year's lag. In the past five years, agricultural production shrank in three years and industrial growth in the following two years.

On many occasions when two successive drought occur or even with two successive deficient monsoons when industry is almost forced into recession, agricultural exports decline, inflation flare up, fiscal deficit also rise, interest rates harden and consequently investment will fall, leading to lower GDP growth.

Deficient rains in three consecutive years, especially in the western region, meant not only a drop in crop production and consequently incomes, but also a serious shortage of drinking water and animal fodder, all resulting in migration of the rural population, aggravating the demographic pressure on the urban areas.

The timely arrival of the monsoon and sufficient rains are thus key for India's economic growth because farmers' success depends heavily on the monsoon rains. Good rains are needed for India's economy is projected to expand at the official forecast of 8% in the current financial year.

However, it is also experienced that the Indian vulnerability to monsoon rain has declined considerably in recent time. Agricultural production and the rural consumer aren't as vital to the Indian economy as they were some years ago. With vital industries such as software, telecom, automobile, engineering goods and pharmaceuticals powering India's growth, people don't have to worry as much about whether a drought will dry up even village demand for motor cycle, cars or shampoo. There by increasing the flexibility of adjusting economic growth both for rural and urban India. It could well be said that new growth drivers have emerged in the Indian economy. On the other hand as 70% of India population lives in rural areas and their economy is directly linked with monsoon, the monsoon does matter much even today. A bad monsoon can quickly knock a percentage point or so off India's GDP growth, like it did in many years. Analysts still worry that a really bad monsoon could be even worse for the economy. Thus with a good monsoon rainfall, as predicted for 2010, the Indian economy is expected to be one of the fastest growing economies of the world after China.

12.3.2. Impact of El Niño-southern oscillation on Indian food grain production

The impact of El Niño-southern oscillation (ENSO) on Indian food grain production was analyzed for the period 1950-99 (Selvaraju, 2003). The inverse relationship between sea-surface temperature (SST) anomalies from June to August (JJA) over the NINO3 sector of the tropical Pacific Ocean and Indian food grain production anomalies (r = -0.50) was significant at 1% level. During the warm ENSO phase, the total food grain production decreased (12 out of 13 years) by 1 to 15%. In 10 out of 13 cold ENSO-phase years, the total food grain production increased from normal. The relationship between the SST-based NINO3 ENSO index and the Kharif season (June-September) food grain production anomalies (r = -0.52) was greater than for the Rabi season (October-February) food grain production (r = -0.27). The ENSO impact on rice production was greatest among the individual crops. The average drop in rice (Kharif season crop) production during a warm ENSO-phase year was 3.4 million tonnes (7%). In a cold ENSO-phase year the average production increase was 1.3 million tons (3%). Wheat (Rabi season crop) production was also influenced by ENSO, as it depends on the carryover soil water storage from the Kharif season. Sorghum and chickpea production are not significantly influenced by ENSO extremes. Inter-annual fluctuation of the gross value of Indian food grain production was found to be very large. The cumulative probability distributions of food grain production anomalies during cold and warm ENSO phases are shifted positively or negatively, relative to the neutral distribution. The warm ENSO forcing significantly (1% level) reduced the probability of above-average production. The cold ENSO forcing moderately increased the above-average food grain production over the neutral ENSO phase (5% level). A simple conditional probability forecast based on annual and JJA NINO3 SST predicted the category of food grain production in 11 of the 14 years. The results demonstrated that the relationship between NINO3 ENSO index and food grain production could be used for agricultural applications and policy decisions on food security for the rapidly growing population in India

12.3.3. Cropping pattern in relation to monsoon in India

A broad picture of the major cropping patterns in India can be presented by taking the major crops into consideration. To begin with, the South westerly monsoon crops popularly known as kharif crops such as rice, sorghum, bajra, maize, ragi, groundnut and cotton, etc. Among the post-monsoon crops (rabi), wheat, sorghum, mustard and gram, etc. are the major base crops.

12.3.3.1. The kharif-season cropping patterns

Rice is grown in the high-rainfall area or in areas where supplemental irrigation is available to ensure good yields. If the crop has to depend solely on rainfall, it requires not less than 30 cm per month of rainfall over the entire growing period. However, only 9 percent of the area in the country comes under this category, and it lies in the eastern parts. Nearly 45 percent of the total rice area in India receives normally 30 cm per month of rainfall during at least two months (July and August) during the south west monsoon and relatively less during other months. In contrast, the eastern and southern regions comprising Assam, West Bengal, coastal Orissa, coastal Andhra Pradesh, Karnataka (most part), Tamil Nadu and Kerala receive rainfall of 10 to 20 cm per month in four to eight consecutive months, starting earlier or going over later than the south west monsoon months. Nearly 80% of rice is grown during June-September and the rest (20 percent) during the other season.

In the most humid areas of eastern India comprising Tripura, Manipur and Mizoram, rice is the exclusive crop. In Meghalaya, rice is alternated with cotton, vegetable and food-crops, whereas in Arunachal Pradesh, where rice is not grown exclusively, the alternative crops being maize, small millets and oilseeds. In parts of Assam, West Bengal, Bihar, Orissa and northern coastal districts of Andhra Pradesh, jute forms an important commercial crop alternative to rice. In Bihar, rice is grown over 49 percent (5.3 M ha) of its cropped area (14.2 percent of all-India area), whereas pulses, wheat, jute, maize, sugarcane and oilseeds are the alternative crops. In Uttar Pradesh rice is grown on 19 percent (4.6 M ha) of its cropped area and represents about 12.4 percent of the all-India area under this crop. In Orissa and

Chhattisgarh rice is grown on more than 50 and12 percent respectively of the area, whereas the alternative crops are: pulses, ragi, oilseeds, maize and small millets. In the southern states, namely Andhra Pradesh, Tamil Nadu and Kerala rice is grown in more than one season and mostly under irrigation or under sufficient rainfall. Together, these three states have over 6.0 M ha, representing over 17 percent of the all-India area under rice. Important alternative crops in Andhra Pradesh are: pulses, groundnut, jowar, maize, sugarcane and tobacco. In Karnataka crops alternative to rice are: ragi, bajra, cotton, groundnut, jowar and maize. In Kerala crops and tapioca form the main alternative to rice. In Maharashtra rice is grown mostly in the Konkan area over 1.3 M ha, along with ragi, pulses, Rabi jowar, sugarcane, groundnuts and oilseeds. In other states, namely Gujarat, Jammu and Kashmir, Rajasthan and Himachal Pradesh, rice forms a minor crop and is mostly grown with irrigation. However, in Punjab and Haryana and to some extent in western Uttar Pradesh owing to high water-table during this monsoon season and irrigation facilities, rice has become a major crop in such areas.

The maize is a crop grown commonly in high-rainfall areas or on soils with a better capacity for retaining moisture, but with good drainage. Next is jowar in the medium rainfall (500-600 mm) regions whereas bajra has been the main crop in areas with low or less dependable rainfall and on light textured soils. Bajra is more drought-resistant crop than several other cereal crops. The area under the bajra crop in India is about 12.4 M ha and Rajasthan (4.6 M ha) shares about the one third of the total area. Maharashtra, Gujarat and Uttar Pradesh together have over 4.6 M ha, constituting an additional 1/3 area under bajra. Over 66 percent of this crop is grown in areas receiving 10-20 cm per month of rainfall, extending over 1 to 4 months of the south west monsoon season.

The area under the *kharif* jowar in India is the highest in Maharashtra (2.5 M ha), closely followed by Madhya Pradesh (2.3 M ha), whereas in each of the states of Rajasthan, Andhra Pradesh, Karnataka and Gujarat, the area under this crops is between 1.0 and 1.4 M ha. Jowar is mainly grown where rainfall distribution ranges from 10-20 per month at least for 3 to 4 months during south west monsoon season or is still more abundant.

Groundnut is sown over an area of about 7.2 M ha, mostly in five major groundnut-producing states of Gujarat (24.4 per cent), Andhra Pradesh (20.2 per cent), Tamil Nadu (13.5 per cent), Maharashtra (12.2 per cent) and Karnataka (12.0 per cent). Five other states viz., Madhya Pradesh, Uttar Pradesh, Punjab, Rajasthan and Orissa together have about 17.3 percent of the total area under this crop. The rainfall in the groundnut area ranges from 20-30 cm per month in one of the monsoon months and much less in the other months. In some cases the rainfall is even less than 10 cm per month during the growing season and that too, in a few states only, viz., Punjab (16.4 percent), Tamil Nadu (13.3 percent) and Andhra Pradesh (12.5 percent).

Cotton is grown over 7.6 Mha in India. Maharashtra shares 36 percent (2.8 M ha), followed by Gujarat with 21 percent (1.6 M ha), Karnataka with 13 percent (1 M ha) and Madhya Pradesh with 9 percent (0.6 M ha) of the area. Most of the cotton areas in the country are under the high to medium rainfall zone. The cotton grown in Madhya Pradesh, Maharashtra, Karnataka, and Andhra Pradesh (4.8 M ha) is rainfed, whereas in Gujarat and Tamil Nadu (1.93 M ha) it receives partial irrigation (16-20 percent of the area). The area under cotton in Punjab, Haryana, Rajasthan and Uttar Pradesh (0.8 Mha) gets adequate irrigation, ranging from 71 to 97 percent of the area.

12.3.3.2. The *Rabi* season cropping patterns

Table 12.6: Percent of All-India area of various crops in different States

Crop	Area	Region (per cent of all-India area)
Sugarcane	2.5 M ha	Uttar Pradesh (51), Haryana (6), Bihar (6), Punjab (6), Maharashtra (8), Andhra Pradesh (5), Tamil Nadu (5), Karnataka (3)
Tobacco	0.427 M ha	Andhra Pradesh (48), Gujarat (19.5), Karnataka (8.7), Maharashtra (3.5), Tamil Nadu (3.5)
Potato	0.491 M ha	Uttar Pradesh (33.6), Bihar (20.4), West Bengal (13.3), Assam (5.2), Orissa (4.8)
Jute	0.778 M ha	West Bengal (60), North eastern Region (18.7), Bihar (17.6), Orissa (6.1), Uttar Pradesh (1.7)

Coconut	1.05 M ha	Kerala (68.3), Karnataka (12.4), Tamil Nadu (9.7), Andhra Pradesh (3.5)
Rubber	0.197 M ha	Kerala (92.8), Tamil Nadu (5.0), Karnataka (1.9)
Cashew	0.264 M ha	Kerala (67.4), Karnataka (12.1), Andhra Pradesh (10.8), Tamil Nadu (9.8), Maharashtra (4.8)
Tea	0.35 M ha	West Bengal, Assam and Tripura (77), Kerala, Tamil Nadu and Karnataka (20)
Coffee	0.138 M ha	Kerala , Tamil Nadu and Karnataka (99)
All fruit-crops	1.8 M ha	Spread all-over India
Onion	0.16 M ha	Maharashtra (18.5), Karnataka (11.7), Andhra Pradesh (12.8), Tamil Nadu (11.2), West Bengal (7.6), Madhya Pradesh (7.2), Orissa (6.8), Punjab (6.2)
Chillies	0.733 M ha	Andhra Pradesh (26.9), Maharashtra (20.4), Karnataka(14.5), Madhya Pradesh (5.5), Tamil Nadu (10.1)
Coriander	0.283 M ha	Andhra Pradesh (36), Rajasthan (23.6), Madhya Pradesh (11.1), Tamil Nadu (10.0)

Among the *Rabi* crops wheat together with barley, oats, jowar and gram are the main crops. Generally, wheat and gram are concentrated in the subtropical region and grown by applying supplemental irrigation where winter rain is not adequate whereas the *Rabi* sorghum is grown mostly in the Deccan Plateau. Depending on the season, tobacco, jowar, oilseeds and maize are grown in rotation in the jute-growing areas. Rice is the usual alternative crop.

12.3.3.3. The future of cropping patterns

India is a country of small farmers. With the increase in irrigated area and advances in agricultural science, most of the extensive cropping patterns are giving way to intensive cropping. The development in minor irrigation works has especially provided the farmers with opportunities to crop their land all the year round with high-yielding varieties. This intensive cropping will require an easy and ready availability of balanced fertilizers, well distributed rainfall/irrigation and plant protection chemicals and an appropriate price policy for inputs and agricultural produce. In the future the size of the holdings is expected to diminish further. The country has to produce enough for its people without deteriorating the quality of the environment

and eco-system. This is the challenge of the future for the farmers, agricultural scientists, extension workers and administrators.

12.3.3.4. Contingency crop planning

Indian Agriculture is primarily rain-dependent. While the onset and progress of the monsoon decide the crop planning, the temporal and spatial variability in the monsoon activity influences the crop productivity and requires contingent planning on regular basis. On the other hand earlier or late withdrawal of SW monsoon and onset of NE monsoon, residual soil moisture, winter rain due to passage of western disturbances decide cropping pattern in the Rabi season. Unexpected failure of monsoon activity in early or mid-season, contingent planning is a regular necessity to cope up with the vagaries of the monsoon.

12.4. Monsoon and food grains production

2006-07:

Agriculture's share in India's GDP declined in recent years, thus marking a structural shift in the composition of the GDP. Traditionally, agriculture accounted for two-fifths of the GDP, but in recent times it witnessed a declining trend. Due to the effect of the drought, the agriculture and allied sector growth in fiscal 2002-03 and 2009-10 was negative one and accordingly GDP for the years 1996-97 which was 7.8 percent declined to 4.4 percent in 2002-03 and 7.6 percent in 2009-10 respectively.

Despite some unusual aberrations in the rainfall pattern the year 2006-07 experienced slightly better rainfall when compared with the forecast made in July. The overall monsoon rainfall index for the country as a whole indicates a jump of 2.6 percent above its normal level and was better than the previous year's index. Of the four major zones of the country, two regions - east and west witnessed above normal rainfall as measured by the rainfall indices and the remaining two regions - north and south experienced 11 percent and 6 percent deficiency in monsoon rainfall. From an overall perspective the behaviour of monsoon rainfall during the year 2006-07 was

more or less consistent with the predictions made by the IMD both before the onset of the season and during the middle of the season. This is true for the majority of regions with a few exceptions, which witnessed moderate drought like conditions. These include - Andaman and Nicobar Islands (-26 percent), Arunachal Pradesh (-29 percent), Assam and Meghalaya (-37 percent), west Uttar Pradesh (-43 percent) and Haryana (-39 percent). The other sub-divisions, where deficiency in monsoon rainfall remained between -19 to -26 percent were - Nagaland, Manipur, Mizoram and Tripura (-21 percent), east Uttar Pradesh (-23 percent), Uttaranchal (-22 percent), Himachal Pradesh (-23 percent) and Tamilnadu and Pondicherry (-21 percent).

2008-09:

In the year 2008, the South-West monsoon season rainfall over the country as a whole was 98 percent of its Long Period Average (LPA). During monsoon 2008, two meteorological sub-divisions received excess rainfall, 30 sub-divisions received normal rainfall, and 4 sub-divisions received deficient rainfall. The sowing of kharif crops declined during 2008-09 to 98.9 percent of the normal which was about 2.5 percent lower than the previous year. The sown area declined mainly because rainfall was deficient in most of the weeks of the sowing months, viz., July and August. All food grain crops have shown decline in the area sown except rice. Among non-food grains, while area covered under oilseeds increased, that for sugarcane, cotton and jute declined. Kharif crop production was reduced by 2.8 % during 2008 as compared to previous year.

12.4.1. Indian Monsoon and Agricultural Planning

For growing crops that are entirely rainfed and are of short duration, it is desirable to choose a short time interval of a week instead of a month for analyzing and presenting rainfall data. Information on the quantum of assured water availability over short intervals is essential for developing farming technology, particularly in dryland areas. Over these areas of scanty rainfall, the distribution of rainfall over short periods is not of the normal type and is highly variable from one year to another. It is made up of a large number of small amounts of rainfall and a very small

number of higher amounts of rainfall. The variability increases with the shortness of the interval and the average of weekly rainfall becomes less representative and not of much use. In the circumstances, the fitting of the series into a theoretical distribution of the gamma type appears to be more useful.

The analysis of weekly totals of rainfall over long periods gives information on the lower limits of rainfall, week by week, that are likely to be expected for various percentage probability ranges from 90 percent to 10 percent. It is expected that these probability figures will be useful in the identification of periods favourable for sowing different crops and for determining the quantum of moisture that would be available during different phases of crop growth. Again, where the amount of assured moisture in a particular probability range is deficient to meet the needs of a crop in a phase, the extent to which this deficiency has to be made up by irrigation can also be determined.

To ensure sustainable agriculture in a region, knowledge of the local climate is vital. Climatic limitations are a strong indicator of agronomic potential and can be used to determine which crops are best suited for a region, as rainfall and temperatures are two major variables affecting crop type and yield. Planning is especially critical in monsoon regions which experience distinct wet and dry seasons. Soil moisture prior to the beginning of the rainy season is usually negligible, a situation exacerbated by the preceding heat buildup and high evaporative losses. Except where irrigation is available, planting is consequently restricted to the beginning of the wet season.

In various parts of India, the onset of the south-west monsoon for a particular area is expected between June to mid July, depending on its geographic location. The highest concentration of non-irrigated agriculture happens to be in western and southern oilseed, grain and cotton areas while in the East, much of the rice is rain-fed. These crops suffer most from a late or weak start to the rainy season and could be considerably affected during an extended break in monsoon rains. Also, if the southwest monsoon withdraws from the region earlier than expected, late planted crops may be damaged during the grain filling stages due to lack of moisture. A late

withdrawal or late-season rains are on the other hand considered detrimental to maturing crops, especially cotton.

During the drought of 1987, kharif crop production was down as rainfall was the lowest on record in central and northern rainfed areas. Rabi crops, those planted primarily in southern India and areas with adequate irrigation reserves, partially made up for the shortfall but also suffered some losses. In winter planting of wheat was delayed considerably due to insufficient moisture availability for germination in northern India (although most of the crop is irrigated, low reservoirs and fuel shortages hampered irrigation efforts).

Conversely, a strong monsoon circulation can bring flooding, especially along the Ganges and Indus Rivers. Bangladesh encompasses most of the area considered the Mouths of the Ganges, with other major rivers (primarily the Brahmaputra and the Meghna) converging within its boundaries. Eastern India and Bangladesh are the least drought prone areas, indicating the consistency of the monsoon in that region. In fact, a certain level of flooding is expected each year, and local rice cropping patterns are dependent upon the seasonal abundance. In September 1992, a late surge in the monsoon flooded cotton areas of Pakistan and generally soaked crops mostly in the open boll stage. Disease was also a problem as the spraying patterns schedules were disrupted by the unseasonable heavy showers. However, while the rainfall was unusually heavy, it occurred only a few weeks later than normal, highlighting the hazard for the cotton, which ripens so close to the end of the rainy season.

Over the past 20 years, course grain production has risen steadily despite dramatic decreases in area. Much of this area, predominately rainfed acreage in western India, is increasingly used for oilseed production, which has increased dramatically over the same period. This shift in agriculture, vital in meeting the nation's nutritional requirements, was possible due to the region's versatile climate. As stated earlier though, the rainfed crops are at greatest risk of failure in times of drought, especially in the drought-prone west. Advances in genetic research have been important in developing cultivars which would thrive in certain areas, balancing drought resistance and yield. In fact, some of the more drastic increases in

production stem from improved germplasm, most notably wheat production, which has more than quadrupled over the last 30 years.

The monsoon climates are especially vulnerable to disruptions in global weather which could result in drought, flooding or both in any given year. Decades of research on the driving forces behind the Indian monsoon circulation has resulted in a better understanding of weather extremes experienced and their impact on agriculture. The benefits of good monsoon are certainly tempered by the risks of farming in such a volatile area, although forecast techniques currently being developed are helping to mitigate the impacts of poor monsoon performance using contingency planning through Integrated Agromet Advisory Services.

Examination of summer total food grain production in India during the last five decades shows significant relation with the total monsoonal rainfall in the country. In addition, many crops show significant relation with the local/regional rainfall and temperature variability at different crop stages. While technological innovations have dramatically improved the agricultural production, climate-related anomalies still continue to cause severe hardship to a large population in South Asia. Advance information on the possible climatic impacts on food production is crucial to ensure proper food security in the region (Krishnakumar et. al., 2004).

Aggregate food grain production is strongly correlated (at 1% significant level) with monsoon rainfall. Most of the food crops are grown during the kharif season when correlation with monsoon rainfall is particularly strong. Production of food grains, cereals and pulses show strong association with NINO3 and Indian Ocean SST anomalies during the kharif but not during the Rabi season. The food grains production increased to an all-time record level of 230.67 million tonnes during 2007-08. Similarly, oilseeds, milk, fruits, vegetables and fish production has been growing over the past few years to reach record levels.

Observed variation of the monsoon food grain production and GDP

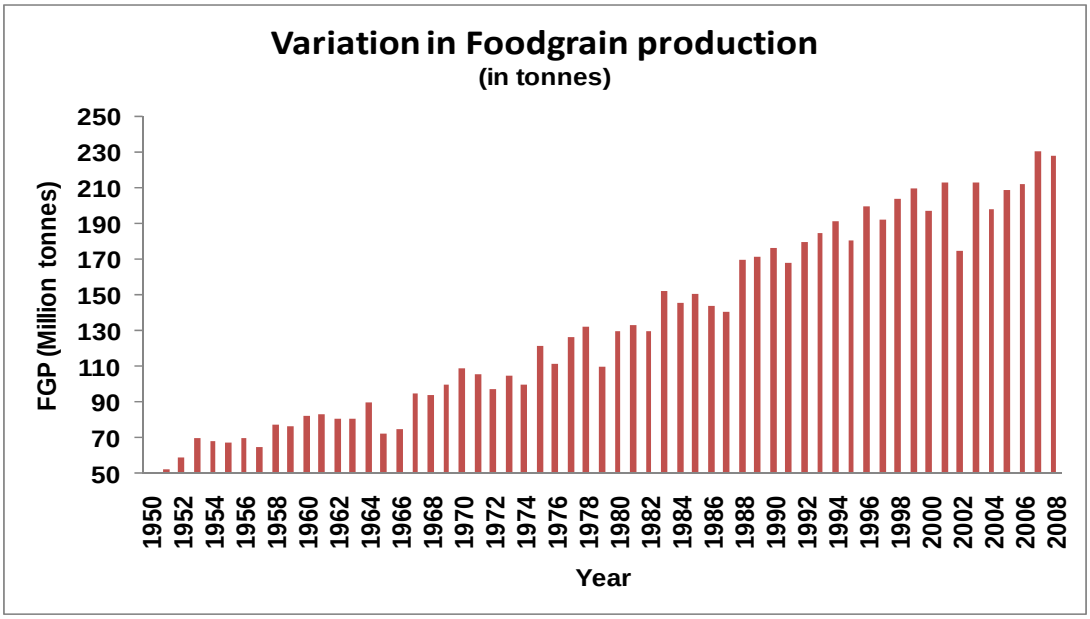

Fig.12.11: Variation in food grain production (in million tonnes)

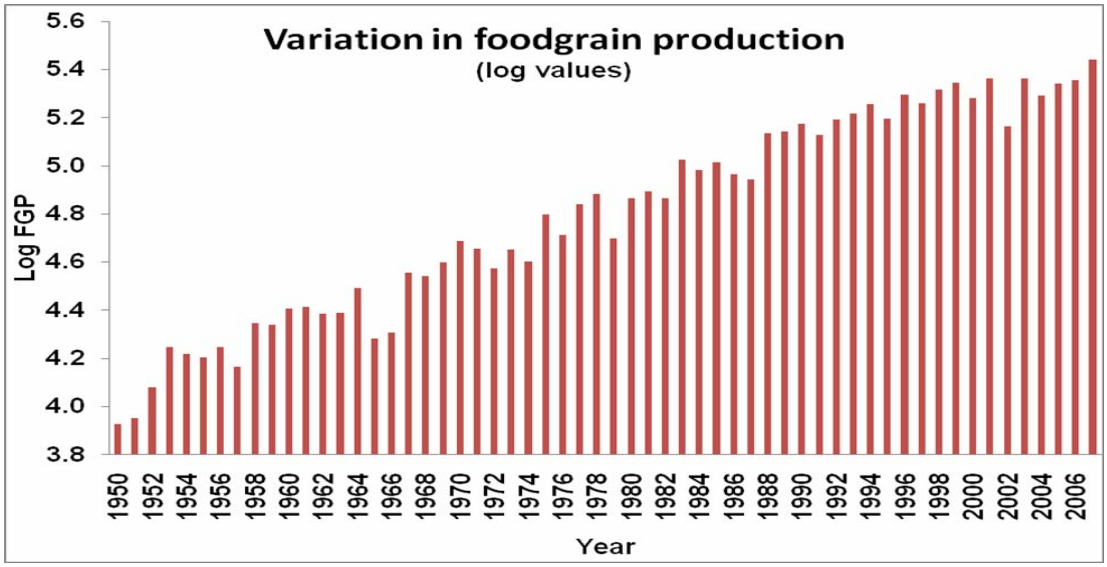

Fig.12.12: Variation in foodgrain production (log values)

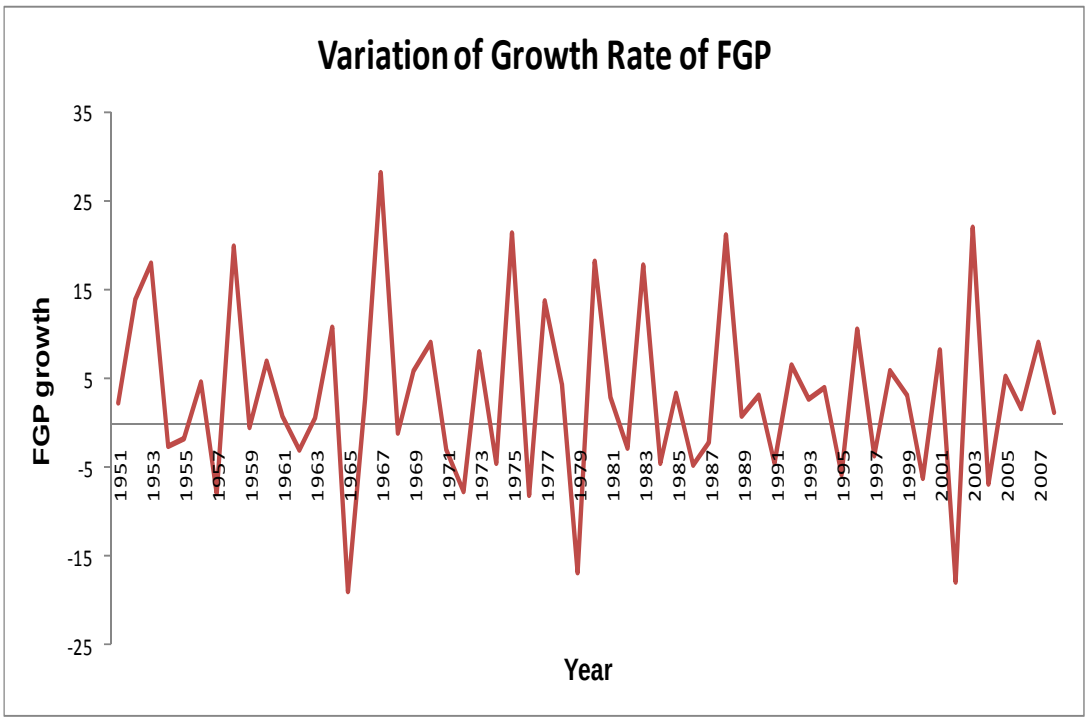

Fig. 12.13: Variation of Growth Rate of FGP

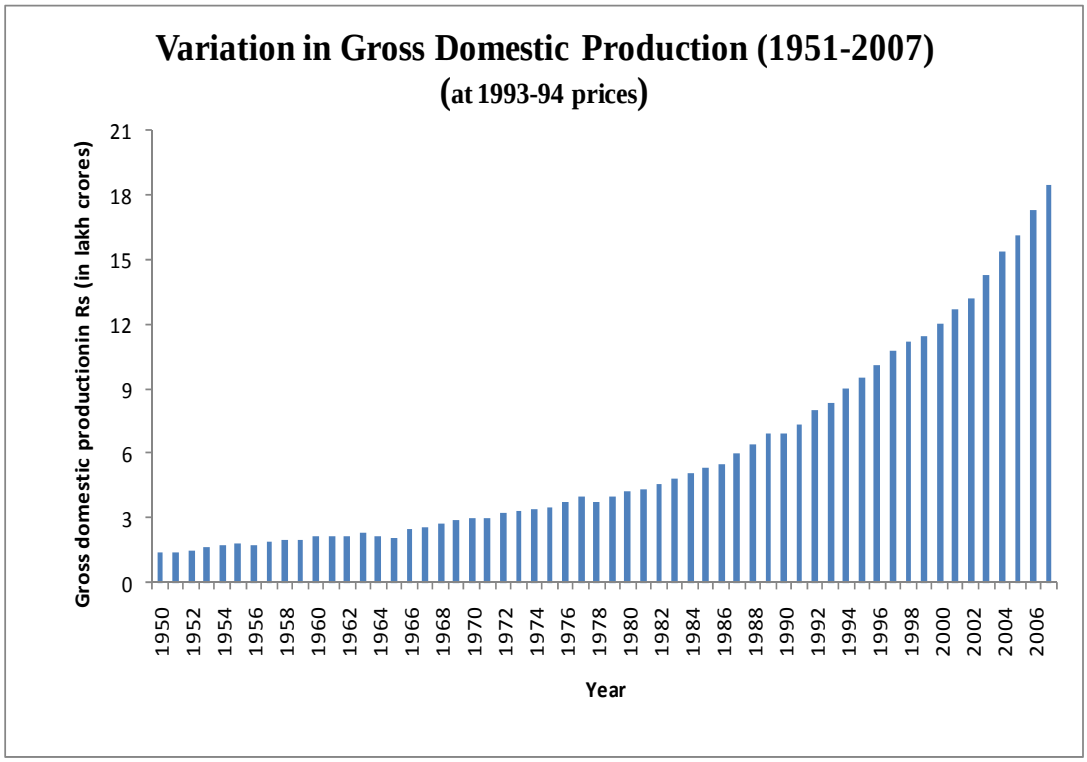

Fig. 12.14: Variation in GDP (1951-2007)

