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Six decades of research in diagnostic meteorology of the Asian tropics

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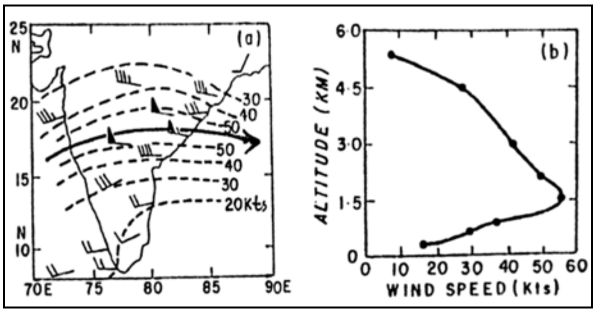
सार – इस शोध पत्र में 1960 के आरम्भ से छः दशकों के दौरान लेखन द्वारा किए गए शोध कार्य को संक्षेप में बताया गया है जिसमें उष्णकटिबंध में मौसम प्रणालियों और वायुमंडलीय महासागरीय प्रक्रियाओं का अध्ययन किया गया है। उन्होंने परम्परागत नाम "सिनॉप्टिक मोटीरियोलॉजी" के स्थान पर इसे "डायग्नॉस्टिक मीटीरियोलॉजी" का नाम दिया है। भारत में मॉनसून संचरण में निम्न स्तरीय जेट स्ट्रीम देखा गया है। मॉनसून के सक्रिय विरल चक्र में और एल निनो वर्षों के दीर्घ विरल मॉनसून दौर में महासागर वायुमंडल कपलिंग में मुख्य भूमिका है। मॉनसून ऋतु के दौरान मई के महीने में उत्तरी हिंद महासागर में उष्ण हिस्सा और बंगाल की खाड़ी में ठंडा हिस्सा मॉनसून के दौरान मुख्य भूमिका निभाते हैं। यह पता चला है कि मॉनसून से पूर्व के महीनों में हिंद और प्रशांत महासागर में सतह तापमान विसंगतियाँ भारत और ऑस्ट्रेलिया दोनों पर ग्रीष्मकालीन मॉनसून के आरंभ की तारीखों को निर्धारित करती हैं। पश्चिमोत्तर भारत में ऊपरितन क्षोभमंडलीय पवनों के दक्षिण की ओर पछुआ में छह नंबर की तरंग की द्रोणी मानसून ऋतु के दौरान अनावृष्टि लाती है। यह द्रोणी स्टेशनी रॉसबाए वेव ट्रेन का एक भाग है जिसे एशिया पैसिफिक वेव का नाम दिया गया है। मॉनसून उष्ण स्रोतों द्वारा परिकल्पित की गई है। इन पश्चिमी पवनों के आने से वायुमंडलीय महासागर में अस्थिरता पैदा होती है जिसके परिणाम स्वरूप भारत में अक्सर अनावृष्टि होती है जैसे कि 30 वर्षों के समय 1961-1990 जो एटलांटिक मल्टी डिकेडल दोलन की शीत अवस्था के साथ उष्णकटिबंध में SST- कनवेक्शन संबंध मानसून और मॉनसूनेतर (ITCZ भूमध्य रेखा के समीप) क्षेत्रों में भिन्न रूप में पाई गई है। दिल्ली हवाई अड्डे पर संवहनी आँधी का अध्ययन करते हुए यह देखा गया कि दोपहर और शाम के समय चली आँधी की अपेक्षा रात के समय क्षैतिजीय दृश्यता में स्थानिक परिवर्तनशीलता काफी धीमी थी। आँधी के समय दृश्यता परिवर्तनशीलता को समझने के लिए गर्ज भरे तूफान डाउनड्राफ्ट मॉडल का उपयोग किया गया है जिसमें हवाई जहाज के उड़ने और उतरने के समय ऐपलीकेशन थी।

ABSTRACT. This paper briefly describes the research work done by the author during the six decades beginning in early 1960s studying weather systems and atmosphere-ocean processes in the tropics. He named the science behind as “Diagnostic Meteorology” instead of the conventional “Synoptic Meteorology”. A Low Level Jet stream was shown to exist in the monsoon circulation over India. It has a major role in the ocean – atmosphere coupling in the Active-Break cycle of the monsoon and in long break monsoon spells of El Nino years. A “warm pool” in north Indian Ocean in May and a “cold pool” in Bay of Bengal in the monsoon season were shown to have important roles in monsoon. It was shown that surface temperature anomalies over Indian and Pacific oceans of the pre-monsoon months have control on the dates of summer monsoon onsets over both India and Australia. Southward intrusion of upper tropospheric winds into northwest India as a wave number six trough in westerlies was found to cause monsoon season droughts. This trough is part of a stationary Rossby Wave train named Asia Pacific Wave, hypothesised to be induced by monsoon heat sources. These westerly wind intrusions were shown to have caused an atmosphere-ocean instability resulting in frequent monsoon droughts in India in 30 year epochs like 1961-1990 coinciding with the cold phase of the Atlantic Multi decadal Oscillation. SST – Convection relation in the tropics was found to be different in monsoon and non-monsoon (ITCZ close to the equator) regions. Studying convective dust-storms (Andhi) at Delhi airport it was found that temporal variations in the horizontal visibility was much slower at night compared to afternoon and evening Andhi occurrences, A thunderstorm downdraft model was used to explain the visibility variations in Andhi which has applications in aircraft take-off and landing operations.

Key words – Low Level Jet stream (LLJ), Asian tropics, Asian summer monsoon.

1. Introduction

I was invited by the Editor Mausam to contribute a review article highlighting and synthesising my research contributions on weather and climate variability done during the last six decades. Of this period for two decades I did operational weather forecasting, cyclone warning and aviation meteorology and two decades worked as a teacher of tropical meteorology and global climate. During all these six decades I did research in Diagnostic



Figs. 1(a&b). Low level jet stream of a day in July (a) winds, isotachs in knots (1 knots = 0.51 m/s) and jet axis at 850 hPa, (b) vertical profile of wind speed at a point on the jet axis. Figure adapted from Joseph and Raman (1966)

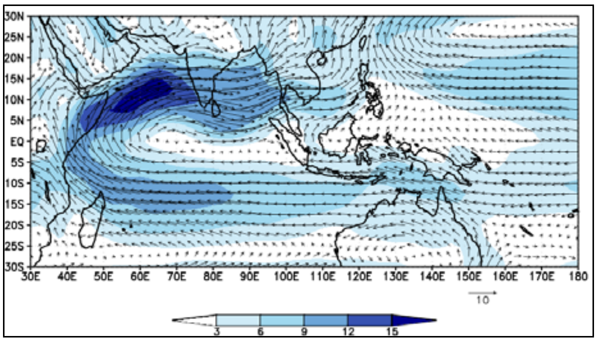


Fig. 2. Mean wind at 850 hPa of July and August (1950-2010) showing the cross-equatorial Low Level Jet stream

Meteorology of the Asian tropics. Diagnostic Meteorology is the new name I am giving to Synoptic Meteorology, a 200 year old science which studied the structure and evolution (diagnostics) of weather systems like thunderstorms, cyclones, monsoon, waves in the wind flow & jet streams. Forecasting of the associated weather was done using simple principles of dynamics and thermodynamics of the atmosphere and a wealth of empirical rules derived from analysis of atmospheric data at several successive synoptic observational timings. The name Diagnostic Meteorology for such a science is better than Synoptic Meteorology which implies only that measurements are made at the same time over large areas. As one of the scales of weather systems however the name Synoptic Scale may be used respecting the history of the science of meteorology which began with Synoptic Observations and study of Synoptic Systems; these names may continue to be used.

2. Low level Jet stream of Asian summer monsoon

Bunker (1965) from the aircraft wind measurements made during the International Indian Ocean Expedition found that high winds of about 50 knots (1 knot = 0.51 m/s) occurred during the summer monsoon season over south-west Arabian Sea at about one kilometer above sea level and these winds showed strong shear in the vertical both below and above this wind maximum. The existence of a Low Level Jet stream (LLJ) over peninsular India during the summer monsoon month of July with strong shears both in the vertical and horizontal was established by Joseph and Raman (1966). Fig. 1(a) gives the LLJ axis at 850 hPa on a typical active monsoon day in July and the isotachs of the wind showing strong horizontal wind shears north and south of the jet axis. The vertical wind speed profile at Visakhapatnam on the east coast of India close to the axis of the LLJ may be seen in Fig. 1(b). Analysing wind data of 5 July months they found that (a) such an LLJ axis passes through peninsular India on many days in July with its core at about 1.5 kms above sea level (850 hpa) and core wind speeds of the order of 40-60 knots (20-30 m/s) and (b) In the vertical LLJ has strong shears above and below the jet axis, the wind shear in the vertical below the jet core being more than that above it.

Findlater (1969) established the existence of a cross equatorial LLJ during the Asian Summer Monsoon. The wind maximum found by Bunker (1965) and the LLJ over peninsular India discovered by Joseph and Raman (1966) are parts of this cross equatorial jet stream. Fig. 2 gives the mean wind flow of the period 1950 to 2010 at 850 hPa over South Asia of July and august based on NCEP/NCAR reanalysis - Kalnay *et al.* (1996). The mean LLJ crosses the equator along the east African coast with strong southerly winds and has its maximum wind at 850 hPa of more than 15 m/s over the Arabian Sea. The jet axis passes through peninsular India and the westerly winds of the LLJ are seen extending into the West Pacific Ocean.

