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THE GLOBAL MONSOON SYSTEM: RESEARCH AND FORECAST

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South Asian Summer Monsoon : An Overview

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1 The south Asian Monsoon System

As the word 'monsoon' indicates, the south Asian summer monsoon is part of a seasonally reversing wind system (Fig.1b,e) [*Ramage*, 1971; *Rao*, 1976] characterized by wet summers and dry winters (Fig.1a,d). The winds during summer monsoon season are associated with a large scale cyclonic vorticity at 850 hPa and the low level westerly jet (LLJ) over the Arabian Sea (Fig.1e) and an anticyclone (the Tibetan anticyclone) at the upper level (200 hPa) with the monsoon easterly jet (Fig.1f). Wet summer and dry winter conditions (Fig.1a,d) associated with the seasonal changes of low level winds are crucial for agricultural production [*Parthasarathy et al.*, 1988; *Webster et al.*, 1998; *Abrol and Gadgil*, 1999] and economy of the region. During northern winter, the equatorial easterlies are weak and confined between 5°N and 10°S in the upper atmosphere (200 hPa),while the sub-tropical westerlies intrude all the way to 10°N (Fig.1c). The sub-tropical westerlies recede to north of 30°N during northern summer and a strong easterly jet characterizes the equatorial region (Fig.1f). The monsoon or the seasonal changes of winds and rainfall in the region could be interpreted as a result of northward seasonal migration of the east-west oriented precipitation belt (tropical convergence zone, TCZ) from southern hemisphere in winter to northern hemisphere in summer [*Gadgil*, 2003]. The largest northward excursion of the rain belt takes place over the Indian monsoon region where it moves from a mean position of about 5°S in winter (Fig.1a) to about 20°N in northern summer (Fig.1d). In addition to the primary TCZ around 20°N during the summer season, a secondary TCZ exists over the eastern equatorial Indian Ocean (IO). The large scale summer mean monsoon circulation is characterized by a deep baroclinic vertical structure with low level cyclonic vorticity (and convergence) and upper level anticyclonic vorticity (and divergence).

The south Asian monsoon (SAM) climate is , therefore, a phase of the strong annual cycle that exist in this region. The amplitude of the climatological mean annual cycle (AC) of precipitation and that of the zonal and meridional winds at 850 hPa (the standard deviation of anomalies of climatological monthly means after removing the annual mean) are shown in Fig.2. The largest amplitude of the AC of precipitation occur over the Asian and the Australian monsoon regions. The largest amplitude of the AC for zonal winds at low level occur over the low level westerly jet (LLJ) region of central and western Arabian Sea while that for the meridional winds occur over the western equatorial Indian Ocean and Somalia.

The climatological mean monsoon annual cycle (MAC) is not sinusoidal but is characterized by a sharp transition at the beginning of the summer monsoon season known as the 'onset' of the SAM. This can be seen in the time-latitude section of precipitation averaged over 70°E−100°E (Fig.3a) as well as in the kinetic energy(K.E.) of the LLJ averaged over 50°E−65°E, 5°N−15°N (thick solid line). A characteristic feature of the Indian monsoon is the onset associated with rapid transition of the high precipitation zone from near the equator to about 15°N towards the end of May or beginning of June. Dramatic increase of K.E. of the LLJ by a factor of more than 6 takes place within about a week around the onset. The seasonal migration of the TCZ is associated with a seasonal evolution of the meridional gradient of tropospheric heating illustrated by the evolution of the mean temperature of the tropospheric layer between 200 hPa and 700 hPa averaged between 30°E and 110°E (Fig.3b)

A wide spectrum of variability characterize the south Asian monsoon system. Taking daily precipitation over the central India for three summer seasons, sub-seasonal and interannual variability are illustrated in Fig.4. In Fig.4a, day-to-day fluctuations (thin solid line) are caused by synoptic disturbances (lows and depressions) while the slow variability within the season (thick dashed line) are intraseasonal oscillations (ISOs). The ISOs result in wet and dry spells within the season known as 'active' and 'break' conditions (Fig.4a) [*Webster et al.*, 1998; *Gadgil*, 2003; *Goswami*, 2004b]. The seasonal mean (horizontal bar in Fig.4a) has year-to-year variations which is further illustrated in Fig.4b over a long period for precipitation over the whole country. This figure also indicates existence of significant interdecadal variability of the south Asian monsoon.

The year-to-year variations of the long term seasonal mean precipitation over the Indian region is strongly correlated with food production in the region. Even as the green revolution steadily increased food productivity over the last four decades, a modest decrease in the monsoon rainfall (e.g. 10% of long term mean) leads to a significant decrease in rice production over India [*Swaminathan*, 1987; *Parthasarathy et al.*, 1988; *Webster et al.*, 1998; *Gadgil*, 1995; *Abrol and Gadgil*, 1999]. The interannual variability of the monsoon rainfall may have even larger impact on agricultural productivity in the coming years, as the growth rates of agricultural production have decreased in recent years in association with the fatigue of the green revolution [*Gadgil et al.*, 1999]. Frequency of occurrence of active and break spells within the season influences the seasonal mean and hence agricultural production. Regional distribution of precipitation and knowledge of the wet and dry spells within the season are also important for sowing, harvesting of crops and water management. Further, long breaks in critical growth periods may lead to substantially reduced yield of certain crops [*Gadgil and Rao*, 2000]. Therefore, 'long range' forecasting of seasonal mean monsoon at least one season in advance and 'extended range' forecasting of dry and wet spells of the sub-seasonal variability $2-3$ weeks in advance are of great importance.

Understanding gained over the past few decades for the maintenance of the MAC as well as physical processes responsible for intraseasonal and interannual variabilities of the SAM are reviewed. Forecasting needs are identified and challenges faced by the community in simulation and prediction of the seasonal mean ASM and potential for extended range prediction of dry and wet spells are highlighted. Outstanding unresolved issues are also identified.

2 The Monsoon Annual Cycle (MAC)

The AC of the monsoon is a manifestation of the seasonal migration of the zonally oriented belt of precipitation or the tropical convergence zone (TCZ) [*Gadgil*, 2003]. An interesting aspect of the MAC is the asymmetry in seasonal transition of the TCZ from Asian summer monsoon to Australian summer monsoon and back to the Asian summer monsoon. The evolution of the AC of precipitation (deviation of climatological monthly mean from annual mean) averaged over $80°E-120°E$ shows [*Hung et al.*, 2004] that the high precipitation zone moves smoothly southward from NH to SH (Fig.5a). However, the TCZ is not allowed to move smoothly from boreal winter to summer and the transition takes place only through the abrupt Asian monsoon onset. The evolution of surface wind speed and sea surface temperature (SST) associated with the evolution of precipitation (Fig.5b,c) indicates interaction between the ocean and the atmosphere in the evolution of the MAC. Due to the orientation of the Asian and Australian land masses, global scale mass redistribution facilitates southward march of maximum convection from the Asian summer monsoon to the Asian winter (Australian summer) monsoon, but deters the reverse march in boreal spring [*Chang et al.*, 2004].