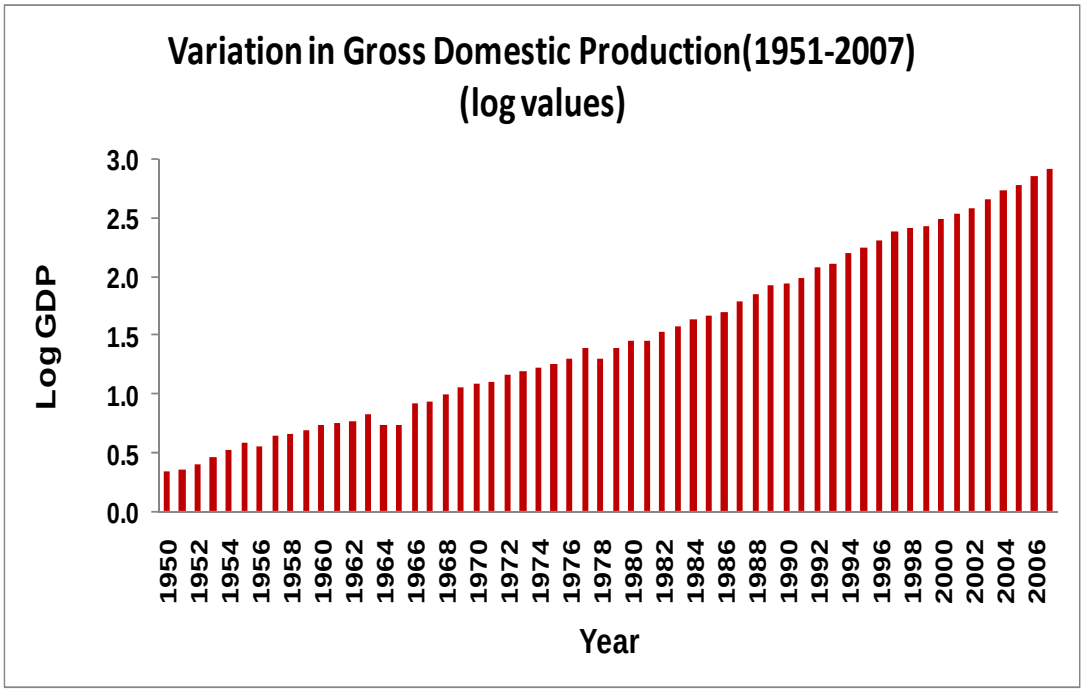

Fig.12.15: Variation in GDP (log values)

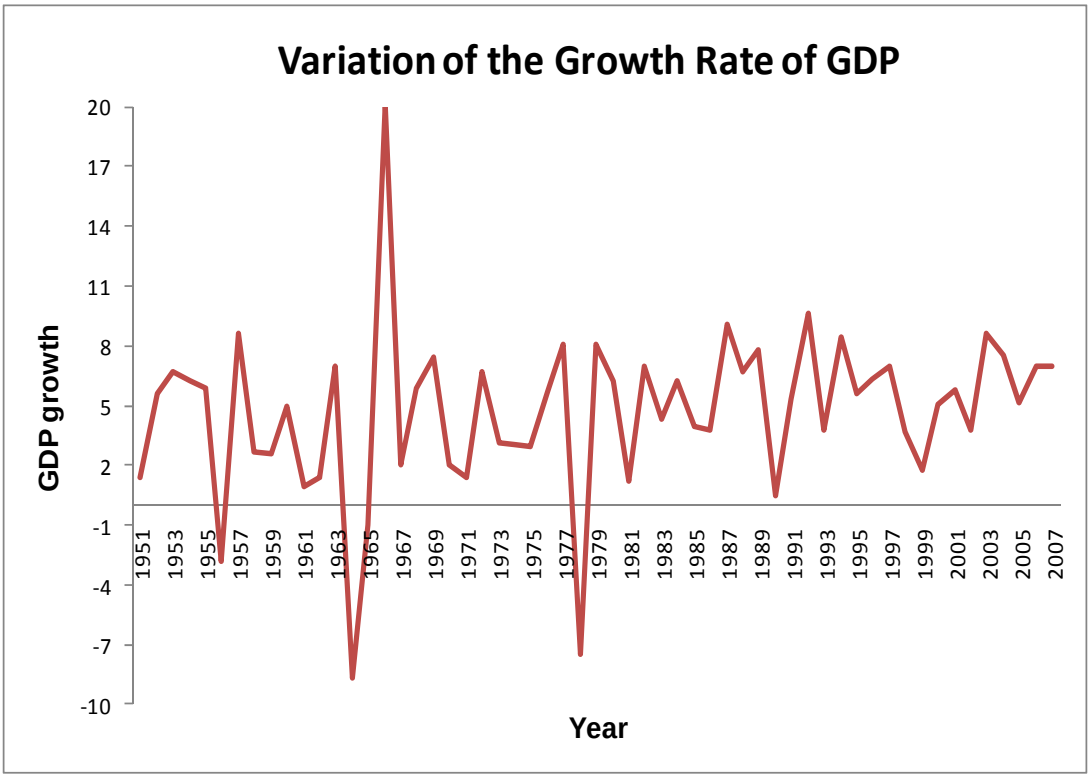

Fig. 12.16: Variation of Growth Rate of GDP

Fig. 12.14 shows the variation of GDP during 1951-2007 which is dominated by sustained increase. It is clearly seen from the figs.11-13 that there is decrease in the food grain production during the years 1965, 1966, 1968, 1972, 1974, 1979, 1982, 1986, 1987, 2002, 2004 which correspond to the large deficits in monsoon rainfall. This is a manifestation of the well known impact of the monsoon on Indian food grain production (Hanumantha Rao et al., 1988; Parthsarathy et al., 1992; Krishnakumar et al., 2004).

The monsoon was good in 1964, which led to the spurt in agricultural output, but this was put into reverse the next year. In 1965-66 and 1966-67 there was a severe drought with the monsoon of 65-66 being probably the worst of the twentieth century. Food grain production fell by 27% (from 89 million tons to 65 million tons) between 64-65 and 65-66 and rose only slightly out of this trough the next year. Food prices started to soar and even with the increased food imports, there were severe shortages. Food grain prices rose about 30% relative to industrial prices between 1964-65 and 67-68.

For the years when the ISMR anomaly is large, most of the country experience a similar anomaly (drought or excess rainfall as the case may be) whereas, for years for which the ISMR is near negative or near its average value, there is a large spatial variation in the rainfall anomalies over the country, with deficit rainfall over some parts of the country and normal or excess rainfall over other parts. The variation of the all India food grain production depends on the production in different agroclimatic zones of the country which in turn depends on the rainfall distribution in that zone. So large magnitude of the ISMR anomaly will have large impact on the food grain production. This can be seen from Fig. 12.13. During the last five decades there were large fluctuations in FGP.

With planned development since independence, the contribution of agriculture to Gross Domestic Product (GDP) decreased substantially and led to the expectation that the impact of monsoon on the economy would have also decreased. However, a recent analysis of the variation of GDP and the monsoon has revealed that the impact of severe droughts on GDP has remained between 2 to 5% of GDP throughout. The large impact of droughts on GDP can be attributed to the indirect

impact on the purchasing power of large fraction of the population dependent on agriculture. It has been shown that while the magnitude of the adverse impact on food grain and rainfall is not large. In other words, there is asymmetry in the response of the food grain production after 1980, is that the strategies that would allow farmers to reap benefits of the good rainfall years (use of fertilizers and pesticides) are not economically viable in the current milieu.

12.4.2. Gross Domestic Product (GDP)

Parthsarathy et al., (1988) have shown that an exponential function is also a good fit for the trend on the growth of FGP. The exponential growth curves to the food grain production and the GDP were fitted taking into consideration of latest data set upto 2007.

12.4.3. Trends

The variation of the natural logarithm of the FGP and GDP is shown in Fig. 12.12 and Fig. 12.15 respectively. The variation of FGP (tonnage) and GDP (value) is shown in Fig. 12.11 and Fig. 12.14 respectively. The best fit curve equations are as follows:

GDP

Y=A+B*X +C/X ; Year 1950-1980

where

$A=0.5368*10^{-13}$

$B = -0.5080*10^{6}$

$C = 0.7691*10^{-1}$

$R^2 = 0.9877$

FGP

Y = A+B*lnX ; Year 1950-1980

where

$A=-0.1812*10^{-1}$

$B = 0.1005*10^{1}$

$R^2 = 0.9839$

Y=A*X^B; Year 1981-2007

where

$A=0.1165*10^{2}$

$B = 0.7357*10^{-1}$

$R^2 = 0.9955$

Y=A*B*lnX; Year 1981-2007

where

$A=0.6846*10^{-1}$

$B = 0.9865$

$R^2 = 0.9455$

The root mean square error for FGP is 0.98 million tons while that of GDP is around 1900 crore. The growth rate of GDP had been about 3.9 percent during 1951-80 and since 1980s it has increased more rapidly (at the rate of 5.6%).

The decrease in the growth rate of food grain production has been contributed to the decrease in the growth rates of (i) irrigated land and also other associated problems viz., salinity, water logging, etc. and (ii) the yield because of steady decrease in fertility (nutrient availability) of the lands due to intensive agriculture in the previous three decades (Abrol, 1996). Variation of monsoon rainfall and early or midseason droughts has added to such fluctuation. Change in the cropping pattern has contributed to the decrease in the area under cultivation of food grains.
The growth rate is determined by:

GDP_{gr}(year)=100*(GDP(year)-GDP(year-1))/GDP(year-1)
FGP_{gr}(year)=100*(FGP(year)-FGP(year-1))/FGP(year-1)

The growth rate of FGP has increased steadily from 1951, well before the green revolution in the mid 1960s. Hanumantha Rao et al (1988) have shown that there was no significant change in growth rate with the green revolution due to the fact that the growth rate of wheat though enhanced due to new technology that of several other food crops decreased due to several other reasons.

12.4.3.1. Wheat

The annual total wheat production comes from a single growing season. It is planted during October-December and harvested during March-May. The chief wheat producing states are Uttar Pradesh, Punjab, Haryana, Madhya Pradesh, Bihar, West Bengal and Assam account for 95% of the total area under wheat in India. More than 80% of the area under wheat is irrigated.

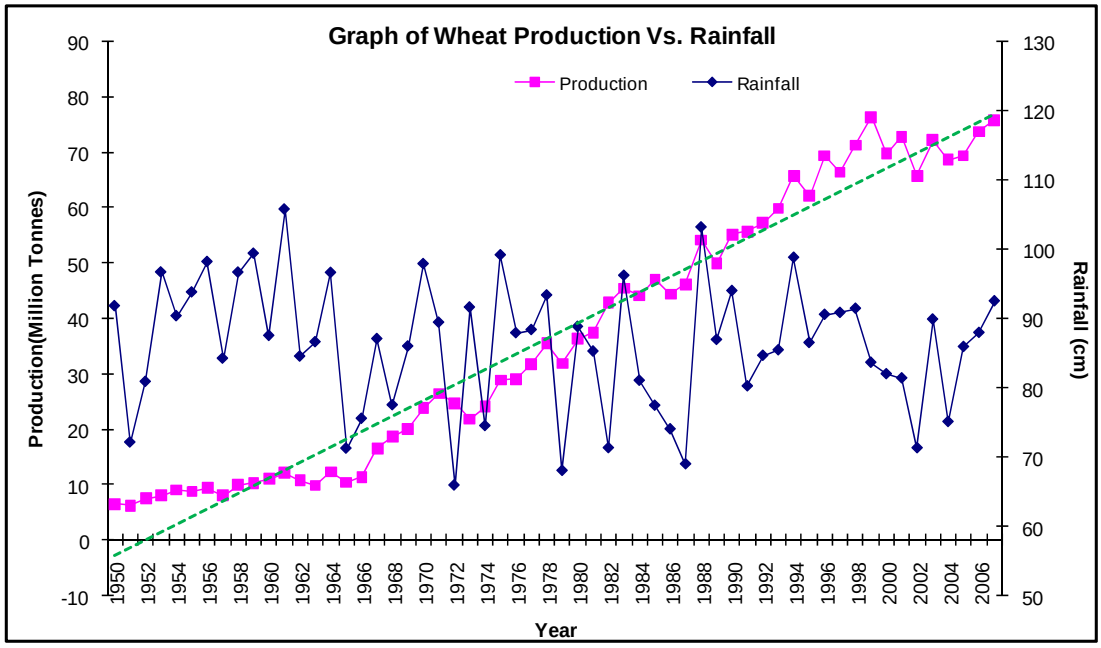

Fig.12.17: Wheat Production Vs. Rainfall

The time series of all-India wheat production (Fig. 12.17) shows a sudden increase after the mid-1960s that can be attributed to the Green Revolution. The year to year variability of wheat production lacks a strong association with the monsoon rainfall. Although it is grown in non-monsoon months, its production shows a rather weak but significant correlation with the monsoon rainfall in the months of June and July and is correlated with monsoon rainfall (c.c. = 0.29 and 0.16 for June and July respectively).

Apart from bio-physical factors such as fluctuations in temperatures, lack of access to assured irrigation, timely availability of inputs, especially seeds and fertilizers, and low seed replacement are the main causes of low yield in Madhya Pradesh and Bihar. The production rate during 1990s was positive across states and at the national level. During 1990s wheat production increased at higher rate (6-7%) in Madhya Pradesh and Rajasthan than other states. Also, area and yield equally contributed to the growth of wheat production in Madhya Pradesh while in Rajasthan the production increased mainly due to increase in area (above 5%). During 2001-06, the decline in yield is much more in Bihar than Punjab. Evidence has shown that delay in sowing by 25-30 days leads to a grain yield loss of 25% (Sant Kumar, 2008).

12.4.3.2. Rice

Kharif rice is grown in most parts of India. Orissa, West Bengal and Assam in the east, coastal Andhra Pradesh and Coastal Tamil Nadu in the southern peninsula, Madhya Pradesh in the central region and parts of Uttar Pradesh and Punjab in the northwest form the major rice-growing areas. Minor rice growing areas include Kerala and Karnataka in the south. Kharif rice is grown in July-August and harvested between October and January. The major rice-growing states account for 85% of the total area under kharif rice. The rice grown during Rabi season is mainly confined to the region affected by the northeast monsoon. Andhra Pradesh, West Bengal, Tamil Nadu, Karnataka and Orissa account for more than 90% of the area under rabi rice. It is planted in November-December and harvested in March-April.

All-India rice production (Fig. 12.18) from 1950 to 2007 shows a strong trend and high year to year variability. The strong increasing trend is the result of both an increase in cultivated area and the influence of improved production technology. Series of all-India rice production and monsoon rainfall shows correlation of 0.32 in June and -0.14 in August.

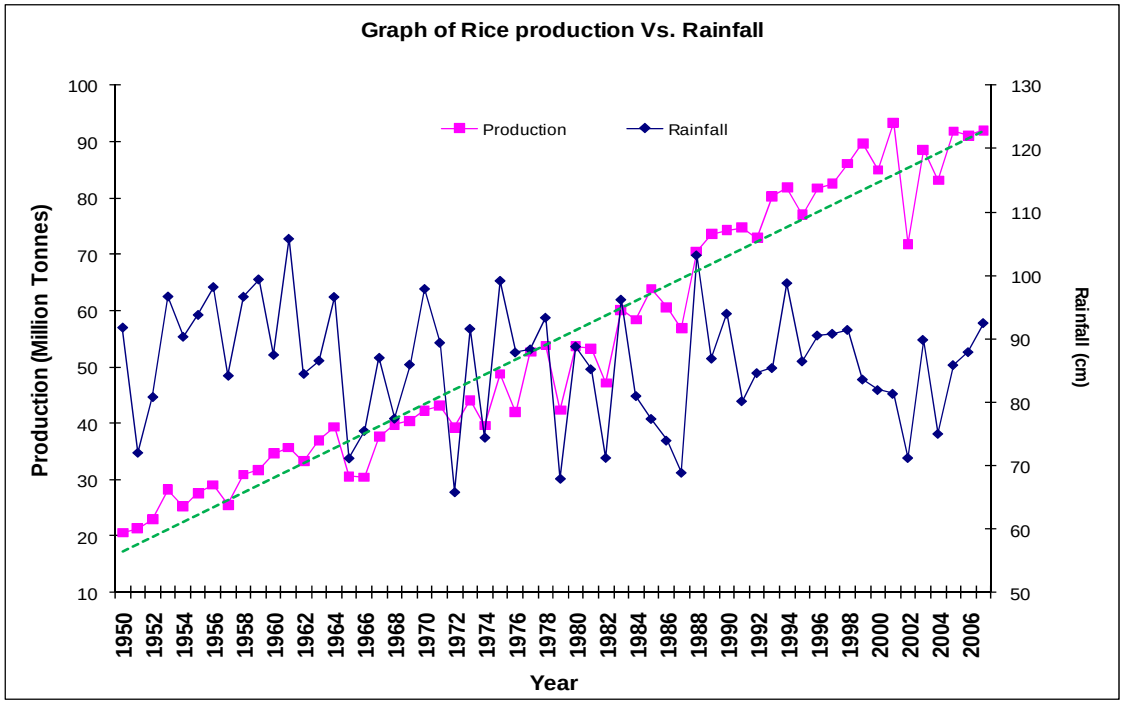

Fig.12.18: Rice Production Vs. Rainfall

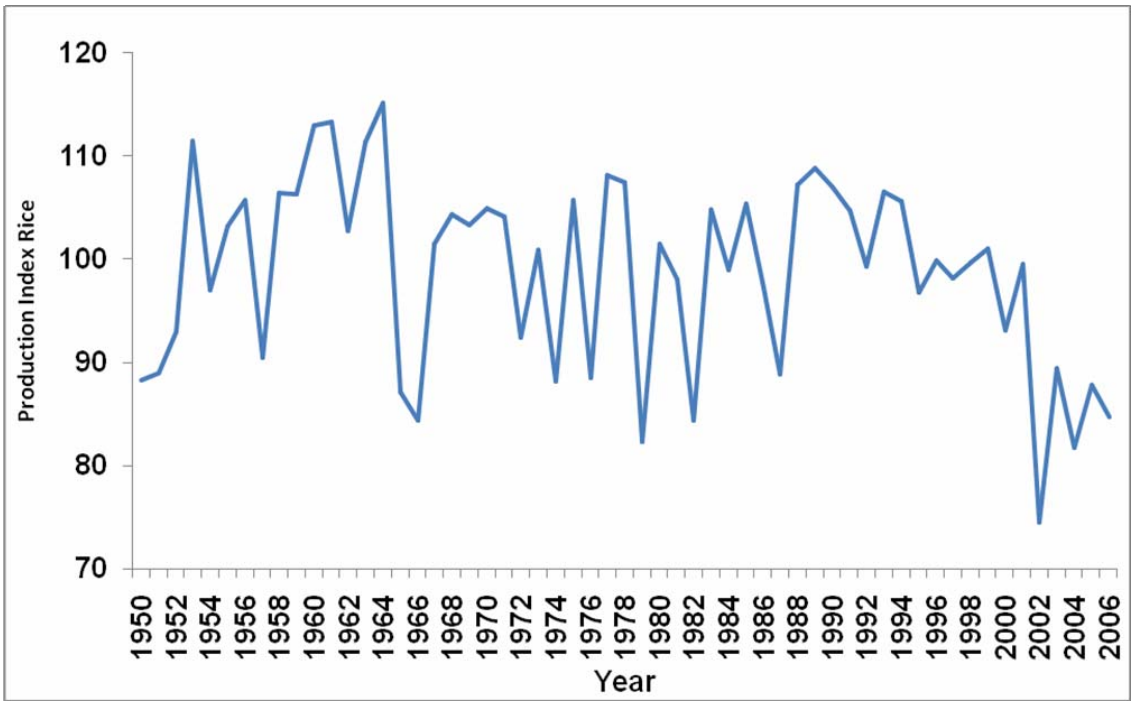

Fig.12.19: Production Index of Rice

The increasing trend which is attributed to non-meteorological factors such as increased gross sown area, improved technology, fertilizer application, pest and disease control, etc. has been estimated by fitting an exponential equation of the following form into the historical annual rice production data for the period 1950-2007:

Trend = EXP (0.0272931*Year)*1.79309E-22

Production indices (PI) of Rice have been obtained by eliminating the trend component from actual production in each year

PI= (Production/trend)*100

Mooley (1981) and Parthsarathy (1988) have also observed the similar trend with rice and also found the exponential relationship.

12.4.3.3. Sorghum

Sorghum is grown in both the kharif and rabi seasons, although kharif sorghum contributes more than 70% of the total production. Kharif sorghum is planted during June-July and harvested in December-February. Rabi Sorghum is planted after the monsoon season (October-November) and harvested in March-April. Sorghum is grown mainly in the semi-arid belt spanning from the southern peninsula to western India is seldom irrigated. About half of the sorghum production in India comes from Maharashtra while Karnataka (13%) and Madhya Pradesh (12%) also make significant contribution. Andhra Pradesh, Uttar Pradesh, Tamil Nadu, Rajasthan and Gujarat produce Sorghum in smaller amounts.

The time series of all India sorghum production (Fig. 12.20) shows linear trend. Series of all-India sorghum production and monsoon rainfall shows correlation of 0.17 in June, -0.21 in July, -0.15 in August and -0.16 during June to September which are significant at 5% levels respectively**.**

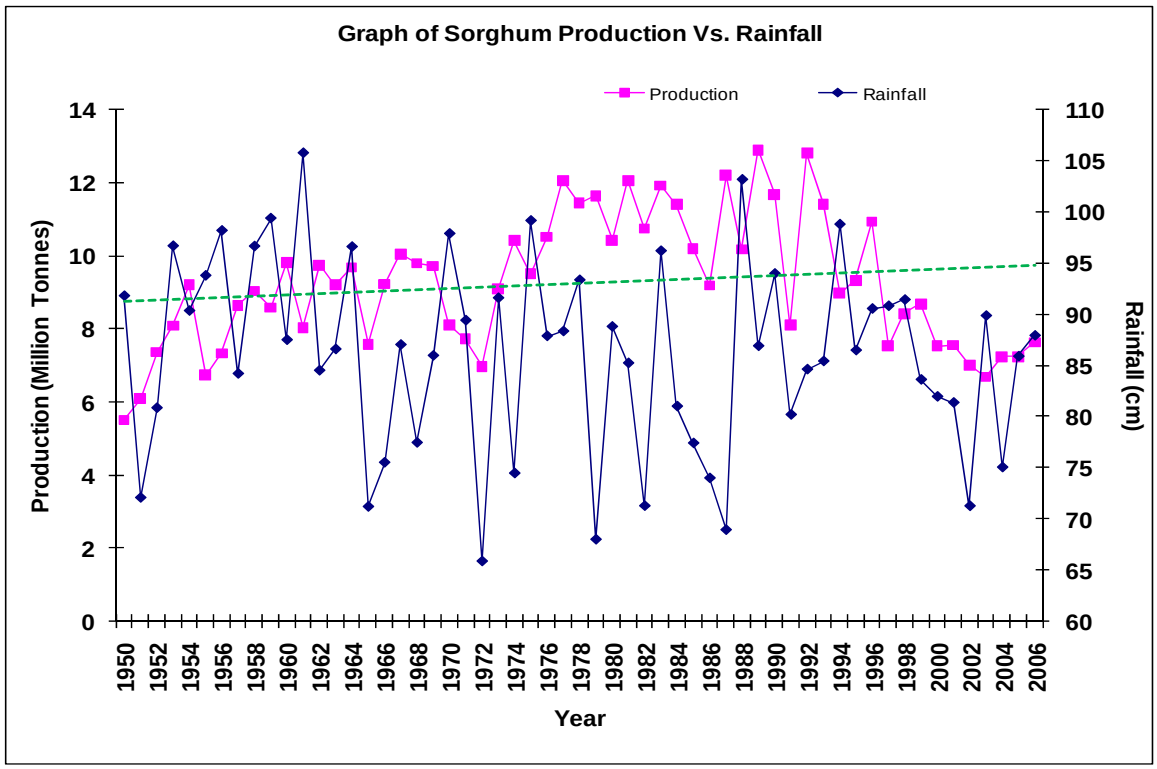

Fig. 12.20: Sorghum Production vs. Rainfall

12.4.3.4. Maize

Maize is grown in both the kharif and rabi seasons, although Rabi maize contributes more than 70% of the total production. In India, maize is grown in a wide range of environments, extending from extreme semi-arid to sub-humid and humid regions. The crop is also very popular in the low and mid-hill areas of the western and northeastern regions. Broadly, maize cultivation can be classified into two production environments: (1) traditional maize growing areas, including Bihar, Madhya Pradesh, Rajasthan and Uttar Pradesh and (2) non-traditional maize areas, including Karnataka and Andhra Pradesh.

Production of cereals other than rice and wheat stagnated during the 1980s and declined marginally during the 1990s. During these two decades, maize performed better than other important coarse cereals (barley, sorghum, and pearl millet). Production of maize continued to increase and reached 11.5 million tons in 1999/2000, from a mere 4.1 million tons in 1960/61 and 7.5 million tons in 1970/71, mainly due to a notable rise in its yield levels (Fig. 12.21). Maize yields went up from 1.1 t/ha in the triennium average ending (TE) 1981/82 to 1.7 t/ha in TE 1998/99. The maize area also gradually expanded from about 4.4 M ha in TE 1960/61 to 5.9 M ha in TE 1980/81 and 6.2 M ha in TE 1998/99. This is a clear indication that maize is gradually spreading to new areas and to some extent, also replacing barley, sorghum and pearl millet as a feed and fodder crop.

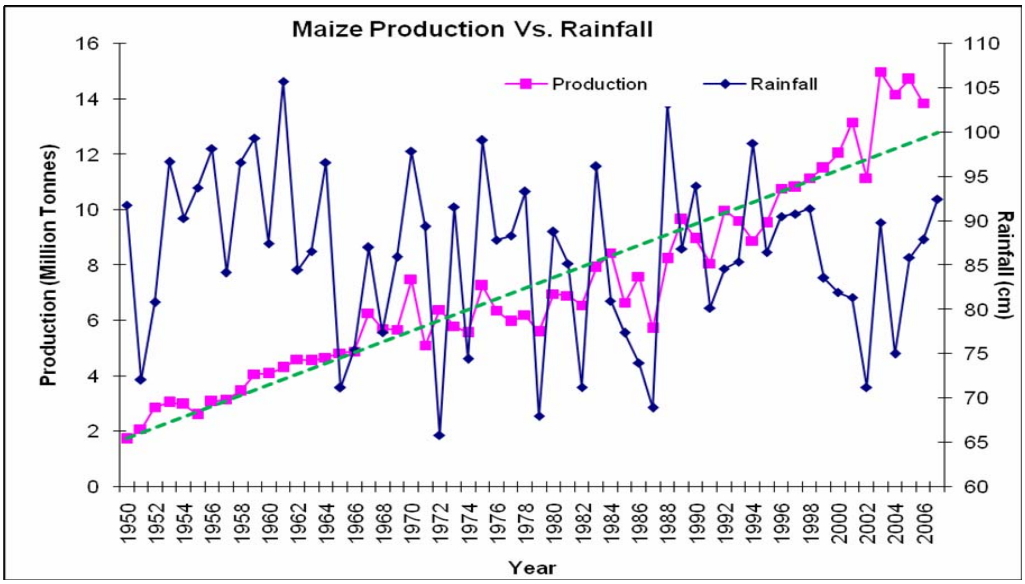

Fig. 12.21: Maize Production vs. Rainfall

During 2001-02, as much as 70% of the maize grown in India was cultivated in six states (Andhra Pradesh, Bihar, Karnataka, Madhya Pradesh, Rajasthan and Uttar Pradesh). In 1999/2000, the national average maize yield (1.8 t/ha) was far behind the world average of 4.86 t/ha. During this period, the average maize yield on about 45% of the total maize area in India was less than 1.5 t/ha and on only 15% it was slightly more than 3 t/ha. Lower yields and higher production costs in India, as compared to other countries, made maize non-competitive on the international market. In a globally competitive environment, maize yields in India need to increase to protect the maize producer. In 1999/2000, maize yield levels across states ranged from less than 1.5 t/ha in Madhya Pradesh, Rajasthan and Uttar Pradesh to more than 3 t/ha in Andhra Pradesh, Karnataka and West Bengal. In Bihar, where a sizable area of maize was cultivated under irrigation, yield levels were still low, approaching 2 t/ha. These four traditional maize growing states (Bihar, Madhya Pradesh, Rajasthan and Uttar Pradesh) have huge potential to raise maize production through increasing yield levels and intensifying cultivation in upland areas, provided that existing constraints are alleviated. In 1998/99 these states accounted for nearly 60% of the total maize area and about 40% of total production in India. An increase in average maize yields of about 25% in these states would result in additional maize production equal to more than 1 million tons throughout the country.

