Research Article **Open Access**

Inter-annual variability of Summer Monsoon Rainfall over Myanmar in relation to IOD and ENSO

Zin Mie Mie Sein[1,2], Bob Alex Ogwang[1,3], Victor Ongoma[1,4,*], Faustin Katchele Ogou[1], Kpaikpai Batebana[1]

[1]Nanjing University of Information Science and Technology, Nanjing, P.R. China
[2]Department of Meteorology and Hydrology, Myanmar
[3]Uganda National Meteorological Authority, Kampala, Uganda
[4]South Eastern Kenya University, Kitui, Kenya

Article History
Received
April 30, 2015

Published Online
June 17, 2015

Keywords:
El Nino,
ENSO,
EOF,
IOD,
La Nina,
Myanmar,
Rainfall

Abstract: Monsoon rainfall has significant impact on socio-economic welfare of Myanmar. This calls for need to understand rainfall variability in order to minimize socio-economic losses associated with extreme weather events. This study investigates the inter-annual variability of summer monsoon rainfall over Myanmar in relation to Indian Ocean Dipole (IOD) and El Niño-Southern Oscillation (ENSO).The dominant modes of variability of rainfall were identified by performing Empirical Orthogonal Function (EOF) analysis, using rainfall data from Global Precipitation Climatology Centre (GPCC) spanning from 1971 to 2010.The rainfall over Myanmar is negatively correlated with the positive phase of the IOD and positively correlated to the positive phase of the Southern Oscillation Index (SOI),implying that El Nino events result into drought events in the region, while La Nina events result into floods. The year 2001 was identified to be the wettest while 1972 is the driest period of study. The influence of SOI on rainfall over Myanmar is more than that of IOD. The moisture convergence characterizes circulation in wet years, while divergence is dominant during dry years. The findings of this study are important in improving the accuracy of seasonal forecast over Myanmar.

***Corresponding authors:** Victor Ongoma; Email: victor.ongoma@gmail.com

Cite this article as: Sein, Z.M.M., B.A. Ogwang, V. Ongoma, F.K. Ogou and K. Batebana. 2015. **Inter-annual variability of summer monsoon rainfall over Myanmar in relation to IOD znd ENSO**. *Journal of Environmental & Agricultural Sciences*. 4:28-36.

1. Introduction

Large-scale air-sea interactions have a significant role in the generation of inter-annual weather variability in the tropical region. Myanmar is located in the Southeast Asia; latitudes 9˚32'N - 28º31'N and longitudes 92˚10'E - 101˚11'E (Fig. 1). The predominant climate system over the country is the Asian summer monsoon. According to Parthasarathy et al. (1994) and Gadgil et al. (2004), the Asia summer monsoon rainfall accounts for almost 90% of the observed rainfall over Myanmar, and supports more than a half of the human population globally. The economy of Myanmar is mainly dependent on rain fed agriculture; the country stands out as one the world's leading rice producing. The country's socio-economic welfare is thus adversely affected by both spatial and temporal variability of monsoon rainfall, calling for better understanding of summer monsoon over the country.

Studies (e.g. Qian and Lee, 2000; Sen Roy and Kaur, 2000; Lwin, 2000) show that Myanmar experiences five distinct weather seasons: pre-monsoon season is observed in mid-April to mid-May, the main monsoon season is experienced in mid-May to mid-October while the post monsoon season is reported from mid-October to end-November. The studies have observed that the country experiences a dry and cool season towards the end of November to mid of March, a hot season is observed from mid-March to mid-April.

A relatively large section of the South Asian monsoon variability is known to be associated with El Nino Southern Oscillation (ENSO) (Ju and Slingo, 1995; Shukla and Paolino, 1983). During an El Nino phenomenon, there is an observed increase in Sea Surface Temperature (SST) in the central-eastern equatorial Pacific, shifting the deep convection center from the Maritime continent eastward into the central Pacific. The process is associated with a reduction in rainfall over the western Pacific and South Asia (Latif et al., 2001). Shukla (1987) attributed the tendency of negative correlation between South Asian monsoon rainfall with the eastern Pacific SST to El Nino process. In Thailand; a neighboring country to Myanmar, a recent study has shown that both Indian Ocean Dipole (IOD) and ENSO are dominant climate modes that impact its rainfall. However, the effect of the IOD and ENSO, and the rainfall interaction is still unclear since it varies from region to region (Muangsong et al., 2014; Cherchi and Navara, 2013). In Myanmar, Sen Roy and Kaur (2000) observed an estimated 10% below average rainfall during El Nino years over the country.

There has been a lot of interest on the strongest coupled system of the ENSO, which exerts significant impacts on climate extending to the global extent (Philander et al., 1990). Researchers (e.g. Webster and Yang, 1992; Chen and Yen, 1994; Yang and Lau, 1998; Lau and Nath, 2000; Wang et al., 2001) have observed that ENSO can influence Indian summer monsoon directly through large-scale circulation changes over the Indo-western Pacific. The influences of ENSO on the rainfall varies, it can either suppress or enhance Indian summer monsoon. The indirect influence leads to an anomalous Indian summer monsoon by modulating the north Indian Ocean SST (Kawamura et al., 2001; Wu and Kirtman, 2004), the meridional thermal difference over South Asia (Yang and Lau, 1998; Kawamura, 1998), and the atmospheric moisture content over the tropical Indian Ocean (Ashok et al., 2004; Wu and Kirtman, 2004).