The geographic location of the Indian continent surrounded on three sides by warm waters of the IO and by the Himalayan mountain range to the north play a pivotal role on the annual evolution of the SAM. A model of the SAM must explain the seasonal migration of the TCZ and the existence of a double TCZ during the summer season. It must also explain the deep baroclinic nature of the SAM circulation. Followed by the pioneering suggestion of Halley (1686), a paradigm for MAC has been a giant land-sea breeze driven by land-ocean thermal contrast. The land-ocean surface temperature contrast theory for the SAM is used even today, for example in explanation of paleoclimatic monsoon variabilities (e.g.[*Anderson et al.*, 2002; *Gupta et al.*, 2003]). However, the traditional land-ocean contrast model of the monsoon is inadequate to explain the SAM. Firstly, after the onset of the monsoon during June- September, the surface temperature over the continent is always colder than the ocean to the south. Secondly, surface heating to the north alone would produce only a shallow circulation confined to the lower atmosphere [*Schneider and Lindzen*, 1977] and could not explain the deep vertical circulation associated with the ASM. However, the seasonal migration of the TCZ is indeed a result of seasonal evolution of the meridional gradient of tropospheric heating illustrated by the evolution of the mean temperature of the tropospheric layer between 200 hPa and 700 hPa averaged between 30°E and 110°E (Fig.3b). The onset or the ushering in of the south Asian summer monsoon takes place with the reversal of the meridional temperature gradient. The problem then boils down to understanding the generation and maintenance of the deep tropospheric heat source in the north. The heating over the Tibetan Plateau plays an important role in this seasonal evolution of the meridional gradient of heating [*Wu and Zhang*, 1998; *Wu et al.*, 2002; *Yanai et al.*, 1992] and in triggering the onset of the SAM. It may be noted that after the onset, the tropospheric temperature (TT) in the north may be affected by the convective heating associated with the TCZ. The reversal of meridional temperature before the onset of the monsoon is, however, not affected by the TCZ heating.

The change in the sign of meridional gradient of TT ushers in the setting up of an off-equatorial large scale deep heat source. The atmospheric response to such a heat source [*Gill*, 1980] leads to cross equatorial flow and strengthening of the low level south-westerlies above the planetary boundary layer (PBL). This leads to the zero absolute vorticity line at lower atmosphere (e.g. 850 hPa) to move north of the equator [*Tomas and Webster*, 1997] to about 5°N facilitating symmetric inertial instability [*Tomas and Webster*, 1997; *Krishnakumar and Lau*, 1997, 1998]. The 'potential convective instability' builds up during the pre-monsoon season but can not be realized due to the subsidence above the PBL. The symmetric instability forces frictional boundary layer convergence, overcomes the inhibition and explosive development of off-equatorial convection over India and the Bay of Bengal takes place. Therefore, change in the sign of the meridional gradient of the TT may be used as a thermodynamic definition of onset of SAM [*Goswami and Xavier*, 2004a]. After the onset, the large scale flow at lower levels (above PBL) produces a large scale cyclonic vorticity (the monsoon trough) through interactions with the mountains to the north. This large scale low level cyclonic vorticity helps organize convection and helps maintain the northern TCZ over the continent.

The second TCZ over the equatorial IO is controlled by SST over the northern IO. The climatological mean SST during January and July and corresponding depth-longitude section along the equator are shown in Fig.6. During the northern summer season, the secondary TCZ over the ocean is maintained by the meridional gradient of SST at the location of SST maximum over the eastern equatorial IO [*Goswami et al.*, 1984]. The annual cycle of the SST is very important for the annual evolution of the monsoon precipitation [*Fennessy and Shukla*, 1994b]. The SST in the IO is determined by the net heat flux at the surface and the wind stress forcing [*Murtugudde et al.*, 1996; *Murtugudde and Busalacchi*, 1999]. The wind stress, on the other hand, is largely forced by the latent heating associated with the precipitation in the TCZ (Fig.7). Also the cloudiness associated with the TCZ contributes significantly to the net heat flux at the surface. The net heat flux over the northern IO (40°E-110°E, Eq.-25°N) is about +80 Wm^{-2} during northern spring and summer (April-October) while that over the southern IO (40°E-110°E, 30°S-Eq.) it is -10 Wm^{-2} [*Josey et al.*, 1998, 1999]. Similarly during northern winter (November-February) it is about +10 Wm^{-2} in the northern IO while it is about +90 Wm^{-2} in the southern IO. An annually reversing upper ocean meridional circulation is required for maintaining the annual cycle of the SST [*Loschnigg and Webster*, 2000]. While SST has a strong control on the annual evolution of the TCZ, the precipitation in the TCZ has a strong control on the evolution of the SST. Thus, significant air-sea interaction is involved with the MAC. Unlike in the Pacific, this air-sea interaction is modified by the interaction with the land surface processes to the north and the east. How unstable is this air-sea interaction? What is the role of external forcing (e.g. Solar forcing) in limiting this air-sea interaction? Can rectification of the air-sea interaction on the annual time scale lead to interannual variability of the ASM? Answers to questions like these are not available as the coupled GCMs are not yet able to simulate the mean annual cycle of the precipitation and SST in this region with reasonable fidelity.

3 Synoptic Disturbances and Medium Range Prediction

The synoptic activity during the summer monsoon season consists of lows and depressions that determine the day to day variability of precipitation. Efforts to improve GCMs for improved medium range prediction of these events must continue. However, it has been noted [*Kumar and Dash*, 2001] that the number of occurrence of lows during the summer monsoon season has steadily increased over the last two decades while that of depressions has steadily decreased over the same period keeping the sum of lows and depressions during the summer monsoon season approximately unchanged. There is need to understand this interdecadal change in the characteristic of monsoon synoptic variability. There is some suggestion [*Dash et al.*, 2004] that change in the background mean flow associated with tropical interdecadal variability [*Krishnamurthy and Goswami*, 2000; *Goswami*, 2004b] may be responsible for this. This implies that the predictability monsoon weather may have gone down in the recent decades making the prediction of day to day fluctuations more difficult in recent years. This conjecture needs to be tested through predictability experiments with GCMs.

4 Intraseasonal Variabilities of the SAM

With the availability of the satellite data (e.g. NOAA OLR, TMI SST, QuikSCAT surface winds etc.) and reanalysis products (NCEP/NCAR and ERA), better description of spatio-temporal characteristics of monsoon intraseasonal oscillations (ISOs) have evolved over the past decade or so. Such observations have revealed that the active and break of SAM or the wet and dry spells over the Indian continent, are manifestation of repeated northward propagation of the TCZ from the equatorial position to the continental position [*Sikka and Gadgil*, 1980; *Yasunari*, 1979] and results from superposition of a 10-20 day and a 30-60 day oscillations. Both the 10-20 day oscillation and the 30-60 day oscillation contribute roughly equally to the total intraseasonal variability (ISV) in the SAM region. While 30-60 day oscillation has a very large zonal scale encompassing both the SA and the EA/WNP monsoon regions, the 10-20 day oscillation has a smaller zonal scale and is regional in character. The 30-60 day mode is characterized by a northward propagation while the 10-20 day mode is characterized by a westward propagation. The spatial scale of the dominant ISO and the meridional circulation associated with it are shown in Fig.8 and compared with spatial structure of the seasonal mean and the mean regional monsoon Hadley circulation. The large scale spatial of the dominant ISO mode and its similarity with that of the seasonal mean is evident from Fig.8. It indicates that relative frequency of occurrence of active and break phases could influence the seasonal mean and contribute to the interannual variability (IAV) of the SAM.

Major advances have been made during the past two decades in understanding the temporal scale selection and northward propagation of the 30-60 day mode and temporal scale selection and westward phase propagation of the 10-20 day mode (see *Goswami* [2004a]; *Wang* [2004] for detail). Two mechanisms seem to contribute to the temporal scale selection of the 30-60 day mode. One is a 'convection-thermal relaxation feedback mechanism' [*Goswami and Shukla*, 1984], according to which convective activity results in an increase of static stability which depresses convection itself. As convection dies, dynamical processes and radiative relaxation decreases moist static stability and brings the atmosphere to a new convectively unstable state. This mechanism does not involve wave dynamics and may be responsible for the northward propagating 30-60 day oscillations not associated with eastward propagation of convection in the equatorial region. The other mechanism involves eastward propagation of convection the equatorial IO in the form of a Kelvin wave and west-northwest propagation of Rossby waves emanated over the western Pacific [*Wang and Xie*, 1997]. The time scale is determined in this case by propagation time of the moist Kelvin wave from the eastern IO to western Pacific, the moist Rossby waves from western Pacific to the Arabian Sea where they decay and a new equatorial perturbation is generated. Until recently no clear physical mechanism for the selection of period, wavelength and westward phase propagation of the quasi-biweekly mode was known. A unified model now explains the spatial structure (wavelength), period and westward phase speed of both summer and winter 10-20 day oscillation or the quasi-biweekly mode (QBM). It is demonstrated that the QBM is a n = 1 equatorial Rossby wave with about 6000 km wavelength and period of 14-16 days driven unstable by convective feedback involving boundary layer convergence that is shifted to the north (south) of the equator by about 5 degrees by the summer (winter) background mean flow [*Chatterjee and Goswami*, 2004].