All the Kharif crops except sorghum are strongly associated with ENSO conditions (i.e. NINO3.SST and Darwin SLP anomalies) (Krishnakumar et al., 2004). But not valid for Rabi crops. When examining the annual statistics, the influence of ENSO is apparent only for rice, groundnut, oilseeds and food grains. The influence of NINO3.SST anomalies on kharif food grain production is strongest in the western and central peninsula (Gujarat, Rajasthan, Uttar Pradesh, Punjab and Andhra Pradesh) and to a lesser extent in Karnataka and Tamil Nadu.

The rapid increase in food grain production during the green revolution occurred primarily in the areas with irrigation and relatively few soil and climatic constraints on production. In contrast, the progress has been rather slow in the rainfed belt which accounts for about 70% of the area and almost half the total crop production in the country. The rapid increase in production during the green

revolution was associated with (i) a large increase in yield due to adoption of new dwarf, high yielding varieties (of rice and wheat, in particular) and fertilizer (ii) marked increase in the area under cultivation. This increase in average yield was made possible by substantial increase in fertilizer application, irrigation and pesticide application.

It has been noticed that the growth rate of food grain production has in fact decreased over last few years. This decrease is sharper in the production of the kharif season. The declining growth rates are nothing but the fatigue of green revolution. The fertility of the lands is also lost to some extent. For example, for an enhancement of production by 15 kg of grain, 1 kg fertilizer nutrients were adequate in seventies but at present 1.5 kg of fertilizer nutrient are required.

Furthermore due to the presence of large irrigated areas under monoculture, the level of incidence/infestation of pests/diseases has increased enormously. The production was maintained at high level by increasing the use of pesticides. The consumption of pesticides increased from about 2000 tons during 1950-51 to 24,000 tons during 1970-71, 45,000 tons during 1980-81 and is now leveling off at about 70, 000 tons.

The impact of monsoon vagaries has remained very large throughout the period. It is necessary to assess the extent to which the potential crop production is achieved in drought and good monsoon years. More emphasis on the management of non-drought years may result in higher overall production.

It is surprising that in the rainfed belt, with low level yield which are highly variable, farmers operating under constraints of limited resources (for fertilizers, pesticides and soil and water conservation measures) have been reluctant to adopt the strategies worked out at research stations and the ‘Lab to Land’ transfer of technology approach has not been very successful.

12.5. Crop insurance schemes in India

In order to provide a boost to the agriculture in India, a number of experimental crop insurance schemes have been introduced in the country. Considerable attention has been given to extension of insurance services to the rural sector, particularly to agricultural activities. It is realized that agricultural insurance can help insurers in broad-basing their activities. It can also provide a measure of support to a vital segment of the national economy. Agricultural insurance, thus, has an important role in shaping not only the financial stability of the farming community but also the national economy.

A well-devised agricultural insurance programme can be an effective tool of implementation of the country's agricultural policies and help in promotion and stabilization of the agricultural sector. Above all, it enables introduction of risk management practices to the rural setting. Decision making and handling of agricultural risks must go through the risk management process.

12.5.1. National Agricultural Insurance Scheme (NAIS)

(Rashtriya Krishi Bima Yojana – RKBY)

The **NAIS** itself was coined in order to overcome the many shortcomings — primarily huge losses —incurred by the earlier Comprehensive Crop Insurance Scheme (CCIS).

Table 12.7: Performance of the National Agricultural Insurance Scheme

Particulars	Rabi 1999-2000	Kharif 2000	Rabi 2000-2001
Farmers covered (No)	579940.00	8409419.00	2079109.00
Sum Insured (Rs Crore)	356.41	6903.47	1525.15
Premium (Rs Crore)	5.42	206.51	27.45
Area Coverage (in million ha)	7800.00	13000.00	3092.00
Claims (Rs Crore)	**7.69**	**1179.49**	**41.90**

12.5.2. Nature of coverage and indemnity

If the 'Actual Yield' (AY) per hectare of the insured crop for the defined area on the basis of requisite number of Crop Cutting Experiments (CCEs)] in the insured season, falls short of the specified 'Threshold Yield' (TY), all the insured farmers growing that crop in the defined area are deemed to have suffered shortfall in their yield. The Scheme seeks to provide coverage against such contingency.
'Indemnity' shall be calculated as per the following formula:

$$\frac{\textbf{Shortfall in Yield}}{\textbf{Threshold yield}} \times \textbf{Sum Insured for the farmer}$$

{Shortfall in Yield = 'Threshold Yield - Actual Yield' for the Defined Area}.

12.5.3. Piloting weather insurance in India

Weather insurance is a tool for all farmers (big and several hundred smallholders). These insurance policies protect them against extreme aberration in weather patterns. The pilot program has been launched for non-irrigated farmers in developing countries for their livelihoods. Weather insurance does not suffer from the usual moral hazard and adverse selection and high administration cost problems of traditional crop insurance, and it is therefore better suited to small farmers in rainfall-dependent regions.

As has been noted, the spread of agricultural insurance in India is restricted, inter alia, because of lack of knowledge on the subject. Basic research on several aspects of agricultural insurance is also needed.

12.6. Challenges ahead

The challenges are (i) to enhance the crop production in rainfed areas and (ii) to attempt to overcome fatigue of the green revolution and enhance the growth rate using climate information.

12.6.1. Rainfed areas

The major problems in the rainfed regions are (i) the large fluctuation in production in response to the variability in rainfall (ii) losses due to infestation of pests/diseases. The task in the rainfed region is to identify strategies which can enhance the overall production, with appropriate management of pests/diseases, in the face of climate variability. In semi-arid regions the emphasis should be on effective management of years of normal and good rainfall, rather than droughts.

12.6.2. High potential areas

The impact of intensive agriculture, with an efficiency of utilization of irrigated water only about 30-40% has led to degradation of land and large areas have become uncultivable due to water logging and salinity. Typical example of salinization caused by rise in groundwater was observed in Uttar Pradesh, Haryana, Rajasthan, Maharashtra and Karnataka. It has been observed that the high potential areas comprise large relatively homogeneous regions under monoculture. The presence of host plants and habitats, which are continuously favourable for growth of pests/diseases and led to, attacks of ever increasing intensity. Management in this situation requires pesticides in large quantities, which increase cost of farming and also pollute the environment. The adverse effects of pesticides on soil microflora and hence soil health has been well known. Excessive use of pesticides is also undesirable if opportunities to export global markets are to be used.

In rainfed areas also, the cropping patterns have changed from the traditional multiple cropping to monoculture over a large fraction of the area. There are patches of irrigated land in the midst of the rainfed belt, in which crops are grown round the year. Thus conditions favourable for the maintenance and growth of pests/diseases akin to the high potential areas have been created. Hence management of pests and diseases has become critical for the rainfed regions as well.

Efficient management of natural resources with use of soil, water resources and chemical inputs (fertilizers and pesticides) at the optimal level is needed for the high potential areas. In order to determine the optimal level, crop growth simulation

models and decision support system have to be used. For determining the optimum dosages, mixes, timing, etc. of fertilizer applications they have to be modified to incorporate the relevant facets of plant physiology and advances in areas such as nitrogen fixation, bio-fertilizers, etc. on the specific varieties grown under Indian conditions.

Decision support systems have to be developed for the complex and diverse ecologies of the rainfed belt. These decision support systems have to harness the advances made over last two decades in understanding the variability of the monsoon and generating predictions over short and medium range with the advances in agricultural sciences. Active participation of the experts in agricultural, atmospheric sciences and farming system is essential.

Some problems such as management of pests/diseases have to be tackled with development of models as well as setting up a network of special observatories for generation of micro-meteorological data as input to model of pests and diseases. It may also be possible to tailor varieties and management practices for avoidance of some pests and diseases.

Studies on what are the optimal cropping patterns in specific agroclimatic zones are also of great interest. Farmers have adopted cropping patterns on the basis of economic pasture/market forces. But by matching the soil and climate variability regime with the requirements of crops a optimal land use system can be derived. With the concerted efforts by the interdisciplinary groups with active participation of the farmers to generate the science that can lead to the enhancement of growth rates of food grain production is needed.

The thrust areas in agrometeorology are:

(i) Integration of results from time-series analysis based on historical crop-weather data, with current crop condition and forecast weather situations both at local and regional level for formulation of weather-based agro advisories. (ii) Development of models for early warning systems to cover all types of stress caused due to aberrant weather. (iii) Organization of crop data collection and on-line transmission facilities from a network of crop observatory (similar to weather

observatory network stations) to be operated from selected spots in the cropped fields; to relate satellite observations validated by ground truth. (iv) These should lead to preparation of “Agro-weather Charts” including information on both crop conditions and weather on a single map, as a routine. (v) Collection and compilation of agrometeorological information for each of the major crops from available literature supported by laboratory and field research to fill up gaps in such information where required. (vi) Monitoring weather systems on synoptic scale and crop conditions for development of region/zone based risk management strategies and operations in agriculture. (vii) Studies on impact of climate change and variations on crop planning and productivity need to be intensified. (viii) Development of early warning system models for identifying incidence, growth, intensification and spread of insect pest and diseases in crops/animals in relation to weather and plant factors. (ix) Use of crop-weather simulation models and development of EXPERT Systems for day-to-day weather based agricultural operations using real-time data. (x) Strengthening of education and in-service training in agricultural meteorology is essential. (xi) Development of strategies and contingency plans for mid-season corrections in irrigation scheduling, integrated pest and disease management and any other related operations during extreme events like droughts and floods, heat and cold stress.

12.6.3. Adaptation strategies in Agriculture

Any perturbation in agriculture can considerably affect the food systems and thus increase the vulnerability of a large fraction of the farming community. There is a need for developing adaptation strategies for minimizing the negative impact of climate change. We need to identify ‘no-regret’ adaptation strategies for sustainable development of agriculture. These adaptations can be at the level of individual farmer, society, farm village, watershed, or at national level. Some of the possible adaptation options are discussed below:

(i) Change in land use and management

It is possible by altering dates of planting, spacing and input management.

(ii) Development of resource conserving technologies

Recent researchers have shown that surface seeding or zero-tillage establishment of upland crops after rice gives similar yield to when planted under normal conventional tillage over a diverse set of soil conditions. These restrict release of soil carbon thus mitigating increase of CO_2 in the atmosphere.

(iii) Improved land use and natural resource management policies and institutions

This includes crop insurance, subsidies, pricing policies and change in land use. Policies are needed that would encourage farmers to conserve water, energy and soil resources.

(iv) Improved risk management through early warning system and crop insurance

The increasing probability of floods and droughts increase the vulnerability of poor farmers to global climate change. Early warning systems and contingency plans can provide support to regional and national administration, as well as local bodies and farmers to adapt. Policies that encourage crop insurance can provide protection to the farmers in the event their farm production is reduced due to natural calamities. Weather based index for crop insurance would help to protect the interest of the farmers especially in dry farming region.

(v) Reducing dependence on agriculture

Although the share of agriculture in gross domestic product in India has declined to less than 20% but large population continues to remain dependent on this. Such trends have resulted in fragmentation and decline in size of holdings leading to inefficiency in agriculture compared to world (Table 19.8 and 19.9) and rise in unemployment and low volume of marketable surplus and therefore increased vulnerability to global change.

Table12.8: India's Position in World Agriculture

Item	India	World	India's		Next to
			Share %	Rank	
1. Area(million ha)**					
Total area	329	13428	2.4	Seventh	Russian Federation, Canada, U.S.A., China, Brazil, Australia
Land area	297	13067	2.3	Seventh	Russian Federation, China, Canada, U.S.A., Brazil, Australia
Arable land	162 F	1404	11.5	Second	U.S.A.
Irrigated area	55 F	277	20.6	First	
2. Population (million)					
Total	1050	6225	16.9	Second	China
Agriculture	755	3234	23.3	Second	China
3. Economically active population (million)					
Total	460	3037	15.1	Second	China
Agriculture	270	1333	20.3	Second	China
4.Crop production (million t)					
Total Cereals	232	2075	11.2	Third	China, USA
Wheat	65	556	11.7	Second	China
Rice (Paddy)	132	589	22.4	Second	China
Coarse grains	35	930	3.8	Forth	U.S.A., China, Brazil
Total pulses	12	57	21.1	First	
Groundnut	8*	36	22.2	Second	China
Rapeseed and Mustard	4	36	11.1	Third	China, Canada
5. Fruits and Vegetable (million t)					
Vegetables and Melons	82F	842	9.7	Second	China
Fruits excluding Melons	46	480	9.6	Second	China
Potatoes	23	311	7.4	Third	China, Russian Federation
Onion (Dry)	5F	53	9.4	Second	China
6. Commercial crops (million t)					
Sugarcane	290	1333	21.8	Second	Brazil
Tea	0.89	3.21	27.7	First	
Coffee (green)	0.28	7.20	3.9	Sixth	Brazil, Vietnam, Indonesia,

Item	India	World	India's		Next to
			Share %	Rank	
					Colombia, Mexico
Jute and allied fibers	1.98	3.23	61.3	First	
Cotton(lint)	2.10	19.53	18.8	Third	China, USA
Tobacco leaves	0.60	6.2	9.7	Third	China, Brazil
7. Livestock (million head)					
Cattle	226 *	1371	16.5	First	
Buffaloes	97 *	171	56.7	First	
Camels	0.90 F	19.07	4.7	Fourth	Somalia, Sudan, Mauritania
Sheep	59 F	1024	5.8	Third	China, Australia
Goats	124 F	768	16.1	Second	China
Chicken	824 F	16605	5.0	Fifth	China, USA, Indonesia, Brazil,
8. Implements (000' numbers)**					
Tractors- in –use	1525 F	26704	5.7	Fourth	USA, Japan, Italy
9. Animal Products					
Milk (000'Mt)	86960*	599600	14.5	First	
Eggs (000'Mt)	2200 F	60469	3.6	Fourth	China, U.S.A., Japan
Total meat (000'Mt)	6038F	253528	2.4	Sixth	China, U.S.A., Brazil, Germany, France

Note : FAO Estimates, * Unofficial Figures, ** Figures relate to

Table 12.9: Production of Major Crops (Million Tonnes)

Group	1970-71	1980-81	1990-91	1999-00	2000-01	2001-02	2002-03	2003-04	2004-05	2005-06	2006-07
Food grain	108.4	129.6	176.4	209.8	196.8	212.9	174.8	213.2	198.4	208.6	216.1
Kharif	68.9	77.7	99.4	105.5	102.1	112.1	87.2	117.0	103.3	109.9	110.5
Rabi	39.5	51.9	77.0	104.3	94.7	100.8	87.6	96.2	95.1	98.7	105.6
Cereals	96.6	119.0	162.1	196.4	185.7	199.5	163.6	198.3	185.2	195.2	201.9
Kharif	65.0	73.9	94.0	100.7	97.6	107.2	83.1	110.8	98.6	105.0	105.8
Rabi	31.6	45.1	68.1	95.7	88.1	92.3	80.5	87.4	86.6	90.2	96.1
Pulses	11.8	10.6	14.3	13.4	11.0	13.3	11.1	14.9	13.1	13.4	14.2
Kharif	3.9	3.8	5.4	4.8	4.4	4.8	4.2	6.2	4.7	4.9	4.7
Rabi	7.9	6.8	8.9	8.6	6.6	8.5	6.9	8.7	8.4	8.5	9.5
Rice	42.2	53.6	74.3	89.7	85.0	93.3	71.8	88.5	83.1	91.8	92.7
Kharif	39.5	50.1	66.3	77.5	72.8	80.5	63.1	78.6	72.2	78.3	80.1
Rabi	2.7	3.5	8.0	12.2	12.2	12.8	8.7	9.9	10.9	13.5	12.6
Wheat	23.8	36.3	55.1	76.4	69.7	72.8	65.8	72.2	68.6	69.4	74.9
Jowar	8.1	10.4	11.7	8.7	7.5	7.6	7.0	6.7	7.2	7.6	7.4
Kharif	5.8	7.5	8.3	4.8	4.5	4.2	4.2	4.9	4.0	4.1	3.7
Rabi	2.3	2.9	3.4	3.9	3.0	3.4	2.8	1.8	3.2	3.6	3.7
Maize	7.5	7.0	9.0	11.5	12.0	13.2	11.2	15.0	14.2	14.7	15.0
Bajra	8.0	5.3	6.9	5.8	6.8	8.3	4.7	12.1	7.9	7.7	8.6
Gram	5.2	4.3	5.4	5.1	3.9	5.5	4.2	5.7	5.5	5.6	6.3
Tur	1.9	2.0	2.4	2.7	2.2	2.3	2.2	2.4	2.3	2.7	2.4
Oilseeds	9.6	9.4	18.6	20.7	18.4	20.6	14.8	25.2	24.4	28.0	23.8
Kharif	7.0	5.0	9.8	12.5	11.9	13.2	9.0	16.7	14.1	16.8	13.9
Rabi	2.6	4.4	8.8	8.2	6.5	7.4	5.8	8.5	10.2	11.2	9.9
Groundnut	6.1	5.0	7.5	5.3	6.4	7.0	4.1	8.1	6.8	8.0	4.9
Kharif	na	3.7	5.1	3.8	4.9	5.6	3.1	6.9	5.3	6.3	3.3
Rabi	Na	1.3	2.4	1.5	1.5	1.4	1.0	1.2	1.5	1.7	1.6
Rapeseed and Mustard	2.0	2.3	5.2	5.8	4.2	5.1	3.9	6.3	7.6	8.1	7.1
Sugarcane	126.4	154.2	241.0	299.3	296.0	297.2	287.4	233.9	237.1	281.2	345.3
Cotton[c]	4.8	7.0	9.8	11.5	9.5	10.0	8.6	13.7	16.4	18.5	22.7
Jute and	6.2	8.2	9.2	10.6	10.5	11.7	11.3	11.2	10.3	10.8	11.3

Mesta[d]											
Jute[d]	4.9	6.5	7.9	9.4	9.3	10.6	10.3	10.3	9.4	10.0	10.3
Mesta[d]	1.3	1.7	1.3	1.2	1.2	1.1	1.0	0.9	0.9	0.9	1.0
Plantation crops											
Tea[e]	0.4	0.6	0.7	0.8	0.8	0.9	0.9	0.9	0.9	0.9	0.9
Coffee	0.1	0.1	0.2	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3
Rubber	0.1	0.2	0.3	0.6	0.6	0.6	0.6	0.6	0.6	0.6	0.6
Potato	4.8	9.7	15.2	24.7	22.5	23.9	23.3	23.1	23.6	23.9	na

Sources: 1) Directorate of Economics and Statistics, Department of Agriculture and Cooperation

2) Ministry of Commerce & Industry

na: Not available

[a] 4th advance estimates

[b] include groundnut, rapeseed & mustard, sesamum, linseed, castorseed, nigerseed, safflower, sunflower and soyabean

[c] Bale of 170 Kgs.

[d] Bale of 180 Kgs

[e] Calendar Year

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CHAPTER 13

INDIAN NORTHEAST MONSOON

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13.1 Introduction

The Indian northeast monsoon (NEM), also known as the retreating southwest monsoon (SWM), is a small scale monsoon confined to parts of southern Indian peninsula with duration October-November-December (OND). It is associated with the seasonal reversal of surface and lower tropospheric winds from southwesterlies (during the SWM season of June–July-August-September (JJAS) to northeasterlies which set in over the Indian region in October (India Meteorological Department (IMD), 1973). In a broader perspective, it is also associated with the northern hemispheric winter circulation dominated by a strong surface high pressure region over Siberia, a primary low over eastern equatorial Pacific region and secondary shallow lows over the north Indian ocean. Though the SWM is the most important weather phenomenon for the country providing it with nearly 75% of the normal annual rainfall of 115 cm (IMD, 2010), the NEM is also an important rainfall season for parts of southern peninsular India. The state and meteorological sub division of Tamil Nadu (TN) is the major beneficiary of the NEM rainfall, which is prodigious over the belt of Coastal Tamil Nadu (CTN). Over CTN, the onset of NEM, preceded by a well defined reversal of low level winds from southwesterlies to northeasterlies, is clearly marked. A synoptic / climatological overview of Indian NEM is given in IMD(1973). In this chapter we present a brief description of a few aspects of Indian NEM *viz*., rainfall climatology, flow pattern, onset and withdrawal, synoptic scale systems, certain thermodynamic features, relation with global features and seasonal forecasting.

13.2. Regions affected by NEM and their rainfall climatology

Geographical location

Fig.13.1 presents the geographical locations of five meteorological sub divisions of India *viz*., Coastal Andhra Pradesh (CAP), Rayalaseema (RYS), South Interior Karnataka (SIK), Kerala (KER) and TN (including the union territory of Pondicherry as well). Here, TN and KER are separate states, SIK is a part of Karnataka state, RYS and CAP are parts of the state of Andhra Pradesh. Fig.13.2 presents the boundaries of Bay of Bengal (BOB) which lies east of the southern Indian peninsula and the Arabian Sea (AS) that lies to the west. Both BOB and AS together constitute what is known as the North Indian Ocean.

Normal rainfall

The seasonal normal rainfall distribution over the above region for the season OND is depicted in Fig.13.3. Table 13.1a presents the normal monthly, seasonal and annual figures of the above five sub divisions based on the 50 year data of 1951-2000 (IMD, 2010). The rainfall figures expressed as percentage of annual rainfall are also given. In Table 13.1b are presented similar rainfall figures for 14 stations which by and large represent the various regions influenced by NEM. The inferences which could be drawn from the normal monthly and seasonal rainfall figures are described herein below, subdivision wise:

CAP: This sub division receives 1024 mm of annual normal rainfall out of which 327 mm (32%) is contributed by NEM. The month of October is the major rainy season for CAP. However, there is wide latitudinal variation of rainfall. Visakhapatnam, located on the north CAP (NCAP) at about 17.5˚N receives 302 mm during OND, October alone contributing 219 mm. Machilipatnam, located on the central stretch at about 16˚N receives 355 mm during OND. Nellore situated on the southern parts at 14˚N receives 656 mm with a breakup of 260 mm in October, 295 mm (November) and 104 mm (December) thus experiencing rainfall throughout the season.

RYS: Rainfall over this sub division during OND is 219 mm (31% of annual rainfall of 706 mm). Anantapur representing northwestern dry parts of RYS receives annual rainfall of 567 mm with NEM contribution at only 154 mm. Tirupati located in the southern parts of RYS receives 934 mm of annual rainfall with 467 mm realised during OND.

SIK: SIK receives 1019 mm of annual rainfall and 210 mm during OND (21% of annual). Bangalore located in SIK receives 977 mm of annual rainfall and the NEM seasonal contribution is 245 mm only. Most of this rain is received in October itself (171 mm).

KER: This sub division receives 478 mm of rainfall during OND. But the NEM rainfall is only 16% of the annual rainfall of 2928 mm, wherein, the main contribution comes in during the southwest monsoon season of JJAS (2051 mm, 70%). Kochi, located on the west coast of India in central Kerala at 10˚N, receives 497 mm of rainfall during OND, but this becomes insignificant compared to its annual rainfall of 3147 mm. Thiruvananthapuram, situated at 8˚N in southern Kerala, receives 528 mm of rainfall during OND, the annual rainfall being 1792 mm.

TN: NEM rainfall of Tamil Nadu (NRT) accounts for 48% (OND, 438 mm) of its annual rainfall of 914 mm making this state / sub division, the major beneficiary of NEM. This is also the lone meteorological sub division of India where rainfall realised during NEM is significantly more than that during the southwest monsoon season (JJAS, 317 mm). There is however wide spatial variation of rainfall within the state. At Salem, located in the northwestern parts of the state, annual rainfall is 1020 mm, JJAS rainfall is 516 mm and OND rainfall, 332 mm only. Coimbatore, located in the southern portion of the northwestern parts is drier than Salem, with OND contributing 309 mm which is nearly 50% of the annual rainfall of 614 mm. Tiruchirapalli, located in the central interior parts receives 872 mm (annual) with NEM contributing to 393 mm. Chennai city located at 13˚N receives 857 mm during OND and 1403 mm annually. Vedaranyam, located at 10˚N, at the tip of the north-south oriented coastal Tamil Nadu receives prodigious rainfall of 1038 mm during OND making this station the wettest in India during the NEM season. At 347.4 mm, the rainfall realised in the month of December is quite substantial. Tuticorin, located at 8.8˚N on the southeast

coast of Tamil Nadu abutting Gulf of Mannar, receives 440 mm of its annual rainfall of 631 mm during OND.

A reference to the daily variation of normal rainfall for the month of October for the stations whose normal rainfall figures are presented in Table 13.1b provides some more insight into the extent of southwest monsoon rainfall in the northeast monsoon season of OND. The SWM, by and large, continues up to the first half of October in most of the five sub divisions benefitted by the NEM (IMD,1973). The normal rainfall for 1-15 October (Oct-I) and 16-31 October (Oct-II) are also presented in Table 13.1b. As seen, VSK and MPT situated on the east coast of the Indian peninsula receive nearly same rainfall during Oct-I and Oct-II. In respect of the coastal stations of NLR, MDS, VRM and TTC, Oct-II (110-191 mm) receives substantially more rainfall than Oct-I (39-74 mm). At BNG, SLM and ANT, Oct-I rainfall exceeds that of Oct-II. However, at CMB, despite its relative dryness, Oct-II rainfall (93mm) far exceeds Oct-I rainfall (54mm). At CHN, both the figures are nearly equal whereas, at TRV, Oct-II receives higher rainfall than Oct-I.

The above description of spatial distribution of monthly and seasonal rainfall, especially, the rainfall during the NEM season, clearly brings forth the following features:

(i) The Oct-Dec NEM rainfall is highest along South CAP (SCAP) and North CTN (NCTN) with spatial rainfall variation of 70-100 cm.

(ii) Over CAP, RYS and KER, OND rainfall decreases from north to south.

(iii) Over TN, CTN receives more rainfall than Interior TN (ITN) for a given latitude. Over CTN also, OND rainfall increases from Chennai to Vedaranyam but decreases further south due to the sheltered (by Sri Lanka lying east) nature of the extreme south eastern peninsula (Fig.13.1).

(iv) During OND, the month of October is the rainiest over NCAP, North RYS, North Kerala and ITN. In these regions, rainfall during December is not very high. In most parts of CTN, November is the rainiest month. Over SCAP, KER and CTN, rainfall continues in December also.

(v) Over parts of CTN, the rainfall during the month of January is also high (VRM: 64mm).

(vi) In October which experiences both the monsoons, rainfall contribution from southwest and northeast monsoons appears to be the same over north and central CAP. Over the North RYS, SIK, northern parts of ITN, north / central Kerala, SWM rainfall exceeds NEM rainfall in October. Over ITN, NEM rainfall is moderately higher. However, over SCAP and the entire CTN, the Oct NEM rainfall far exceeds Oct SWM rainfall.

13.3. Statistical properties and Inter annual variation of NEM rainfall

The long term monthly and seasonal rainfall data of sub divisions benefitted by NEM can be obtained from two readily available sources *viz*.(i) the web site of the Indian Institute of Tropical Meteorology (IITM), Pune which provides such data for the period since 1871 and (ii) the National Data Centre (NDC), IMD, Pune which provides data from 1901. (i) is a homogeneous series derived from the data of a few fixed stations and (ii) is based on a larger dataset but the number of stations used vary from year to year. Using the normals presented in Table 13.1 and the percentage departure from normal (PDN) values from the source (ii), the actual rainfall for a given year for any month/season could be derived. This methodology is used to generate the rainfall data of the five sub divisions considered, for the 110 year period of 1901-2010, for the OND season. The NEM rainfall of the Southern Region (SR) defined as the union of the five sub divisions (NRS) considered is also derived for each year by area weighted averaging of the rainfall of the five sub divisions. Table 13.2a presents the rainfall and the PDN of NRS and NRT for the period 1901-2010. Fig.13.4a&b depict the two series graphically. It must be stated here that rainfall figures taken from other sources might differ slightly in individual years, due to differing basic databases, method of computation adopted to derive the rainfall etc.