The birth of the LLJ coincides with the onset of monsoon over Kerala according to Joseph *et al.* (2006) and Boos and Emanuel (2009). Fig. 3(a) shows composites of OLR (Outgoing Longwave Radiation, a proxy for rainfall) and 850 hPa wind flow around the time of monsoon onset over Kerala. The rainfall area is just north of the axis of the LLJ where in the atmospheric boundary layer there is strong cyclonic vorticity which is the cause of the dynamic component of the rainfall associated with LLJ. Where there is mountainous terrain facing the strong winds of the LLJ we get heavy rains which is the orographic component of LLJ rainfall.

After its formation at the time of monsoon onset over Kerala, LLJ has a life duration the same as that of the monsoon, with major fluctuations in the

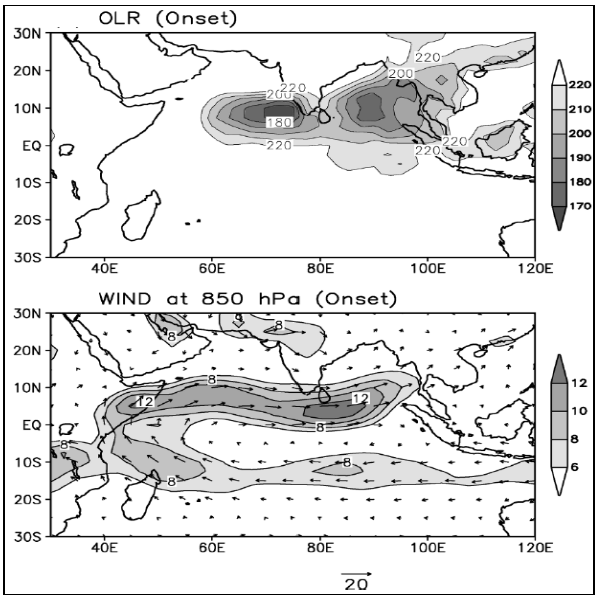


Fig. 3(a). Composite of several cases of monsoon onset over Kerala - OLR in watts/sq. m (top) and 850 hPa wind in m/s (bottom)

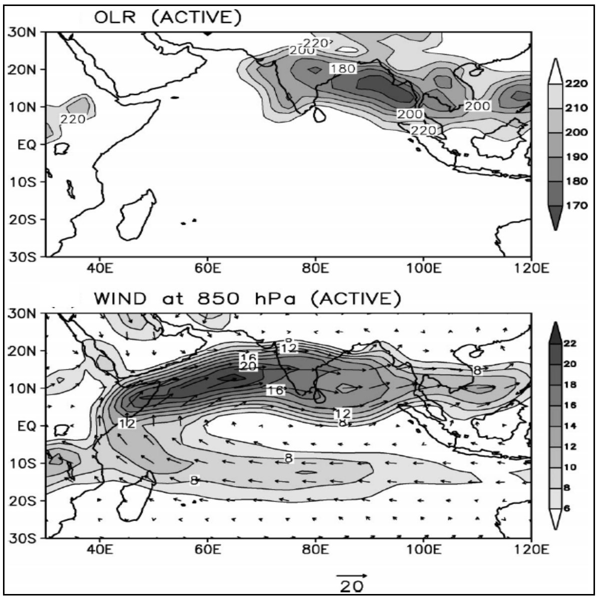


Fig. 3(b). Composite of several active monsoon days - OLR in watts/sq.m (top) and 850 hPa wind in m/s (bottom

Active-Break cycle of the monsoon. The LLJ passing through India has different locations for its axis in the active and break phases of monsoon as shown by Joseph and Sijikumar (2004). In the active phase, LLJ axis passes through peninsular India along latitude close to 15° N [Fig. (3b)]. In break monsoon, the LLJ axis shifts to a position south of peninsular India and close to the equator [Fig. (3c)].

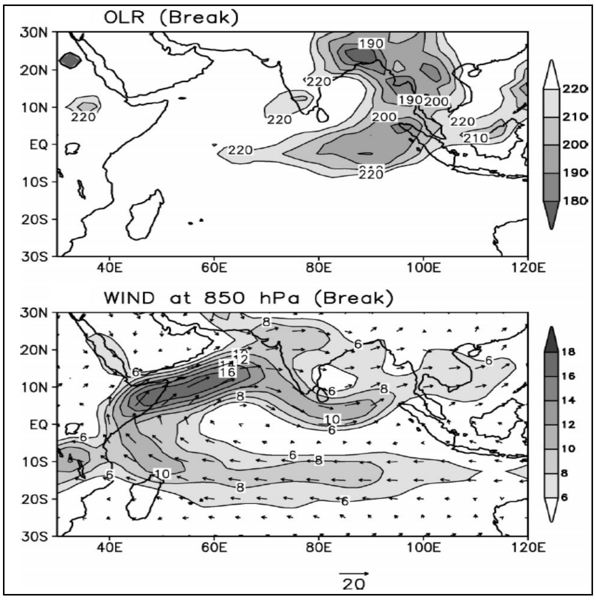


Fig. 3(c). Composite of several break monsoon days – OLR in watts/sq.m (top) and 850 hPa wind in m/s (bottom)

LLJ has two main functions (Joseph, 2014): (*i*) It acts as a conduit carrying the moisture generated by the trade winds over South Indian Ocean and the evaporative flux from the Arabian Sea to the areas of monsoon rainfall generation over South Asia including India and (*ii*) the area of cyclonic vorticity in the atmosphere boundary layer close to the LLJ axis is a dynamic forcing for the generation of upward motion of the moist monsoon air for the production of monsoon rainfall. In the active phase of the monsoon, deep convection occurs north of the axis of the LLJ between longitudes 70° E and 120° E. Joseph and Sijikumar (2004) found that the linear correlation coefficient between the convective heating of the atmosphere in this area (as represented by the OLR) and the strength of the zonal winds of the LLJ through peninsular India is very high and statistically significant at a lag of 2-3 days, convective heating leading. Srinivasan and Nanjundiah (2002) had shown earlier that on synoptic time scales the speed of LLJ over the Arabian Sea lags latent heating over Bay of Bengal by about 3 days. The atmospheric heating by convection north of the LLJ axis is able to accelerate the LLJ winds through peninsular India in 2-3 days. When the convection (rainfall) north of the LLJ in the Bay of Bengal decreases, the LLJ turns clockwise over the Arabian sea conserving potential vorticity as shown by the modelling studies by Hoskins and Rodwell (1995) and Rodwell and Hoskins (1995) & it flows along a latitude south of India. A good part of the moisture carried by the LLJ is then taken to the Pacific Ocean to nourish the tropical cyclones there. In the break monsoon phase that follows, deep

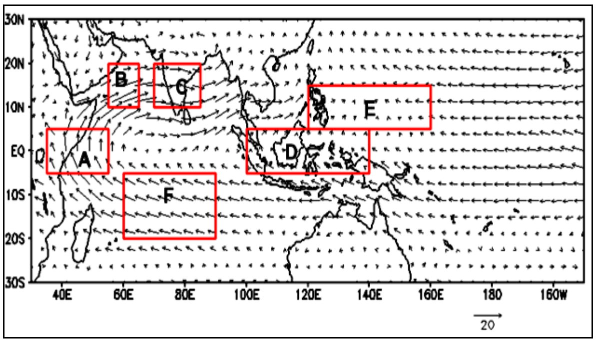


Fig. 4(a). Map showing the six zones (boxes) of 850 hPa wind of LLJ of monsoon season (JJAS), namely A (5° S-5° N, 35° E-55° E), B(10° N-20° N, 55° E-65° E), C(10° N-20° N, 70° E-85° E), D(5° S-5° N, 100° E-140° E), E(5° N-15° N, 120° E-160° E) and F(20° S-5° S, 60° E-90° E)

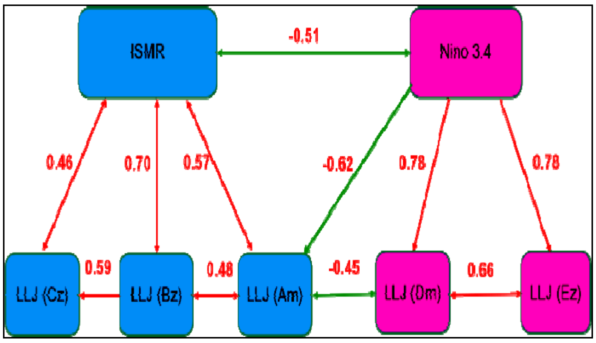


Fig. 4(b). Schematic diagram representing Inter-annual linear correlation coefficients (LCC) among ISMR, Nino 3.4 and LLJ components (positive LCC by red arrows and negative LCC by green). A one way relation that is statistically significant is marked by an arrow with single head. If there is a high and significant LCC between two parameters with two way interaction, it is represented by lines with arrows on both ends. All LCCs marked are statistically significant. The minimum LCC for statistical significance of 99% is 0.35 and for 99.9% is 0.44 (data length - 1958 to 2010)

convection shifts to an area close to and south of the equator in the Indian Ocean.