With the availability of high resolution reliable SST from satellite on daily time scale and time series data from some moored buoys in the Bay of Bengal [*Sengupta and Ravichandran*, 2001] it became clear that the ISV of SST over the north IO has large amplitude and large spatial scale similar to that of the atmospheric ISV [*Sengupta et al.*, 2001]. Coherent northward propagation of intraseasonal SST, surface wind speed, net heat flux at the surface and OLR (or precipitation) are found [*Sengupta et al.*, 2001]. It is also noted that there exist a quadrature phase relationship between northward propagation of SST and precipitation on the 30-60 day time scale. A reasonable estimate of intraseasonal variation of net heat flux at the surface (made possible by availability of reliable SST from TMI, surface winds from QuikSCAT and NOAA OLR on daily time scale) showed that ISV of heat flux is a major driving for the ISV of SST over most of the tropical IO although advection and entrainment play roles in the equatorial IO and the Somali current region. Coupled modeling studies [*Fu et al.*, 2003; *Zheng et al.*, 2004; *Rajendran et al.*, 2004] demonstrate that air-sea interaction is required to explain the observed space-time spectra of summer ISO in SST and precipitation.

The monsoon ISOs are a crucial building block of the ASM. Through multi-scale interactions with synoptic activity on one hand and the seasonal cycle on the other, they determine not only the probability of occurrence of daily precipitation but also the IAV of the seasonal mean. The horizontal structure of low level winds associated with the summer monsoon 30-60 day mode (Fig.8b) also have large scale and are similar to that of the seasonal mean (Fig.8a). Therefore, the meridional shear of the low level zonal winds and cyclonic vorticity at 850 hPa are significantly enhanced (weakened) during an active (break) phase of the ISO. Hence, conditions for cyclogenesis are much more favorable during an active phase compared to a break phase. Like MJO, do the monsoon ISOs too modulate the synoptic activity in the region during northern summer? Using genesis and track data for low pressure systems (LPS) for 40 years (1954-1993), *Goswami et al.* [2003] show that genesis of an LPS is nearly 3.5 times more favorable during an active condition (147 events corresponding to normalized index $>$ +1) compared to a break condition (47 events corresponding to normalized index $<$ -1) of the monsoon ISO. They also show that the LPS are spatially strongly clustered to be along the monsoon trough region during an active condition (Fig.9). Since the day to day fluctuation of precipitation is essentially governed by these synoptic activity, the ISO phase modulates the probability of occurrence of daily precipitation. We noted earlier that the spatial structure of the 30-60 day mode is similar to that of the seasonal mean (Fig.8), strengthening (weakening) the seasonal mean in its active (break) phases. Thus, the ISOs also have the potential to produce IAV of the seasonal mean precipitation. This point will be discussed in detail in the next section.

The amplitude of the ISV (e.g. the coefficient of variability) is much larger than the amplitude of IAV of the SAM. This fact (the intraseasonal signal being strong) and the fact that the monsoon ISOs are associated with quasi-periodic oscillations, indicate potential predictability of the ISO phases beyond the medium range weather prediction. Estimates made by *Goswami and Xavier* [2003]; *Waliser et al.* [2003] show potential predictability of the break phase of monsoon ISO to be about three weeks while that of the active phase being smaller. Earlier, *Lo and Hendon* [2000]; *Mo* [2001]; *Jones et al.* [2004] demonstrated usefulness of empirical techniques in making useful prediction of the ISO phases. Built upon these work and using only precipitation (including contribution from 10-20 day mode) as predictor *Goswami and Xavier* [2003] developed a simple empirical model and show that monsoon breaks could be predicted up to 20 days in advance with useful skill. *Webster and Hoyos* [2004] also use an empirical technique but use a number of other predictors and demonstrate highly significant skill of prediction of the ISO phases up to 25 days in advance. This demonstration of ability to predict the dry and wet spells of the SAM three weeks in advance represents an important advance SAM research.

5 Interannual Variability of the SAM

The interannual variability of the SAM monsoon is indicated in Fig.4b. The interannual of the SAM is rather modest with the interannual standard deviation being about 10% of the seasonal mean. However, larger excess or deficit of all India rainfall are associated with large spatial scale covering most of the country [*Shukla*, 1987]. Extremes in monsoon rainfall leads to devastating floods and droughts [*Shukla*, 1987; *Mooley and Shukla*, 1987; *Webster et al.*, 1998; *Sikka*, 1999] leading to enormous economic loss and human misery. Therefore, understanding of the physical processes responsible for the observed IAV of SAM is crucial for advancing the capability for predicting the IAV.

One notable connection with the IAV of the SAM is that with the ENSO. There is a tendency for the El Nino's to be associated with droughts and La Nina's to be associated with above normal conditions over India . While a connection between the SAM and the ENSO exits, it is not very strong. It is worth noting here that many droughts and floods of the SAM occur without El Nino or La Nina. Since the pioneering work of Sir Gilbert Walker [*Walker*, 1924], this influence of the ENSO on the SAM has been investigated [*Sikka*, 1980; *Rasmusson and Carpenter*, 1983; *Shukla*, 1987]. This interaction is primarily through the change in the equatorial Walker circulation influencing the regional Hadley circulation associated with the Asian monsoon [*Webster et al.*, 1998; *Goswami*, 1998; *Lau and Nath*, 2000]. The lag correlation between SAM rainfall and ENSO SST (Fig.10) shows that the ENSO affects the SAM rainfall in its evolving phase, the maximum correlations being with eastern Pacific SST following the SAM . This prompted the possibility of SAM influence on the ENSO itself. The strong heat source associated with the ASM could indeed influence the atmospheric circulation in a significant way and could modify the surface stresses over the central and western Pacific and influence the strength and evolution of the ENSO [*Yasunari*, 1990; *Chung and Nigam*, 1999; *Kirtman and Shukla*, 2000]. These independent studies of ENSO influence on the ASM and ASM influence on the ENSO, made it clear that the ENSO and the ASM are not independent phenomena but part of a coupled ocean−atmosphere oscillation. This recognition led to recent studies with coupled ocean−atmosphere GCMs [*Loschnigg et al.*, 2003; *Wu and Kirtman*, 2004; *Yu et al.*, 2003] and demonstrated that air−sea interaction involving ASM and the ENSO actually leads to a tropical biennial oscillation (TBO) of the ASM. A plausible mechanism through which ocean-atmosphere coupling leads to a TBO may be described as follows [*Wu and Kirtman*, 2004]. A strong ASM during JJA can enhance surface easterlies in the central equatorial Pacific, induces an eastward propagating upwelling Kelvin wave and gives rise to negative SST anomalies in the eastern Pacific that amplifies through air−sea interactions. Colder SST in the eastern Pacific is also associated with warmer

SST in the western Pacific. A strong ASM also cools the Indian Ocean through enhanced evaporation and upwelling. Associated intensification of the Walker circulation leads to divergence of moisture supply in the western Indian Ocean. Reduced moisture supply at low levels together with upper level subsidence leads to a weaker ASM during the next summer. A weak ASM induces opposite affects and can lead to a stronger monsoon next year. Thus, the ocean-atmosphere interaction generates IAV of the ASM via generation of TBO signal.

The biennial tendency of IAV of the ASM also has been known for a long time [*Mooley and Parthasarathy*, 1984; *Yasunari*, 1990; *Clarke et al.*, 1998; *Webster et al.*, 1998; *Meehl and Arblaster*, 2002] and recognized as manifestation of the TBO. Various mechanisms have been proposed for the TBO (e.g. *Nicholls* [1978]; *Meehl* [1987, 1993]; *Clarke et al.* [1998]; *Goswami* [1995]; *Chang et al.* [2000]). The modeling studies mentioned above provides a synthesis of the TBO studies and the ENSO-monsoon connection studies and shows that they are linked and part of the same air-sea coupled oscillation involving both the IO and the Pacific basins.