The mean rainfall, standard deviation (SD) and coefficient of variation (CV) have been derived for the rainfall series of each of the five sub divisions and for NRS. The mean rainfall for 1901-2010, SD, CV and the range for all the five sub divisions and the SR are presented in the Table 13.2b. The difference between the normal based on IMD(2010) and the long period average (LPA) based on 1901-2010 data is only marginal. As shown, the CV of OND rainfall is lowest for Kerala and

Tamil Nadu at 27.3% and varies in the range of 36-40% for the remaining three sub divisions. The CV of NRS is 25.3%. The range of rainfall is also quite substantial as could be seen from Table 13.2b.The correlation coefficient (CC) between NRS and NRT during the period 1901-2010 is 0.84 at 0.1% level of significance (LS). During our subsequent discussions, frequently NRT will be taken as the index of NEM activity.

For both NRS and NRT, a year is defined as an excess NEM rainfall year if the PDN is ≥20% and deficient if the PDN is ≤-20%, in accordance with the criteria followed by IMD. For TN, an overall below par performance of NEM was experienced during 1947-1976 (mean PDN: -9.1%) with 4 consecutive deficient years during 1949-1952. The period 2004-2010 did not register any negative departure of NRT at all but experienced as many as 4 excess rainfall years with a high of 79% PDN in the year 2005. This prodigious performance of NEM over TN continued in the year 2011 as well which registered a PDN of 23% during OND. This run of positive PDN for 8 consecutive years is unprecedented in the time series of NEM rainfall of TN, even if the time series is extended back to the year 1871, i.e., for the period 1871-2011.

Table 13.2c presents the excess and deficient NEM years for SR and TN. As shown, during the 110 year period of 1901-2010, there have been 26 and 27 excess and deficient years over SR. For TN, the corresponding figures are 26 and 25. Thus the frequencies corresponding to excess and deficient years are by and large same showing good symmetry. During the 19 year period 1947-1965, no excess NEM rainfall was recorded in TN. The NRT was excess during three consecutive years in 1930, 1931 and 1932 and again in 1977, 1978 and 1979. Table 13.2d presents the decadal frequencies of excess and deficient years for NRS and NRT, for the period 1901-2010. It is clearly seen that the last two decades, *viz*., 1991-2000 and 2001-2010 experienced more excess rainfall years than deficient rainfall years for both SR and TN.

The power spectra of NEM rainfall of SR and TN are presented in Fig.13.5a&b. The values of first auto CC r_1 for the series of NRS and NRT are 0.02 and 0.l3 respectively thus showing insignificant persistence. The power spectra of NRS and NRT reveal periodicities of 14-15 years and 12-13 years respectively which

could perhaps be attributed to the sun spot cycle. It is interesting to note that the well known 2-3 year periodicity present in the SWM rainfall of India (Asnani, 2005) is not so conspicuous in the case of NEM rainfall. Though the spectra for NRS and NRT rainfall peak corresponding to periodicities of 3 and 2.9 years respectively, both the peaks fall short of being significant.

13.4. Month wise wind flow pattern over India during the period September-February

The IMD has defined four seasons of India *viz*., winter (January-February (JF)), pre-monsoon or hot weather (March-April-May (MAM)), southwest monsoon (JJAS) and post monsoon or northeast monsoon (OND). During winter for which January is the representative month, the surface pressures which are high over northwestern parts of India decrease towards south. In July however, the pattern reverses with low pressure developing over northwest India and increasing pressures towards south (Fig.13.6a-d) (Rao and Ramamurti, 1983). This reversal is caused by the intense surface heating that takes place over India and neighbouring countries during the MAM season. The southwest monsoon sets in over India by the end of May/ beginning of June and continues up to September / early October. In October the surface pressure pattern changes with the establishment of a dumb bell shaped isobar. The low level winds at 850 hPa during the months of September, October and November are presented in Fig.13.7a-c (Source: Asnani, 2005). As seen, the seasonal trough at 850 hPa is located near 20˚N in September but gradually moves southwards to 16-17˚N in October and thence to 12-13˚N in November. The reversal of winds from southwesterlies to northeasterlies that takes place in October in the peninsula can be clearly observed. This is a prelude to the establishment of northeast monsoon over the southern Indian peninsula. Rao (1976) and Asnani (2005) could be referred which provide excellent description of the wind flow over India during southwest monsoon and as to how the flow changes into northeasterlies after the monsoon's retreat.

13.5 Onset, withdrawal and duration of northeast monsoon

The southwest monsoon which is the overriding weather phenomenon during June-September starts withdrawing from northwest India in the beginning of September (Rao, 1976). By 15th October, the monsoon withdraws up to 15˚N over the peninsula, though the dates may show wide variation in individual years. North of 20˚N and in most of the areas north of 17˚N, the withdrawal of southwest monsoon is associated with near cessation of rainfall and rise of maximum temperature in October. However south of 15˚N, there is a marked rise in rainfall followed by increased clouding and continuation of rainfall in the eastern parts of the peninsula. This marked increase in rainfall occurring after the reversal of winds from southwesterlies to northeasterlies is considered as the northeast monsoon onset.

Reference to old issues of Indian Daily Weather Reports (IDWR) published by IMD revealed that northeast monsoon onset was first mentioned in the year 1923 by IMD. Subsequently, the onset was mentioned during some years but not mentioned during several other years. However, since 1977, the setting in of northeast monsoon has been mentioned in the IDWRs every year though phraseology sometimes differed from year to year. There was no well defined and standard criteria to declare the NEM onset as mentioned in IMD (1973).

Determination of the dates of NEM onset

The 1987 conference of forecasting officers of IMD defined a set of criteria to declare NEM onset over southern peninsula on real time basis (IMD, 1987). These criteria were slightly modified to determine NEM onset over CTN in a diagnostic study (Raj, 1992). The NEM onset dates over CTN were derived for the period 1901-1990 in this study based on daily rainfall of six stations located in CTN *viz*., Chennai, Cuddalore, Nagapattinam, Vedaranyam, Pamban and Tuticorin. Fig.13.8a presents the geographical locations of these stations.

Criteria for determination of dates of NEM Onset

The following five rules (R_1 to R_5) formulated by Raj(1992) were used:

R_1: Southwest monsoon should have withdrawn up to coastal Andhra Pradesh

R_2: Deep easterlies should have set in over Tamil Nadu or seasonal low should have established in south BOB adjacent to Tamil Nadu coast.

R_3: After R_1 and R_2 are satisfied, the first day of *Fairly widespread* (FW) rainfall or a higher category over CTN would be the day of northeast monsoon onset.

R_4: If the date arrived at by R_3 happens to be earlier than 10 October, the winds / surface charts are to be scrutinised to decide as to whether the onset of easterlies are temporary or permanent. If it is permanent, then the date of R_3 should be taken as the onset date. If the easterly onset is temporary and if westerlies appear again in the lower troposphere over CTN, the date of permanent onset of easterlies is to be determined and R_3 to be applied again.

R_5: In case, the date fixed is completely unsatisfactory, a review is to be made and the next date of FW rainfall is to be considered as onset date.

The onset dates of NEM over CTN were derived for the period 1901-90 in Raj(1992) and for 1991-2000 in Raj(2003). Table 13.3a presents the dates of (i) easterly onset over CTN, (ii) onset of NEM over TN whenever declared by IMD and (iii) onset of NEM determined by Raj(1992 and 2003).

Determination of dates of Withdrawal

Reversal of lower tropospheric winds from southwesterlies to northeasterlies over CTN is taken as a pre-requisite for declaration of onset of northeast monsoon. But, no such conspicuous or discernible change in the flow pattern either at lower or upper levels of the atmosphere over the Indian region is known to be associated with the withdrawal of northeast monsoon over CTN. The low level easterlies continue over the NEM region and the south BOB up to around mid-April. By this time, air temperature substantially increases over the region due to increased insolation leading to occasional thunderstorm activity over southern Indian peninsula. Obviously, this type of summer rainfall activity cannot be considered as NEM though the low level winds might be easterlies. Thus, the NEM withdrawal criteria has to be

based not on the wind pattern but on the spatial and temporal variation of rainfall. Hence, the rules for determining the date of withdrawal of northeast monsoon have been formulated based on *rainfall*, which is the primary index of monsoon activity (Ananthakrishnan et al, 1967). For this purpose, a parameter *Daily Rainfall Index* (DRI) has been carefully defined to quantify the spatial and temporal persistence of rainfall over CTN (Raj, 1998b).

Daily Rainfall Index

The DRI for a day has been defined as the percentage number of rainy days over a five day pentad, with the day in question as the central day, the number of rainy days counted over all the stations with available data. Thus, for the Nth day, $(N-2)$, $(N-1)$, N, $(N+1)$ and $(N+2)^{th}$ days are considered. Out of $5M$ rainfall observations of M stations, if $M1$ observations correspond to a *rainy day* (day with at least 2.5 mm of rain), then $DRI= (M1/5M)100$ for the Nth day. Obviously, DRI varies between 0 and 100. A threshold value of 40 has been chosen as the minimum DRI value a day must have for the northeast monsoon to be prevalent. This threshold value has been arrived at after a detailed examination of the intra seasonal variation of DRI for all the years. As it is well known that NEM rainfall occurs in spells with long dry spells in between (IMD,1973), the withdrawal of NEM should be identified with the end of the ultimate wet spell which marks the cessation of the season. For identification of the date of withdrawal, the same methodology adopted by Raj (1998b) is followed.

Rules for determination of date of withdrawal of northeast monsoon over CTN

The rules for determination of date of withdrawal of northeast monsoon which were followed in the above study are as under:

R1: The DRI is to be computed for every day for the period September to February (of the next year). If $DRI>40$ for a day, the day would be deemed to belong to a significant rain spell.

R2: If no significant rain spell commenced on or after 1 January, the first day with $DRI<40$ which is not succeeded by dates of $DRI>40$ in the calendar either until 31 December or up to the end of the significant rain spell which may have commenced

on or before 31 December but continued thereafter, would be deemed to be the mid-date of the withdrawal pentad.

R3: If a significant rain spell commenced on or after 1 January, the withdrawal pentad is to be determined by critically studying the JF rainfall. The various rain spells, intensity and length of each, duration of dry spells in between – all are to be considered before concluding whether the JF spell could be considered as continuation of northeast monsoon.

R4: Once the withdrawal pentad is located, the precise date of withdrawal could be selected from the 5 days of the pentad by studying the spatial distribution of daily rainfall. Preferably, date of withdrawal should be a dry day over CTN.

The withdrawal dates for NEM over CTN as per the criteria defined above were derived for the period 1901-90 in Raj(1998b) and for 1991-2000 in Raj(2003). Table13.3a lists the withdrawal dates for the period 1901-2000.

IMD Criteria

The criteria for declaring onset of NEM was set by IMD in August 1988 by means of an official circular, which was amended further in August 2006 (IMD, 2008). The criteria for commencement of NEM rains as per the latest circular are:

(i) Withdrawal of SWM up to Latitude 15˚N.

(ii) Onset of persistent surface easterlies over Tamil Nadu coast.

(iii) Depth of easterlies up to 850 hPa over Tamil Nadu coast.

(iv) Fairly widespread rainfall over coastal Tamil Nadu, south coastal Andhra Pradesh and adjoining areas.

As seen, we have also adopted the above criteria only for determining the onset dates save for the provision of review in case the determined date turns out to be completely unsatisfactory. The criterion (i) above, *viz*., withdrawal of SWM up to 15˚N latitude proved to be a constraint in some years. If due to operational constraints the SWM is not withdrawn, NEM could not be declared on real time basis even if conditions (ii) - (iv) are satisfied. The determination of NEM onset dates in the various studies – Raj(1992 and 2003), were basically diagnostic and not carried

out on real time basis and it has been taken that (i) above always preceded or coincided with (ii).

As for NEM withdrawal, IMD started announcing withdrawal dates of NEM only from the year 1993 and prior to that no such declaration appears to have been made. A sub-committee constituted by the IMD's Annual Monsoon Review meet in 2006 submitted a detailed report recommending that real time declaration of NEM rainfall cessation may not be accurate and so may not be resorted to and that the date may be fixed diagnostically by considering rainfall data up to 31st January and a statement to the effect could be released on 31st January (IMD, 2006). This recommendation is still under consideration. IMD has declared cessation of NEM rainfall by considering rainfall and may be a few other parameters such as depth of moisture, temperature etc. in a subjective way but no objective criteria or guidelines to determine NEM have so far been published.

Table 13.3a enumerates the dates of NEM withdrawal for the period 1993-94 to 2010-11, as announced by IMD in its daily weather bulletins.

Redetermined / determined dates of onset and withdrawal of NEM

The process of NEM onset and withdrawal were found to be well defined mainly over CTN and SCAP. In other sub divisions, increase in rainfall associated with onset was not that conspicuous and in NCAP and North Kerala, the southwest monsoon activity in October itself overshadowed the NEM onset. The SCAP and CTN were found to be the most suitable regions to determine NEM onset and withdrawal.

In the most recent and extensive study undertaken on this topic, Geetha and Raj, (2011) re-determined the onset and withdrawal dates of NEM for the 100 year period of 1901-2000 based on the daily rainfall data of large number of raingauge stations numbering about 25. Stations of south CAP and CTN which were located within a distance of 100 kms from the coast were considered. The geographical locations of the stations considered for the study are presented in Fig.13.8b. The study also freshly determined the dates for the 19th century 30 year period of 1871-

1900 and the dates for the recent decade of 2001-2010. The methodology used for determination / redetermination of onset and withdrawal dates in the above study are the same as the methodologies described earlier except that instead of 6 stations of CTN, a larger network of 25 raingauge stations of SCAP and CTN have been used.

The onset and withdrawal dates of NEM over SCAP and CTN for the period 1871-2010 determined / re-determined as in the above study are listed in Table 13.3b. Comparison between onset / withdrawal dates as declared by IMD on real time basis and the dates as determined in the study on diagnostic mode reveal the following facts: Whereas there is negligible difference in respect of onset dates save for a few years, the difference in respect of withdrawal dates is quite substantial in several years. This aspect has been discussed in detail in IMD (2006) in the AMR subcommittee report on NEM withdrawal.

The statistics such as normal date, SD, range etc., are presented in Table 13.3c. In Fig.13.9(a-b) are presented the mean superposed epoch profile of rainfall over CTN and SCAP with reference to onset and withdrawal dates for each year for 50 days prior to the onset date and after the withdrawal date, based on the dates for the period 1901-2000.

In the above study, separate withdrawal dates for SCAP/NCTN and Central and South CTN (CSCTN) for the 50 year period 1961-2010 were also derived. Table 13.4 presents the dates determined and their statistical parameters. The major results on NEM onset and withdrawal dates determined in the various studies and mentioned in this section are stated herein below:

(i) The normal date of NEM onset over SCAP / CTN has remained steady over the long period 1901-2010 with a normal date of 20 October and a SD of 7-8 days.
(ii) The daily rainfall at the time of onset increases from 2-4 mm prevailing 50 days before onset to 10-13 mm up to 15 days after onset and remains above 8 mm even after 50 days of onset, as shown by the superposed epoch analysis.
(iii) The normal date of withdrawal (re-determined) over CTN during 1901-2000 is 30 December (SD of 14 days) which is 3 days later than the earlier determined normal date of 27 December for the same period.

(iv) During the 110 year period 1901-2010, the normal date of onset and withdrawal are 20 October and 29 December with SD of 8 and 14 days respectively.

(v) The range of revised onset dates during 1901-2010 is 4 October (2000) to 11 November (1915). The withdrawal range is 27 November (1910) to 22 January (1994-95). The mean duration of NEM for the same period is 70 days, SD 16 days with a range of 26-107 days.

(vi) Over SCAP/NCTN, the normal date of NEM withdrawal is 17 December and that for CSCTN, it is 31 December thus showing that NEM lingers on for nearly a fortnight over the CSCTN strip of CTN after its retreat from SCAP/NCTN.

The onset and withdrawal phenomena of NEM, as took place in two different years are pictorially illustrated in Fig.13.10a-b which present the daily rainfall over CTN during 1 October – 31 January for two years *viz.* 1898-99 and 2002-03. In 1898, though the date of low level easterly onset is not available, the commencement of NEM rains on 22 October is clear and precise. In 2002-03, the easterlies set in on 1 October, but onset of NEM took place on 9 October only. In both the years, the dates of withdrawal are also clearly defined at 1 Jan 1899 and 12 Dec 2002 respectively. Out of 140 years considered in this study, the NEM onset took place in November during 12 years. In nearly one-third of the years, the withdrawal of NEM spilled over to the next calendar year.

Direction of movement of clouding over BOB at the time of NEM onset

Raj et al (2007), based on INSAT Outgoing Longwave Radiation (OLR) data for the period 1981-2000, showed that the movement of low level clouds from BOB into the SCAP and CTN area is from southeast to northwest direction and *not* from the northeast direction from which the low level winds blow which is also the reason behind the naming of this monsoon as northeast monsoon. Fig.13.11a-d presents the spatial variation of mean OLR for -4, -2, 0 and 2 days with respect to the NEM onset dates. As shown, the 230 Wm^{-2} OLR isopleth (which is taken as associated with clouding) moves from 8ºN (-4 days) to 10.5ºN (-2 days), 15ºN (0th day, the date of onset) and 17ºN (2 days after onset) along the east coast of the peninsula clearly suggesting advancement of low level clouding from south to north. In Fig.13.12a-d, OLR distribution during the onset phase of 1997 is presented. In this

year, the NEM set in on 13 October. The OLR distributions are presented for 6, 9, 13 and 16 October. The movement of clouds from southeast to northwest is clearly evident.

13.6. Synoptic systems that form over Bay of Bengal during NEM season

With the shift of the equatorial trough (ET) southwards from September to December, the latitude of formation of low pressure areas also shifts southwards over the BOB. IMD classifies low pressure systems (LPS) based on the maximum wind speed (when they are located over the sea). The classifications and the corresponding wind speeds are presented in Table 13.5.

Raj(2011) presents a detailed climatology of several aspects of depression (D) / cyclonic storm (CS) / severe cyclonic storm (SCS) which form and move over Indian seas. Most of the statistics have been derived from the data for the 50 year period 1961-2010 which has been extracted from IMD's *Cyclone eAtlas* (2008) and its updates. As presented in Raj(2011), during OND, the probability of a BOB depression intensifying into a CS is 61% and that for CS to SCS is 65%, thus exhibiting high probability of intensification. The monthly frequencies of formation of D+CS+SCS, CS+SCS and SCS over BOB during 1961-2010 are 77, 70 and 37 for October, 34, 57 and 22 for November and 19, 42 and 13 for December respectively. The normal area of formation of LPS over BOB is (82.6ºE, 15.1ºN) in October, (86.2ºE, 14.4ºN) in November (84.9ºE, 11.6ºN) in December, shifting southwards with the advancement of the season . The frequencies of D+CS+SCS, CS+SCS and SCS which crossed the Tamil Nadu coast during OND 1961-2010 are 34, 24 and 19 respectively; for CAP, 46, 29 and 18; for Orissa 14, 10 and 9 and for West Bengal, 10, 7 and 5. For Bangladesh and Arakan (Myanmar) coasts, the frequencies are 36, 25 and 19. Thus large number of BOB LPS move in the northward and northeastward directions. Fig.13.13a presents the tracks of the LPS that formed over BOB during 1961-2010, for the months of October, November and December.

In the recent decades, the frequency of LPS crossing Tamil Nadu coast during OND has shown a slight decrease. The frequencies for the decades 1971-80, 1981-90, 1991-2000 and 2001-10 are 6, 4, 11 and 3 respectively. However, the

mean rainfall for the decade 2001-2010 for Tamil Nadu is excess by 15% of the LPA thus showing that despite the decrease of frequency of LPS crossing, the NEM sustained an above par activity over the state.

When a D/CS/SCS crosses southeast Indian coast during the NEM season, heavy to very heavy and extremely heavy rainfall could be realised over the coast and the interior regions as well. Generally, rainfall extends more towards northern coast than towards the southern side with reference to the landfall location. The D/CS/SCS are orgainsed transient systems which generally originate over the BOB, cross the coast, cause rainfall and then dissipate over the land. Occasionally, such systems reemerge into the AS and intensify again. Fig.13.13b presents the track of CS *Nisha* which formed over Sri Lanka, emerged into BOB and crossed Tamil Nadu coast and gave prodigious rainfall during 26-28 November 2008. Isohyetal maps of these three days are presented in Fig.13.13b.

Easterly wave (EW) is another transient synoptic scale feature which gives rise to good rainfall during the NEM season. An EW may not always be detected with clarity and precision as normally happens with LPS. However it is possible to track the movement of an EW using vertical time section plots and also by critically studying 24 hr pressure changes (IMD, 1973). Geetha and Raj (2011) studied the EW activity over the southern peninsula for the NEM season of 2010. In this year three EWs were detected and the period of the waves was determined as 4-5 days, the speed of movement as nearly 26 kmph and the amplitude, 2800 km.

Aside from the transient systems such and LPS and EW, weak and feeble low pressure areas (LOPAR) which do not have clearly defined centres, but oscillate from one location to other over the BOB off the southeast coast of peninsular India and troughs off the east coast can also cause active monsoon conditions over the NEM region. On some occasions low level (900-850 hPa) winds strengthen up to 40 knots over the BOB and the resultant moisture incursion and frictional convergence can cause active NEM with heavy rainfall. Generally, the NEM rainfall is heavier along the east coast and decreases inland. Late night and early morning is the preferred period for NEM rainfall when the rainfall is not associated with any LPS.

13.7. Variation of thermodynamic parameters during NEM

A study on thermodynamic parameters was undertaken based on 00 UTC (0530 IST) upper air data of Chennai Radio sonde / Radio wind data for the period 1971-80 (Raj,1996). Some of the results derived in this study are presented below:

(i) During active NEM conditions, an average east to west moisture flux of 21.1 x 10^8 metric tons per day (mtpd) over an one degree latitudinal wall (approx. 110 km) is transported across CTN.

(ii) During an excess NEM year, the mean moisture flux is 16.5 x 10^8 mtpd whereas the normal flux is 13.3 x 10^8 mtpd for OND.

(iii) The mean liquid water content over CTN for the entire season does not show much interannual variability.

(iv) During active NEM conditions, the meridional wind which is northerly in the lower levels veers to southerlies and during dry spells, the meridional winds are northerly at all levels up to mid troposphere.

13.8. Relation between NEM and ENSO/Siberian High

El Nino and Southern Oscillation (ENSO) are two inter-related global parameters displaying wide and sustained relations with several meteorological features obtained over various parts of the globe. There are numerous studies on the effects of these parameters, reference could be made to Asnani (2005) for a review. The ENSO parameters exhibit considerable influence on Indian SWM which has been extensively studied. Relation between ENSO and the Indian NEM has been studied recently in a few studies [Sridharan and Muthuchamy (1990), De and Mukhopadhyay (1999), Jayanthi and Govindachari (1999), Khole and De (2003), Raj and Geetha (2008), Geetha and Raj (2011)]. Most of the earlier studies propounded a theory of negative relation between Indian NEM and the southern oscillation index (SOI) and positive relation between the NEM and El Nino, which is opposite to the relation that ENSO exhibits with the Indian monsoon rainfall (IMR). However, in Raj and Geetha (2008) and Geetha and Raj (2011) the relation between SOI and NEM was studied much more critically and it has been found that this relation itself underwent significant intra seasonal variation and also reversed towards the end of the season.

Table 13.6a presents the CCs between SOI and North east monsoon rainfall of Tamil nadu in antecedent and concurrent modes based on data of a long 104 year period of 1901-02 to 2004-05. As shown, SOI(JJAS) is related to NRT(OND) with a CC of -0.38 (significant at 1% level); CC between SOI(OND) and NRT(OND) is -0.33 (1% LS). There are CCs indicating stronger negative relation between SOI and NRT during various periods. However, CC between SOI (Jan) and NRT(Jan) is +0.18 (10% LS). Table 13.6b, which presents the conditional means (CMs) of NRT given SOI, clearly authenticates this reversal of relationship. When SOI (JJAS) values are <-8, -8 to +8 and >+8, the CMs of PDN of NRT (OND) are 16, -1 and -11 respectively. But when SOI (Jan) lies in the above three intervals, the CMs of PDN of NRT (Jan) are -47, 4 and 37 respectively clearly showing how the relation has reversed.

This reversal of relation could be convincingly explained by invoking the concept of the transposition of the sub tropical ridge (STR) at 200 hPa level (STR200) over the Indian peninsular region and the ET at 850 hPa (ET850) during the initial and final stages of the NEM season. As shown in Fig.13.14, the STR200 and ET850 move across the southern peninsula from north to south during September to December / January. But SOI is positively related to the latitudinal position of STR throughout the year, the CCs being 0.68, 0.54, 0.66 and 0.58 (all CCs significant at 1% level) for the four seasons of JF, MAM, JJAS and OND respectively. Now, in October, when SOI is negative, STR200 is likely to be located in a southern position compared to its normal position of 16.5ºN (Fig.13.14). However, in December / January, a positive SOI could be associated with a northern position of STR200 compared to its normal position of 9.7ºN / 8.7ºN in view of the geographical location of peninsular region. Extensive application of Australian *Rainman* software (Clewett et al 2002) also confirmed this interesting and important nature of SOI-NRT relationship.

Influence of Siberian High

The Siberian High (SH) is taken as the counterpart of the Mascarene High located in the Indian Ocean when analogies between northeast and southwest monsoons are drawn. In the same way as outflow from the Mascarene High plays an important role in generating southwesterly winds over the Indian Ocean and their

transport to Arabian Sea, the cold outflow from the SH in northern hemispheric winter plays an important role in maintaining the Southeast Asian northeast monsoon (Das, 1986). Hence a relation between Indian NEM and the strength of the SH could be expected to exist conceptually. In Geetha and Raj (2009 and 2011), the relation has been studied in detail and found that the intensity of SH over the box bounded by 87.5ºE & 102.5°E longitudes and 47.5ºN & 52.5°N latitudes (Fig.13.15a) in September and October does exercise some influence over the NEM rainfall though the SH is in its formative stage only, during September-October. Table 13.7 and Fig.13.15b-c depict the type and nature of the relation. As shown, the mean sea level pressure (MSLP) over the above defined SH area during August-September and September relate to NRT of October, November and December in a modest way showing that an intense SH is conducive to good NEM over Tamil Nadu. Here again the relation reverses its sign during the fag end of the season. When the MSLP anomaly over the SH region (P_aSH) is >+1 hPa in September, PDN of NRT(ON) is 6%, but that of NRT(Dec) is -21%. In years when NRT(OND) is <-25%, the P_aSH values are negative during September to November but positive in December. When NRT(OND) >+25%, P_aSH values are positive during September-November, but strongly negative in December. Though the relation between the Indian NEM and SH is not clearly defined as that between NEM and SOI, the overall conceptual relation that an intense SH during September is a favourable factor for good Indian NEM has been shown to hold true at least in the beginning of the season. In analogy with SOI, again the relation reverses in December, the reasons being the same as those given for SOI.

It also has been shown that a strongly negative SOI is favourable for a slightly early onset of NEM, but persistence of negative SOI throughout the season brings NEM to an abrupt and early withdrawal. A positive SOI in December and January is favourable for a prolonged NEM spilling into January of next year.

Based on results derived in the above studies, the relation between SOI /SH and Indian NEM could be summarised as under:

(i) Relation between SOI of JJAS / OND and NRT of October-December of southern Indian peninsula is negative. However, the relation is stronger in the antecedent

mode and slightly weaker in the concurrent mode. This is in contrast to the existence of positive relationship between SOI and the Indian southwest monsoon rainfall during JJAS which builds up in March-May and becomes stronger in the concurrent mode.

(ii) An analysis of seasonal /monthly NRT in relation to seasonal / monthly SOI based on linear correlation and conditional means in the antecedent and concurrent modes has revealed that the good negative relationship existing between SOI and Indian northeast monsoon rainfall at the beginning of the season (October/November) does not continue throughout the season. The magnitude of negative relationship decreases with the advancement of the season. During November/December no clear relationship exists. In January, the relationship turns modestly positive.

(iii) Such a changing nature of relationship between SOI and Indian northeast monsoon rainfall is shown to be associated with the modulation in the latitudinal position of the STR over India at 200 hPa level and the surface ET.