There is considerable variation in rainfall of India and LLJ winds of the monsoon season (1 June to 30 September) from one year to the next (inter-annual variability). El Nino is found to weaken the monsoon circulation and monsoon rainfall of India. It is also associated with increased westerly wind anomalies and weakening of the trade winds at 850 hPa level in the west / central Pacific Ocean - Webster and Yang (1992). Interannual variability of LLJ and its relation with Indian Summer Monsoon Rainfall (ISMR) and El Nino was studied by Wilson *et al.* (2018). ISMR was taken as the mean rainfall of 306 climatic rain gauge stations well distributed over India (Parthasarathy *et al.*, 1994). Wind data of 850 hPa are from Japanese Reanalysis (JRA-55) dataset at 1.25° × 1.25° latitude/longitude horizontal resolution (Kobayashi *et al.*, 2015) Mean SST anomalies in the Nino 3.4 (5° S-5° N, 120° W-170° W) region of the season October to December (hereafter called N3.4 index) represented El Nino of a year. Most of the research work on LLJ reported in the literature considered the whole LLJ as one entity, but in this study the domain of LLJ at 850 hPa covering a large area of Indian and west Pacific oceans was divided into 6 zones represented by boxes A to F as shown in Fig. 4(a). Winds of boxes A and D are predominantly meridional and winds in other boxes zonal. Suffixes z and m denote zonal (u) and meridional (v) wind components respectively.

The Linear Correlation Coefficients (LCC) between pairs were calculated using data of 1958-2010. LCC between N3.4 on one side and Ez and Dm on the other are very high at 0.78. Increased Ez creates increased cyclonic vorticity in the boundary layer causing increased convection (rainfall) in box E. Convection in box E on one side and Ez and Dm on the other grow in a positive feedback. It is well known that El Nino is associated with dry monsoon. Nino 3.4 index and ISMR have a LCC of -0.51. The LCC between N3.4 and Am is negative and very high at -0.62 and statistically significant at 99.9%. This shows that when N3.4 is high (strong El Nino), Am becomes weak which is associated with a weak ISMR. When Am is weak the along-shore winds off the coasts of Somalia and Arabia become weak reducing the coastal upwelling there, generating warm SST anomalies. This explains the finding of Babu and Joseph (2002) that monsoon droughts accompanied by El Nino creates stronger warm SST anomalies over western Indian Ocean than drought monsoon not accompanied by El Nino. The meridional wind over boxes A and D are negatively correlated with a LCC of -0.45. A schematic diagram showing these relationships is given in Fig. 4(b).

An important finding is that the LLJ component south of the equator in the Indian ocean (Fz) has very low LCC with ISMR and the wind components Am, Bz and Cz on the inter-annual time scale. There is a general concept that the trade winds of the Fz region turn right to become the cross-equatorial LLJ (Am) and again turn right to become the monsoon flow (Bz and Cz). If it is so, the correlation between Fz on one side and Am, Bz and Cz on the other side should have been high. But other features like interaction between convective heating of the atmosphere and the LLJ components Bz, Cz and Am (convection leading by 2-3 days) in a positive feedback as described in Joseph and Sijikumar (2004) and Joseph and Sabin (2008) become important. The strength of the LLJ zonal wind components Bz and Cz has no relation with the strength of the trade winds over the south Indian

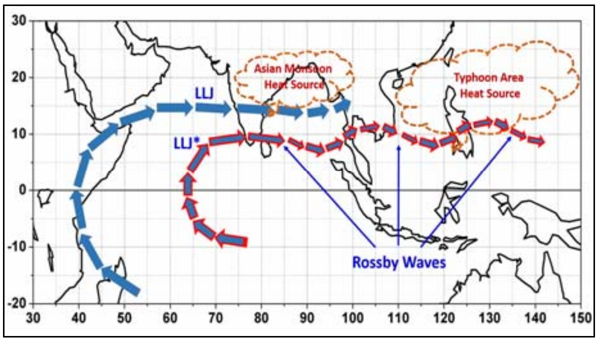
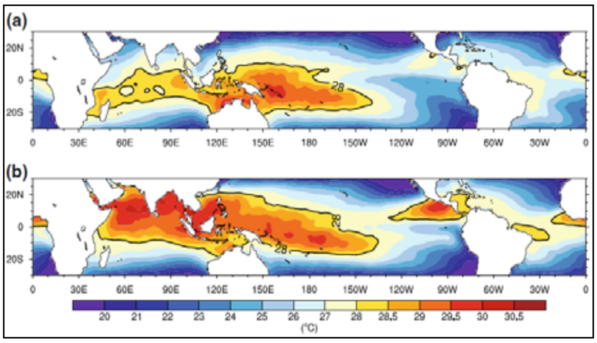


Fig. 5. Schematic of the air flow at 850 hPa showing the Active monsoon LLJ flowing through peninsular India and the newly discovered LLJ* flowing through Kerala in the two heavy rain spells of Kerala of July and August, 2018



Figs. 6(a&b). Warm Pool SST in (a) January and (b) may in HadISST data of 1961-1990. Note the northwestward shift of the center of the warm pool from January to may - figure taken from Joseph (2014)

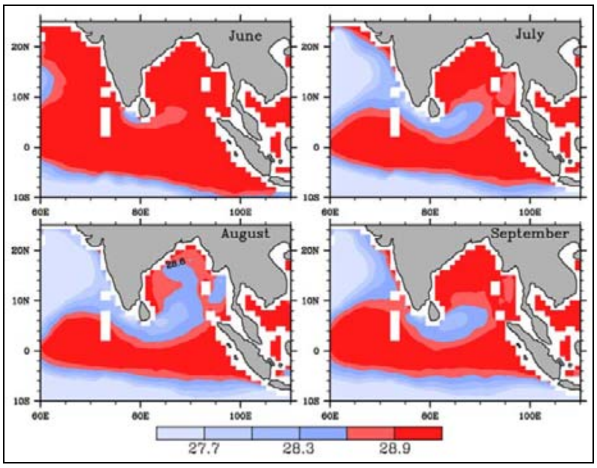


Fig. 7. Mean TMI SST in °C of june, july, august and September as mean of 1998 to 2005 showing the evolution of the Cold Pool of the Bay of Bengal - figure taken from Joseph and Sabin (2008)

ocean (box F) which remains a passive component of the LLJ on the interannual time scale.

Monsoon season of 2018 had devastating floods in Kerala. This was caused by two spells of heavy rains one from 7-19 July and the other from 7-17 August. Rainfall on 14-16 august was exceptionally heavy, several times more than the normal for those three days for Kerala. A research study to find the cause of these two rains spells by the author. During these ten day long spells of heavy rainfall in Kerala, Monsoon of India was in active phase and west Pacific ocean was producing considerable amounts of rainfall in super cloud clusters there which gave genesis to unusually large numbers of tropical cyclones some of which reached typhoon intensity. They hypothesised that the heating of the atmosphere by the latent heat released in the intense rainfall in west Pacific ocean has accelerated the weak winds over Kerala south of the active monsoon LLJ passing through peninsular India and produced a short lived LLJ* passing through Kerala that extended upto the convective heating zone in the west Pacific ocean. We have already seen that LLJ has three types - those at monsoon onset over Kerala and in active and break monsoon spells. This newly found LLJ* is the fourth type and it has a short life span of less than ten days. Examination of a few decades of data has shown that the occurrence of LLJ* is rare as when monsoon is in active phase over India, generally rainfall is suppressed over western Pacific Ocean. Fig. 5 gives a schematic diagram showing the features associated with the formation of this newly discovered Jet stream LLJ*.

3. Warm and cold pools of the Indian Ocean

In January, the centre of the warm pool of the tropical oceans lies over southwest Pacific Ocean. In the annual cycle, the central region of the warm pool gets shifted to the north Indian Ocean by May as may be seen from Figs. 6(a&b), when north Indian ocean becomes the warmest ocean globally (center of the warm pool) - Joseph (1990 a,b) and Joseph (2014) . This large-scale change in the ocean SST is found necessary for monsoon onset over India (Kerala). The warm pool of May attracts moisture convergence and over a period of about a month builds up the vertically integrated moisture content of the atmosphere over south Asia and the oceans around to about 45 kilograms per square metre which is needed for monsoon onset to take place over India (Kerala) - Pearce and Mohanty (1984) and Joseph *et al*. (2006). In the annual cycle the centre of the warm pool shifts from western Pacific Ocean south of the equator (December to March), northwestwards to north Indian ocean in may, eastwards to west Pacific ocean north of the equator in July to October and back to its December-March position. It will be interesting to study the interannual variability of the Warm Pool's month to month movement in the annual cycle and its global impacts.