The IOD has a strong influence on the Indian Ocean and on the larger neighboring spatial scale. According to Saji et al. (1999) and Webster et al. (1998), the IOD index is calculated by considering SST difference between the western (60°-80°E, 10°S-10°N) and eastern (90°E-110°E, 10°S-0°N) of the tropical Indian Ocean. Studies (e.g. Yamagata et al., 2002; Ashok et al., 2003; Saji and Yamagata, 2003) have shown that IOD is an inherent mode of the tropical Indian Ocean. The IOD is one feature of the general cycle of global climate, interacting with related phenomena like the ENSO; a common occurrence in the Pacific Ocean. The IOD – ENSO interaction involves the forcing of the remote atmospheric circulation by SST anomalies in the tropical Pacific, and the ensuing response of the remote Indian Ocean to such atmospheric changes (Venzke et al., 2000; Lau and Nath, 2003). A recent study by Cherchi and Navara (2013) has shown that some positive IOD events occur simultaneously with an El Nino event in the Indo-Pacific ocean sector or over India, others happen either in the absence of El Nino or in some cases even with a La Nina phenomena. Previous studies noted a reasonable impact on the Asian monsoon variability from local climate variability over the tropical Indian Ocean (Ashok et al., 2001; Li et al., 2003). The SST gradient during a positive IOD phase forces and extreme southerly monsoon wind over the northern Indian Ocean, which flows into the South Asian continent, enhancing rainfall there (Guan et al., 2003; Li et al., 2003; Ashok et al., 2004). This is linked to a positive correlation between the South Asian rainfall and IOD (Webster et al., 1999; Ashok et al., 2001; Wu and Kirtman, 2004). These studies have suggested that the South Asian monsoon can be affected significantly by both ENSO and IOD (Ashok et al., 2004). Observational studies show that substantial IOD events can be triggered by ENSO (Allan et al., 2001; Baquero-Bernal et al., 2002). Contrary, there are also evidences to support that a significant number of IOD events are generated locally in the Indian Ocean climate system, notwithstanding the remote Pacific ENSO (Ashok et al., 2003; Saji and Yamagata, 2003; Black et al., 2003; Lau and Nath, 2004). A recent research conducted on Indo-China monsoon indices showed that south Myanmar and Thailand are little influenced by tropical climate phenomena such as ENSO and IOD (Tsai et al. 2015).

Almost all of the previous studies observed and reported complex relationship between the South Asian monsoon and ENSO, and the IOD. The studies also focused on the entire south East Asia, in support, Sen Roy and Kaur (2000) and Sen Roy and Sen Roy (2011) noted that studies on monsoon climate dynamics over Myanmar are limited. This study narrows down on the possible effects of ENSO and IOD on monsoon rainfall over Myanmar. Ummenhofer et al. (2013) noted that little research has focused on the Indian Ocean in the context of inter-annual rainfall variability over the entire Asia continent. This study comes in handy to fill the gap; the understanding of the Inter-annual variability of May - October rainfall over Myanmar in relation to IOD and ENSO will help to improve the quality of the seasonal weather forecast in the country.

2. Data and Methodology

The study utilized both observed and reanalyzed rainfall data. Observed daily rainfall data was sourced from the Department of Meteorology and Hydrology of Myanmar, for 35 stations spanning from 1981 – 2010. This study only considered rainfall behavior in the months of May to October. The reanalyzed monthly precipitation data spanning from 1971 to 2010 was obtained from Global Precipitation Climatology Centre (GPCC), a centre that provides precipitation data sets from 1901 to present, calculated from the global station data (Schneider et al., 2013).

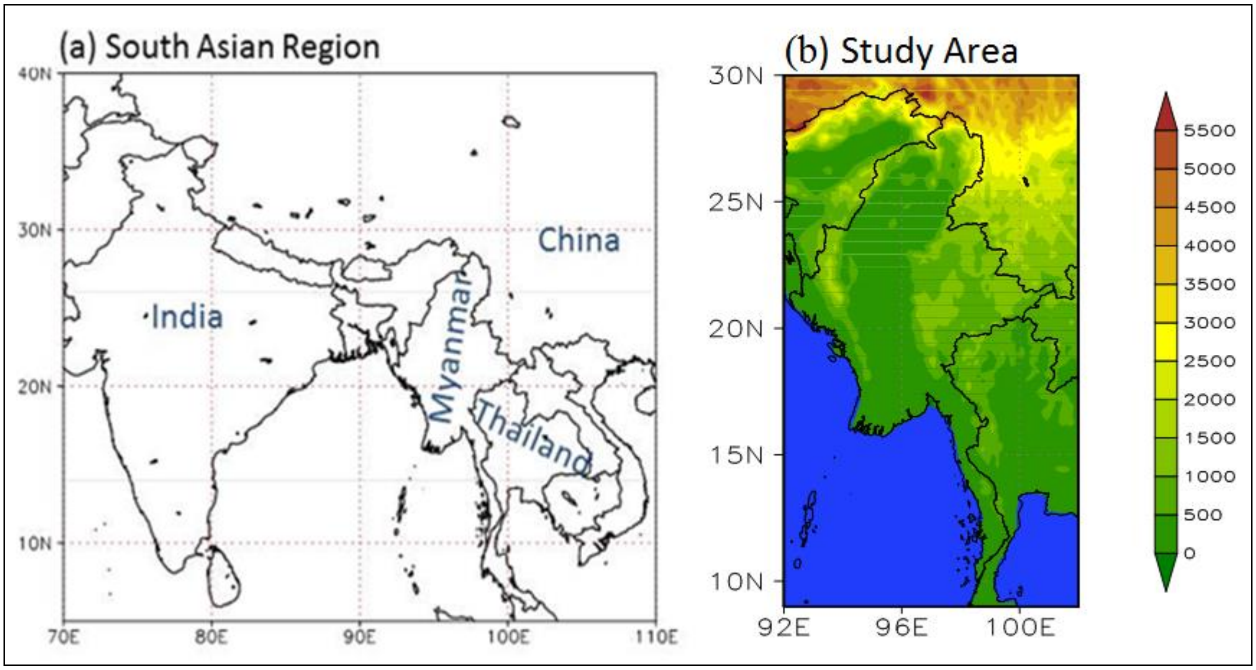

Fig. 1: Area of study (a) South Asian Region, (b) Orography over Myanmar (m)

The GPCP version 2.2 combined precipitation dataset, gridded at 2.5 degree resolution (Adler et al., 2003; Huffman et al., 2011) designed by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, available from their Web site at http://www.esrl.noaa.gov/psd/.

Moisture transport was computed at 850 hPa using datasets from the National Center for Environmental Prediction/National Center for Atmospheric research (NCEP/NCAR) reanalysis datasets (winds, relative humidity and temperature) at a resolution of 2.5 degree (Kalnay et al., 1996).

The SST data used is the Extended Reconstructed Sea Surface Temperature (ERSST) version3b, sourced from the National Oceanic and Atmospheric Administration/National Climatic Data Center (Smith et al., 2008). The data is available in their website at http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCDC/.ERSST/.version3b/.sst/ .