Another large scale forcing that seem to influence the IAV of the SAM is the snow cover over Eurasia. Starting with [*Blanford*, 1884], several studies [*Hahn and Shukla*, 1976; *Dey and Kumar*, 1982, 1983; *Dickson*, 1984; *Ropelewski et al.*, 1984; *Yang et al.*, 1996; *Rao et al.*, 1996] find a weak negative correlation between snow cover over Eurasia and intensity of the SAM. However, due to the brevity of available records, the statistical significance of these results is marginal as some depend on data for a decade or less. The physical basis for such a relationship is considered to be the following. The net effect of the increased snow cover generally over the central and southern Eurasia is to lessen the land-ocean temperature contrast and decrease the strength of the SAM. However, as we mentioned in Section 2 the land-ocean surface temperature contrast paradigm is inadequate to explain either the SAM precipitation or the deep vertical structure of its circulation. Using surface temperature, soil moisture and snow cover record from 1870 to 2000, *Robock et al.* [2003] find that Indian monsoon rain fall and snow cover anomalies over Eurasia to be positively correlated. The SAM is not driven by the land-sea surface temperature gradient but by the tropospheric temperature gradient. In fact *Liu and Yanai* [2001] find a significant positive correlation between March-April-May (MAM) upper tropospheric temperature over western Europe and All India monsoon rainfall (AIR). Thus, neither the observational evidence of the Eurasian snow-monsoon connection is robust nor the physical mechanism through which Eurasian snow influence the SAM is well understood.

In addition to the ENSO related ocean-atmosphere interaction, local warm-ocean atmosphere interaction over the Indian Ocean (IO) and western north Pacific can also give rise to IAV of the MAC. Recently discovered Indian Ocean dipole mode (IODM, *Saji et al.* [1999]; *Webster et al.* [1999]) is a good example of manifestation of such air-sea interaction. This mode is not an equatorially confined zonal mode. The SST dipole is coupled with the south IO anticyclonic anomalies. In the presence of the summer monsoon background flow, the ocean to the east of the anticyclone near Sumatra cools due to coastal upwelling, evaporation and entrainment. Reduction of convection associated with the cooling excites westward propagating descending Rossby waves and reinforces the anticyclone [*Li et al.*, 2003; *Wang et al.*, 2003]. This air-sea interaction also contributes to a quasi-biennial signal of the monsoon [*Loschnigg et al.*, 2003; *Li et al.*, 2003]. Similar warm ocean-atmosphere interaction involving the western north Pacific (WNP) anticyclone leads to IAV of the EAM [*Wang et al.*, 2003]. The direct contribution of the IODM to the IAV of the SAM is unclear at this time. However, through the generation of the TBO, the IODM may indirectly contribute to the IAV of the SAM.

6 Interdecadal Variability of the SAM

As seen in Fig.4 (dotted line), the SAM precipitation does not seem to have any climatic trend but has epochs of roughly three decades when the precipitation has the tendency to be more above than below normal followed by a roughly three decades when it has the tendency to be more below than above normal. The large scale circulation changes associated with the interdecadal variability may lead to change in teleconnection patterns in these time scales. For example, the ENSO-monsoon relationship is known to undergo low frequency variations in these time scales. It may be noted from Fig.10b that the simultaneous as well as the lag relationship between AIR and Nino3 SSTA has undergone major changes during the recent years compared to earlier decades. Our understanding of the interdecadal variability of the SAM remains much poorer than our understanding of the MAC and its intraseasonal and interannual variabilities. These issues will be discussed in some detail in 'Interdecadal Variability' under scientific topics (Part-II) and hence we shall refrain from discussing them in detail here.

7 Simulation and Prediction of the Seasonal Mean SAM

Over the last couple of decades, the climate models have improved steadily in simulating the mean global climate [*Gates et al.*, 1999] and a conceptual framework for predicting the tropical climate has also been established [*Charney and Shukla*, 1981; *Shukla*, 1981, 1998]. Following the seminal work of *Charney and Shukla* [1981], a series of sensitivity studies with climate models [*Lau*, 1985; *Kumar and Hoerling*, 1995; *Fennessy and Shukla*, 1994a; *Shukla and Wallace*, 1983; *Anderson et al.*, 1999] have established that the interannual variability of the tropical climate is largely driven by slowly varying anomalous boundary conditions (ABC) and is much less sensitive to initial conditions and hence more predictable compared to the extra-tropical climate. While this conclusion is generally true over large part of the tropics, it has also come to light over the last decade that the summer climate over the Asian monsoon region appears to be an exception within the tropics, simulation of which is quite sensitive to initial conditions [*Sperber and Palmer*, 1996; *Krishnamurthy and Shukla*, 2000; *Cherchi and Navarra*, 2003; *Brankovic and Palmer*, 2000; *Sperber et al.*, 2001]. The current skill of most atmospheric GCMs with perfect boundary conditions in predicting Asian summer monsoon precipitation is, however, insignificant (e.g. *Wang et al.* [2004]; *Kang et al.* [2004]). The skill of simulating JJA precipitation by one AGCM based on 10 member ensemble prediction for each of 21 summer seasons is shown in Fig.11a. Similarly, correlation between observed and simulated JJAS precipitation based on 5 ensemble of 20 year simulations with observed SST as boundary conditions is shown in Fig.11b. It is clear from both figures that while there is considerable skill in simulating and predicting the summer precipitation over a large part of the deep tropics, the skill is nearly zero or negative over the Asian monsoon region. Sensitivity of simulation of the seasonal mean to initial conditions indicate that the interannual variability of the seasonal mean is governed partly by 'internal' low frequency (LF) variability in addition to contribution from the ABC.

In light of some of the major achievements in climate modeling, the failure in predicting the mean south Asian summer monsoon has remained the single major hurdle. Why is dynamical prediction of seasonal Asian monsoon so difficult, when the skill of dynamical prediction of seasonal climate over a large part of the tropics has steadily improved over the last decade and showing good promise? Is it because the simulation of the mean Asian summer monsoon by GCMs are still too poor and the systematic biases are still too large? Most climate models still have large systematic bias in simulating the seasonal mean SAM precipitation [*Gadgil and Sajani*, 1998; *Kang et al.*, 2002]. Or, is there a more fundamental reason for the poor predictability of the Asian monsoon?

Several recent studies attempted to estimate the potential predictability of the summer monsoon by making estimate of 'internal' variability from modeling studies [*Goswami*, 1998; *Cherchi and Navarra*, 2003; *Kang et al.*, 2004; *Molteni et al.*, 2003] and from observations [*AjayaMohan and Goswami*, 2003]. Almost all these studies indicate the contribution of the 'internal' component to IAV of the summer monsoon is as large as or larger than that from the 'external' component over the Asian monsoon region. We use zonal winds at 850 hPa from NCEP/NCAR reanalysis data for the period between 1979 and 2002 and estimate the 'internal' variance of the monthly means during northern summer using the method described in *AjayaMohan and Goswami* [2003] and potential predictability is estimated as a ratio (F) between 'total' interannual variance and 'internal' variance and is shown in Fig.12a. The predictability over the Asian summer monsoon region is poor as the F ratio is less than or close to 2 over the region indicating that the 'internal' variance is comparable to or larger than 'external' variance. A similar F ratio between the the 'total' interannual variance of the seasonal mean (JJAS) precipitation and the 'internal' variance simulated by an AGCM (the LMD GCM, Sadourny and Laval 1984) from a five member ensemble simulation of 20 year duration, all forced by observed SST as boundary condition but differing only on initial conditions, is shown in Fig.12b [*Xavier et al.*, 2004]. This figure also indicates that the F ratio for precipitation during northern summer (JJAS) over the Asian monsoon region is less than or close to 2. Thus, indeed there is a fundamental reason for poor predictability of the the Asian summer monsoon. This is because as the predictable signal ('external' variance coming from slow oscillations of the coupled system) is weak over this region and comparable to the unpredictable noise ('internal' variance) over the region.