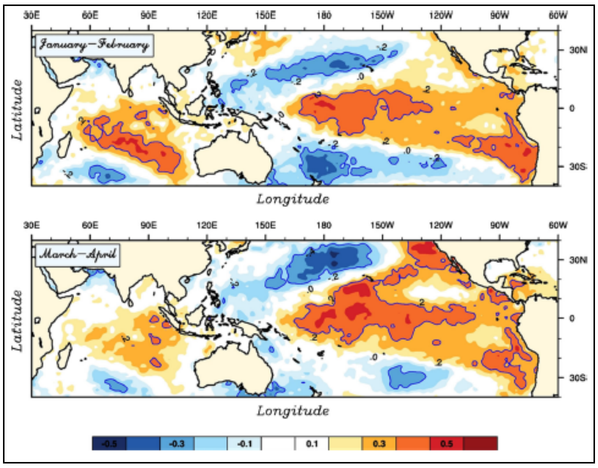


Fig. 8. Linear correlation coefficient between HadISST and IMD's objective dates of MOK from 1971 to 2014 for January-February (above) and March-April (below). The significant values are marked by thick blue line. Linear correlation coefficient is 0.3 for a significance level of 95% by a t-test and r = 0.3 is marked by a blue line. (Fig. taken from Preenu *et al*., 2017)

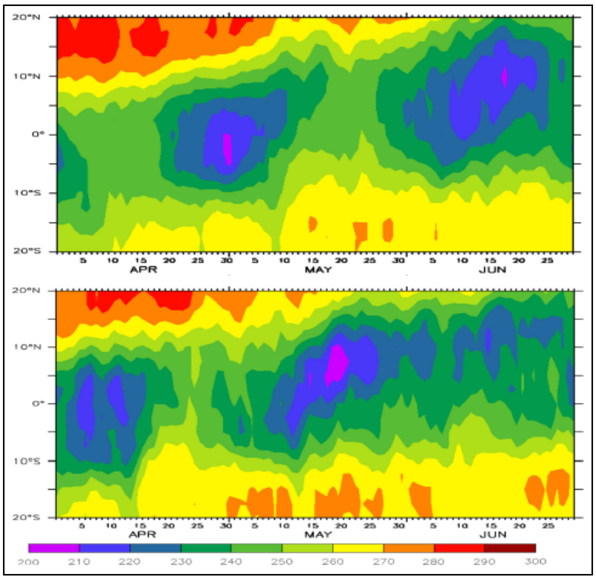


Fig. 9. Hovmoller diagrams of the daily OLR (Wm^{-2}) averaged between longitudes 30° E and 120° E (Indian Ocean) as composites of delayed MOK (above) and early MOK (below). [Fig. taken from Preenu *et al*. (2017)]

Using TMI SST Joseph *et al.* (2005) reported the existence of a large tongue of cold water in the Bay of Bengal during the monsoon season in the latitude belt 3° N to 10° N. It is in between two warm pools one around the equator and the other in the North Bay of Bengal. Composite monthly SST of the eight years 1998-2005 of June, July, August and September are shown in Fig. 7. In the latitude belt 3° N to 10° N a narrow tongue of low SST is seen in June, which expands in length (east -west) and width (north-south) during the following July and August. They named this the 'Cold Pool' of the Bay of Bengal. From June to August the temperature of the Cold Pool falls They suggested that the cold pool could be due to the spreading eastwards of the cold up welled water off Kerala and Sri Lanka coasts by the prevailing ocean currents into the Bay of Bengal. That suitable currents exist there during the monsoon season has been shown by Vinayachandran *et al.* (1999). Open Ocean upwelling during break monsoon spells may also contribute as shown by George *et al*. (2017). Oceanographic aspects of this Cold Pool have been discussed by Rao *et al.* (2006 a,b) renaming it as the "mini cold pool off the southern tip of India". The Cold Pool of the Bay of Bengal is important in the ocean - atmosphere interactions that maintain the active - break cycle of the monsoon - Joseph and Sabin (2008).This aspect will be discussed in a later section

4. Summer monsoon onset over India & Australia

The mean date of monsoon onset over Kerala at the southern tip of India is 1 June with a standard deviation of

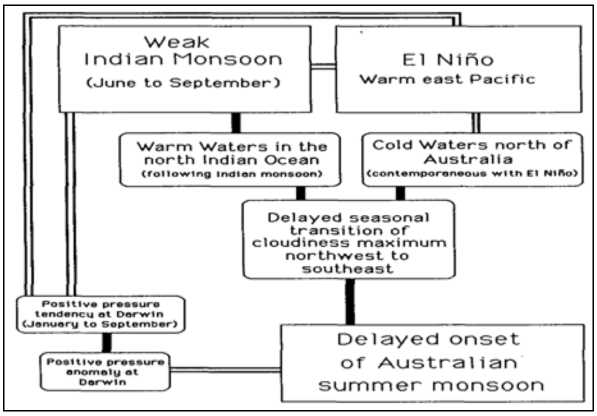
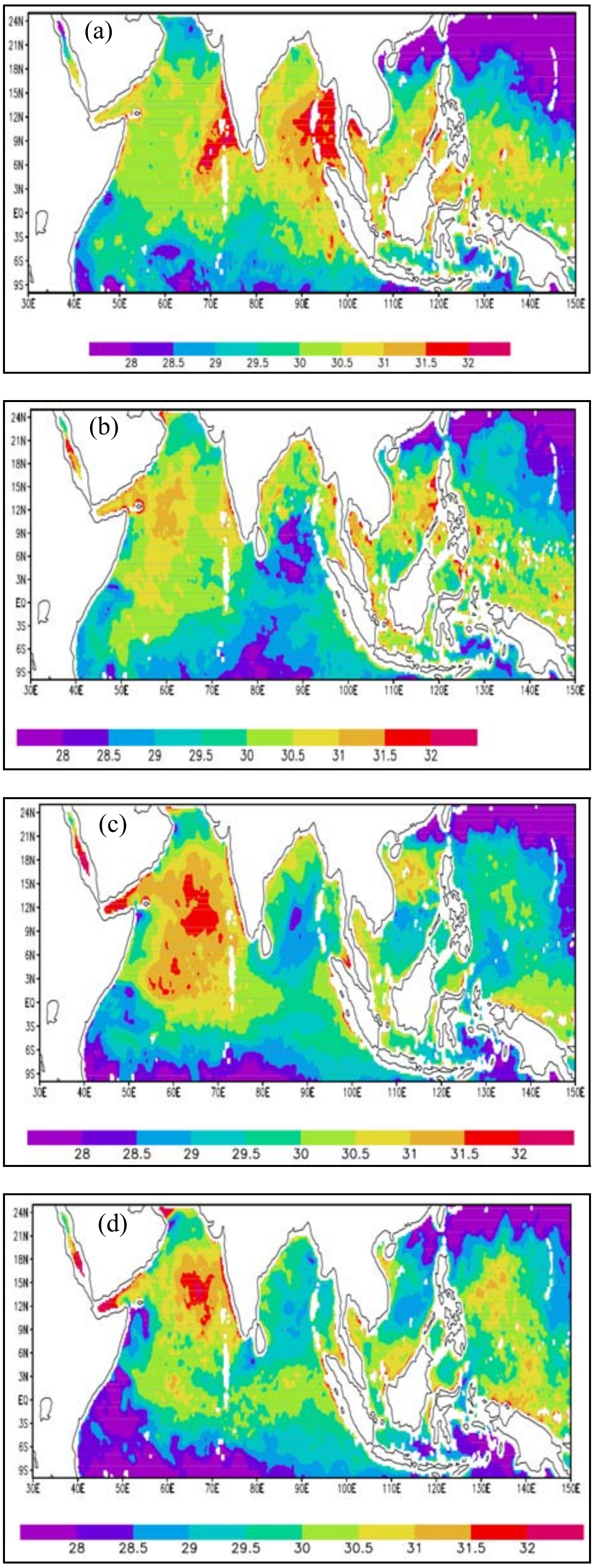


Fig. 10. Schematic of hypothesized sequence of events leading to a delayed monsoon onset over Australia. Events that have a causal connection are connected by solid lines. Events that are simultaneous with no causal mechanism implied are connected by double lines (taken from Joseph *et al*., 1991)

about 8 days. The earliest and the most delayed dates of monsoon onset over Kerala (MOK) have been 11 May and 18 June respectively according to records of India Meteorological department which uses a wind and rainfall criteria to determine objectively the date of MOK - Pai and Rajeevan (2009). The mean date of Monsoon Onset over Australia (MOA) was shown as 25 December with a standard deviation of 16 days by Hendon and Liebmann (1990) who also used wind and rainfall criteria. Joseph *et al.* (1994) and on similar lines Preenu *et al.* (2017) using more recent data, studied the interannual variability (IAV) of the date of MOK. Linear correlation coefficient



Figs. 11(a-d). TMI SST in °C averaged over pentads (a) P-7, (b) P-5, (c) P-2 and (d) P0 with respect to MOK which was on 8 June 2003 (P0). A pentad is 5 days

between HadISST and IMD's objective dates of MOK (Pai and Rajeevan, 2009) from 1971 to 2014 for January - February and March - April are given in Fig. 8 which shows that delayed MOK is associated with warm SST anomalies south of the equator and cold SST anomalies north of the equator over Indian and Pacific oceans. This SST signal has persistence of nearly six months prior to MOK and the anomaly pattern is similar to that of an El Nino. Joseph *et al*. (1991) studied the IAV of the date of MOA. Delayed MOA is associated with warm SST anomalies north of the equator particularly in the Indian Ocean and cold SST anomalies in the oceans around Australia during the September to November season prior to MOA and the SST anomaly pattern is similar to that soon after a drought Indian monsoon accompanied by El Nino.