Summer monsoon rainfall characteristics during ENSO and IOD years are analyzed through composites calculated over a 40 year period. The study used Empirical Orthogonal Function (EOF) method (Weare et al., 1976) to identify the dominant modes of the inter-annual rainfall variability. The variability is analyzed based on seasonal mean data for the monsoon season (May - October) rainfall.

The EOF analysis involves computation of the eigen modes for a correlation matrix based on the normalized rainfall data for each season for all grid points over the Myanmar stacked into a single vector. The normalizing for each season helps to eliminate the intraseasonal variability related to Inter-tropical Convergence Zone (ITCZ). Lorenz (1956) gives a detailed discussion on EOF technique.

3. Results and discussion

Myanmar receives rainfall during the months May to October (Fig. 2), in agreement with other studies such as Qian and Lee (2000). According to Sen Roy and Kaur (2000), Myanmar's summer monsoon season rain season is experienced in June - September. However, the study further acknowledges that the southern parts of the country start receiving rain in mid-May; monsoon onset and there is some rainfall (RF) in October as well. The two datasets, observed rainfall (OBS RF) and the reanalyzed (GPCC) exhibit similar rainfall pattern and thus one can be used in place of the other. The seasonal (summer monsoon) total rainfall accounts for about 43% of the total annual rainfall over Myanmar, based on the GPCC dataset used in this study.

The observed rainfall pattern is associated with the movement of the ITCZ that reaches the north most end in June - August, which coincides with the rainfall peak. The phenomena; north-south seasonal march of the ITCZ causes cross hemispheric flow of winds transporting water vapour from high pressure regions over the relatively colder Indian and Pacific oceans to reach on land.

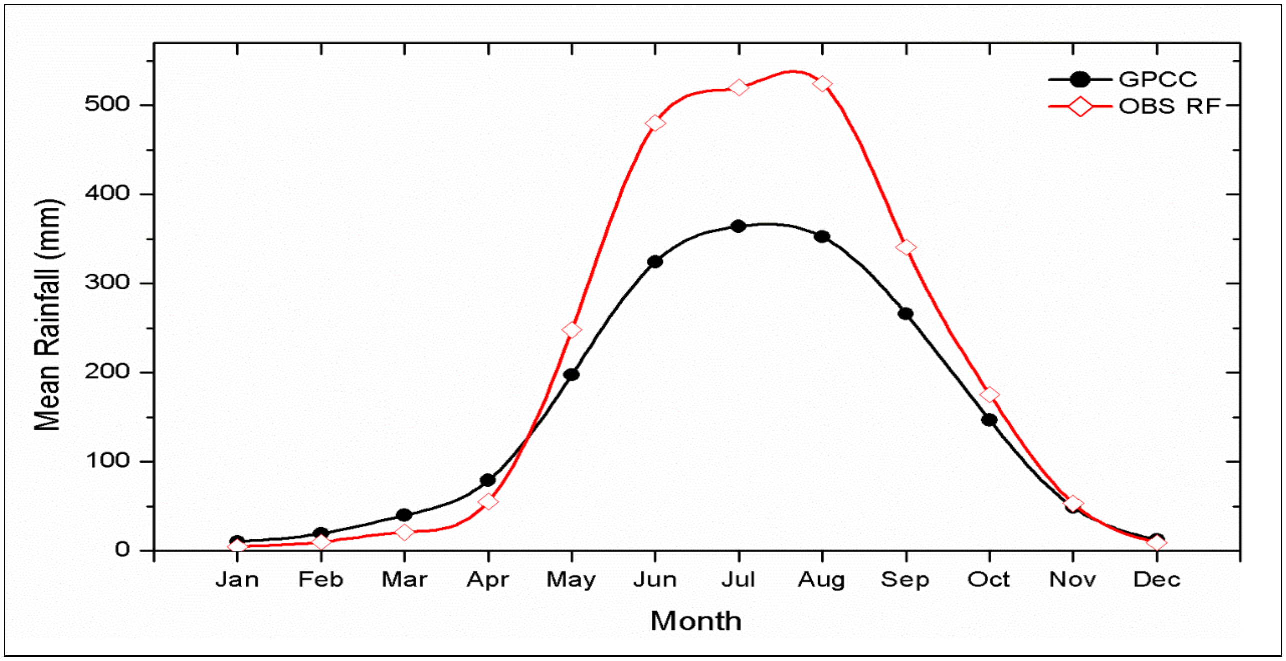

Fig. 2: The mean annual cycle of rainfall based on GPCC and observed rainfall (OBS RF) datasets, averaged over longitudes 92ºE - 102ºE and latitudes 9.5ºN - 30ºN for the period 1981-2010.