(iv)The SOI and the latitudinal position of STR are positively correlated throughout the year which implies that a positive (negative) SOI shifts the STR/ET northwards (southwards) through out the year.

(v) The Indian northeast monsoon rainfall especially that of November is negatively correlated with the latitudinal position of the STR (Oct). But, the December rainfall is correlated positively with the STR (Dec) i.e., a southward shift of STR in October and northward shift in December are associated with good northeast monsoon rainfall activity.

(vi) During October/November when STR and ET are situated to the north of Tamil Nadu, negative SOI shifting them southwards is conducive for good rainfall activity. During November/December, when the ET is right over the Tamil Nadu latitudes, neither a northward shift nor a southward shift of STR/ET, but a normal position is favourable for enhanced monsoon activity. In December/January, when the STR/ET has moved south of Comorin, positive SOI shifting the STR/ET northwards is favourable for good northeast monsoon.

(vii) An analysis of the onset and withdrawal dates of the Indian northeast monsoon in relation to SOI has revealed that negative SOI in September leads to normal or slightly earlier onset. But, continuation of negative SOI throughout the season leads

to an early and abrupt withdrawal of the monsoon in December. Prevalence of positive SOI in September leads to a late onset.

(viii) The intensity of high pressure over the Siberian region (87.5˚E-102.5˚E, 47.5˚N-52.5˚N) is associated with intra seasonal variations of the NRT. An intense MSLP over SH region (PSH) in September is associated with a slightly early onset, a good NRT during ON and a poor NRT during December. A weak PSH (Sep) is associated with a late onset, a poor NRT(ON) and a good NRT(Dec). The relationship is better defined for NRT(ON) than for NRT(Dec).

(ix) A positive PSH profile throughout the NEM season (OND) results in a good NRT(ON) but a poor NRT(Dec) with an early and abrupt withdrawal. A negative PSH profile during OND is associated with a late onset, poor NRT(ON) and a good NRT(Dec).

(iii) An excess NRT(OND) is associated with the PSH anomaly profile remaining positive during September to November, but changing sign in December. But a deficient NRT(OND) is associated with a PSH anomaly profile that is negative during September to November but changing to positive in December.

(x) The manifestation of PSH on NRT appears to be by way of modulating the strength of low level easterlies over the BOB off the southeast coast of peninsular India as well as the latitudinal positions of STR at 200 hPa and ET at 850 hPa over India.

(xi) During December, a weaker than normal PSH(Dec) is associated with northward location of ET from its normal latitudinal position near the equator which becomes conducive for good NRT(Dec).

13.9. Seasonal forecasting of NEM rainfall

The CVs of NRS and NRT are nearly 30% and 25% respectively. The large CV is a manifestation of frequent occurrences of large excess and deficient NEM rainfall during individual years. Failure of NEM affects the state and sub division of Tamil Nadu all the more in view of the substantial dependence of this state on NEM rainfall for agricultural and hydrological sustenance. Seasonal forecasting of NEM rainfall assumes importance for the southeast peninsula especially for Tamil Nadu.

The first known attempt on seasonal forecasting of Indian NEM rainfall was made by Doraiswamy Iyer (1941). Raj (1989), carrying out an extensive search of relation between Indian upper air parameters and NRT identified a few predictors belonging to the preceeding southwest monsoon season for predicting the NEM rainfall of CAP, Rayalaseema, Tamil Nadu and Kerala The upper tropospheric (150 hPa) zonal wind over the region represented by Thiruvananthapuram during August-September, which is the immediate preceding period of the NEM season emerged as a predictor for NRT with a CC value of 0.77 based on a data sample of 17 years and verified over a test sample of 12 years. The interpretation is that a weak and disintegrating Tropical Easterly Jet Stream (TEJ) over Thiruvananthapuram during August-September favours a good NEM season during the succeeding period of October-December whereas a strong and persisting TEJ is unfavourable. Despite the fact that the high value of the CC realised from such a small sample did not get maintained, the relation held true even with a larger dataset and manifested strongly if the concept of CMs was used for analysis.

In Raj (1998a), another attempt was made to identify more predictors for prediction of NRT by including parameters of April which is a representative month for pre-monsoon season. Table 13.8 presents the list of predictors indentified and the CCs obtained. The 200 hPa zonal wind over the Indian peninsula exhibited a CC of 0.61 (1% LS) showing that strong westerly zonal winds even in April, favour a good NEM five months later and that weak westerly winds are associated with a weak NEM, based on data for the period 1965-87.

The mean temperature at 150 hPa level over Hyderabad and Port Blair during JJAS exhibited a CC value of -0.80 (1% LS) showing that a colder / warmer upper troposphere during the southwest monsoon season over the Indian region favoured good / poor NEM. The zonal wind over Thiruvananthapuram at 150 hPa during August and at 300 hPa level during September displayed significant positive CCs and this is evidently another manifestation of relation between TEJ and NEM. A statistical model combining all the predictors listed in Table 13.8 and tested in an independent sample of 7 years (1988-1994), yielded reasonably correct forecasts with a mean absolute error of 18%.

Now, strong 200 hPa westerly jet in April over India is associated with poor SWM rainfall over India whereas weak westerly jet with good SWM as shown in several studies (Asnani, 2005) Considering the relation SOI exhibits with both SWM and NEM rainfall, it could be concluded that, by and large, parameters that favour good SWM over India do not favour a good NEM and vice versa. Fig.13.16 depicts schematically the influence of April 200 hPa zonal winds, JJAS 150 hPa temperature, August-September 150 hPa zonal winds over extreme southern peninsula on NEM.

As discussed in Section 13.8, global parameters such as ENSO and Siberian High do exercise some influence over Indian NEM, but apparently through the via media of STR at 200 hPa over the Indian peninsula. As STR is transient during the NEM season, the relation NRT has with STR200 and hence with ENSO/SH also undergoes intra seasonal change.

In recent years, experimental forecasts have been generated at Regional Meteorological Centre, Chennai on NEM rainfall especially for Tamil Nadu based on the predictors which are listed in Table 13.9. Some of the predictors were added in the later years and some of them slightly redefined. The signal that each predictor provides is evaluated and a combined ensemble forecast is then derived from the aggregate of individual forecasts. The resulting forecast which is in descriptive terminology is termed as '*Outlook*' rather than as forecast. The performance of this experimental forecast system for the period 2001-2011 is presented in Table 13.10. The system yielded 4 correct, 3 partially correct and 3 incorrect forecasts as per a qualitative evaluation of forecast performance. In one year, there was no clear signal. The overall performance of the system as evident from Table 13.10 could be termed as modest only.

The Long range forecasting division of IMD at National Climate Centre, Pune has derived a forecasting system for NEM rainfall of southern peninsula which forecasts rainfall for a large part of southern peninsula and not for individual states / sub divisions (Pai, 2011).

Complete reliance on ENSO to predict NEM rainfall would not obviously work as manifested in the modest CCs and also the fact that frequently in individual years,

ENSO failed to provide proper clue to the ensuing NEM. For e.g., the substantially large excess in 2005 was not at all predicted by SOI (OND NRT 79% excess; SOI: Aug: -6.9 , Sep: +3.9, Oct: +10.9, Nov: -2.7 and Dec: 0.6). Similarly, SOI failed to predict the excess NRT in some of the years during the period 2007-2010, though in 2006, SOI gave correct indication.

To summarise the progress achieved in India on seasonal forecasting of NEM rainfall, it can be stated that though some modest success has been achieved, still considerable road has to be covered to achieve reliability needed to issue forecasts on operational basis.

13.10. Other large scale influences

Rao (1963) has shown that the location of the western portion of the sub tropical high at 500 hPa level near the Indian east coast is favourable for good NRT while the location of the eastern portion of the sub tropical high cell near the east coast of India is not conducive for good NRT.

Dhar and Rakhecha (1983) have shown that the southwest monsoon rainfall over Tamil Nadu could indicate to some extent the performance of Indian NEM. An excess or deficient southwest monsoon rainfall over Tamil Nadu is generally followed by an opposite tendency in the NRT.

Singh (1995) has studied the oceanic forcings on northeast monsoon rainfall and shown that high evaporation rate over BOB area bounded by 10-20˚N, 80-90˚E, higher sea surface temperature (SST) over western BOB and instability in the surface layer over north and adjoining central BOB (between 15-20˚N) preceding the NEM season are associated with good rainfall and low evaporation rate over 10-20˚N, 80-90˚E, lower SST over western BOB and stability in the surface layer over BOB between 15-20˚N are associated with poor rainfall.

The influence of the Indian Ocean Dipole Mode (IODM) on the NEM rainfall variability has been examined by Kripalani and Pankajkumar (2004). It has been shown that positive (negative) phase of IODM is associated with enhanced

(suppressed) NEM activity. The enhancement of NEM rainfall by the positive dipole phase has been shown to be due to anomalously warm SST in the western Indian Ocean, cold SST in the eastern Indian Ocean and the associated large-scale convergence extending towards south India. The suppression of NEM rainfall activity during the negative phase is due to the anomalously cold SST in the western Indian Ocean and warm SST in the eastern Indian Ocean and the associated divergent circulation and transport of moisture towards Sumatra, away from south peninsular India.

Balachandran et al (2006) have analysed global surface air temperature anomalies in relation to NEM rainfall and have observed the following: (i) NEM rainfall is related to North Atlantic Oscillation (NAO) index through the meridional gradient in surface air temperature between Europe and North Africa during the month of September preceding the NEM season. During a high (low) phase of NAO, the September temperature gradient between Europe and North Africa is directed from higher (lower) to lower (higher) latitudes. The study has shown that during excess NEM years, the meridional temperature gradient during September is directed from tropics to higher latitudes (sub tropics) which is associated with a low phase of NAO. (ii) The surface air temperature anomalies over Central and Eastern Equatorial Pacific oceanic region are significantly positively correlated with NRT and the temperatures over the Western Equatorial Pacific region are significantly negatively correlated with NRT. (iii) The zonal temperature gradient between Eastern Equatorial Pacific and Western Equatorial Pacific is inversely related to NRT. Difference in the area averaged surface temperature anomalies over Asian and neighbouring region (80-90˚E, 20-25˚N and 80-100˚E, 5˚S-5˚N) during the pre-monsoon month of April have also been found to be inversely related to NRT (Asokan and Balachandran (2008)).

13.11. Concluding remarks

Though a monsoon of smaller scale over Indian region and a bit disorganised, the Indian north east monsoon is characterised by several features normally associated with monsoons of larger scale. It can also be considered as a component of the south east Asian winter monsoon albeit in a restricted way. With

the availability of substantial quantum of satellite data and data generated from modern observing systems such as Doppler Weather radars especially over the Bay of Bengal sea area, there is plenty of scope for further research in Indian NEM to unravel its complexities.

Abbreviations and expansions

AS	:	Arabian Sea
BOB	:	Bay of Bengal
CAP	:	Coastal Andhra Pradesh
CC	:	Correlation coefficient
CM	:	Conditional mean
CS	;	Cyclonic storm
CSCTN	:	Central and South Coastal Tamil Nadu
CTN	:	Coastal Tamil Nadu
CV	:	Coefficient of variation
D	:	Depression
DRI	:	Daily rainfall index
ENSO	:	El Nino and Southern Oscillation
ET	:	Equatorial trough
ET850	:	Equatorial trough at 850 hPa level
EW	:	Easterly wave
FW	:	Fairly widespread
IDWR	:	Indian Daily Weather Report
IITM	:	Indian Institute of Tropical Meteorology
IMD	:	India Meteorological Department
IMR	:	Indian monsoon rainfall
INSAT	:	Indian satellite
IODM	:	Indian Ocean Dipole Mode
IST	:	Indian Standard Time
ITN	:	Interior Tamil Nadu
JF	:	January-February
JJAS	:	June-July-August-September
KER	:	Kerala
LOPAR	:	Low pressure area

LPA	:	Long Period Average
LPS	:	Low pressure system
LS	:	Level of significance
MAM	:	March-April-May
MSLP	:	Mean sea level pressure
NAO	:	North Atlantic Oscillation
NCAP	:	North Coastal Andhra Pradesh
NCTN	:	North Coastal Tamil Nadu
ND	:	November-December
NDC	:	National Data Centre
NEM	:	Northeast monsoon
NRS	:	Northeast monsoon rainfall of Southern Region
NRT	:	Northeast monsoon rainfall of Tamil Nadu
OLR	:	Outgoing Longwave radiation
ON	:	October -November
OND	:	October-November-December
P_aSH	:	Mean sea level pressure anomaly over Siberian High region
PDN	:	Percentage departure from normal
PSH	:	Mean sea level pressure over Siberian High region
RYS	:	Rayalaseema
SCAP	:	South Coastal Andhra Pradesh
SCS	:	Severe Cyclonic Storm
SD	:	Standard deviation
SH	:	Siberian High
SIK	:	South Interior Karnataka
SOI	:	Southern Oscillation Index
SR	:	Southern Region
SST	:	Sea surface temperature
STR	:	Sub tropical ridge
STR200	:	Sub tropical ridge at 200 hPa level
SWM	:	Southwest monsoon

TEJ	:	Tropical easterly jet
TN	:	Tamil Nadu
UTC	:	Coordinated Universal Time

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Table 13.1a: Normal monthly, seasonal and annual rainfall of five meteorological sub-divisions of India benefited by northeast monsoon

Month / Season	Rainfall in mm					Rainfall as % of annual rainfall				
	CAP	RYS	SIK	KER	TN	CAP	RYS	SIK	KER	TN
Jan	8.3	3.0	1.4	8.5	17.5	0.8	0.4	0.1	0.3	1.9
Feb	10.4	3.6	3.0	14.9	13.4	1.0	0.5	0.3	0.5	1.5
Mar	11.1	6.5	8.5	30.1	18.3	1.1	0.9	0.8	1.0	2.0
Apr	21.8	19.9	43.8	108.5	42.3	2.1	2.8	4.3	3.7	4.6
May	64.1	55.6	92.9	238.1	67.5	6.3	7.9	9.1	8.1	7.4
Jun	103.9	67.7	141.5	652.2	46.0	10.1	9.6	13.9	22.3	5.0
Jul	160.4	94.2	216.1	733.6	68.0	15.7	13.3	21.2	25.1	7.4
Aug	157.7	103.3	161.4	422.0	87.4	15.4	14.6	15.8	14.4	9.6
Sep	159.1	133.1	141.0	242.9	115.8	15.5	18.9	13.8	8.3	12.7
Oct	193.2	129.4	147.7	290.9	180.2	18.9	18.3	14.5	9.9	19.7
Nov	106.6	66.1	49.2	149.5	170.0	10.4	9.4	4.8	5.1	18.6
Dec	27.6	23.7	12.7	37.1	88.0	2.7	3.4	1.2	1.3	9.6
JF	18.7	6.6	4.4	23.4	30.9	1.8	0.9	0.4	0.8	3.4
MAM	97.0	82.0	145.2	376.7	128.1	9.5	11.6	14.2	12.9	14.0
JJAS	581.1	398.3	660.0	2050.7	317.2	56.7	56.4	64.8	70.0	34.7
OND	327.4	219.2	209.6	477.5	438.2	32.0	31.0	20.6	16.3	47.9
Annual	1024.2	706.1	1019.2	2928.3	914.4					

Based on 1951-2000 data . For abbreviations refer consolidated list

Table 13.1b: Normal monthly seasonal and annual rainfall (in mm) of selected stations of the five meteorological sub divisions benefited by northeast monsoon

Month / Season	Station and region/sub region represented													
	VSK	MPT	NLR	ANT	TPT	BNG	CHN	TRV	SLM	CMB	TRP	MDS	VRM	TTC
	NCAP	SCAP	SCAP	RYS	RYS	SIK	KER central	KER south	ITN	ITN	ITN	CTN	CTN	CTN south
Jan	11.0	4.2	13.8	0.9	8.8	2.7	24.4	22.9	5	8.3	17.0	22.6	63.5	17.5
Feb	11.0	5.7	9.8	1.0	7.0	7.1	33.1	24.8	6.3	7.8	7.5	15.8	28.2	20.3
Mar	11.2	6.6	3.9	4.7	5.5	9.4	45.5	33.8	13.2	14.6	8.8	3.3	21.4	33
Apr	25.0	12.5	12.3	17.1	24.4	44.0	126.1	115.6	44.7	47.5	43.8	12.2	25.9	60.2
May	55.7	27.1	40.1	53.4	51.8	115.9	347.2	211.5	105.7	73.0	62.6	49.5	34.8	24.4
Jun	110.5	109.2	47.7	58.5	60.6	87.0	717	331.5	82	31.7	35.5	72.6	34.6	3.5
Jul	135.1	190.6	80.1	68.7	103.8	109.5	666.3	213.1	112.2	42.7	63.7	100.9	52.7	10.5
Aug	150.0	176.5	93.3	73.9	109.2	141.6	381.7	141	135.3	26.3	76.8	143.0	93.7	6.4
Sep	181.5	168.2	102.0	134.8	113.7	211.5	309	170.1	186.6	53.0	163.5	126.1	83.9	14.8
Oct	219.1	207.7	260.0	107.8	195.0	170.6	291.5	270.1	194.7	146.6	165.2	274.1	254.6	149.3
Nov	75.5	124.2	295.1	34.1	183.0	57.2	147.1	192	96.6	116.5	134.1	419.7	436.3	198.3
Dec	7.5	23.3	104.4	11.7	90.9	20.9	58.3	65.6	40.8	45.8	93.8	162.9	347.4	92.3
JF	22.0	9.9	23.6	1.9	15.8	9.8	57.5	47.7	11.3	16.1	24.5	38.4	91.7	37.8
MAM	91.9	46.2	56.3	75.2	81.7	169.3	518.8	360.9	163.6	135.1	115.2	65.0	82.1	117.6
JJAS	577.1	644.5	323.1	335.9	387.3	549.6	2074.0	855.7	516.1	153.7	339.5	442.6	264.9	35.2
OND	302.1	355.2	659.5	153.6	468.9	248.7	496.9	527.7	332.1	308.9	393.1	856.7	1038.3	439.9
Annual	993.1	1055.8	1062.5	566.6	953.7	977.4	3147.2	1792.0	1023.1	613.8	872.3	1402.7	1477.0	630.5
Oct-I	111	89	74	65	65	105	141	111	106	54	74	83	88	39
Oct-II	108	118	186	42	130	66	150	159	89	93	91	191	167	110

Based on 1951-2000 data; Oct-I: 1-15 October; Oct-II: 16-31 October

Table 13.2a: Time series of NRS and NRT, 1901-2010

Year	NRS		NRT		Year	NRS		NRT	
	ACT (mm)	PDN (%)	ACT (mm)	PDN (%)		ACT (mm)	PDN (%)	ACT (mm)	PDN (%)
1901	346.7	5	420.7	-4	**1931**	404.5	17	596.0	36
1902	452.5	37	631.0	44	**1932**	419.4	25	574.0	31
1903	439.8	40	521.5	19	**1933**	353.1	10	407.5	-7
1904	193.8	-42	236.6	-46	**1934**	280.4	-15	390.0	-11
1905	244.2	-28	376.9	-14	**1935**	293.3	-14	438.2	0
1906	337.3	2	495.2	13	**1936**	308.9	-7	385.6	-12
1907	269.2	-24	420.7	-4	**1937**	326.3	-3	482.0	10
1908	188.3	-47	315.5	-28	**1938**	154.6	-57	192.8	-56
1909	163.5	-52	227.9	-48	**1939**	442.0	35	508.3	16
1910	367.1	15	390.0	-11	**1940**	427.9	31	609.1	39
1911	316.9	-7	442.6	1	**1941**	356.6	6	473.3	8
1912	361.4	9	512.7	17	**1942**	272.2	-21	359.3	-18
1913	329.0	-7	534.6	22	**1943**	400.2	25	473.3	8
1914	298.7	-16	547.8	25	**1944**	420.6	26	587.2	34
1915	341.3	5	376.9	-14	**1945**	267.0	-25	390.0	-11
1916	385.1	25	363.7	-17	**1946**	507.9	53	727.4	66
1917	334.8	5	346.2	-21	**1947**	212.8	-34	232.2	-47
1918	333.3	-1	486.4	11	**1948**	300.7	-9	394.4	-10
1919	387.6	14	534.6	22	**1949**	231.9	-27	236.6	-46
1920	356.9	-2	609.1	39	**1950**	246.4	-25	302.4	-31
1921	294.4	-8	372.5	-15	**1951**	222.0	-36	302.4	-31
1922	420.9	28	587.2	34	**1952**	289.2	-8	324.3	-26
1923	279.7	-23	407.5	-7	**1953**	348.1	13	394.4	-10
1924	264.8	-21	311.1	-29	**1954**	309.9	-5	411.9	-6
1925	414.7	21	591.6	35	**1955**	355.0	7	433.8	-1
1926	177.8	-49	289.2	-34	**1956**	439.9	41	495.2	13
1927	219.4	-34	280.4	-36	**1957**	300.8	-9	455.7	4
1928	365.6	10	495.2	13	**1958**	344.8	9	315.5	-28
1929	315.7	-7	455.7	4	**1959**	246.1	-30	398.8	-9
1930	495.6	49	644.2	47	**1960**	333.0	-2	503.9	15

TABLE 13.2a ….(contd.)

Year	NRS		NRT		Year	NRS		NRT	
	ACT (mm)	PDN (%)	ACT (mm)	PDN (%)		ACT (mm)	PDN (%)	ACT (mm)	PDN (%)
1961	265.8	-16	289.2	-34	**1986**	267.2	-19	324.3	-26
1962	371.9	19	385.6	-12	**1987**	473.8	49	464.5	6
1963	346.0	4	495.2	13	**1988**	166.9	-51	271.7	-38
1964	277.6	-15	368.1	-16	**1989**	228.1	-34	359.3	-18
1965	245.8	-34	464.5	6	**1990**	332.9	3	385.6	-12
1966	410.8	25	565.3	29	**1991**	380.0	22	433.8	-1
1967	250.4	-28	460.1	5	**1992**	354.9	7	411.9	-6
1968	251.0	-23	306.7	-30	**1993**	485.4	48	666.1	52
1969	443.3	36	578.4	32	**1994**	436.0	38	442.6	1
1970	260.7	-19	416.3	-5	**1995**	255.5	-20	236.6	-46
1971	285.4	-15	464.5	6	**1996**	411.1	25	525.8	20
1972	442.1	33	587.2	34	**1997**	460.5	38	666.1	52
1973	303.1	-8	411.9	-6	**1998**	409.2	23	490.8	12
1974	205.2	-32	175.3	-60	**1999**	337.9	3	411.9	-6
1975	358.8	16	337.4	-23	**2000**	238.3	-27	341.9	-22
1976	369.0	10	438.2	0	**2001**	345.2	4	341.8	-22
1977	502.3	51	683.6	56	**2002**	299.1	-10	403.1	-8
1978	370.1	9	582.8	33	**2003**	340.4	3	407.5	-7
1979	377.9	14	578.4	32	**2004**	264.1	-20	442.6	1
1980	264.0	-21	333.0	-24	**2005**	550.0	66	784.4	79
1981	279.6	-17	403.1	-8	**2006**	322.1	-3	503.9	15
1982	265.7	-20	359.3	-18	**2007**	337.1	2	525.8	20
1983	303.8	-11	490.8	12	**2008**	334.2	1	574.0	31
1984	204.9	-39	271.7	-38	**2009**	330.4	0	490.8	12
1985	308.1	-10	447.0	2	**2010**	512.0	55	617.9	41

NRS: OND northeast monsoon rainfall of Southern Region;
NRT: OND northeast monsoon rainfall of Tamil Nadu;
ACT : Actual, PDN : Percentage departure from normal
Seasonal PDN data taken from NDC dataset for the period 1901-2000 and extracted from *Mausam* for the period 2001-2010.

Table 13.2b: Statistical parameters of OND northeast monsoon rainfall of five sub divisions and the SR, 1901-2010

	CAP	RYL	SIK	KER	TN	SR
LPA (mm)	330.5	217.0	207.5	469.6	440.4	330.7
SD (mm)	125.0	86.1	74.2	128.1	120.3	83.8
CV (%)	37.8	39.7	35.8	27.3	27.3	25.3
Range (mm)	75.3(1970) to 638.4(1987)	26.3(1938) to 436.2(2005)	54.5(1923) to 383.6(1956)	181.5(1974) to 797.4(2010)	175.3(1974) to 784.4(2005)	154.6(1938) to 541.7(2005)

SR: Southern Region;
LPA: Long period average
SD: Standard Deviation; CV: Coefficient of Variation

Table 13.2c: List of excess (PDN ≥20%) and deficient (PDN ≤-20%) years of NRS and NRT during the period 1901-2010

NRS				NRT			
Excess		Deficient		Excess		Deficient	
Year	PDN	Year	PDN	Year	PDN	Year	PDN
1902	37	**1904**	-42	**1902**	44	**1904**	-46
1903	40	**1905**	-28	**1913**	22	**1908**	-28
1916	25	**1907**	-24	**1914**	25	**1909**	-48
1922	28	**1908**	-47	**1919**	22	**1917**	-21
1925	21	**1909**	-52	**1920**	39	**1924**	-29
1930	49	**1923**	-23	**1922**	34	**1926**	-34
1932	25	**1924**	-21	**1925**	35	**1927**	-36
1939	35	**1926**	-49	**1930**	47	**1938**	-56
1940	31	**1927**	-34	**1931**	36	**1947**	-47
1943	25	**1938**	-57	**1932**	31	**1949**	-46
1944	26	**1942**	-21	**1940**	39	**1950**	-31
1946	53	**1945**	-25	**1944**	34	**1951**	-31
1956	41	**1947**	-34	**1946**	66	**1952**	-26
1966	25	**1949**	-27	**1966**	29	**1958**	-28
1969	36	**1950**	-25	**1969**	32	**1961**	-34
1972	33	**1951**	-36	**1972**	34	**1968**	-30
1977	51	**1959**	-30	**1977**	56	**1974**	-60
1987	49	**1965**	-34	**1978**	33	**1975**	-23
1991	22	**1967**	-28	**1979**	32	**1980**	-24
1993	48	**1968**	-23	**1993**	52	**1984**	-38
1994	38	**1974**	-32	**1996**	20	**1986**	-26
1996	25	**1980**	-21	**1997**	52	**1988**	-38
1997	38	**1984**	-39	**2005**	79	**1995**	-46
1998	23	**1988**	-51	**2007**	21	**2000**	-22
2005	66	**1989**	-34	**2008**	31	**2003**	-25
2010	55	**2000**	-27	**2010**	42		
		2004	-20				
Total frequency	**26**		**27**		**26**		**25**

NRS, NRT & PDN: As in Table 13.2a

Table 13.2d: Decade wise Frequency (No. of years) of excess and deficient years of NRS and NRT, 1901-2010

Decadal period	NRS		NRT	
	Excess	Deficient	Excess	Deficient
1901-1910	2	5	1	1
1911-1920	1	0	4	3
1921-1930	3	4	3	3
1931-1940	3	1	3	1
1941-1950	3	5	2	3
1951-1960	1	2	0	3
1961-1970	2	3	2	2
1971-1980	2	2	4	3
1981-1990	1	3	0	3
1991-2000	6	1	3	2
2001-2010	2	1	4	1

NRS & NRT : As in Table 13.2a

Table 13.3a: Dates of onset and withdrawal of northeast monsoon over CTN for the period 1901-2000 as declared by IMD and as determined in diagnostic study-I

Year	Date of Easterly onset	Date of NEM onset		Date of NEM withdrawal	
		IMD	Onset-I	IMD	WD-I
1901-02	24-Oct		27-Oct		9-Jan
1902-03	20-Oct		22-Oct		7-Jan
1903-04	22-Oct		24-Oct		14-Jan
1904-05	15-Oct		21-Oct		23-Dec
1905-06	4-Oct		6-Oct		30-Nov
1906-07	12-Oct		16-Oct		16-Dec
1907-08	19-Oct		20-Oct		22-Dec
1908-09	11-Oct		22-Oct		7-Jan
1909-10	10-Oct		14-Oct		22-Dec
1910-11	15-Oct		21-Oct		26-Nov
1911-12	14-Oct		16-Oct		28-Dec
1912-13	11-Oct		12-Oct		3-Dec
1913-14	18-Oct		24-Oct		23-Dec
1914-15	2-Oct		10-Oct		9-Dec
1915-16	25-Oct		11-Nov		30-Dec
1916-17	12-Oct		15-Oct		10-Dec
1917-18	13-Oct		18-Oct		6-Jan
1918-19	23-Oct		1-Nov		20-Dec
1919-20	14-Oct		18-Oct		10-Jan
1920-21	17-Oct		21-Oct		15-Jan
1921-22	18-Oct		22-Oct		12-Jan
1922-23	11-Oct		11-Oct		17-Jan
1923-24	15-Oct	15-Oct	17-Oct		10-Jan
1924-25	20-Oct	22-Oct	22-Oct		21-Dec
1925-26	19-Oct		23-Oct		15-Dec
1926-27	26-Oct		30-Oct		18-Dec
1927-28	21-Oct	22-Oct	26-Oct		9-Jan
1928-29	13-Oct		16-Oct		31-Dec
1929-30	21-Oct	25-Oct	26-Oct		11-Jan
1930-31	7-Oct		9-Oct		19-Jan
1931-32	27-Oct	29-Oct	29-Oct		28-Dec
1932-33	5-Oct		9-Oct		19-Dec
1933-34	26-Oct		27-Oct		28-Jan
1934-35	17-Oct	20-Oct	24-Oct		18-Dec
1935-36	9-Oct		13-Oct		28-Dec