Studies also found that dates of both MOK and MOA are related to the timing of the movement across the equator of the ITCZ cloud band, south to north for MOK in April-May and north to south for MOA in October-November. Delays in these equator crossing timings were associated with delays in the date of onset. Hovmoller diagrams of the daily OLR (Wm^{-2}) averaged between longitudes 30° E and 120° E (Indian Ocean) as composites of delayed MOK and early MOK are given in Fig. 9. It was found by Joseph *et al*. (1994) that delayed MOK occurred in El Nino (0) years and more often in El Nino (+1) years. Joseph *et al*. (1991) found that monsoon rainfall of India (ISMR) was negatively correlated to date of MOA in the following winter. Schematic of the hypothesized sequence of events leading to a delayed monsoon onset over Australia is given in Fig. 10. The mechanisms for delay in onset are thus the same for MOK and MOA. Many El Nino years are known to be years of drought in ISMR. Thus El Ninos are associated with delays in dates of both MOK and MOA.

There is a fine structure in the development of the warm pool of the north Indian Ocean associated with MOK which is illustrated for the year 2003 in Figs. 11(a-d). Seven pentads or 35 days before MOK (P-7) the Bay of Bengal SST is very warm while Arabian Sea at the same latitude is much colder [Fig. 11(a)]. In the SST gradient area to the south of the warm pool of Bay of Bengal a convective cloud band forms which generates a low level wind flow through Kerala, accelerating it to a LLJ crossing the equator in the longitude belt 60-70° E. This rain/wind band which gives monsoon onset like conditions in Kerala was named by Joseph and Pillai (1988) as the Pre Monsoon Rain Peak (PMRP) and by Flatau *et al*. (2001) as Bogus Monsoon Onset. This wind and cloud band moving north across Bay of Bengal cools the SST there by P-5 [Fig. 11(b)]. Arabian Sea with no cloud and light wind condition warms fast and at P-2

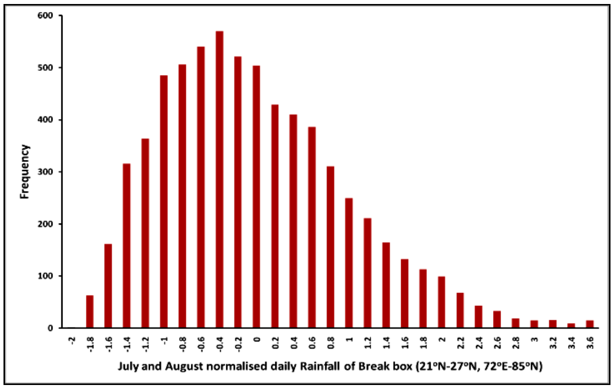


Fig. 12(a). Frequency distribution of the daily normalised rainfall anomaly of the break box (bounded by latitudes 21° N and 27° N and longitudes 72° E and 85° E) at intervals of 0.2 for the July and August months of the whole period 1901-2009

Year	El Nino index	Break days	Monsoon rain (mm)	Departure (percent)
1902	2.1	16	792	-9
1905	1.6	17	718	-17
1918	1.9	24	736	-25
1925	2.0	8	906	-3
1930	2.2	11	876	-5
1940	1.6	1	905	-3
1957	1.8	7	898	-2
1965	2.0	23	738	-18
1972	2.5	19	697	-24
1982	3.3	10	767	-15
1986	1.7	16	770	-13
1987	1.6	19	737	-19
1991	2.1	10	828	-9
1994	1.6	2	1001	+13
1997	3.4	5	927	+2
2002	1.7	24	737	-19
2009	CPC 1.9	25	698	-22

Fig. 12(b). The June to September CPC index of strong El Ninos (as defined in the paper), June to September Indian Summer Monsoon Rainfall (ISMR) and its anomaly and the number of Break monsoon days of July and August of the period 1901 to 2009 are given in the table

develops a warm pool of high SST [Fig. 11(c)]. In the SST gradient area to its south the MOK cloud and wind band (the real LLJ) forms which crosses the equator near longitude 40° E. Intensification of this LLJ brings about MOK at P-0 [Fig. 11(d)]. Date of PMRP is one of the strong predictors for the forecast of the date of MOK a few weeks ahead - Joseph and Pillai (1988). Studying the spatial evolution of wind and OLR (rainfall) over Indian Ocean, an objective method for the determination of the date of MOK was developed by Joseph *et al.* (2006) which were adapted for operational use in India Meteorological Department - Pai and Rajeevan (2009). This method is able to filter out bogus monsoon onsets.

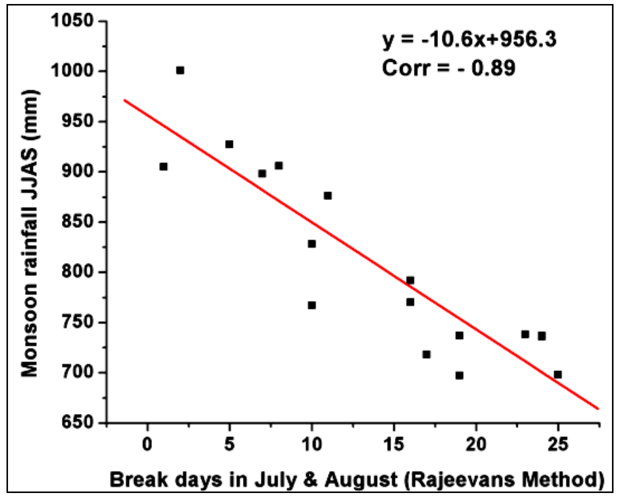


Fig. 12(c). Scatter diagram between Indian Summer Monsoon Rainfall (June to September) and the number of Break monsoon days of July and August for each of the 17 strong El Nino years. The pair has a high linear correlation of 0.89

5. Active break cycle of Asian summer monsoon

Drastic reductions in the monsoon rainfall in major parts of India, particularly in its central and northwest parts, are considered to be the most important feature of the phenomenon "break monsoon". Traditionally breaks have been identified on the basis of the surface pressure distribution and low level wind patterns over the Indian region. For this review breaks are defined based on rainfall following the method of Rajeevan *et al*. (2006). The area mean daily rainfall anomaly (normalised with its standard deviation) of the part of India ("Break box") bounded by latitudes 21° N and 27° N and longitudes 72° E and 85° E, has been derived from the 1°x1° gridded rainfall data of Rajeevan *et al*. (2006). These rainfall data have been used to study the break days in the peak monsoon months of July and August for the years 1901 to 2009 by Joseph *et al*. (2011). A break monsoon day is defined to have standardized negative rainfall anomaly of magnitude greater than 1. An active monsoon day is defined as one in which the standardized positive rainfall anomaly has magnitude greater than one. A review of active and break monsoon studies may be found in Gadgil and Joseph (2003).

Histogram of the daily normalised rainfall anomaly of the break box (at intervals of 0.2) for the July and August months of the whole period 1901-2009 showed that the anomaly varied from -2.0 to +3.6 (long tail to positive values) and skewed to the negative side with frequency maximum for anomaly of -0.4 [Fig. 12(a)]. The total number of break monsoon days in each July and August varied between 0 and 25 and active monsoon days between 1 and 22, respectively. The mean and standard

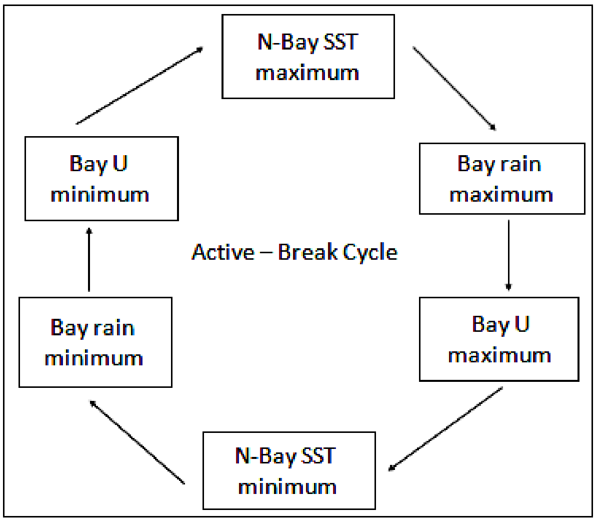


Fig. 13(a). Schematic diagram of the ocean-atmosphere interaction process in an AB cycle of july and august of a La Nina year

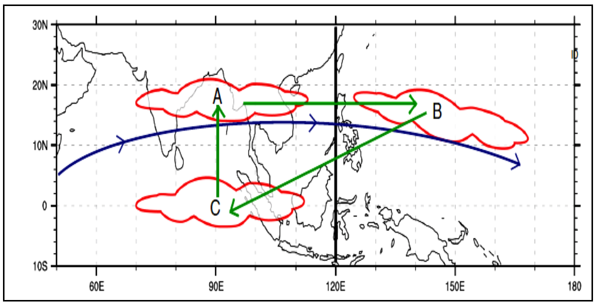


Fig. 13(b). Schematic diagram showing the location A, B and C of convection in the active–break cycle. In El Nino years, convection shifts from A to B and then to C and back to A, in La Nina years from A to C and back to A

deviation of the number of break days in July and August are 10.5 and 6.6 days, respectively. Similar numbers for active monsoon are 9.6 and 3.8, respectively. The linear correlation coefficient between the Indian Summer Monsoon Rainfall (ISMR) of June to September of the period 1901 to 2009 as derived by Parthasarathy *et al.* (1994) and the number of break (active) days in July and August is 0.63 (0.30).