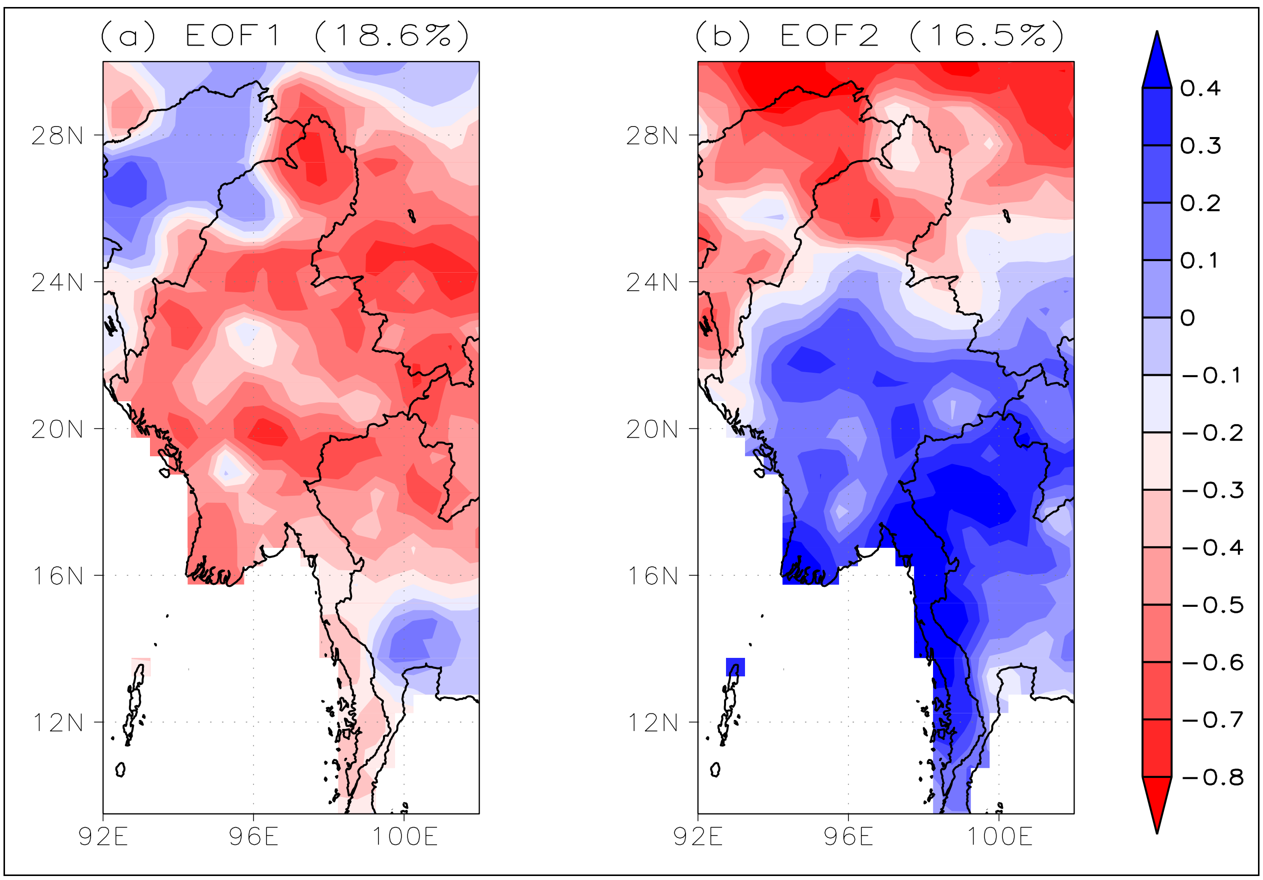

Fig. 3: (a) The first dominant mode of variability (EOF1),explains 18.6% of the total variance of the mean May-October rainfall (b) The second dominant mode of variability (EOF2) explains 16.5% of the total variance of the mean May-October rainfall over Myanmar, based on GPCC data.

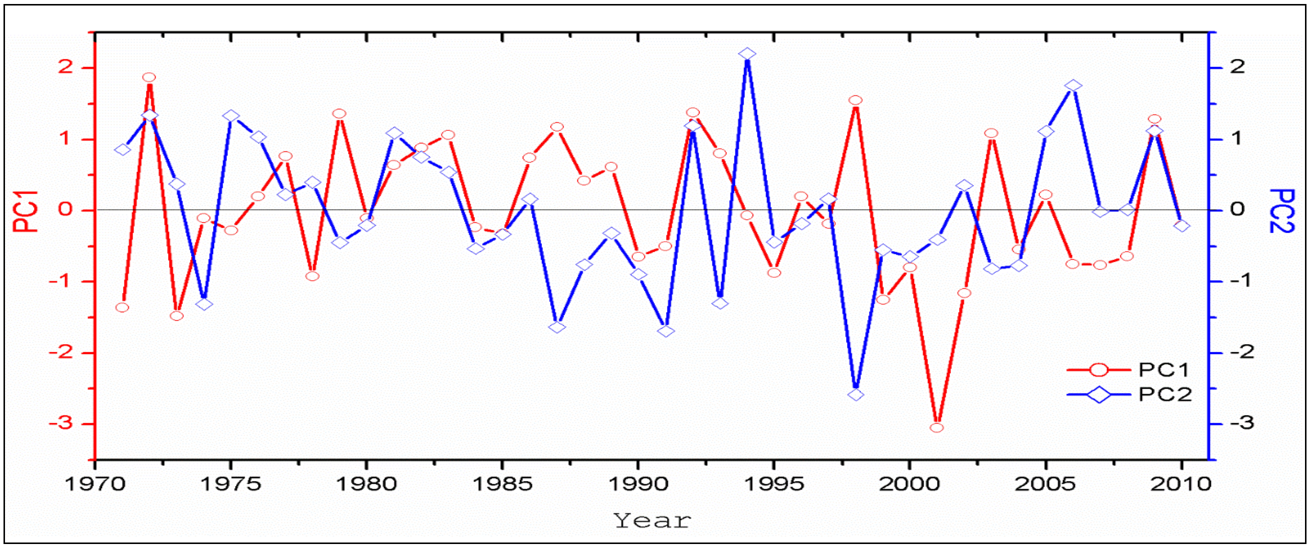

Fig. 4: Variability of the principle component time series PC1 and PC2 associated with EOF1 and EOF2, respectively.