Year	Date of Easterly onset	Date of NEM onset		Date of NEM withdrawal	
		IMD	Onset-I	IMD	WD-I
1936-37	23-Sep	21-Oct	22-Oct		22-Dec
1937-38	17-Oct		27-Oct		19-Dec
1938-39	20-Oct	26-Oct	27-Oct		9-Dec
1939-40	11-Oct	1-Nov	14-Oct		17-Dec
1940-41	23-Oct	28-Oct	27-Oct		9-Jan
1941-42	17-Oct	22-Oct	21-Oct		19-Dec
1942-43	22-Oct		1-Nov		5-Jan
1943-44	2-Oct		5-Oct		13-Jan
1944-45	19-Oct		20-Oct		24-Dec
1945-46	7-Oct		12-Oct		10-Dec
1946-47	8-Oct		16-Oct		8-Jan
1947-48	15-Oct		19-Oct		4-Jan
1948-49	4-Oct	16-Oct	16-Oct		3-Jan
1949-50	15-Oct		20-Oct		1-Dec
1950-51	12-Oct	16-Oct	15-Oct		16-Dec
1951-52	19-Oct	27-Oct	6-Nov		2-Dec
1952-53	1-Oct		15-Oct		15-Dec
1953-54	6-Oct		14-Oct		13-Jan
1954-55	11-Oct	11-Oct	15-Oct		11-Jan
1955-56	13-Oct	6-Oct	20-Oct		10-Dec
1956-57	10-Oct	25-Oct	26-Oct		28-Dec
1957-58	9-Oct		16-Oct		3-Jan
1958-59	25-Oct		29-Oct		19-Dec
1959-60	14-Oct		21-Oct		4-Dec
1960-61	13-Oct		21-Oct		18-Jan
1961-62	20-Oct		24-Oct		15-Dec
1962-63	1-Oct		9-Oct		13-Jan
1963-64	13-Oct		20-Oct		10-Dec
1964-65	12-Oct		29-Oct		27-Dec
1965-66	4-Oct		15-Oct		12-Dec
1966-67	2-Oct		6-Oct		22-Dec
1967-68	4-Oct		15-Oct		20-Dec
1968-69	12-Oct		15-Oct		21-Dec
1969-70	11-Oct		14-Oct		3-Jan
1970-71	10-Oct		14-Oct		12-Jan

Year	Date of Easterly onset	Date of NEM onset		Date of NEM withdrawal	
		IMD	Onset-I	IMD	WD-I
1971-72	17-Oct		19-Oct		19-Dec
1972-73	9-Oct		21-Oct		27-Dec
1973-74	14-Oct		17-Oct		31-Dec
1974-75	28-Oct		5-Nov		30-Dec
1975-76	25-Oct		25-Oct		20-Dec
1976-77	1-Oct		15-Oct		26-Dec
1977-78	5-Oct	14-Oct	10-Oct		6-Dec
1978-79	20-Oct	23-Oct	21-Oct		1-Jan
1979-80	6-Oct	22-Oct	22-Oct		15-Dec
1980-81	3-Oct	11-Oct	10-Oct		6-Dec
1981-82	6-Oct	23-Oct	23-Oct		24-Dec
1982-83	2-Oct	19-Oct	18-Oct		19-Dec
1983-84	17-Oct	24-Oct	24-Oct		18-Jan
1984-85	24-Oct	5-Oct	3-Nov		7-Jan
1985-86	19-Oct	25-Oct	25-Oct		17-Jan
1986-87	11-Oct	27-Oct	26-Oct		30-Dec
1987-88	19-Oct	30-Oct	20-Oct		26-Dec
1988-89	21-Oct	2-Nov	3-Nov		6-Jan
1989-90	20-Oct	27-Oct	29-Oct		10-Jan
1990-91	17-Oct	19-Oct	18-Oct		27-Dec
1991-92	15-Oct	20-Oct	19-Oct		23-Dec
1992-93	23-Oct	2-Nov	2-Nov		9-Dec
1993-94	12-Oct	20-Oct	13-Oct	11-Jan	1-Jan
1994-95	17-Oct	18-Oct	18-Oct	14-Dec	24-Dec
1995-96	22-Oct	23-Oct	23-Oct	22-Dec	13-Dec
1996-97	9-Oct	11-Oct	10-Oct	23-Dec	19-Dec
1997-98	5-Oct	13-Oct	13-Oct	31-Dec	23-Dec
1998-99	22-Oct	28-Oct	28-Oct	22-Dec	4-Jan
1999-2000	1-Oct	21-Oct	4-Oct	28-Dec	12-Jan
2000-01	1-Nov	2-Nov	5-Nov	6-Jan	2-Jan
2001-02	15-Oct	16-Oct		11-Jan	
2002-03	1-Oct	25-Oct		27-Dec	
2003-04	13-Oct	19-Oct		8-Jan	
2004-05	13-Oct	18-Oct		25-Jan	
2005-06	6-Oct	12-Oct		18-Jan	
2006-07	15-Oct	19-Oct		28-Dec	
2007-08	14-Oct	22-Oct		10-Jan	
2008-09	12-Oct	15-Oct		31-Dec	
2009-10	15-Oct	29-Oct		18-Jan	
2010-11	26-Oct	29-Oct		17-Jan	

NEM: Northeast monsoon, CTN: Coastal Tamil Nadu

Table 13.3b: Dates of onset and withdrawal of NEM over SCAP and CTN as determined in diagnostic study -II

Year	Date of NEM onset-II	Date of NEM WD-II	Year	Date of NEM onset-II	Date of NEM WD-II
1870-71	17-Oct	30-Dec			
1871-72	16-Oct	16-Dec	**1901-02**	27-Oct	9-Jan
1872-73	10-Oct	23-Dec	**1902-03**	22-Oct	8-Jan
1873-74	4-Oct	5-Dec	**1903-04**	24-Oct	4-Jan
1874-75	20-Oct	16-Dec	**1904-05**	19-Oct	23-Dec
1875-76	16-Oct	11-Dec	**1905-06**	5-Oct	20-Jan
1876-77	8-Nov	28-Dec	**1906-07**	17-Oct	31-Dec
1877-78	22-Oct	19-Dec	**1907-08**	20-Oct	22-Dec
1878-79	1-Nov	20-Dec	**1908-09**	22-Oct	8-Jan
1879-80	18-Oct	24-Dec	**1909-10**	15-Oct	2-Dec
1880-81	18-Oct	27-Dec	**1910-11**	16-Oct	27-Nov
1881-82	1-Nov	5-Jan	**1911-12**	15-Oct	28-Dec
1882-83	11-Oct	30-Dec	**1912-13**	12-Oct	6-Dec
1883-84	12-Oct	26-Dec	**1913-14**	24-Oct	22-Dec
1884-85	14-Oct	25-Dec	**1914-15**	10-Oct	23-Jan
1885-86	28-Oct	8-Jan	**1915-16**	11-Nov	30-Dec
1886-87	28-Oct	11-Dec	**1916-17**	15-Oct	10-Dec
1887-88	8-Oct	1-Jan	**1917-18**	17-Oct	19-Jan
1888-89	15-Oct	20-Dec	**1918-19**	1-Nov	20-Dec
1889-90	13-Oct	21-Dec	**1919-20**	14-Oct	10-Jan
1890-91	26-Oct	8-Dec	**1920-21**	22-Oct	25-Jan
1891-92	12-Oct	30-Dec	**1921-22**	22-Oct	12-Jan
1892-93	13-Oct	26-Dec	**1922-23**	12-Oct	17-Jan
1893-94	11-Oct	20-Dec	**1923-24**	16-Oct	10-Jan
1894-95	17-Oct	26-Dec	**1924-25**	23-Oct	21-Dec
1895-96	11-Oct	1-Jan	**1925-26**	23-Oct	30-Jan
1896-97	26-Oct	29-Dec	**1926-27**	30-Oct	18-Dec
1897-98	13-Oct	21-Dec	**1927-28**	27-Oct	9-Jan
1898-99	22-Oct	1-Jan	**1928-29**	16-Oct	31-Dec
1899-1900	19-Oct	3-Dec	**1929-30**	26-Oct	11-Jan
1900-01	17-Oct	26-Dec	**1930-31**	9-Oct	19-Jan

Year	Date of NEM onset-II	Date of NEM WD-II	Year	Date of NEM onset-II	Date of NEM WD-II
1931-32	29-Oct	28-Dec	**1971-72**	17-Oct	19-Dec
1932-33	10-Oct	19-Dec	**1972-73**	21-Oct	28-Dec
1933-34	27-Oct	28-Jan	**1973-74**	18-Oct	31-Dec
1934-35	24-Oct	24-Jan	**1974-75**	6-Nov	27-Dec
1935-36	13-Oct	28-Dec	**1975-76**	25-Oct	20-Dec
1936-37	22-Oct	22-Dec	**1976-77**	15-Oct	25-Dec
1937-38	24-Oct	19-Dec	**1977-78**	7-Oct	12-Dec
1938-39	26-Oct	14-Jan	**1978-79**	20-Oct	1-Jan
1939-40	11-Oct	17-Dec	**1979-80**	22-Oct	15-Dec
1940-41	27-Oct	9-Jan	**1980-81**	11-Oct	20-Dec
1941-42	19-Oct	19-Dec	**1981-82**	23-Oct	24-Dec
1942-43	1-Nov	5-Jan	**1982-83**	18-Oct	18-Dec
1943-44	5-Oct	13-Jan	**1983-84**	24-Oct	20-Jan
1944-45	21-Oct	24-Dec	**1984-85**	4-Nov	7-Jan
1945-46	13-Oct	10-Dec	**1985-86**	25-Oct	17-Jan
1946-47	16-Oct	8-Jan	**1986-87**	26-Oct	30-Dec
1947-48	19-Oct	19-Jan	**1987-88**	20-Oct	26-Dec
1948-49	16-Oct	22-Dec	**1988-89**	3-Nov	20-Dec
1949-50	18-Oct	1-Dec	**1989-90**	27-Oct	10-Jan
1950-51	15-Oct	16-Dec	**1990-91**	19-Oct	20-Jan
1951-52	6-Nov	2-Dec	**1991-92**	19-Oct	24-Dec
1952-53	4-Oct	15-Dec	**1992-93**	3-Nov	21-Dec
1953-54	15-Oct	14-Jan	**1993-94**	13-Oct	1-Jan
1954-55	12-Oct	11-Jan	**1994-95**	18-Oct	22-Jan
1955-56	19-Oct	10-Dec	**1995-96**	23-Oct	13-Dec
1956-57	26-Oct	29-Dec	**1996-97**	9-Oct	19-Dec
1957-58	15-Oct	4-Jan	**1997-98**	13-Oct	23-Dec
1958-59	30-Oct	20-Dec	**1998-99**	28-Oct	3-Jan
1959-60	21-Oct	22-Dec	**1999-2000**	4-Oct	12-Jan
1960-61	22-Oct	18-Jan	**2000-01**	3-Nov	2-Jan
1961-62	24-Oct	15-Dec	**2001-02**	15-Oct	1-Jan
1962-63	9-Oct	13-Jan	**2002-03**	9-Oct	12-Dec
1963-64	20-Oct	13-Dec	**2003-04**	19-Oct	8-Dec
1964-65	31-Oct	28-Dec	**2004-05**	18-Oct	16-Dec
1965-66	15-Oct	17-Dec	**2005-06**	11-Oct	21-Dec
1966-67	4-Oct	23-Dec	**2006-07**	17-Oct	14-Dec
1967-68	15-Oct	20-Dec	**2007-08**	19-Oct	7-Jan
1968-69	13-Oct	22-Dec	**2008-09**	12-Oct	21-Dec
1969-70	15-Oct	3-Jan	**2009-10**	29-Oct	26-Dec
1970-71	10-Oct	30-Dec	**2010-11**	29-Oct	6-Jan

NEM, CTN: as in TABLE 13.3a SCAP: South coastal AP

Table 13.3c:Statistical parameters of dates of onset and withdrawal of northeast monsoon for the periods 1871-1900, 1901-2000 and 1901-2010

Period	Parameter	Mean date		Standard deviation (days)	
		I	II	I	II
1871-1900	Onset	-	17-Oct	-	7
	Withdrawal	-	23-Dec	-	9
	Duration (days)	-	67	-	11
1901-2000	Easterly onset	14-Oct	--	8	-
	Onset	20-Oct	20-Oct	8	8
	Withdrawal	27-Dec	30-Dec	13	14
	Duration	67	72	15	16
1901-2010	Easterly onset	-	14-Oct	-	7
	Onset	-	20-Oct	-	8
	Withdrawal	-	29-Dec	-	14
	Duration	-	70	-	16

NEM: as in Table 13.3a; SD: Standard Deviation
I: Diagnostic study I for CTN. II: Diagnostic study II for CTN& SCAP

Table 13.4: Dates of withdrawal of northeast monsoon determined separately for two coastal belts of SCAP & NCTN and CSCTN

Year	Revised withdrawal date for SCAP& CTN	Withdrawal date over SCAP& NCTN	Withdrawal date over CSCTN
1961	15-Dec	15-Nov	15-Dec
1962	13-Jan	12-Jan	13-Jan
1963	13-Dec	11-Dec	26-Dec
1964	28-Dec	24-Nov	28-Dec
1965	17-Dec	18-Dec	29-Dec
1966	23-Dec	8-Dec	19-Jan
1967	20-Dec	16-Dec	20-Dec
1968	22-Dec	21-Dec	22-Dec
1969	3-Jan	20-Dec	3-Jan
1970	30-Dec	1-Dec	13-Jan
1971	19-Dec	18-Dec	19-Dec
1972	28-Dec	15-Dec	28-Dec
1973	31-Dec	31-Dec	31-Dec
1974	27-Dec	25-Nov	30-Dec
1975	20-Dec	29-Nov	20-Dec
1976	25-Dec	12-Dec	26-Dec
1977	12-Dec	29-Nov	12-Dec
1978	1-Jan	1-Jan	1-Jan
1979	15-Dec	14-Dec	27-Dec
1980	20-Dec	14-Dec	20-Dec
1981	24-Dec	5-Dec	24-Dec
1982	18-Dec	6-Dec	18-Dec
1983	20-Jan	27-Dec	31-Jan
1984	7-Jan	6-Jan	19-Jan
1985	17-Jan	17-Jan	17-Jan
1986	30-Dec	10-Dec	30-Dec
1987	26-Dec	26-Dec	26-Dec
1988	20-Dec	17-Dec	7-Jan
1989	10-Jan	5-Dec	10-Jan
1990	20-Jan	2-Jan	20-Jan

Year	Revised withdrawal date for SCAP& CTN	Withdrawal date over SCAP& NCTN	Withdrawal date over CSCTN
1991	24-Dec	23-Nov	25-Dec
1992	21-Dec	7-Dec	21-Dec
1993	1-Jan	24-Dec	2-Jan
1994	22-Jan	19-Jan	22-Jan
1995	13-Dec	29-Nov	17-Jan
1996	19-Dec	18-Dec	19-Dec
1997	23-Dec	21-Dec	3-Jan
1998	3-Jan	14-Dec	4-Jan
1999	12-Jan	26-Dec	13-Jan
2000	2-Jan	2-Jan	2-Jan
2001	1-Jan	11-Jan	1-Jan
2002	12-Dec	10-Dec	12-Dec
2003	8-Dec	29-Dec	8-Dec
2004	16-Dec	29-Nov	16-Dec
2005	21-Dec	21-Dec	16-Jan
2006	14-Dec	14-Dec	15-Dec
2007	7-Jan	6-Jan	7-Jan
2008	21-Dec	11-Dec	21-Dec
2009	26-Dec	23-Dec	15-Jan
2010	6-Jan	22-Dec	6-Jan
Mean		17-Dec	31-Dec
SD		15 days	13 days

SCAP: South Coastal Andhra Pradesh
NCTN: North Coastal Tamil Nadu (North of 12˚N);
CSCTN: Central and South Coastal Tamil Nadu (South of 12˚N)
SD: as in TABLE 13.3c

Table 13.5: Various categories of LPS , associated MWS and T.Number

LPS (type)	Abbreviation	MWS (knots)	T.Number
Low Pressure Area	LOPAR	< 17	1.0
Depression	D	17-27	1.5
Deep Depression	DD	28-33	2.0
Cyclonic Storm	CS	34-47	3.0
Severe Cyclonic Storm	SCS	48-63	3.5, 4.0
Very Severe Cyclonic Storm	VSCS	64-119	4.0-6.0
Super Cyclonic Storm	SuCS	≥120	>6.0

LPS- Low pressure system MWS- Maximum wind speed at msl
(applicable only when the LPS is over sea)

Table 13.6a: CCs between SOI and NRT in antecedent and concurrent modes

SOI (period)	NRT (period)	CC
JJAS	**OND**	-0.38**
Oct	**Oct**	0.00
Nov	**Nov**	-0.21*
Dec	**Dec**	-0.03
ND	**ND**	-0.24*
OND	**OND**	-0.33**
JF	**JF**	0.19#
Oct	**OND**	-0.42 **
Oct	**ND**	-0.47 **
Oct	**Nov**	-0.44**
Oct	**Dec**	-0.18#
Jan	**Jan**	0.18#

(based on 104 yr data from1901-02 to 2004-05)
SOI: Southern Oscillation Index ; NRT: As in Table 13.2a
CC: Correlation coefficient;
** /*/# : CC significant at 1/5/10% level

Table 13.6b: *Conditional means of NRT for various SOI intervals in* antecedent and concurrent modes

SOI	NRT	CC	Conditional means of NRT(PDN) for SOI intervals		
Month /season	**Month /season**		**<-8**	**-8 to 8**	**>8**
JJAS	**OND**	-0.38**	16	-1	-11
Oct	**Oct**	0.00	2	-1	0
Nov	**Nov**	-0.21*	10	1	-13
Dec	**Dec**	-0.03	-9	7	-9
OND	**OND**	-0.33**	14	-1	-9
Oct	**OND**	-0.42**	20	-5	-11
Oct	**ND**	-0.47**	32	-7	-19
Jan	**Jan**	0.18#	-47	4	37

Abbreviations: As in Table 13.2a, 13.6a

Table 13.7: CCs between SH and NRT in antecedent and concurrent modes

Month of PSH	Month of NRT	CC
Antecedent SH		
AS	Oct	0.02
	Nov	0.48**
	ON	0.36*
Sep	Oct	0.13
	Nov	0.36*
	ON	0.34*
	Dec	-0.30
Concurrent SH		
Oct	Oct	0.13
Nov	Nov	0.25
ON	ON	0.28
Dec	Dec	-0.21
Dec	ND	-0.35*

(based on 34 year data from 1971 to 2004)
PSH: MSLP over the Siberian High region
AS: Aug-Sep; ON:Oct-Nov; ND:Nov-Dec
Other abbreviations: As in Table 13.6a

Table 13.8: Predictors identified for NRT

S.No	Predictor	Mean	SD	CC	*n*
1	April 200 hPa zonal wind – BMB, MDS, MNC, TRV and HYD	9.2 m/s	3.2 m/s	0.61**	23
2	April, 150 hPa meridional wind – BMB and AHM	2.7 m/s	3.9 m/s	-0.50*	23
3	June-September, 150 hPa temperature – PBL and HYD	-67.6˚C	1.3˚C	-0.80**	17
4	August-September, 850 hPa wind speed – TRV	10.3 m/s	1.2 m/s	-0.54*	23
5	August, 150 hPa zonal wind – TRV	-32.9 m/s	2.6 m/s	0.47*	23
6	September, 300 hPa zonal wind - TRV	-7.8 m/s	1.8 m/s	0.49*	23

(Source: Raj, 1998a)

n: sample size; CC, **, *: as in Table 13.6a, SD: as in Table 13.3d

Table 13.9: List of predictors used for seasonal prediction of NRT on experimental mode

Predictor	**Type of relation with NRT**	**Long term mean** **(based on NCEP reanalysis datasets)**
PR1: Apr, 200 hPa Zonal wind anomaly Over India (areal ave 70-95E, 5-30N)	Strong westerly winds (positive anomaly) favour good NRT; Weak wind, poor NRT	**15.49**m/sec
PR2: JJAS, 200 hPa Temperature anomaly over central India (areal ave 74-85E, 8-20N)	Negative anomaly favours good NRT; Positive anomaly, poor NRT	**-64.38**˚C
PR3: Aug-Sep, 150 hPa Strength of TEJ over the extreme south peninsula (areal .ave 76-79E, 7-10N)	Strong TEJ, poor NRT; Weak TEJ, good NRT	**-31.96**m/sec (upto 16 Sep)
PR4: JJAS, SOI	Negative SOI, good NRT; Positive SOI, poor NRT	
PR5: IMR of JJAS	Slightly discordant negative relationship; Conditional means (CM) give a better indication	
PR6:Aug-Sep, MSLP over Siberian region (87-103˚E; 47-53˚N)	Negative anomaly is associated with slightly deficient NRT(ON) but may lead to an excess NRT(Dec)	**1011.71** hPa

Abbreviations: Please refer previous tables or consolidated list

Table 13.10: Performance of experimental prediction scheme for NRT during 2001-11

Year	No. of predictors	Overall outlook	Realised rainfall	Forecast performance
2001	3	Near normal	Normal negative departure	Correct
2002	3	Normal	Normal negative departure	Partly correct
2003	3	Normal with a reasonable chance of positive departure	Deficient	Incorrect
2004	4	Normal with a reasonable chance of positive departure	Normal	Correct
2005	4	Normal with a reasonable chance of positive departure	Excess	Partly correct
2006	5	Normal with a reasonable chance of positive departure	Normal Positive side	Correct
2007	6	Normal with a slightly negative departure	Excess	Incorrect
2008	5	No clear signal; 3 parameters indicated positive departure and the other 2 indicated negative departure	Excess	--
2009	6	Normal	Normal positive side	Correct
2010	6	Near normal	Excess	Incorrect
2011	6	Near normal to normal	Excess	Incorrect

NRT: As in Table 13.2a

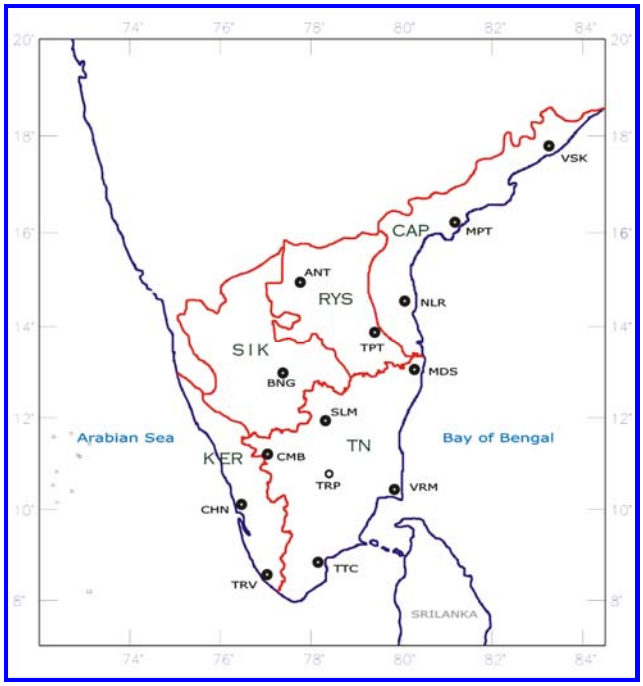


Fig.13.1: Geographical locations of the five meteorological sub-divisions of southern peninsular India influenced by the Northeast monsoon

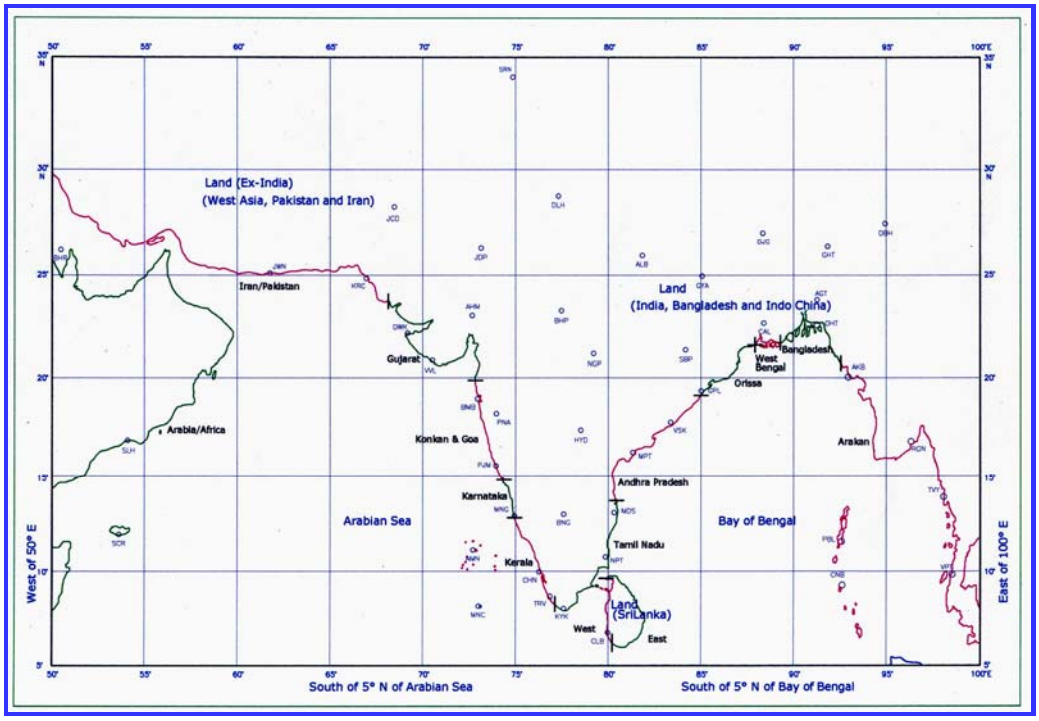


Fig.13.2: Map depicting the boundaries of the Bay of Bengal and the Arabian Sea and the various sub regions of the North Indian Ocean basin