It is well known that El Ninos are associated with monsoon rainfall deficiency in India and that a large percentage of the drought monsoon years are El Nino years. There were 17 strong El Ninos during the period 1901 to 2009 with the associated CPC index (the normalized SST anomaly of June to September of the equatorial Nino 3.4 region) greater than or equal to 1.5 which was taken as the criteria for a strong El Nino. The correlation between ISMR and the number of Break monsoon days (of July and August) of these 17 El Nino years [listed in the table in Fig. 12(b)] is shown by the scatter diagram in Fig. 12(c) (high linear correlation coefficient of 0.89). It is shown that the strong El Nino years 1925, 1940, 1957, 1994 and 1997, which had small number of Break monsoon days in July and August, were not associated with large deficiencies in ISMR. It could be inferred that large number of Break monsoon days in July and August is the most important cause of droughts in ISMR associated with El Nino.

Ocean and atmosphere are closely coupled in the Active-Break (AB) cycle of the Asian Summer Monsoon whose period varies between 30 and 60 days. Joseph and Sabin (2008) studied the AB cycle from an ocean - atmosphere interaction angle. AB cycle begins with maximum Sea Surface Temperature (SST) over north Bay of Bengal (BoB) when in the SST gradent area in central BoB a convective cloud band forms which generates a westerly LLJ through peninsular India. This is the Active phase of the monsoon. Convection and wind cool the SST and when the convection weakens, the LLJ shifts to a loation close to the equator - Joseph and Sijikumar (2004). A convective cloud band now forms close to the equator in the Indian ocean. This is the Break Phase of the monsoon. Clear skies and light winds that follow make the SST of north BoB which has a shallow Mixed Layer Depth (MLD) typically 20 metres reach a maximum and the next AB cycle begins. This description is typical for a La Nina year. Fig. 13(a) gives a schematic diagram of the ocean-atmosphere interaction process in such an AB cycle. The period of the AB cycle is about 30 days in this case.

The eastward extent of LLJ in the monsoon season is only upto longitude 120° E in a La Nina year. LLJ extends further east in west Pacific ocean upto longitude 180° E in El Nino years. Even during the pre-monsoon season of the El Nino year the area between 120° E and 180° E has westerly wind anomalies in low levels. In these years the positive wind stress curl north of the LLJ axis is hypothesised to generate a large area of shallow MLD between longitudes 120° E and 180° E in the west Pacific ocean [Area - B in Fig. 13(b)] - Joseph (2014). Estimation of MLD using available ARGO data support this hypothesis. During the AB cycle Area-B warms fast under the clear sky conditions there and causes the Active phase convection to shift there from BoB (Area-A) and when Area-B cools convection moves to the equatorial Indian ocean (Area-C) as shown by the arrows in Fig. 13(b). The period of the AB cycle is typically 50-60 days with a long Break monsoon spell as defined earlier. In a La Nina year convection moves from Area-A to Area-C and back to Area-A and period of the AB cycle is close to 30 days.

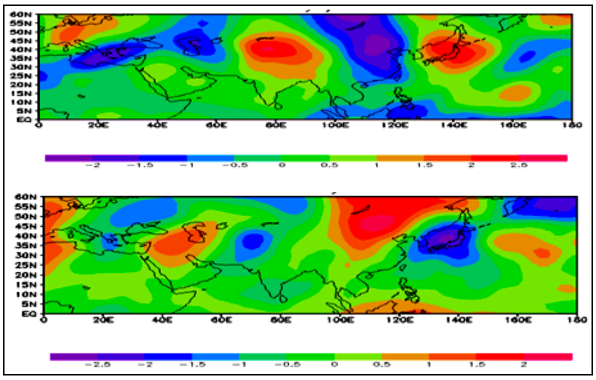


Fig. 14(a). Composite of Asia Pacific Wave in 200 hPa meridional wind anomaly of the monsoon season June to September of the 5 Dry years 1965, 1972, 1979, 1982 and 1987 (top) and 5 Wet years 1961, 1970, 1975, 1983 and 1988 (bottom) of the period 1961-1990

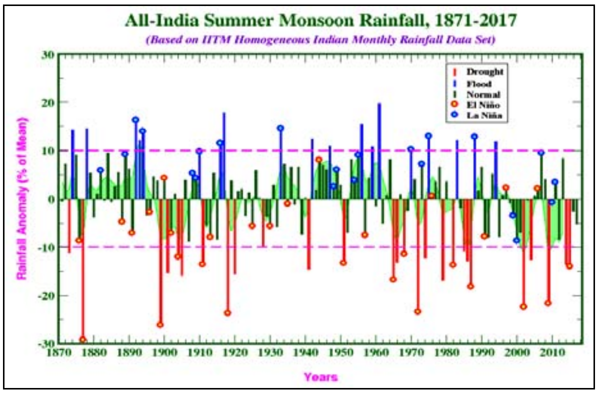


Fig. 15. Time series of ISMR for the period 1871-2017 taken from the website of Indian Institute of Tropical Meteorology www.tropmet.res.in. Years of El Nino and La Nina are marked

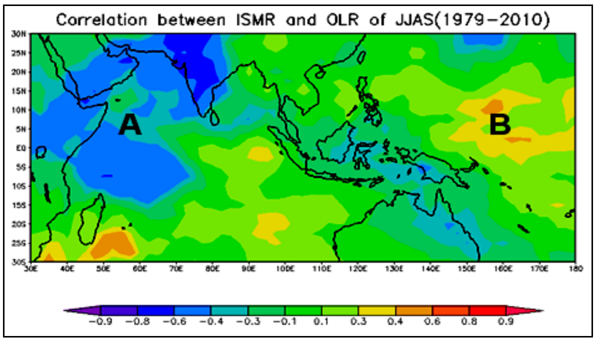


Fig. 14(b). correlation between ISMR and mean olr of the period 1979-2010 showing the two poles of the large convective heating anomaly, A in Wet monsoons and B in Dry monsoons

6. Asia pacific wave and instability in the ocean – monsoon system

It was shown by Joseph (1978a,b) and that Indian summer monsoon droughts were associated with the intrusion into the tropics of south Asia of the middle latitude upper tropospheric westerlies. These westerly wind intrusions were found to have persistence of several months prior to the monsoon season. Prior to a deficient ISMR season, meridional component of upper tropospheric winds were strong southerlies over northwest India in May which showed high and statistically significant correlation with the following ISMR. Joseph *et al.* (1981) suggested the use of this factor in the statistical long range prediction of monsoon rainfall of India. It was shown by Joseph and Srinivasan (1999) that these westerly wind intrusions were part of a stationary Rossby wave train. This wave was named Asia Pacific Wave (APW) by them. APW of May was observed also in the following summer monsoon season with large amplitude but with the same phase spatially as in May. In excess ISMR months of May to September also APW was forced on the mid-latitude westerlies and the sub-tropical Jet stream but its spatial phase was almost opposite, half a wavelength displaced in the east-west direction. APW was shown to affect the ozone distribution in the upper troposphere and lower stratosphere during the monsoon season [Sathiyamoorthy *et al*. (2002)].

ISMR has large interannual variability. Generally a drought monsoon year is followed by a normal or excess monsoon year, a sort of biennial oscillation named Tropospheric Biennial Oscillation (TBO) details of which may be found in the observational study by Meehl (1997) and modelling study by Chang and Li (2000). Joseph (1981) and Joseph and Pillai (1984, 1986) had earlier shown TBO like interactions between the Indian Ocean and the Indian Summer Monsoon. During the 30 year period 1961 to 1990 interannual variability of ISMR was large and TBO was seen prominently in the ISMR time series, but the period of the oscillation was 2 to 4 years, with the occurrence of two consecutive Dry or Wet years occasionally. The wettest monsoons of this period were years A (1961, 1970, 1975, 1983 & 1988) and the driest were years B (1965, 1972, 1979, 1982 & 1987). Fig. 14(a) gives the composites of meridional wind anomalies at 200 hPa averaged for the monsoon season June to September of these Wet (A) and Dry (B) years. APW and their phase shift can be seen in this figure. Fig. 14(b) gives the correlation between ISMR and mean OLR (proxy for rainfall or convective heating of the atmosphere) of the period 1979 - 2010 (a different period was used as 1961-1990 had many years without OLR measurements) showing the two poles of high convective heating anomaly, A in Wet monsoons and B in Dry monsoons. Joseph and Srinivasan (1999) had hypothesised that the longitudinal difference in the location of the heat sources

A and B is the likely cause of the phase shift of APW between Wet and Dry monsoons.