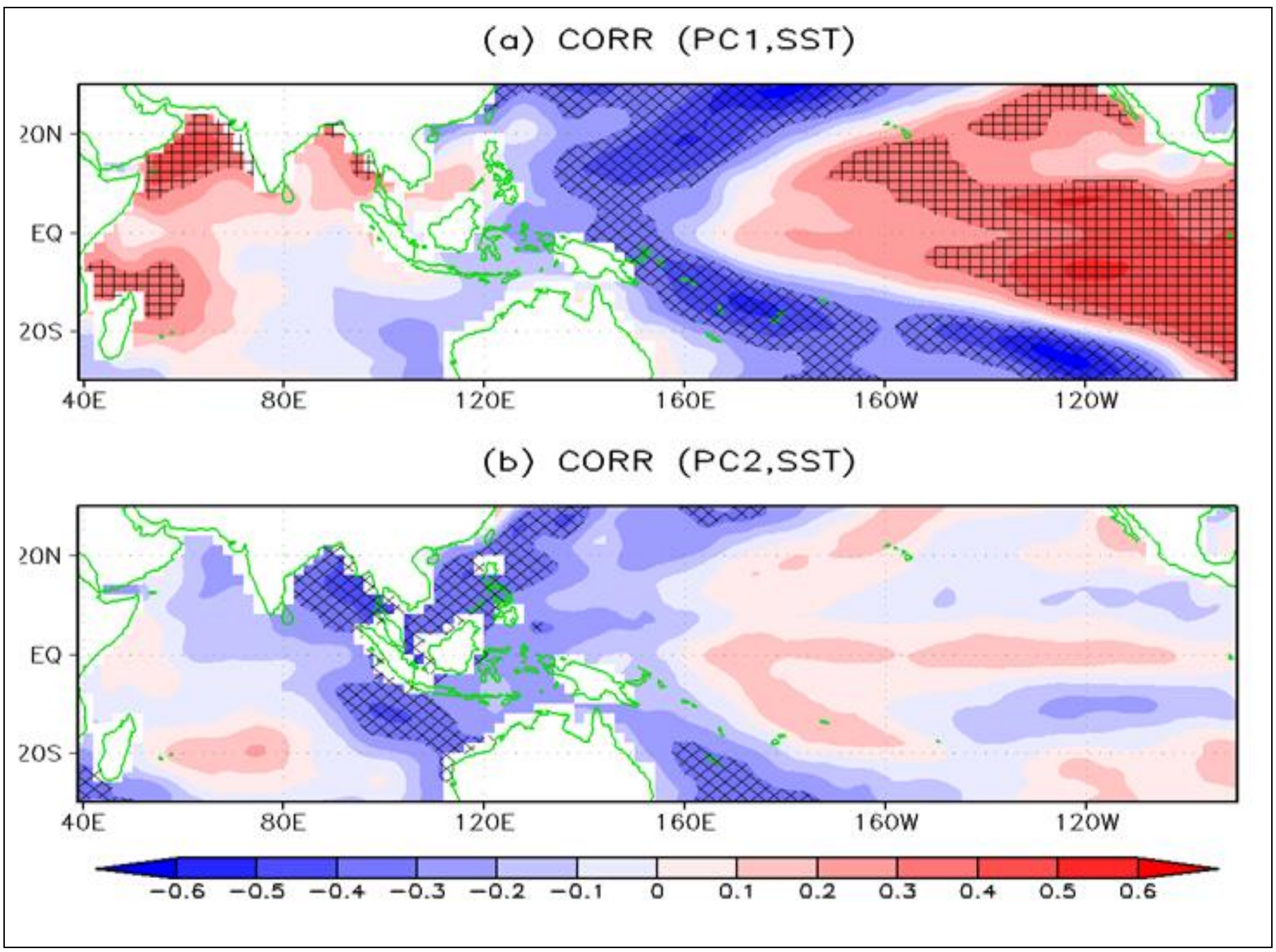

Fig. 5: Correlation map between the mean MJJASO SST (over Indian Ocean and Pacific Ocean) and (a) PC1 (b) PC2; of the mean MJJASO rainfall over Myanmar. A significant correlation (hatched regions) is exhibited at 95% confidence level.

Fig. 3 presents the dominant modes of variability of the mean May - October rainfall over Myanmar. EOF1 exhibits a dipole mode of variability, with positive anomalies in the northern sector and the central and southern sectors are dominated by negative anomalies of rainfall pattern. EOF2 similarly shows a dipole pattern, but with opposite pattern as compared with EOF 1. Their respective principle components are shown in Fig. 4 (i.e. PC1 and PC2)

Fig. 5 gives the correlation maps between the mean MJJASO SST (Indian Ocean (IO) and Pacific Oceans (PO)) and PC1 (Fig. 5a) and PC2 (Fig. 5b) of the mean MJJASO rainfall. The correlation between PC1 and IO SST captures the IOD pattern; where there is a positive (+ive) correlation with SST at the west of the IO and negative (-ive) correlation to the South Asian region during the positive phase of the IOD (Fig. 5a). A strong positive correlation is exhibited between PC1 and the SST over eastern and central Pacific, suggestion ENSO influence over Myanmar region. Fig. 5b shows that PC2 of the mean MJJASO rainfall over Myanmar exhibits a weak positive correlation with PO SST, as well as with IO SST. However, there is a notable significant negative correlation between PC2 and a few areas, including the South Asian sector.

Fig. 6 shows the interannual variability between IOD; measured by Dipole Mode Index (DMI), Southern Oscillation Index (SOI) and rainfall anomaly (RF) over Myanmar. The weak negative correlation of IOD and Rainfall over Myanmar (-0.1), implies that the country receives little rainfall during the positive IOD and enhanced rainfall during negative IOD, similar observation was made by Ashok et al. (2001). On the contrary, the rainfall over Myanmar is enhanced during positive SOI and depressed during negative SOI. During a positive SOI (La Niña), the rainfall is enhanced over the country; depressed rainfall is reported during negative SOI (El Niño).

There exists a significant correlation of 0.4 between SOI and area average rainfall anomaly over Myanmar. Further examination reveals that the correlation between SOI and IOD based on the summer monsoon period is -0.6. In support of the observation, the negative correlation between SOI and rainfall over Japan and adjoining East Asia in the 1994 IOD event which was associated with anomalous drought conditions over the areas (2003).