Then the question naturally arises, what is responsible for the internal IAV of the SAM? The 'internal' variability of the seasonal mean largely arise from variability of ISO activity . How does the ISOs influence the seasonal mean and its IAV? The ISOs could influence the seasonal mean if two criteria are fulfilled. Firstly, the spatial structure of the dominant ISO mode should be similar to that of the seasonal mean and secondly the frequency of occurrence of the positive and negative phases should be asymmetric. If the spatial structure of the dominant ISO mode is similar to that of the seasonal mean and if the ISOs induced IAV of the seasonal mean, the spatial structure of the IAV of the seasonal mean and that of the ISO should also be similar. The possibility is tested with precipitation from NCEP/NCAR reanalysis. Precipitation data from NCEP/NCAR reanalysis between 1948 and 2002 are used in order to get enough statistics of interannual variability of the seasonal mean monsoon. Summer ISOs are extracted from 10-90 day filtered precipitation during June 1 and September 30. A reference time series is created from 10-90 day filtered precipitation between June 1 and September 30 averaged over 70°E-90°E, 10°N-30°N. Active (break) days are identified from the reference time series normalized by its own standard deviation being greater than + 1 (< -1). The spatial structure of the dominant ISO mode is shown in Fig.13a obtained from a composite of active minus break conditions from all years. The spatial structure of the IAV of Asian summer monsoon is shown in Fig.13b constructed from composite of six strong monsoon years minus four weak monsoon years. The similarity between the two patterns (pattern correlation is 0.75) indicates that the intraseasonal and interannual variations of the summer monsoon are governed by a common spatial mode of variability. Several earlier studies [*Fennessy and Shukla*, 1994a; *Ferranti et al.*, 1997; *Ajayamohan and Goswami*, 2000; *Sperber et al.*, 2001; *Molteni et al.*, 2003] have also noted that a common mode of spatial variability governs the interannual and intraseasonal variations of the Indian summer monsoon. We, then, ask whether the probability of occurrence of the positive and negative phases of intraseasonal variability of precipitation is distinctly different in strong and weak monsoon years. Frequency distributions of 10-90 day filtered precipitation anomalies between June 1 and September 30 averaged over 70°E-90°E, 10°N-30°N for six strong monsoon years and for 4 weak monsoon years are shown in Fig.14a,and 14b respectively. The distribution is significantly biased toward positive (negative) side of anomalies during strong (weak) monsoon years. Thus, asymmetry in the frequency of occurrence of the active and break conditions is associated with stronger or weaker monsoons. This is essentially a quasilinear mechanism in which the residual of the ISO anomalies over the season is well correlated with the seasonal mean. However, contributions from individual ISO modes or seasonal mean of 10-20 day filtered or 30-90 day filtered anomalies do not correlate well with the seasonal mean [*Goswami and Xavier*, 2004b]. The aperiodicity or the broad-band nature of the spectrum (10-90 days) of the ISOs is crucial for the asymmetry in the frequency of occurrence of the active/break conditions. The aperiodicity, in turn, arises due to nonlinear interactions among the dominant modes with higher frequency disturbances. Hence, this mechanism implicitly involves some nonlinear interactions. Thus, the monsoon ISOs that are essentially driven by internal dynamics are responsible for generating enough 'internal' IAV of the SAM and limiting its predictability.

In addition to these factor, warm ocean-atmosphere interactions also contribute to the problem of monsoon predictability. Observations indicate that on seasonal time scale, the SST has a negative feedback with precipitation over the eastern Indian and west-north Pacific warm pool. However, AGCMs forced with observed SST produces a positive precipitation response. Therefore, AMIP type models are not really adequate for predicting ASM precipitation. Coupled ocean-atmosphere models would be required for achievement of better prediction skill of seasonal mean SAM precipitation. However, most CGCMs currently have large systematic bias in simulating the annual cycle over the Asian summer monsoon region. Improvement of this systematic bias is crucial before better skill of seasonal prediction could be achieved.

8 Summary of Outstanding Issues

Advances made in our understanding of the seasonal cycle, intraseasonal and interannual variabilities of the SAM have been highlighted. In this section, we summarize some of the outstanding problems and indicate some future directions.

The land-ocean surface temperature contrast model of the SAM is inadequate to explain the sustained precipitation during the SAM season and its vertical structure. A simple diagnostic model of the monsoon has been proposed by *Srinivasan* [2001] based on constraints imposed by energy and moisture balance in a vertical column of the tropical atmosphere that is able to simulate the seasonal variation of rainfall over India, Africa and south America well without explicitly invoking the concept of land-sea contrast in temperature. While such a model may not be useful for predicting the monsoon, it may be useful in diagnosing cause for systematic bias in AGCMs in simulating the SAM climate.

There is indication that the climatological mean MAC is also determined to a large extent by air-sea interactions and modified by land surface processes. However, a clear elucidation of the air-sea interaction process in the climatological mean MAC is not currently available. Careful experiments with AGCMs, OGCMs and coupled GCMs are required to unravel these processes.

A zero-order theory for scale selection and northward propagation of the monsoon ISOs is now available. However, observations indicate that the monsoon ISOs are not regular sinusoidal oscillations. In fact, there is considerable event to event variability as well as year to year variability. For better simulation and prediction of the ISOs, it is important to understand the cause of the variabilities of the ISOs. Is air-sea interaction crucial for the existence of the summer monsoon ISOs? Most AGCMs do simulate some form of summer ISO, although the simulated amplitude and northward propagation may not match with observations. It appears that the summer ISO can be generated by feedbacks within the atmosphere , the space-time spectra of which are modified by air-sea interactions. However, quantitative estimate of this modification of the space-time spectra is not available due to systematic bias of the coupled GCMs. Also it is unclear how the SST feeds back to convection over the warm waters of north IO.

Probably the most important outstanding problem is the extremely poor skill of prediction of the seasonal mean SAM precipitation by all GCMs. Serious effort is required to improve this situation. Part of the problem comes from systematic bias of the models in simulating the summer mean climatology. Some evidence has been shown that skill of predicting SAM precipitation enhances with bias correction [*Kang et al.*, 2004] and through use of super ensemble techniques [*Krishnamurti et al.*, 2003]. However, skill is still not useful. Therefore, vigorous effort in improving the SAM climate needs to be taken up.

The other factor that impedes prediction of seasonal mean SAM precipitation is the fact that the external variability (amplitude of the forced component) is rather small in this region while the amplitude of the internal variability is large. This means that the skill of prediction also depends on the model's ability to simulate the slow 'external' component of the IAV correctly. Since the signal is rather weak, a small error in simulation of the 'external' forced component in either amplitude or location may provide the 'internal' variability undue importance causing degradation of prediction. Currently most AGCMs have large systematic bias in simulating the ENSO related forced variability over the Asian monsoon region. Serious effort is also required to improve this situation.

Since the 'internal' variability is caused by the ISOs, models must simulate the statistics (amplitude, phase propagation and frequency spectra) of the ISOs correctly. It has come to light in the last couple of years that certain amount of air-sea interactions are involved with the summer ISOs. Therefore, ideally a coupled ocean-atmosphere model would be desirable for correct simulation of the observed characteristics of the summer ISOs. However, the coupled models have their own systematic bias in simulating the mean climate. The systematic bias of the coupled model could influence the statistics of the simulated ISOs. Therefore, it is not obvious that a coupled GCM would automatically improve the simulation of the ISOs. However, air-sea coupling associated with the ISOs also raises an interesting issue. If the air-sea coupling introduces certain amount of constraint on the ISOs, would it also introduce constraint on the 'internal' variability and possibly make it more predictable? Systematic studies by coupled and uncoupled AGCMs are required to answer these questions.