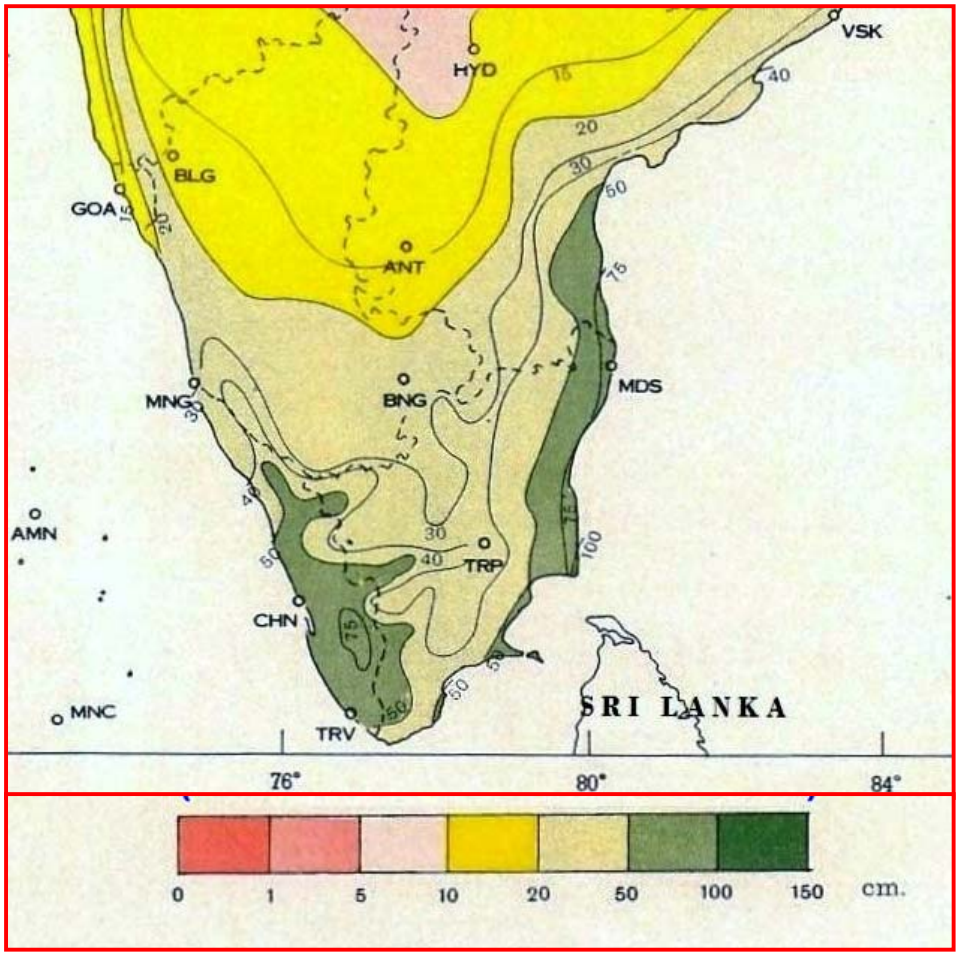
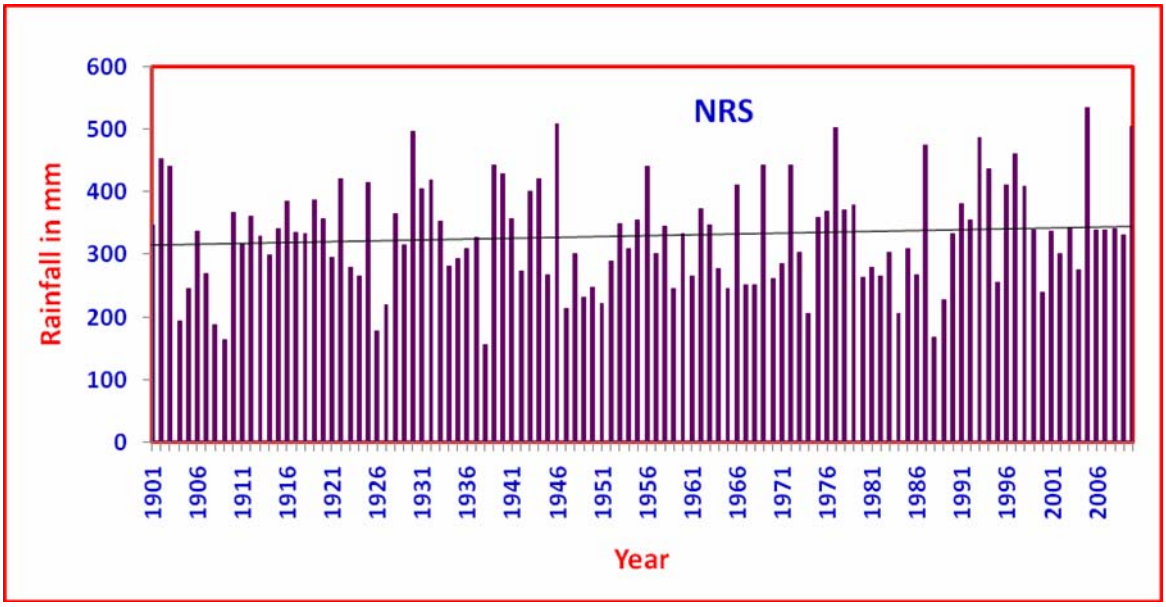
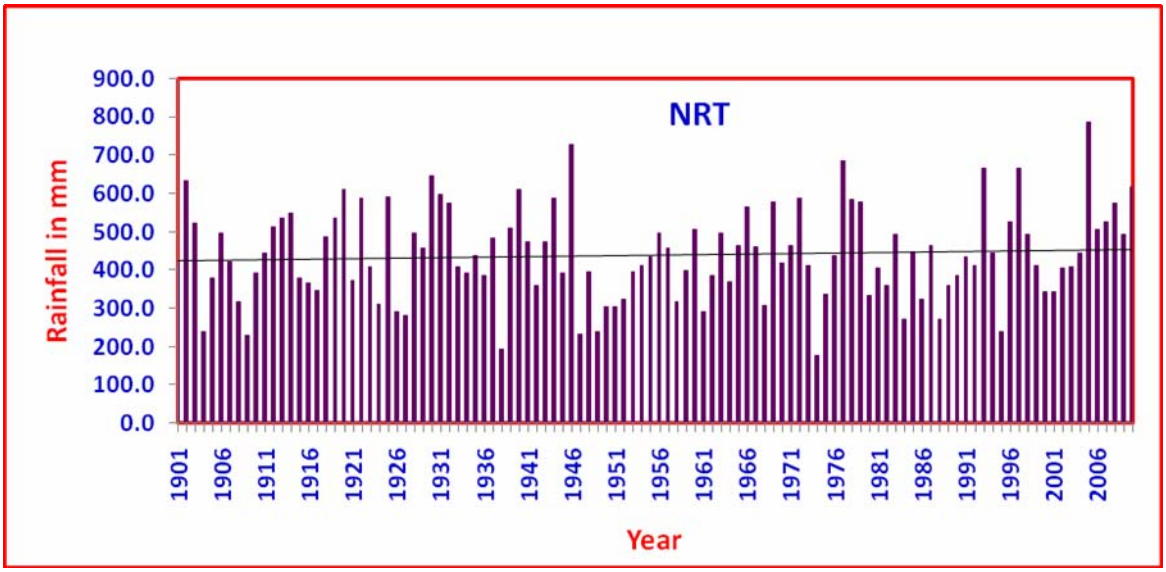


Fig.13.3: Normal spatial distribution of rainfall over southern peninsular India during the Northeast monsoon season (OND).

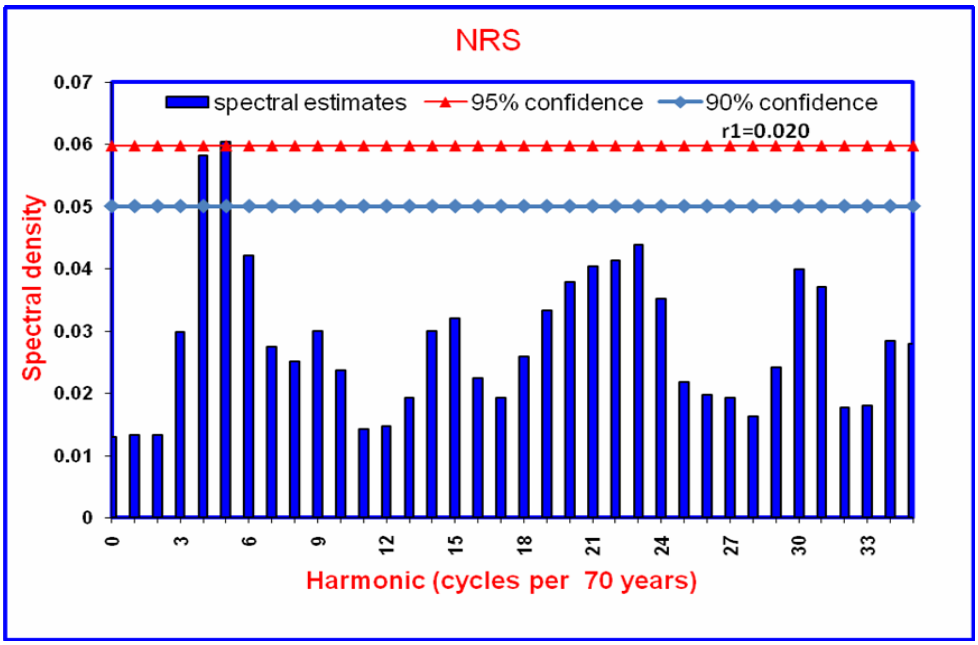


(a)

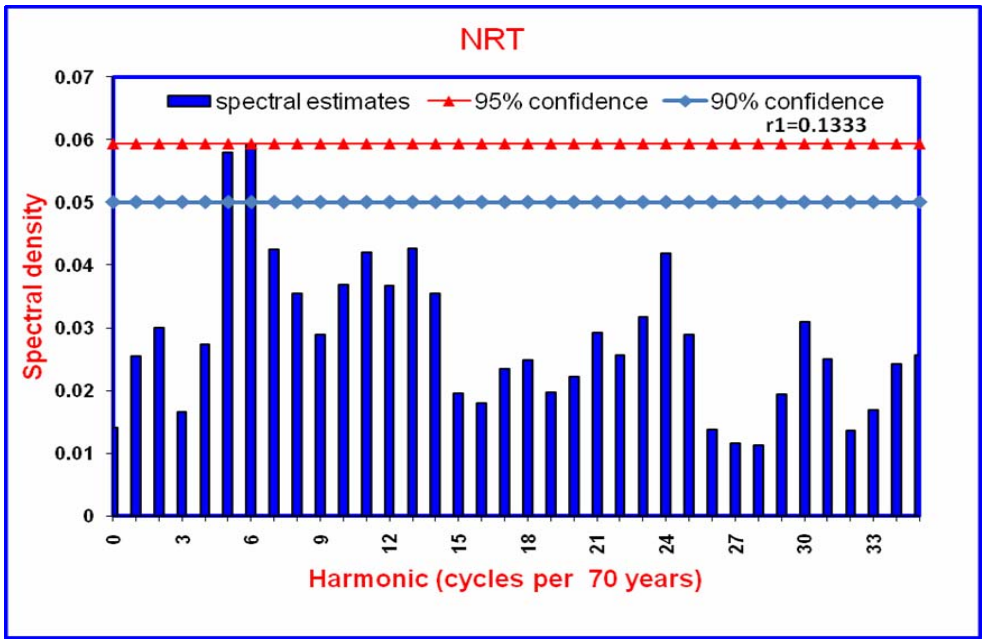


(b)

Fig.13.4a&b: Time series of (a) NRS and (b) NRT for the period 1901-2010.



(a)



(b)

Fig.13.5a&b: Power spectra of (a) NRS and (b) NRT (based on OND rainfall data of 1901-2010)

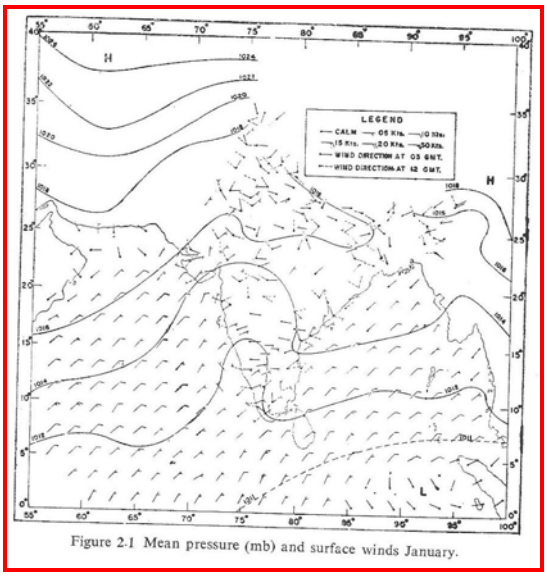


Figure 2.1 Mean pressure (mb) and surface winds January.

(a)

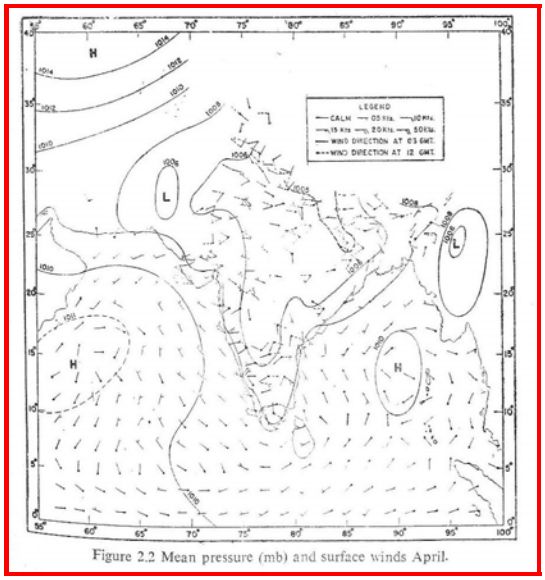


Figure 2.2 Mean pressure (mb) and surface winds April.

(b)

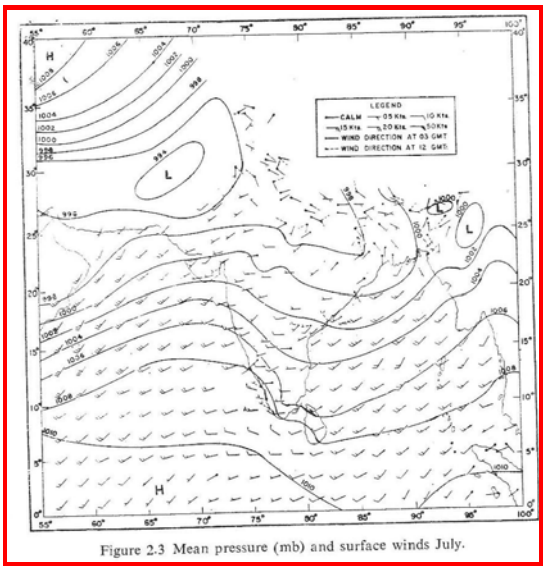


Figure 2.3 Mean pressure (mb) and surface winds July.

(c)

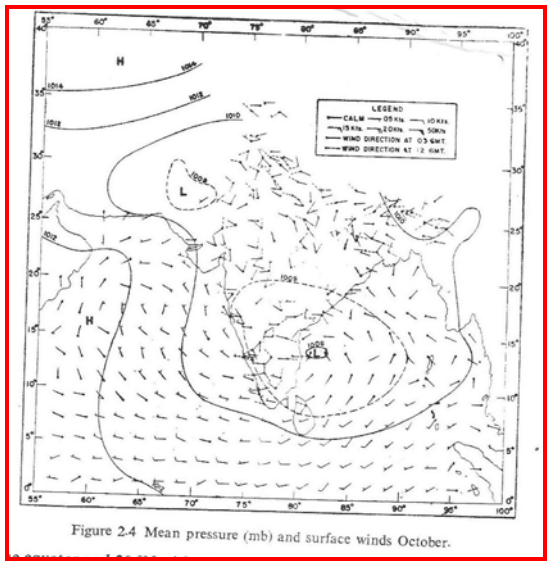
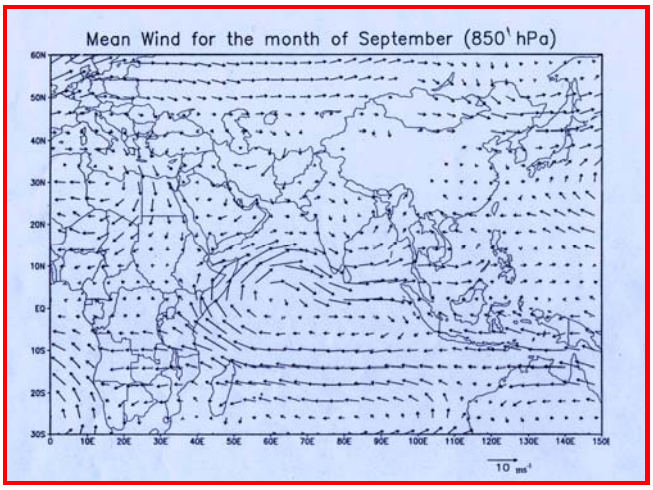


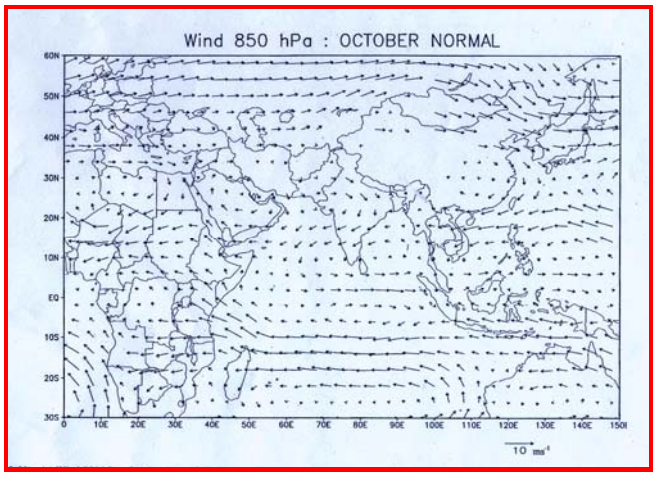
Figure 2.4 Mean pressure (mb) and surface winds October.

(d)

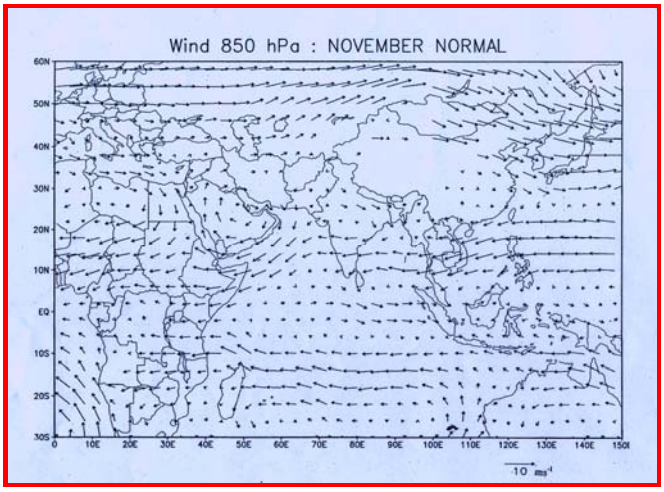
Fig.13.6a-d: Reversal of surface pressure pattern as seen from the mean pattern in (a) January (b) April, (c) July and (d) October.



(a)



(b)



(c)

Fig.13.7a-c: Normal 850 hPa winds over the Asian region during Sep-Oct-Nov (Source: Asnani, 2005)

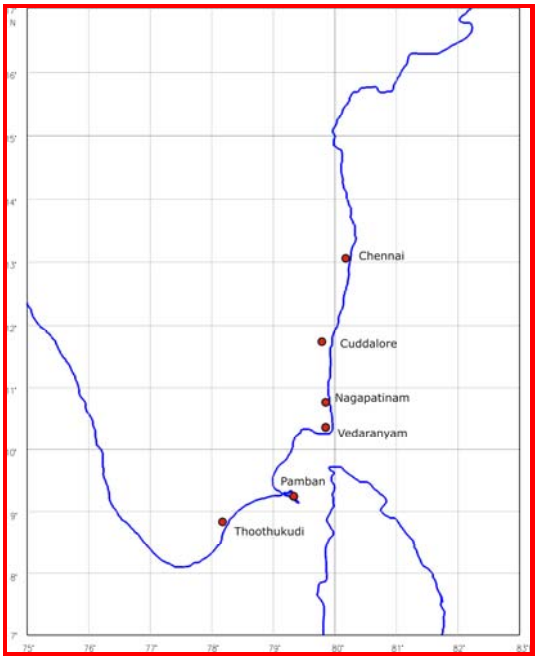


Fig.13.8a: Locations of 6 CTN stations used for determination of dates of onset and withdrawal of NEM Raj(1992&1998).

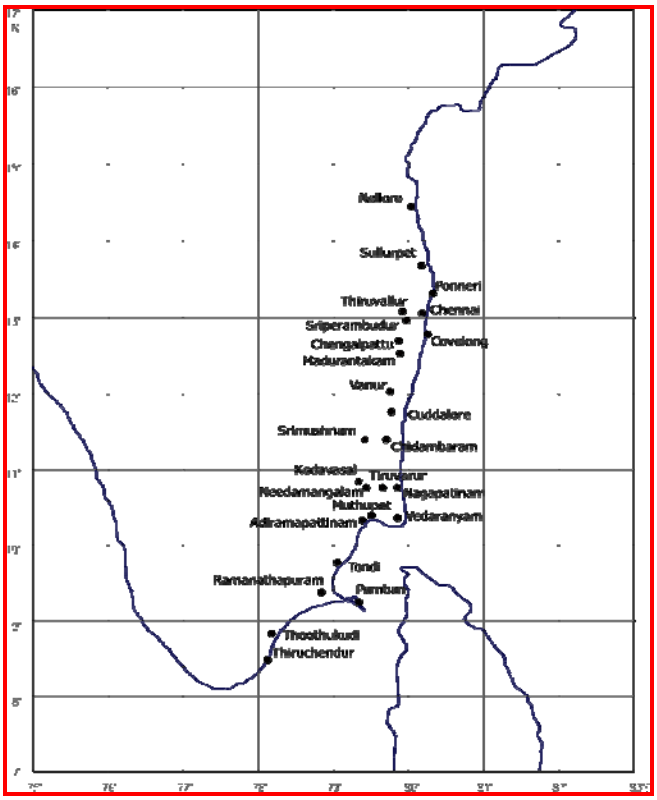


Fig.13.8b: Locations of 25 l stations of CTN/SCAP considered for Re- determination of dates of onset and withdrawal of northeast monsoon

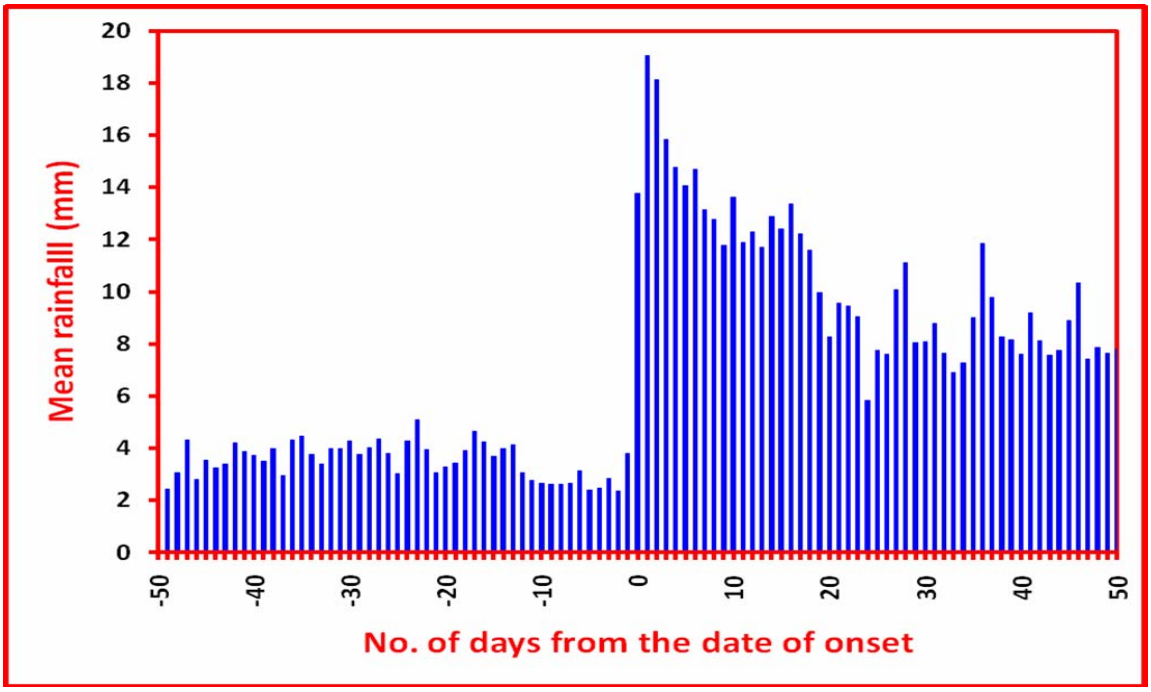


Fig.13.9a: Mean daily rainfall over CTN and SCAP with reference to the onset dates of northeast monsoon (Superposed epoch analysis)

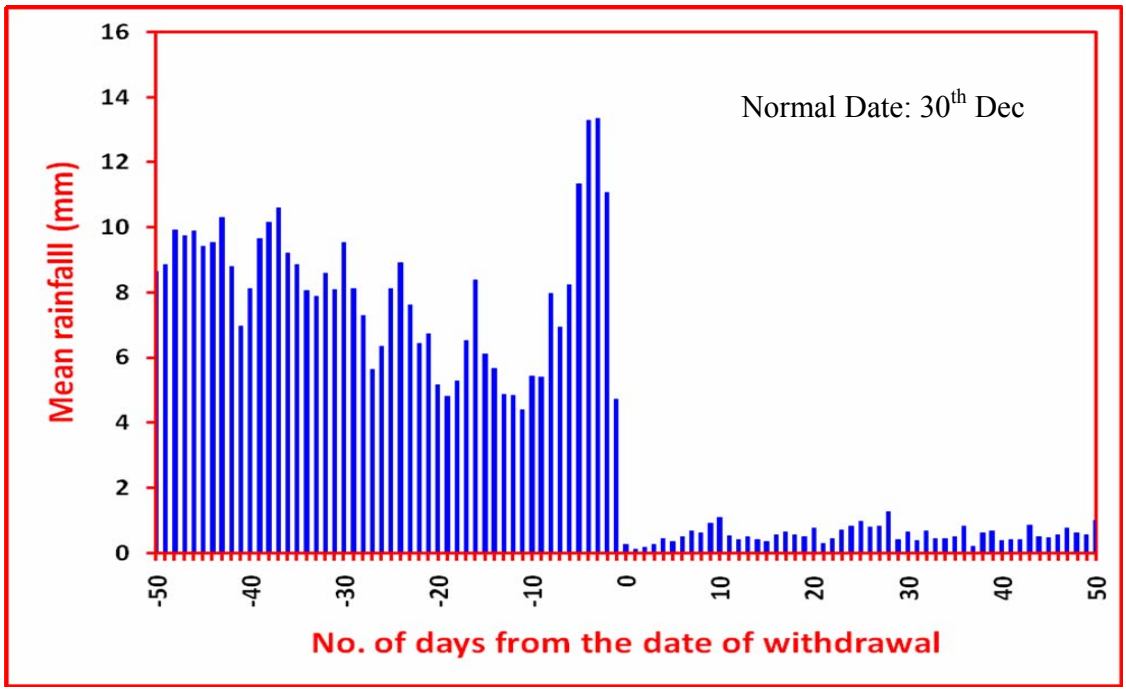
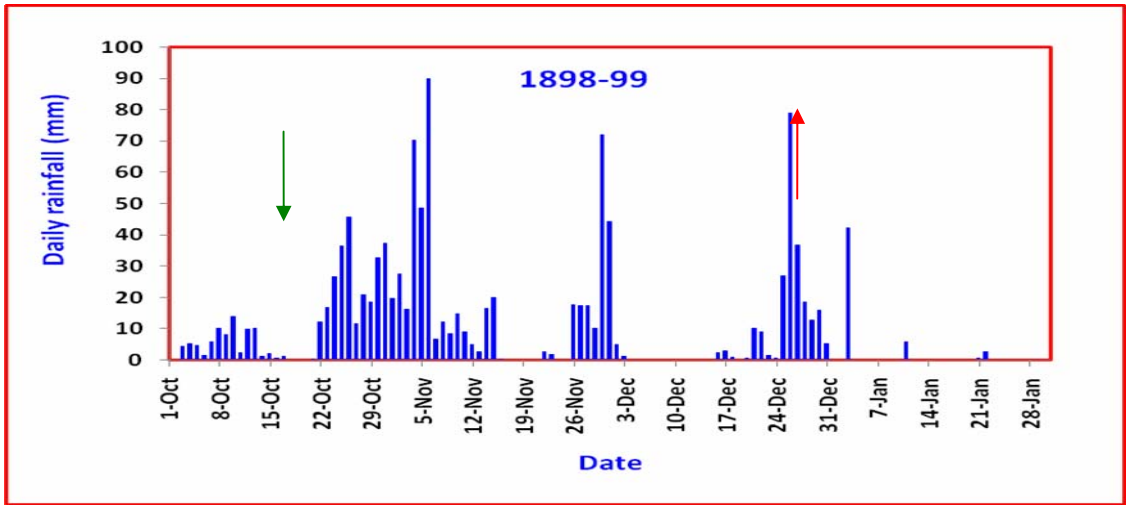
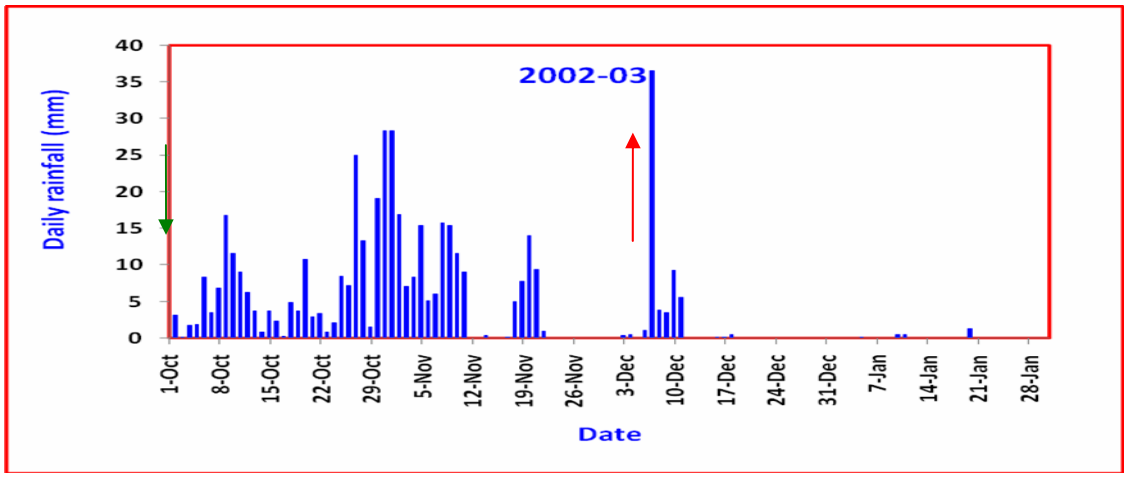