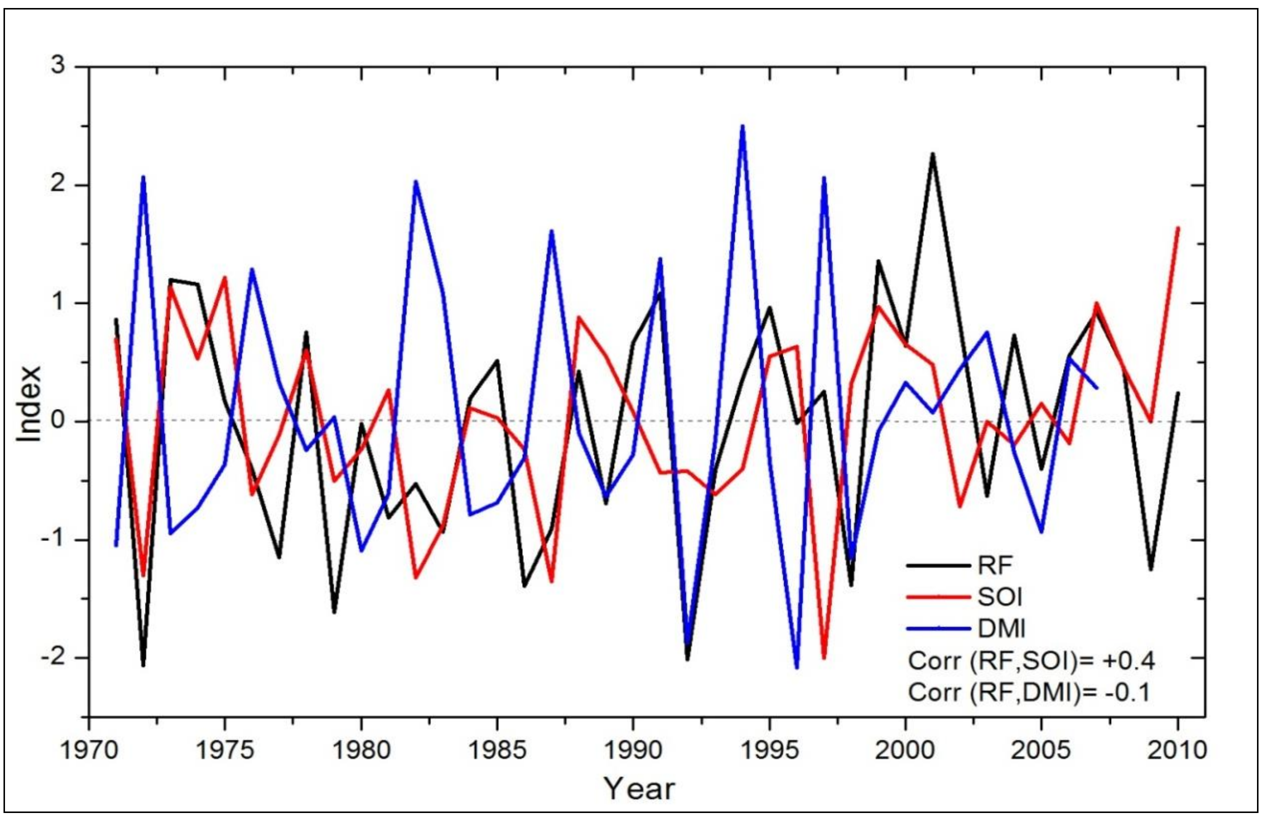

Fig. 6: Interannual variability between SOI over the period 1971-2010, DMI (1971-2007) and area average rainfall anomaly (standardized using standard deviation) (1971 -2010) over Myanmar.

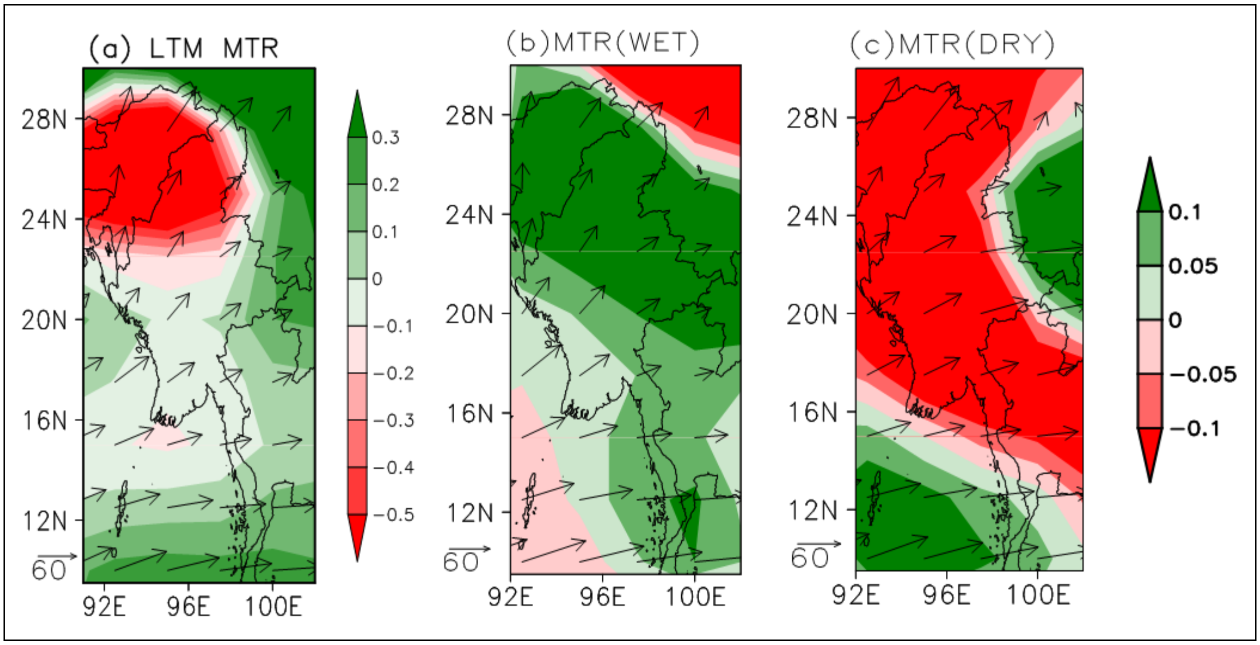

Fig. 7: Moisture transport during May - October (a) Long Term Mean (MTR), (b) Wet year (2001), (c) Dry year (1972) at 850hPa level (The -ive anomaly indicate divergences while +ive anomaly indicates convergence).

Further analysis; a partial correlation between SOI and RF and DMI and RF is summarized in Table 1. Correlation between the mean May - October rainfall (RF) and the datasets; SOI and DMI. CSOI, CSOI(DMI), CDMI and CDMI(SOI) denote correlation between RF and SOI, correlation between RF and SOI, with the influence of DMI removed, correlation between RF and DMI, and correlation between RF and DMI, when the influence of SOI is removed.

Table 1: Correlation between the mean May - October rainfall (RF) and the datasets; SOI and DMI. CSOI, CSOI(DMI), CDMI and CDMI(SOI).

Correlation with RF			
CSOI	CSOI(DMI)	CDMI	CDMI(SOI)
0.4	0.6	-0.1	0.4

The partial correlation between rainfall (RF) and SOI; with DMI removed show that IOD reduces the correlation between rainfall over Myanmar and SOI; the correlation coefficient between SOI and RF reduces from 0.6 to 0.4. Similarly, SOI reduces the correlation between DMI and RF from 0.4 to -0.1. The SOI thus has more influence on rainfall over Myanmar as compared to the DMI. The observation is agreement with a study by Baquero-Bernal et al. (2002) which reported IOD events occur as a part of ENSO events because of the high correlation between the two. Thus, the observed behavior/relationship between the monsoon rainfall over Myanmar and the IOD can partly be attributed to ENSO. In support, Ashok et al. (2001) noted that negative correlation between Indian summer monsoon rainfall and El Niño can be deduced by the IOD during some decades. Similarly, Cherchi and Navarra (2013) showed that El Nino and positive IOD tend to co-occur.