The ISOs emerge as a major building block of the SAM. On one hand, they modulate the day to day precipitation by clustering the synoptic activity and on the other hand they produce 'internal' IAV of the seasonal mean precipitation and limit its predictability. The establishment of extended range predictability of the ISO phases and the demonstration with simple empirical models that the dry and wet phases of ISO could be predicted up to three weeks in advance is a major development. Application of this technique to smaller regions within the country will be highly useful for agriculture planning and water management.

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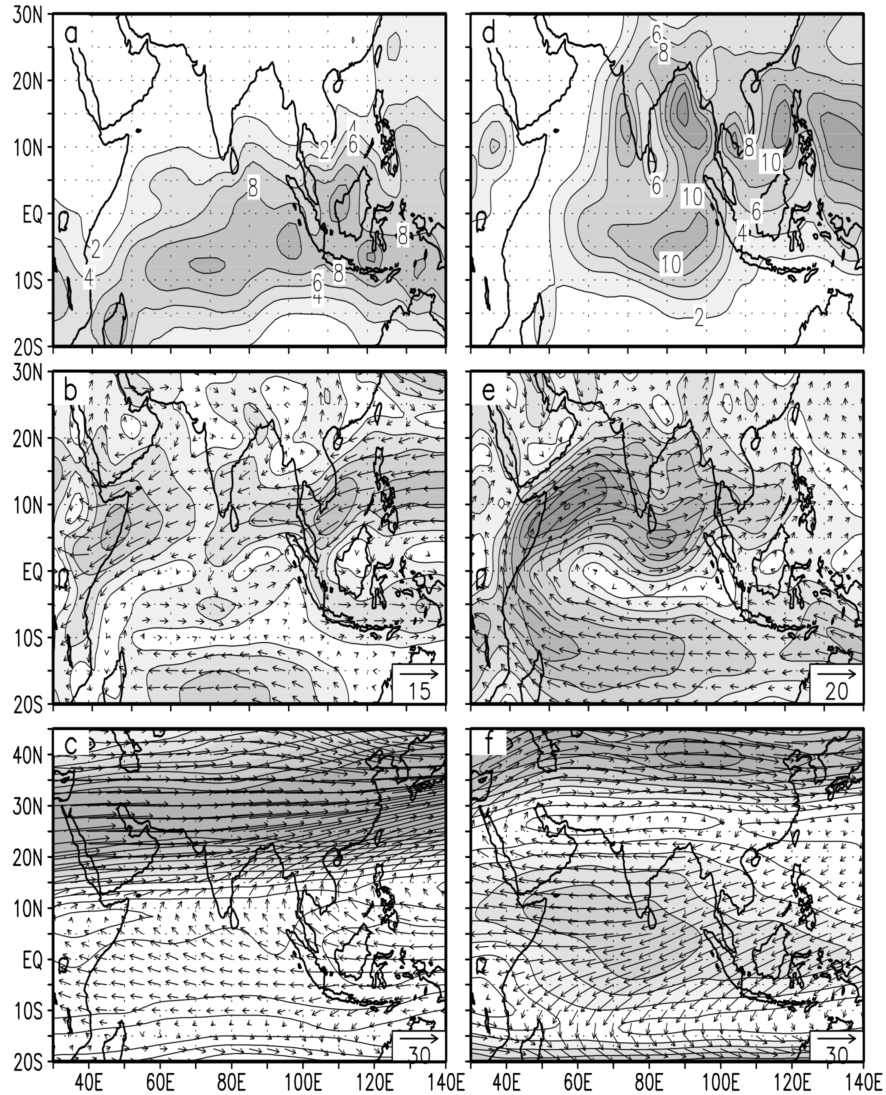

Figure 1: Climatological mean precipitation (mm day^{-1}) based on CMAP during (a) boreal winter (DJF) and (d) summer (JJAS). (b) and (e), same as (a) and (d) but for winds (ms^{-1}) at 850 hPa based on NCEP reanalysis. The contour interval for isotachs is 2 ms^{-1} with minimum contour being 2. (c) and (f) are similar to (b) and (e) but for winds at 200 hPa. The contour interval for isotachs is 5 ms^{-1} with minimum contour being 5. For better depiction of the subtropical westerly jet stream in winter and the Tibetan anticyclone in summer, a larger meridional domain is used for the 200 hPa winds (c,f)

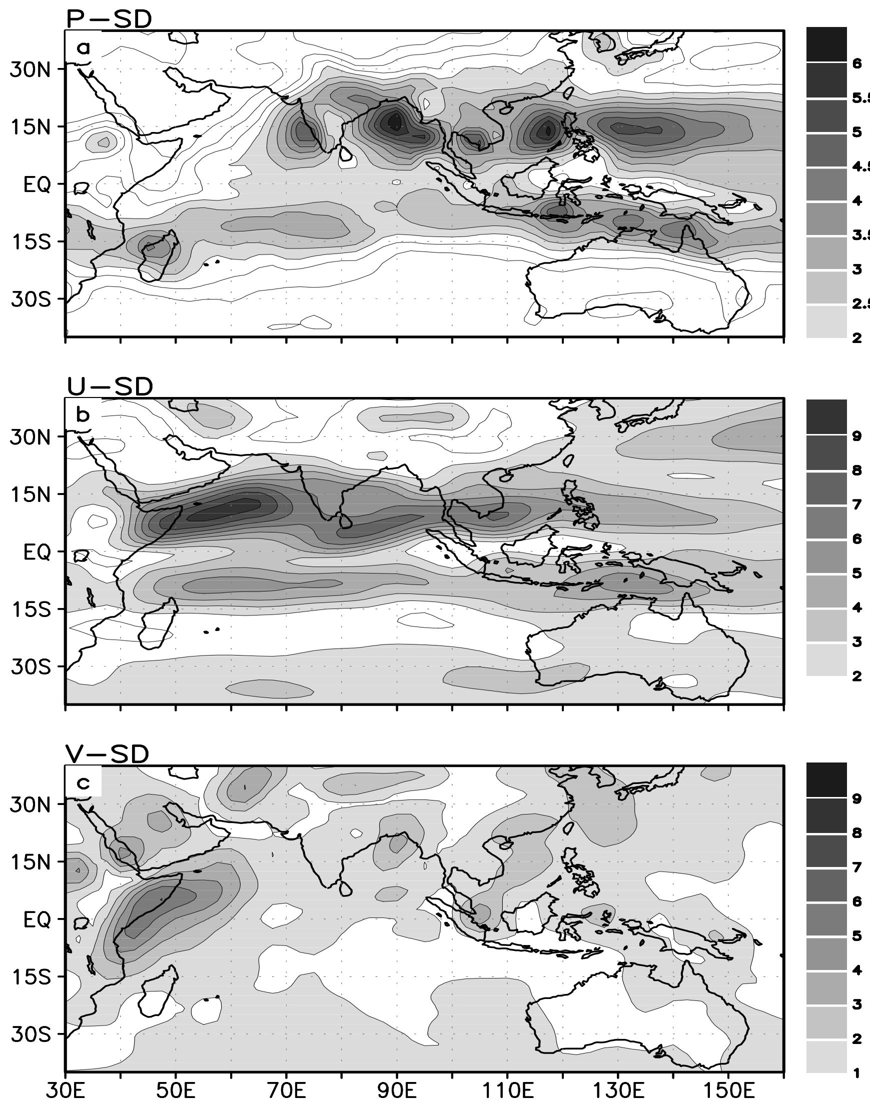

Figure 2: Amplitude of climatological mean annual cycle (AC) as defined by JJA minus DJF climatological means of (a) precipitation ($mmday^{-1}$), (b) zonal wind at 850 hPa (ms^{-1}) and (c) meridional wind at 850 hPa (ms^{-1}).