Indian summer monsoon rainfall had alternating three decade long DRY and WET epochs during the 120 years 1870 to 1990, the 30 year epochs 1870-1900 & 1930-1960 were WET and 1900-1930 & 1960-1990 were DRY. Fig. 14(b) gives the time series of ISMR for the period 1871-2017 taken from the website of Indian Institute of Tropical Meteorology www.tropmet.res.in. Standard Deviation of ISMR is about its 10%. A year with ISMR 10% less is considered a Dry year and year with ISMR in excess by 10% a Wet year. The DRY epochs had frequent drought (Dry) monsoons. A high percentage of severe cyclones of the Bay of Bengal moved northwards in the DRY epochs during 1891 to 1990, the period for which we had good cyclone track information, causing disasters in Bangladesh, Myanmar and in Indian states of Orissa and West Bengal. In the Bay of Bengal north moving cyclones have longer life time over the sea surface compared to the west moving ones and hence reach high intensities, at times becoming super-cyclones. Epochal nature of monsoon rainfall and Bay of Bengal cyclone tracks were first pointed out by Joseph (1976).

The DRY epochs are associated with the cold phase of the Atlantic Multidecadal Oscillation in SST, which are also epochs of high SST gradient between the tropics and northern hemi-sphere mid-latitudes of the global oceans - Joseph *et al*. (2013). Extrapolating this natural cyclicity of the ocean - atmosphere system of period about 60 years into the future, the 3 decades 2020 to 2050 are likely to have frequent disastrous droughts in the Indian Summer Monsoon Rainfall and a large percentage of the severe cyclones of the Bay of Bengal are likely to have northward tracks adversely affecting the coastal region of north Bay of Bengal.

Analysing the available tropospheric temperature (NCEP/NCAR re-analysis - Kalnay *et al.*, 1996) data, Joseph *et al*. (2016) found that the recent. DRY epoch 1960-1990 which had ten monsoon drought (Dry) years and many north moving severe tropical cyclones in the Bay of Bengal was found to have cold 300 hPa temperature anomaly over central Asia. This region of central Asia has also experienced a cooling trend from 1950 to date. It is feared that this cooling trend over the Asian continent is likely to make the expected DRY epoch 2020 to 2050 more severe in its impact than the recent DRY epoch 1960 to 1990. The change in upper tropo-spheric temperature at 300 hPa of the summer monsoon season (June to September), decades (2000-2009) minus (1950-1959) is given in Fig. 16(a). Between latitudes 30 and 60 degrees there is a cold temperature anomaly over

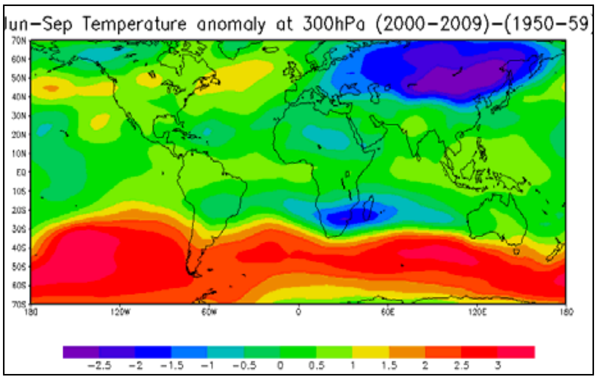


Fig. 16(a). June to September 300 hPa temperature difference between decades (2000-2009) minus (1950-1959) in °C

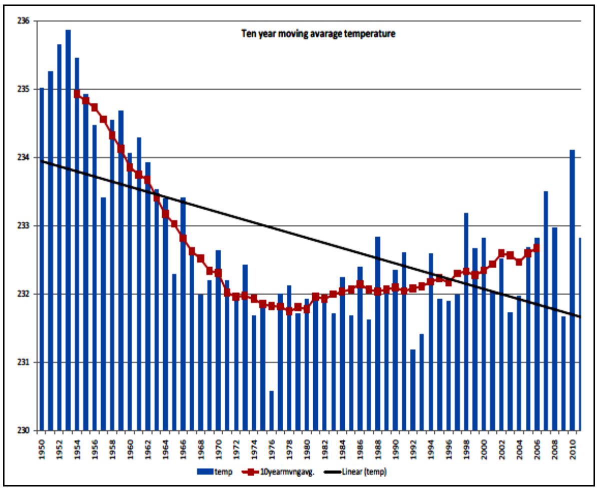


Fig. 16(b). Bars show June-September 300 hPa mean temperature of each year averaged over the box 40° N-60° N;70° E-140° E. Red line shows ten year moving average; linear trend is marked by blue line

the Asian continent while the rest of the globe had warm anomalies in the same latitude belt in both north and south hemispheres. Fig. 16(b) gives the monsoon season mean temperature at 300 hPa over a box bounded by latitudes 40° N and 60° N and longitudes 70° E and 140° E (the core region of the cold anomaly) for each year of the period 1950 - 2011. The 10-year moving average shows the coldness (decadal variation) of the DRY epoch 1960-1990. There is a prominent linear cooling trend also. Joseph *et al*. (2016) has given a hypothesis on the instability (negative feedback) of the ocean - atmosphere - ISMR - El Nino system of the Indo-Pacific ocean basin leading to the occurrence of biennial oscillation of ISMR and Bay cyclone tracks and frequent droughts in ISMR when the upper troposphere over central Asia has a cold temperature anomaly. A schematic diagram showing the details of this instability is given in Fig. 17.

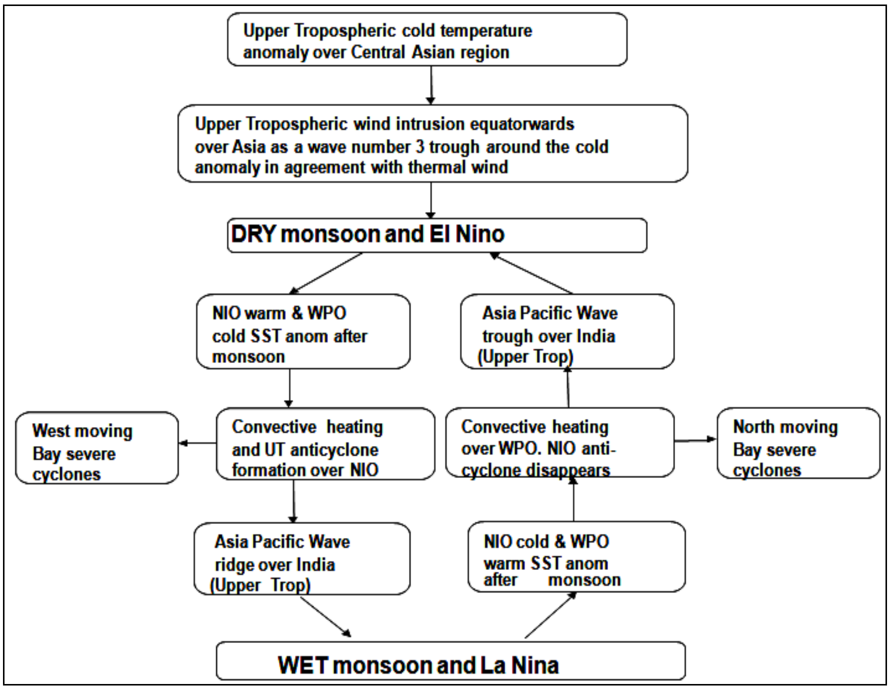
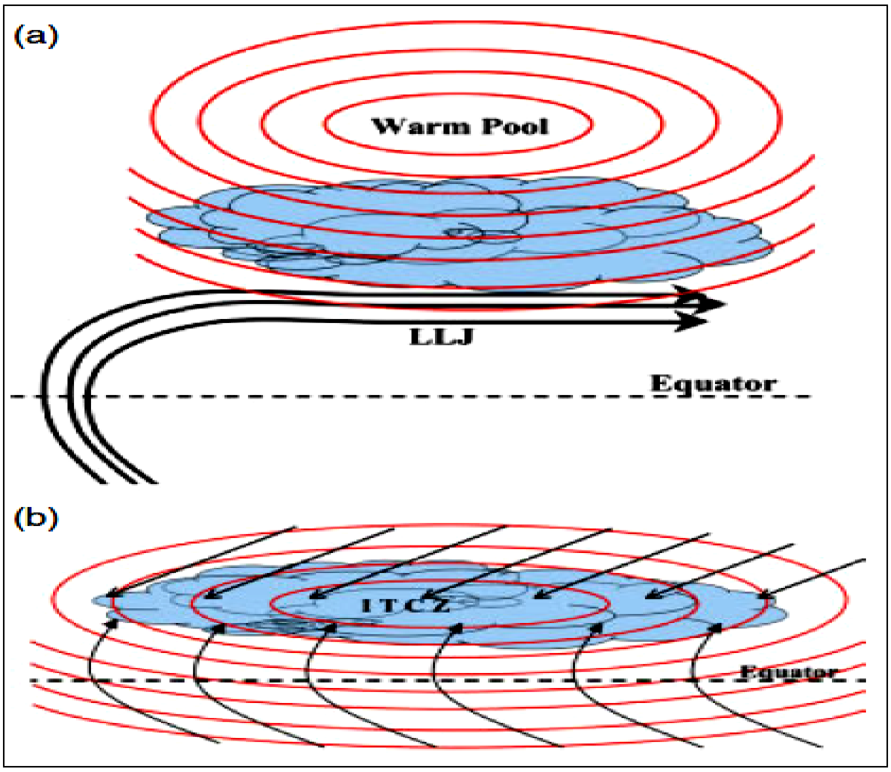