Fig.13.9b: Mean daily rainfall over CTN and SCAP with reference to the dates of withdrawal of northeast monsoon (Superposed epoch analysis)



(a)



(b)

Fig.13.10a-b: Illustrations on determination of dates of onset (denoted by down arrow in green) and withdrawal (up arrow in red) for two years *viz.* 1898-99 & 2002-03

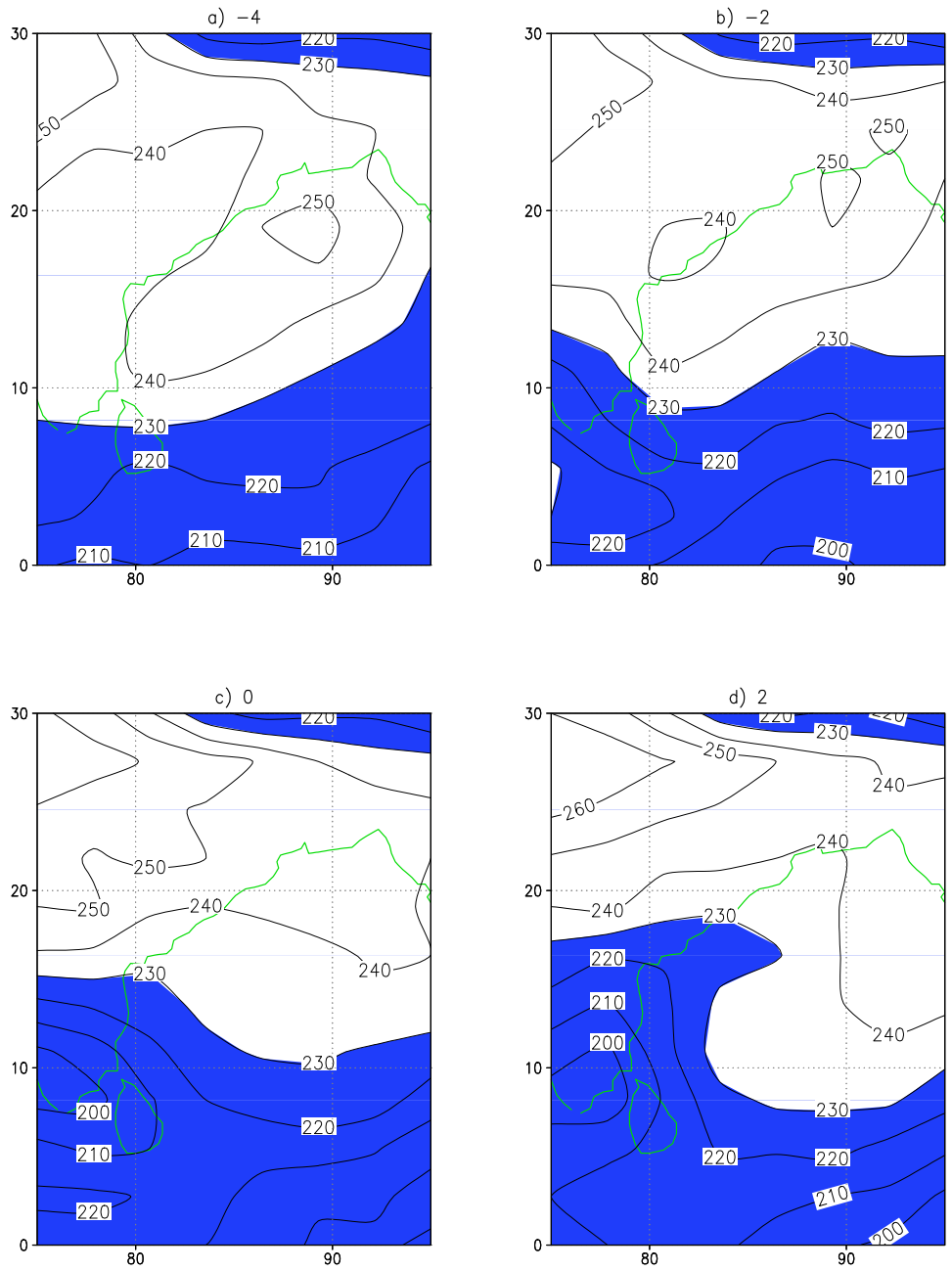


Fig.13.11a-d: Spatial variation of mean OLR for -4, -2, 0 & 2 days with respect to the NEM onset dates (OLR values in W/m^2, shaded area OLR < 230 W/m^2).

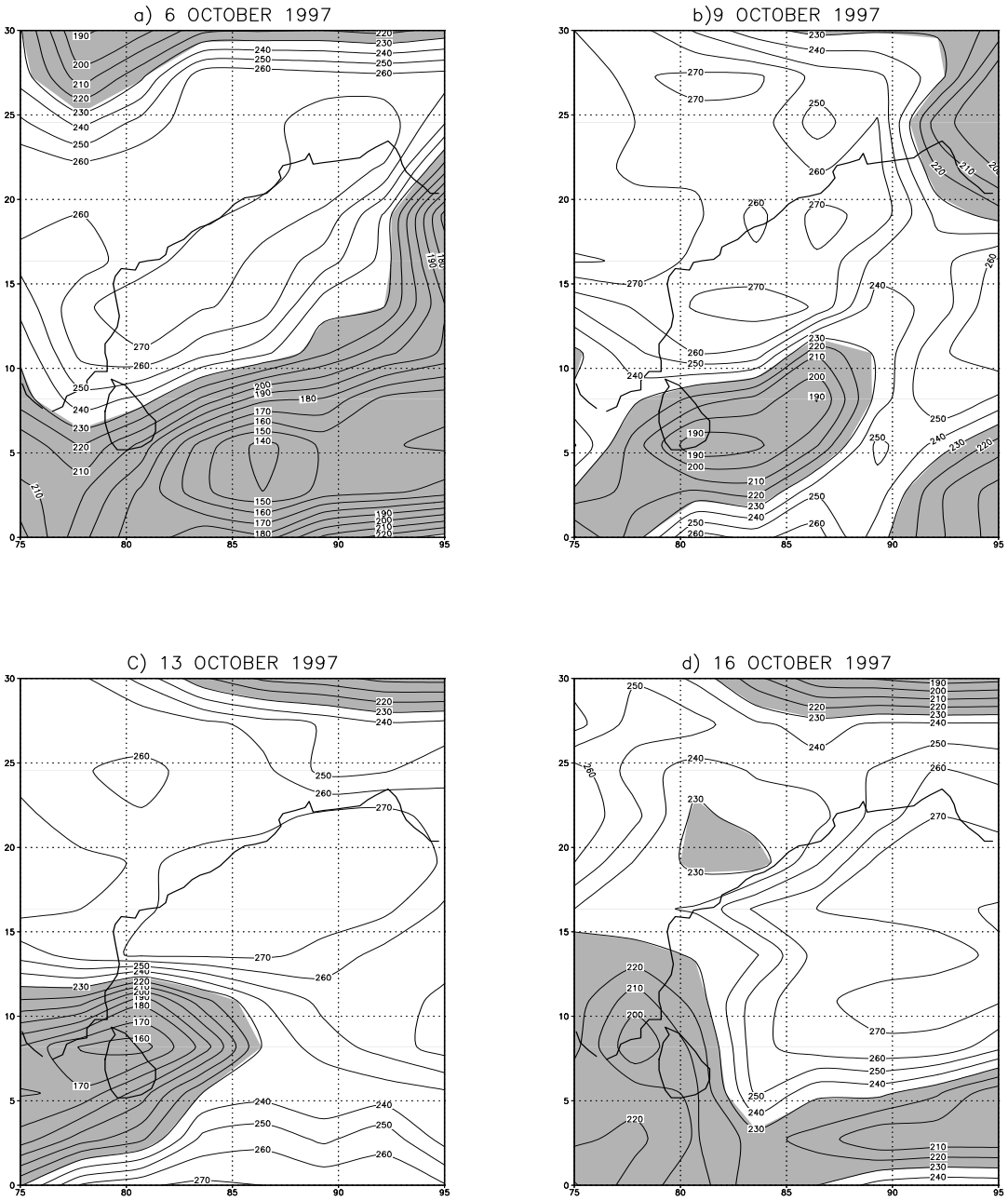
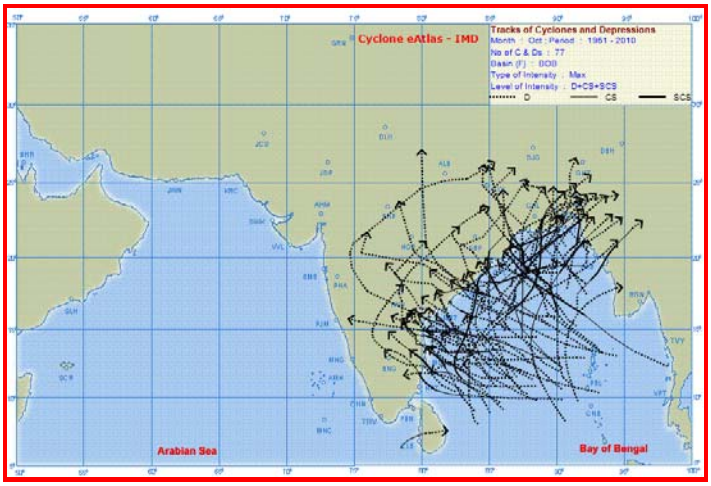
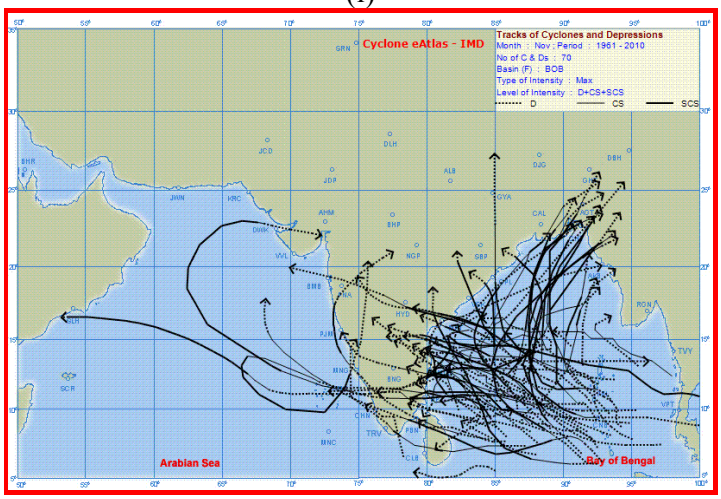


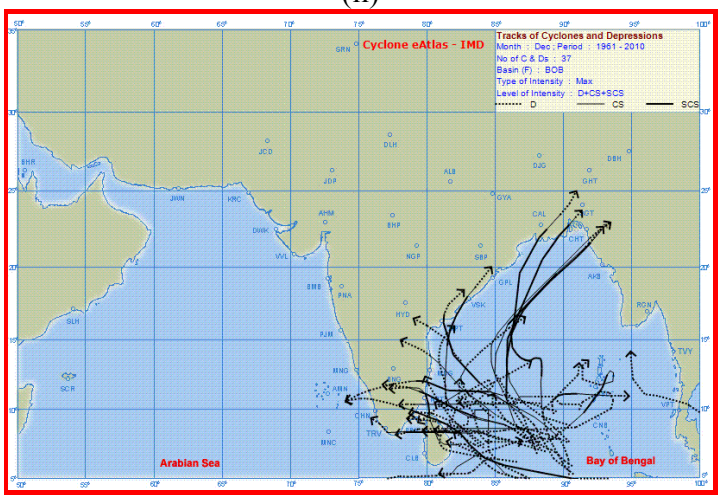
Fig.13.12a-d: OLR distribution during the onset phase of 1997 (6, 9, 13 and 16 October) (OLR values in W/m^2 , shaded area OLR < 230 W/m ^{2}).



(i)



(ii)



(iii)

Fig.13.13a: Tracks of LPS that formed over BOB during (i) October, (ii) November and (iii) December, 1961-2010.

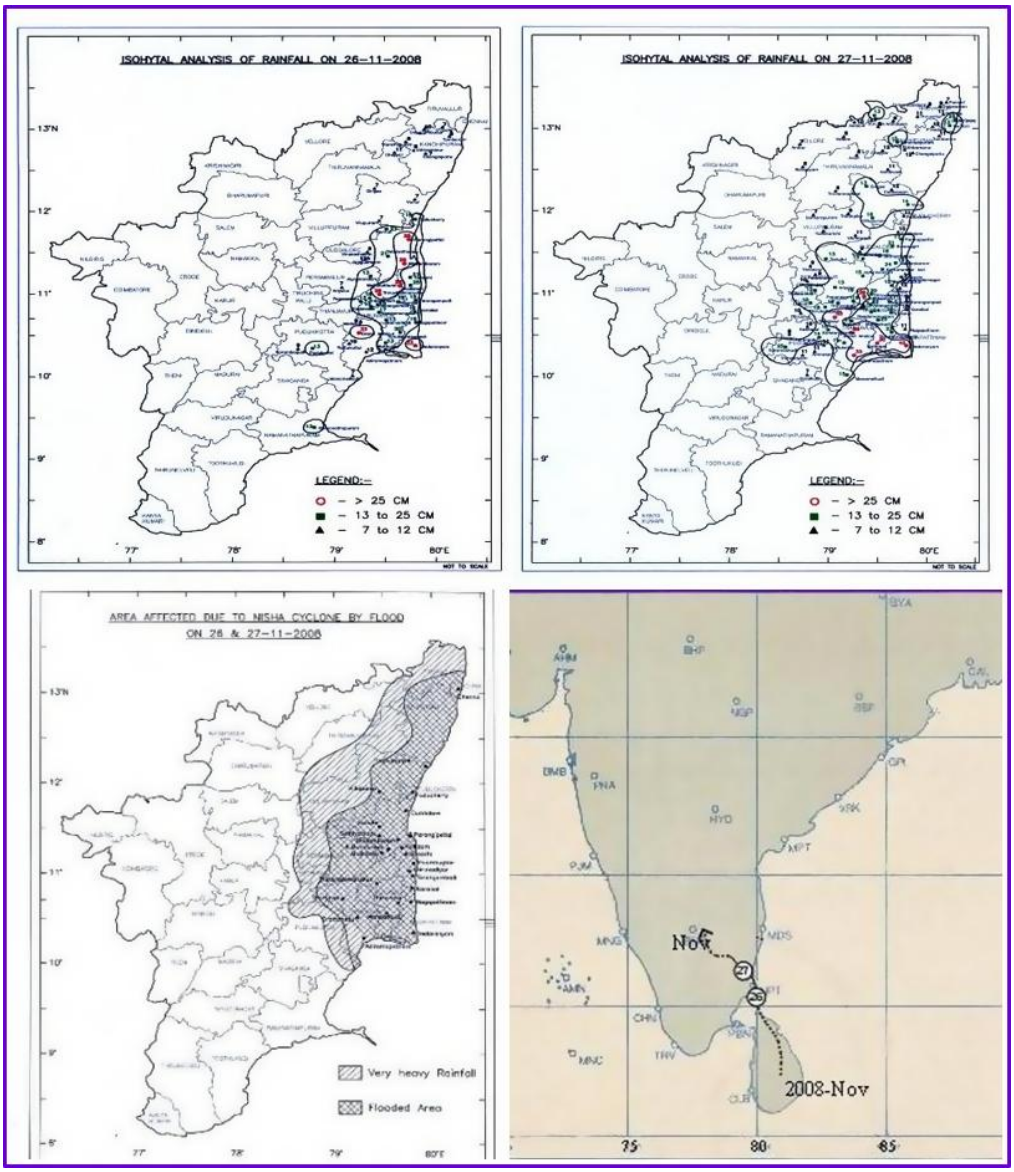


Fig.13.13b: Track and rainfall associated with Cyclonic Storm Nisha (25-28 Nov 2008).

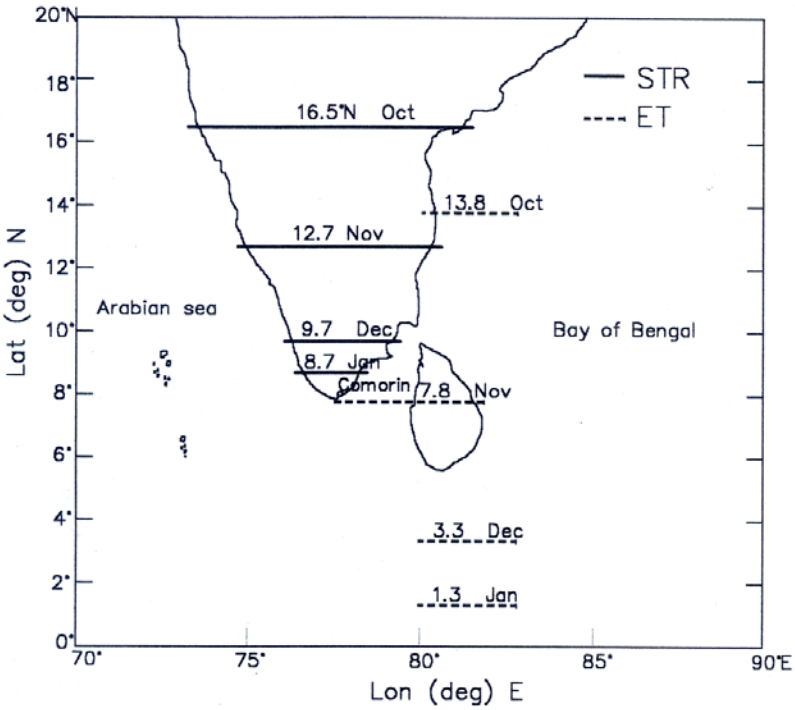


Fig.13.14: Normal latitudinal movement of Sub Tropical Ridge at 200 hPa level over 77.5ºE and surface Equatorial Trough off the east coast of India during October-January

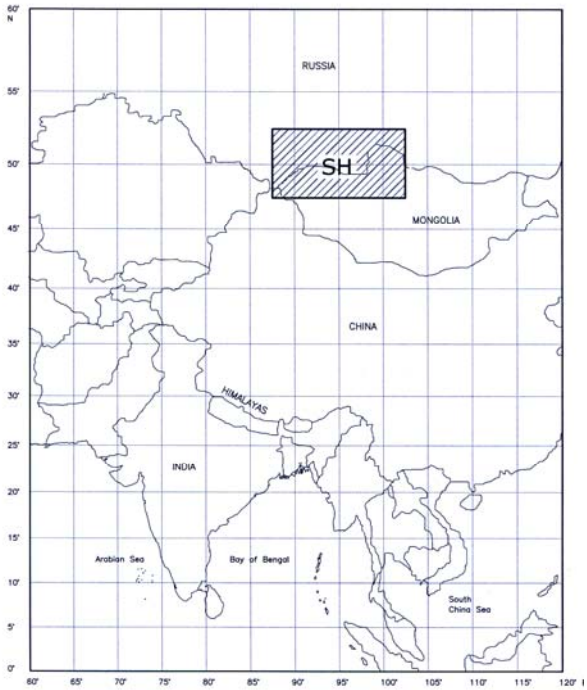


Fig.13.15a: Region of Siberian High shown to influence NEM rainfall

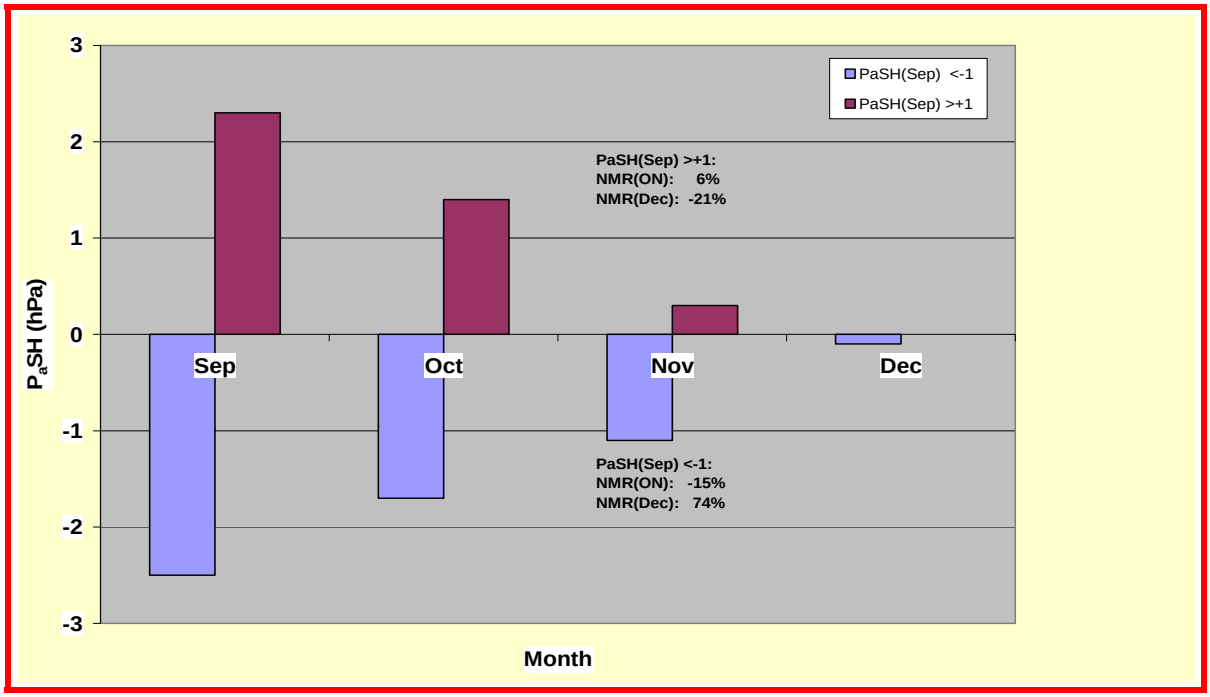


Fig.13.15b: Profiles of P_aSH during the northeast monsoon months given $P_aSH(Sep)$ <-1 & >+1 hPa and the associated performance of Northeast monsoon over Tamil Nadu

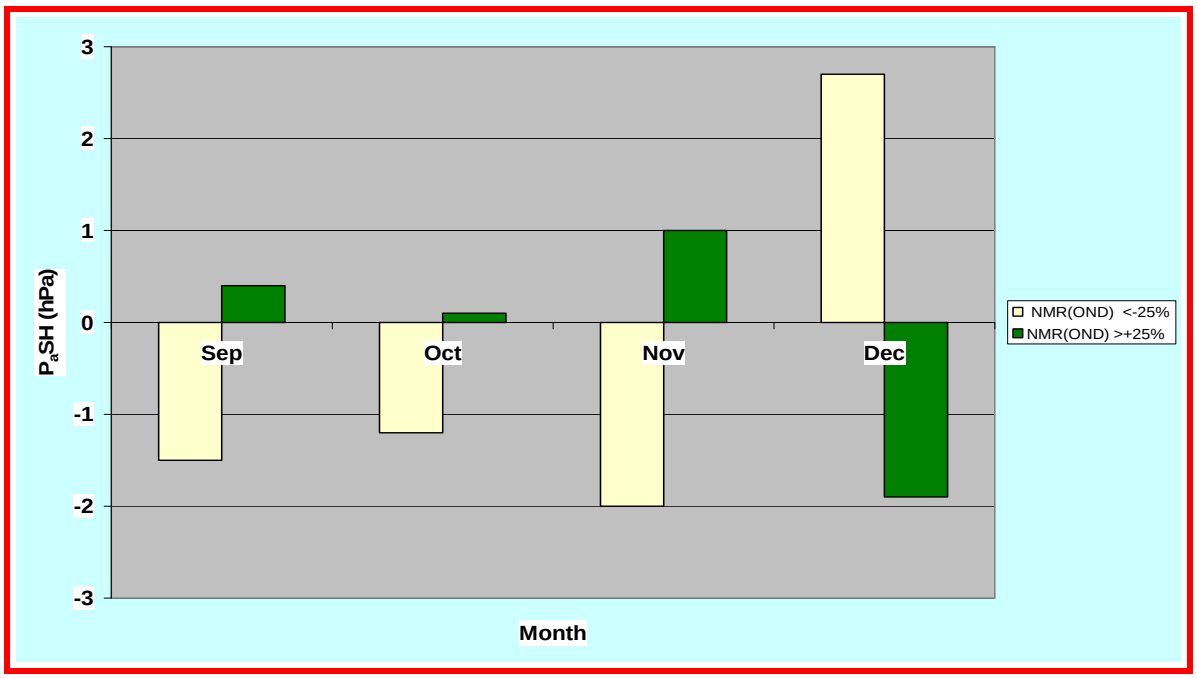


Fig.13.15c: Profiles of P_aSH during the northeast monsoon season for NMR(OND)<-25% and NMR(OND) >+25% [NMR: Northeast monsoon rainfall given as percentage departure from normal]

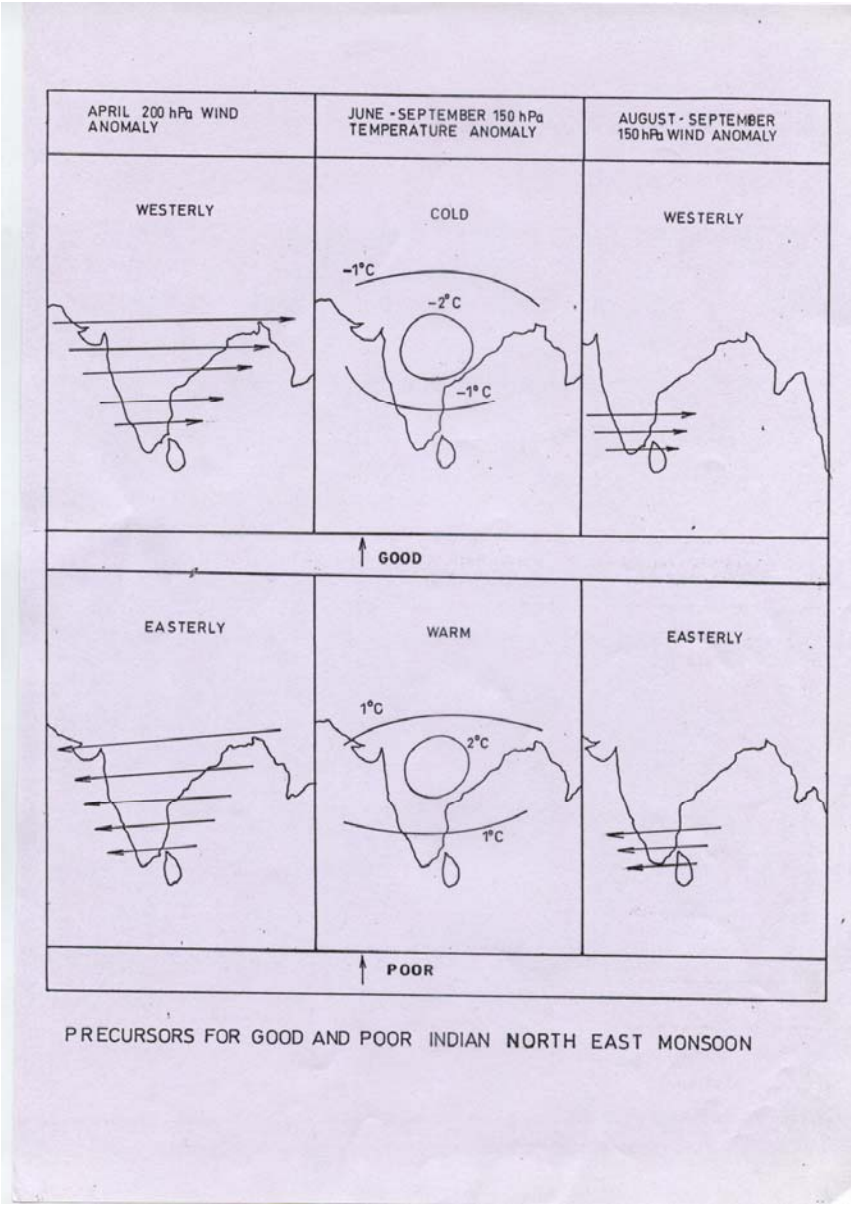


Fig.13.16: Schematic diagram depicting the influence of April 200 hPa zonal winds, JJAS 150 hPa temperature, August-September 150 hPa zonal winds over extreme southern peninsula on NEM.

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