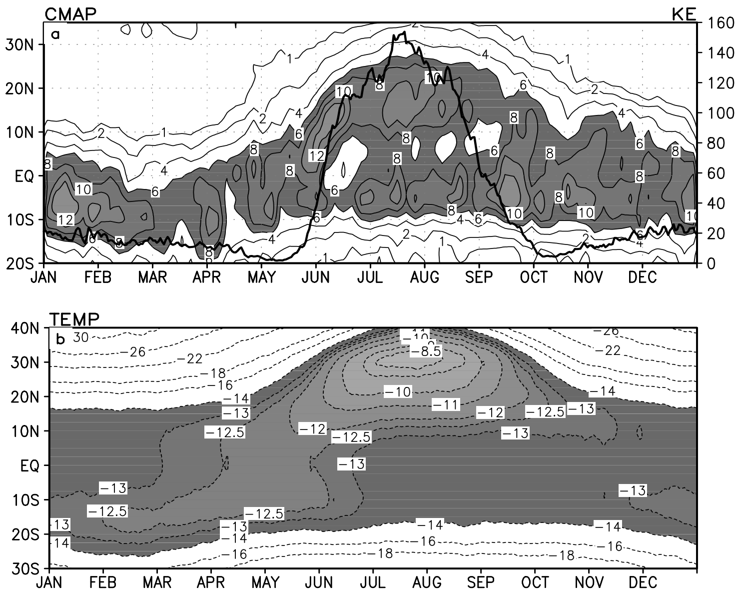

Figure 3: (a) Annual evolution of the ITCZ over the Indian monsoon region, defined by climatological precipitation ($mmday^{-1}$) averaged between 70°E and 90°E (shaded) and kinetic energy (thick solid line) of the LLJ (winds at 850 hPa averaged over 50°E-65°E, 5°N-15°N). (b) Annual evolution of climatological mean temperature averaged between 200 hPa and 700 hPa and averaged over the AA monsoon region between 30°E and 110°E in °C

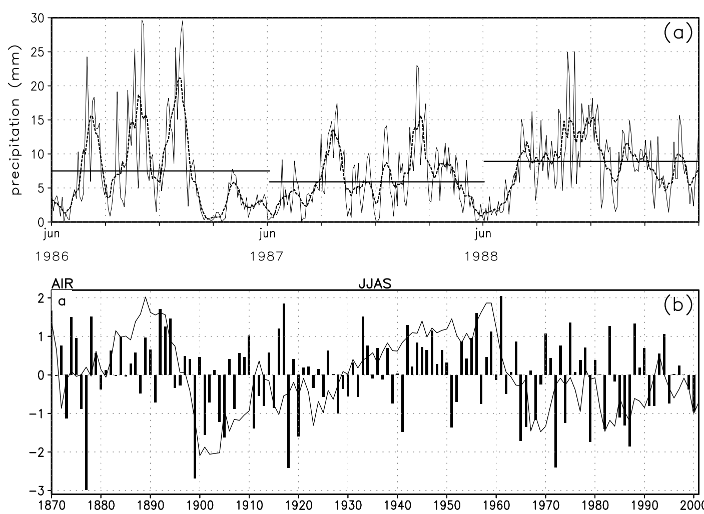

Figure 4: (a) Illustration of synoptic (thin solid) , intraseasonal (thick dotted) and interannual (thick solid) variability with daily rainfall between June 1 and September 30 for three years over central India. (b) Interannual variability of seasonal mean all India rainfall (AIR) over a longer period normalized by its own standard deviation (bar). Normalized interdecadal variability of AIR (solid line). The mean AIR is 85.1 cm and its s.d. is 8.3 cm. The s.d. og the low pass filtered AIR is 2.5 cm.

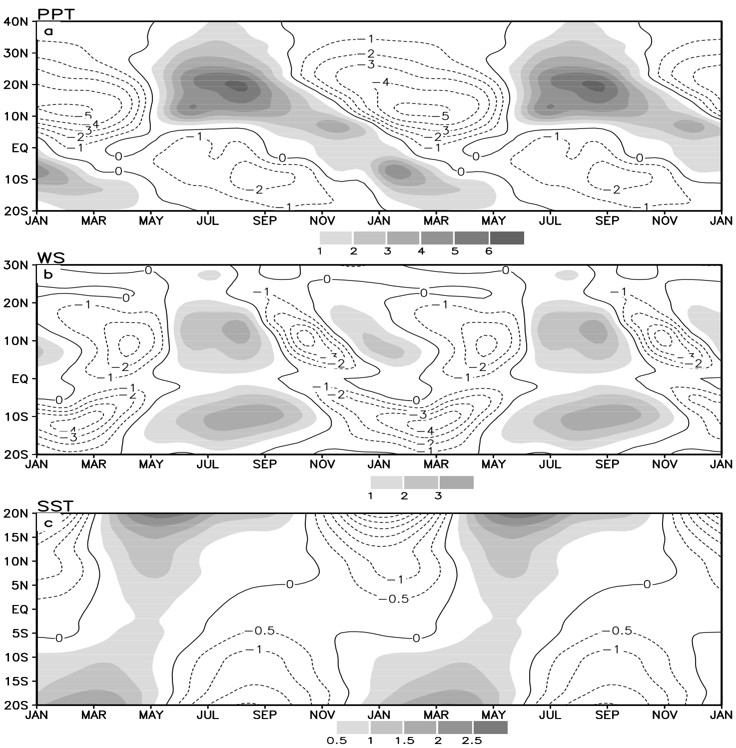

Figure 5: Annual evolution of anomaly of climatological (a) precipitation ($mmday^{-1}$), (b) surface wind speed (ms^{-1}), (c) SST (°C) averaged between 80°E-120°E . All fields are repeated for two years for clarity.

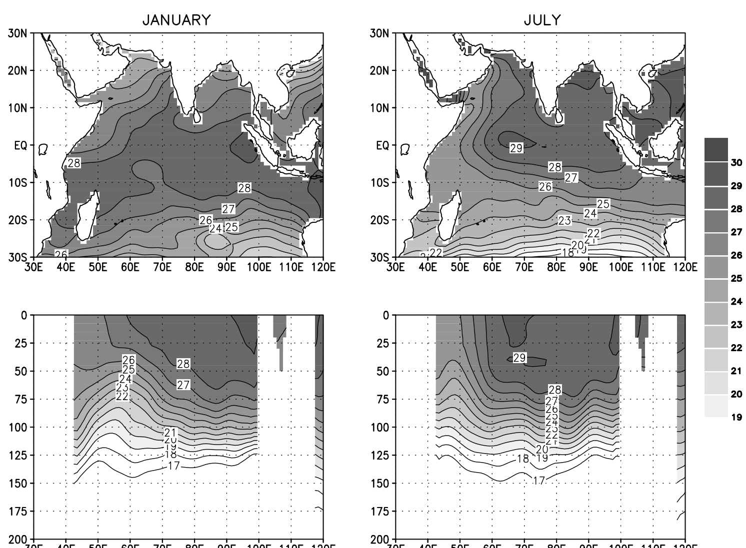

Figure 6: Climatological SST distribution over the north Indian Ocean for (a) January and (b) July. Corresponding vertical cross of temperature along the equator are shown in and (d) respectively.