Fig. 17. Schematic diagram showing the instability of the ocean - atmosphere system (negative feedback) that lead to frequent droughts in ISMR and the tropospheric biennial oscillation in epochs like 1960-1990



Figs. 18(a&b). Schematic diagrams showing the isotherms of SST (lines in red colour), streamlines of wind flow and LLJ and the areas of convection (blue shading) for (a) warm pool and (b) ITCZ cases

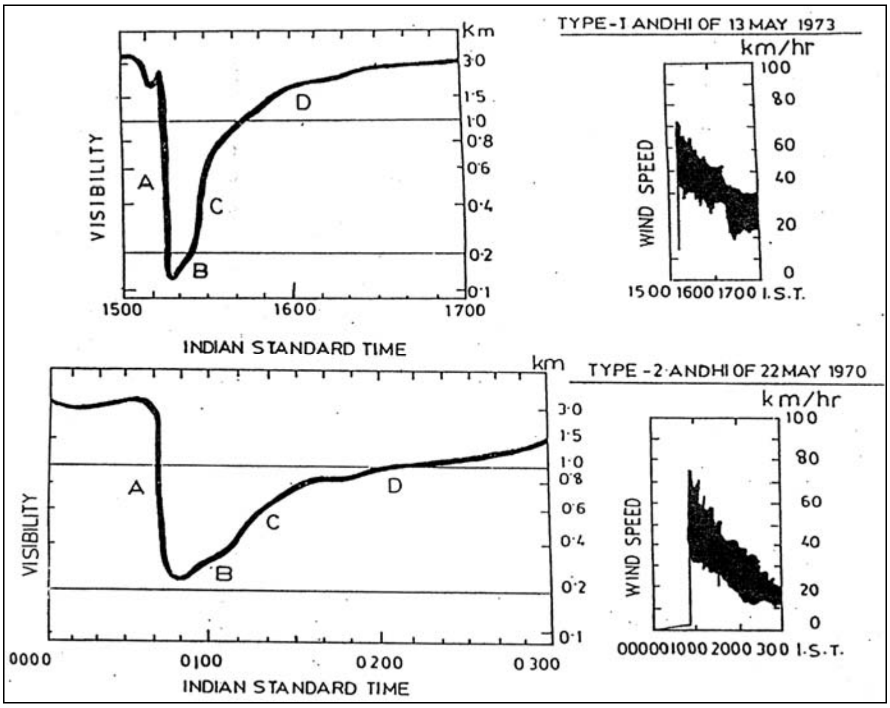


Fig. 19. Horizontal visibility variations (left) and wind speed variations (right) near the ground associated with convective dust storm (Andhi) at Delhi airport during an afternoon / evening case (top) and a case during the night hours (bottom)

7. SST-Convection relation in the tropics

According to Gadgil *et al*. (1984) organised convection over the tropical Indian Ocean occurred when SSTs exceeded a threshold or critical value, but once the threshold is crossed, the intensity of convection was no longer dependant on the SST. Graham and Barnett (1987) while confirming this finding for the Atlantic and Pacific ocean basins also found that when SSTs are above the critical value, surface wind divergence was closely associated with convection. Studies by Waliser *et al*. (1993) showed that convection increased with SST only in a narrow range of SST 26 °C to 29 °C. At SSTs above 29 °C, convection was found to remain nearly steady upto 30 °C and for higher SSTs sharply decreased with increase of SST.

Research by Sabin *et al*. (2012) showed that it is only over the summer warm pool areas of Indian and west Pacific oceans that the Waliser type of SST - Convection relation exists. This has modelling support - Lindzen and Nigam (1987) and Back and Bretherton (2009). Regions of SST maxima have low SST gradients and therefore feeble convection. Convection initiated by SST gradient produces strong wind fields particularly cross-equatorial low-level jet streams (LLJs) on the equator-ward side of the warm pool and both the convection and LLJ grow through a positive feedback process. Thus, large values of convection are associated with the cyclonic vorticity of the LLJ in the atmospheric boundary layer. In the inter-tropical convergence zone (ITCZ) over the east Pacific Ocean low-level winds from north and south hemisphere converge in the zone of maximum SST, which lies close to the equator producing there elongated bands of deep convection, where we find that convection increases with SST for the full range of SSTs unlike in the warm pool regions.

The characteristics of the summer monsoon climate over south Asia, East Asia and the western North Pacific Ocean (warm pool / monsoon areas) were studied using the Meteorological Research Institute-coupled GCM (MRI-CGCM2) by Rajendran *et al*. (2004). The relationship between SST and convection in the model outputs as obtained by them is of Waliser type. SST - convection relation has been reproduced satisfactorily in their modelling study over the entire range of SSTs, including the reduction in convection for high SSTs.

The low-level wind divergence computed using QuikSCAT winds has large and significant linear correlation with convection in both the warm pool and ITCZ areas. But the linear correlation between SST and convection is large only for the ITCZ. These findings have important implications for the modelling of convective rainfall over the tropical oceans. Schematic diagrams showing the isotherms of SST, streamlines of wind flow and LLJ and the areas of convection are given in Figs. 18(a&b) for the warm pool and ITCZ cases. Zone of maximum SST is far away from the equator in the warm pool case and close to the equator in the ITCZ case.

Shankar *et al*. (2007) have shown that convection sets over the Bay of Bengal during the summer monsoon season within a week after the SST difference between north and south Bay of Bengal exceeds 0.75 °C (large SST gradient). Once deep convection is initiated in the SST gradient area south of the central region of the warm pool, the deep tropospheric heating by the latent heat released in the convective clouds produces strong low-level wind fields particularly cross-equatorial LLJ on the equatorward side of the warm pool and both the convection and LLJ are found to grow through a positive feedback mechanism (Joseph and Sijikumar, 2004; Joseph and Sabin, 2008).

8. ANDHI - the convective dust storm

During the months of April to June, dust storms occur frequently over northwest India. They are locally called Andhi. When dust raised by strong wind reduces horizontal visibility to less than 1000 metres we call the phenomena a dust storm. Joseph *et al*. (1980) made a study of 40 cases of Andhi that occurred at the Delhi airport during the period 1973 to 1977, using a transmissometer (to measure the variation of horizontal visibility as the dust wall moved across the airport), a weather radar (to study the movement of the associated thunderstorm cloud) and wind, pressure, temperature and humidity measuring instruments. From the nature of variations of horizontal visibility and wind speed near the ground level associated with these dust storms, it was found that mainly two types of Andhi occur, the first (type-1) occurring in the afternoon / evening when ground is hot and the second (type-2) occurring at night when the air in contact with the ground has cooled.

In strong dust storms at Delhi airport it is observed that horizontal visibility is reduced to less than 100 metres. Fig. 19 gives an example each of visibility variations in a type-1 and type-2 Andhi taken from Joseph *et al*. (1980). The figure also shows the corresponding surface wind variations in the accompanying squall (cold air downdraft from the accompanying thunderstorm). The leading edge of the downdraft air is the gust-front whose speed of movement on the ground depends on the density difference between the downdraft air and the ambient air. Joseph (1982) applied this knowledge to explain the slow improvement of visibility in Andhi occurring at night. The dust wall of the dust storm sloping backwards is at the leading edge of the downdraft air spreading on the ground and along with the gust front this wall of dust moves over the ground slower at night than in the afternoon.

9. Conclusions

In this article I have summarised the main research works done by me, alone or in collaboration with colleagues and Ph. D students, during the last six decades. Isolated items of work done on climate change in jet streams (Low Level, Tropical Easterly and Sub Tropical), monsoon depressions, relation between monsoon and Indian Ocean Dipole and that between MJO and genesis of north Indian Ocean tropical cyclones have not been included in this article. My research was in Diagnostic Meteorology, some on global tropics, but mostly on the Asian tropics particularly on the Asian Summer Monsoon. Both the atmosphere and the oceans were studied using the data sets available. Meteorological satellite data was extensively used. I travelled in time from the age of hand calculators and analysis of synoptic charts using pencil and eraser to the age of personal computers and software like Grads to analyse and digest vast amounts of data. I thank all persons and institutions who supported me in my research efforts. Two decades of operational meteorological work done at decision making levels got me first hand experience on the atmosphere, oceans and weather systems and the two decades of teaching tropical meteorology helped me to get my concepts clear. In this article I have copied and pasted portions from many research papers mostly from the ones where I am the author or one of the co-authors. This may not be considered as plagiarism.

Acknowledgement

The contents and views expressed in this research paper are the views of the authors and do not necessarily reflect the views of their organizations.

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