The year 2001 with anomaly of +2.3 was identified to be the wettest while 1972 is the driest year during the period of study with an anomaly of -2.1 (Fig. 6). Anomalous changes in the summer monsoon moisture transport can lead to significant changes in rainfall over Myanmar. Fig. 7 displays long term moisture transport, and transport during a wet and dry years. The entire country is generally characterized by south westerlies regardless of the years, wet or dry. The southern parts of the country are characterized by convergence in the low levels while the northern is characterized by divergence (Fig. 7a). The observation is in agreement with the previous studies that have observed that the southern parts of Myanmar experience more rainfall as compared to the northern region (Sen Roy and Kaur, 2000). The observation is attributed to the north-south march of the ITCZ. Generally, during wet years, the country is characterized by positive moisture transport anomaly depicting convergence over the entire country (Fig. 7b). The highest positive moisture transport anomaly is noted in the northern parts of the country. During dry years (Fig. 7c), most almost the entire country is characterized by low level divergence.

El Nino is generally associated with dry years in Myanmar and wider parts of Asia. Ummenhofer et al. (2013) noted that El Nino years are characterized by anomalous subsidence over monsoon Asia and reduced moisture flux, leading to drought conditions over India, Southeast Asia and Indonesia.

4. Conclusion

The study investigates the inter-annual variability of summer monsoon rainfall in Myanmar relation to IOD and ENSO. Generally, monsoon rainfall over Myanmar is experienced in the months of May to October, although its onset and cessation highly varies in both space and time. The rainfall is negatively correlated with the positive phase of the IOD but positively correlated to the positive phase of the SOI, implying that El Nino events result into drought events in the region, while La Niña events results into floods. The year 2001 was identified to be the wettest while 1972 is the driest during the period of study. Moisture convergence characterizes circulation in wet years while divergence is dominant during dry years. The findings of this study are important in improving the accuracy of seasonal forecast over Myanmar.

Acknowledgements

The authors express their special gratitude to Nanjing University of Information Science and Technology, for providing them with the necessary facilities to carry out the study. Special appreciation goes to the Department of Meteorology and Hydrology, Myanmar, NOAA and NCEP/NCAR for the provision of the datasets used in the study.

Competing Interests

The authors declare that there is no potential conflict of interest of whatsoever.

References

Adler R.F., G.J. Huffman, A. Chang, R. Ferraro, P. Xie, J. Janowiak, B. Rudolf, U. Schneider, S. Curtis, D. Bolvin, A. Gruber, J. Susskind, and P. Arkin. 2003. The Version 2 Global Precipitation Climatology Project (GPCP) Monthly Precipitation Analysis (1979-Present). J. Hydrometeorol. 4: 1147 - 1167.

Allan R.J., D. Chambers, W. Drosdowsky, H. Hendon, M. Latif, N. Nicholls, I. Smith, R. Stone, and Y. Tourre.2001. Is there an Indian Ocean Dipole, and is it independent of the El Nino-Southern Oscillation? CLIVAR Exchange, 6: 8-22. ISSN 1026 - 0471

Ashok K., Z. Guan and T. Yamagata. 2001. Impact of the Indian Ocean dipole on the relationship between the Indian monsoon rainfall and ENSO. Geophys. Res. Lett. 28: 4499 - 4502.

Ashok K., Z. Guan, and T. Yamagata. 2003. A look at the relationship between the ENSO and the Indian Ocean dipole. J. Meteor. Soc. Japan. 81: 41-56.

Ashok K., Z. Guan, N.H. Saji and T. Yamagata. 2004. Individual and combined influences of the ENSO and Indian Ocean Dipole on the Indian summer monsoon. J. Climate. 17: 3141-3155.

Baquero-Bernal A., M. Latif and S. Legutke. 2002. On dipole-like variability of sea surface temperature in the tropical Indian Ocean. J. Climate, 15: 1358-1368.

Behera S.K, S.A. Rao, H.N. Saji and T. Yamagata. 2003. Comments on “A cautionary note on the interpretation of EOFs.” J. Climate. 16:1087-1093.

Black E., S.J. Slingo and K.R. Sperber. 2003. An observational study of the relationship between excessively strong short rains in coastal East Africa and Indian Ocean SST. Mon.Weather Rev. 131: 74-94.

Chen T.C., and M.C Yen. 1994. Interannual variation of the Indian monsoon simulated by the NCAR community climate model: Effects of the tropical Pacific SST. J. Climate. **7**: 1403-1415.

Cherchi A., and Navarra A. 2013. Influence of ENSO and of the Indian Ocean Dipole on the Indian summer monsoon variability. Clim. Dyn. 41:81-103.

Gadgil S., P.N. Vinayachandran, P.A. Francis and S. Gadgil. 2004. Extremes of the Indian summer monsoon rainfall, ENSO and equatorial Indian Ocean oscillation. Geophys. Res. Lett. 31: L12213.

Guan Z., and T. Yamagata. 2003. The unusual summer of 1994 in East Asia: IOD teleconnections. Geophys. Res. Lett. 30.

Guan, Z., K. Ashok, and T. Yamagata. 2003. Summertime response of the tropical atmosphere to the Indian Ocean dipole sea surface temperature anomalies. J. Meteor. Soc. Japan. 81: 533-561.

Huffman G.J., D.T. Bolvin, and R.F. Adler. 2011. GPCP Version 2.2 Combined Precipitation Data set, WDC-A, NCDC, Asheville, NC.

Ju, J. and J. Slingo. 1995. The Asian summer monsoon and ENSO. Quaterly J. Royal Meteorol. Soc. 121: 1133-1168.

Kalnay E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, L. Iredel, H. Saha, G. White, J. Woolle et al. 1996. The NCEP/NCAR 40-Year Reanalysis Project. Bull. Amer. Meteorol. Soc. 77: 437-471.

Kawamura R. 1998. A possible mechanism of the Asian summer monsoon-ENSO coupling. J. Meteor. Soc. Japan. 76: 1009-1027.

Kawamura R., T. Matsuura and I. Satoshi. 2001. Role of equatorially asymmetric sea surface temperature anomalies in the Indian Ocean in the Asian summer monsoon and El Nino-Southern Oscillation coupling. J. Geophys. Res. 106: 4681-4693.

Latif M., K. Sperber, J. Arblaster, P. Braconnot, D. Chen, A. Colman, U. Cubasch, C. Cooper, P. Delecluse, D. Dewitt, et al. 2001. Ensip: The El Niño simulation intercomparison project. Clim. Dyn. 18: 255-276.

Lau N.C. and M.J. Nath. 2000. Impact of ENSO on the variability of the Asian-Australian monsoons as simulate in GCM experiments. J. Climate. 13: 4287-4309.

Lau N.C. and M.J. Nath. 2003. Atmosphere-ocean variations in the Indo-Pacific sector during ENSO episodes. J. Climate, 16: 3-20.

Lau N.C. and M.J. Nath. 2004. Coupled GCM simulation of atmosphere-ocean variability associated with zonally asymmetric SST changes in the tropical Indian Ocean. J. Climate. 17: 245-265.

Li T., B. Wang, C.P. Chang and Y. Zhang. 2003. A theory for the Indian Ocean dipole-zonal mode. J. Atmos. Sci. 60: 2119-2135.

Lorenz E.N. 1956. Empirical orthogonal functions and statistical weather prediction. Statistical Forecasting Project Rep. 1, MIT Department of Meteorology, 49 pp.

Lwin T. 2000. *The Prevailing Synoptic Situations in Myanmar*. A report submitted at the Workshop in Agricultural Meteorology, Department of Meteorology and Hydrology, Yangon, Myanmar

Muangsong C., B. Cai, N. Pumijumnong, C. Hu and H. Cheng. 2014. An annually laminated stalagmite record of the changes in Thailand monsoon rainfall over the past 387 years and its relationship to IOD and ENSO. Quat. Int. 349: 90-97.

Parthasarathy B., A.A. Munot and D.R. Kothawale. 1994. All India monthly and seasonal rainfall series: 1871–1993. Theor. Appl. Climatol. 49:217-224.

Philander S.G.H. 1990. El Nino, La Nina and the Southern Oscillation. Academic Press, New York. 289pp.

Qian W., and D. Lee. 2000. Seasonal march of Asian Summer Monsoon. J. Climatol. 20: 1371-1386.

Sen Roy, N. and S. Kaur. 2000. Climatology of monsoon rains of Myanmar (Burma). Int. J. Climatol. 20: 913 - 928.

Sen Roy, S. and N. Sen Roy. 2011. Influence of Pacific decadal oscillation and El Nino Southern Oscillation on the summer monsoon precipitation in Myanmar. Int. J. Climatol. 31: 14-21.

Saji N.H., B.N. Goswami, P.N. Vinayachandran and T. Yamagata. 1999. A dipole mode in the tropical mode in the tropical mode in the tropical Indian Ocean. Nature. 401: 360-363.

Saji N.H., and T. Yamagata. 2003. Possible impacts of Indian Ocean Dipole mode events on global climate. Clim. Res. 25: 151-169.

Schneider U., A. Becker, P. Finger, A. Meyer-Christoffer, M. Ziese and B. Rudolf. 2013. GPCC's new land surface precipitation climatology based on quality-controlled in situ data and its role in quantifying the global water cycle. Theor. Appl. Climatol., 115: 15 - 40.

Shukla, J. 1987. Interannual variability of monsoons. In: J.S. Fein and P.L. Stephens (Eds.), *Monsoons*. John Wiley and Sons, Inc., New York, pp. 399 - 464.

Shukla, J. and D.A. Paolino. 1983. The Southern Oscillation and long-range forecasting of summer monsoon rainfall over India. Mon. Weather Rev. 111: 1830 - 1837.

Smith T.M., R.W. Reynolds, T.C. Peterson and J. Lawrimore, 2008. Improvements to NOAA's Historical Merged Land-Ocean Surface Temperature Analysis (1880-2006). J. Climate. 21, 2283-2296.

Tsai C., S.K. Behera and T. Waseda. 2015. Indo-China Monsoon Indices. Sci. Rep. 5 : 8107.

Ummenhofer C.C., R.D. D'Arrigo, K.J. Anchukaitis, B.M. Buckley and E.R. Cook. 2013. Links between Indo-Pacific climate variability and drought in the Monsoon Asia Drought Atlas. Clim. Dyn. 40:1319 - 1334.

Venzke S., M. Latif and A. Villwock. 2000. The coupled GCM ECHO-2. Part II: Indian Ocean response to ENSO. J. Climate. 13: 1371 - 1383.

Wang B., R. Wu and K.M. Lau. 2001. Interannual variability of the Asian summer monsoon: Contrasts between the Indian and the western North Pacific-East Asian monsoons. J. Climate. 14: 4073 – 4090.

Webster P.J., V. Magana, T.N. Palmer, J. Shukla, R.A. Tomas, M. Yanai and T. Yasunari.1998. Monsoons: processes, predictability and the prospects for prediction. J. Geophys. Res. 103:14451 - 14510.

Webster P.J., A.M. Moore, J. Loschnigg and M. Leban. 1999. Coupled ocean atmosphere dynamics in the Indian Ocean during 1997-98. Nature. 401: 356-360.

Webster P.J. and S. Yang, 1992: Monsoon and ENSO: Selectively interactive systems. Quaterly J. Royal Meteorol. Soc. 118: 877-926.

Wu R., and B.P. Kirtman. 2004. The tropospheric biennial oscillation of the monsoon-ENSO system in an interactive ensemble coupled GCM. J. Climate. 17: 1623-1640.

Wu R., and B.P. Kirtman. 2007. Role of the Indian Ocean in the biennial transition of the Indian summer monsoon. J. Climate. 20: 2147-2164.

Yamagata T., S.K. Behera, S.A. Rao, Z. Guan, K. Ashok and H.N. Saji. 2002. The Indian Ocean dipole: A physical entity. CLIVAR Exchanges, No. 24, International CLIVAR Project Office, Southampton, United Kingdom, 15-22.

Yang, S. and K.-M. Lau. 1998. Influences of sea surface temperature and ground wetness on Asian summer monsoon. J. Climate. 11: 3230-3246.