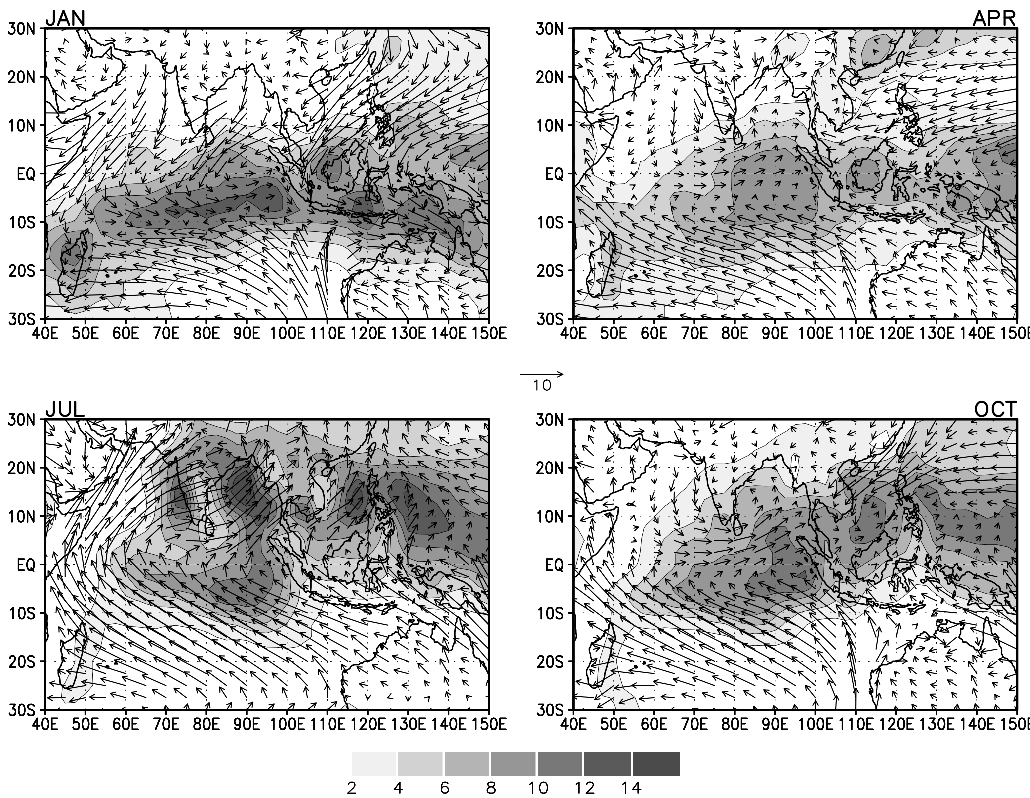

Figure 7: Climatological mean precipitation ($mmday^{-1}$) from CMAP (shaded) and surface winds (ms^{-1}) from NCEP reanalysis during (a) January (b) April , (c) July and (d) October.

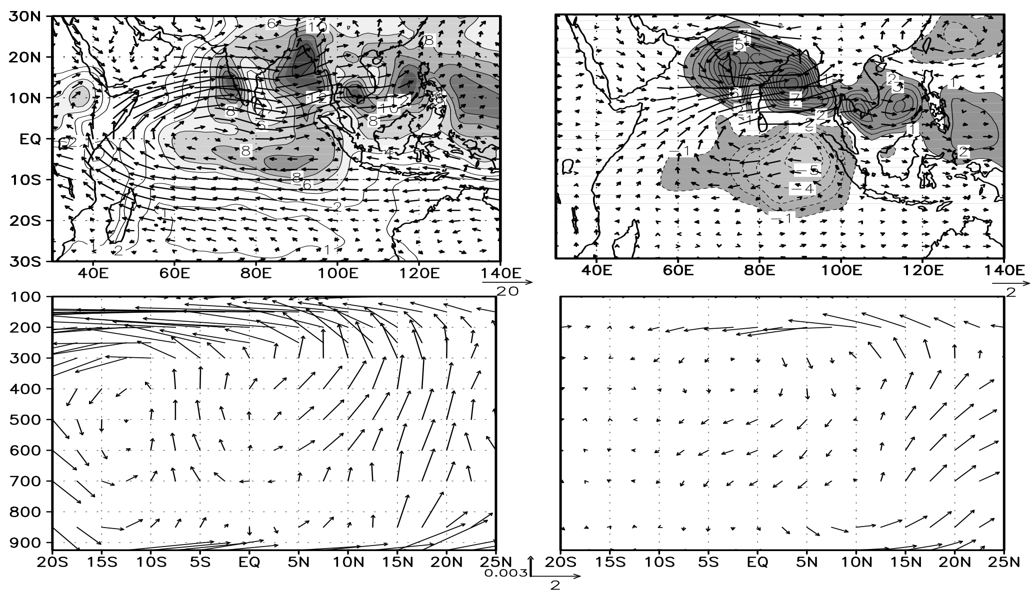

Figure 8: (a) Climatological mean JJAS precipitation and 850 hPa winds , (b) Active minus break composite of intraseasonal anomalies (10-90 day filtered) of precipitation and 850 hPa winds , (c) Climatological mean meridional circulation (monsoon Hadley circulation, MH) constructed with meridional winds and negative of pressure vertical velocity averaged over 70°E-100°E, (d) Anomalous meridional circulation similar to (c) associated with active minus break composites.

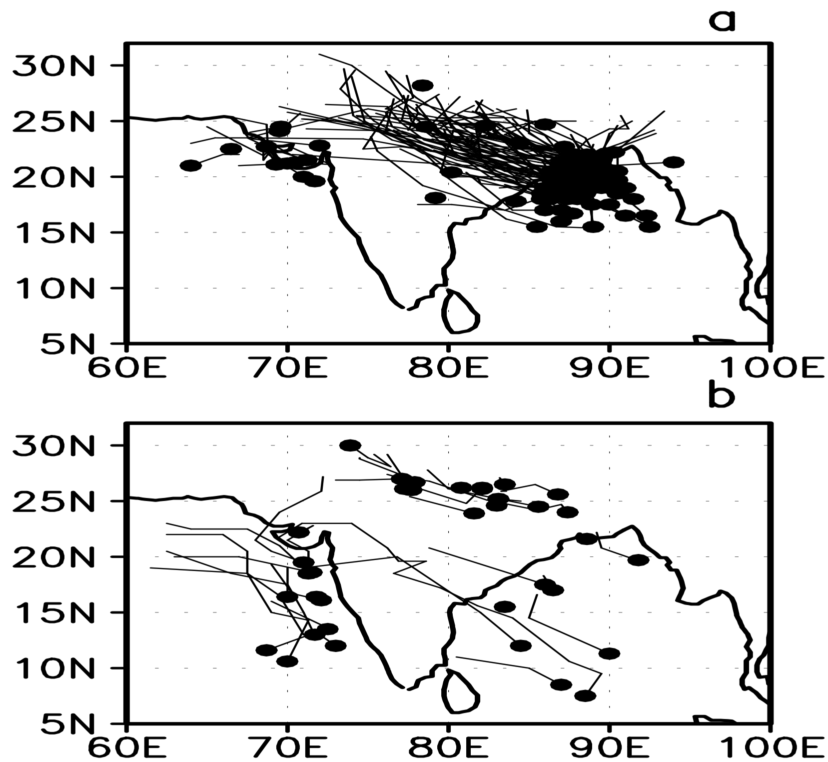

Figure 9: Tracks of LPS for the period 1954-1983 during extreme phases of monsoon ISO.(a) ′Active′ ISO phase (MISI > +1) and (b) ′Break′ ISO phase (MISI < -1). Monsoon ISO index (MISI) used here is 10-90 day filtered relative vorticity during the summer monsoon season (1 June -30 September) averaged over (80°E-95°E, 12°N-22°N). Dark dots represent the genesis point and the lines show their tracks. Large number of LPS during active phase are strongly clustered to be along the monsoon trough (MT). The few LPS that form during breaks clearly avoid the MT region and form either near the foothills of Himalaya or off the western coast and move westward. (after Goswami et al. 2003, (c) American Geophysical Union)

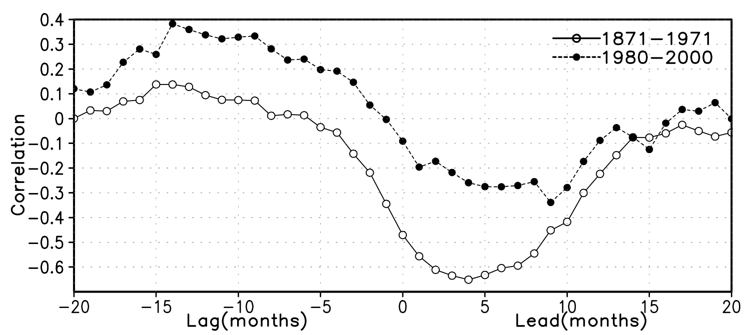

Figure 10: Lag correlations between AIR and Nino3 SSTA. The solid line is based on data from 1871 to 1977 while the dotted line is based on data between 1979 and 2002.

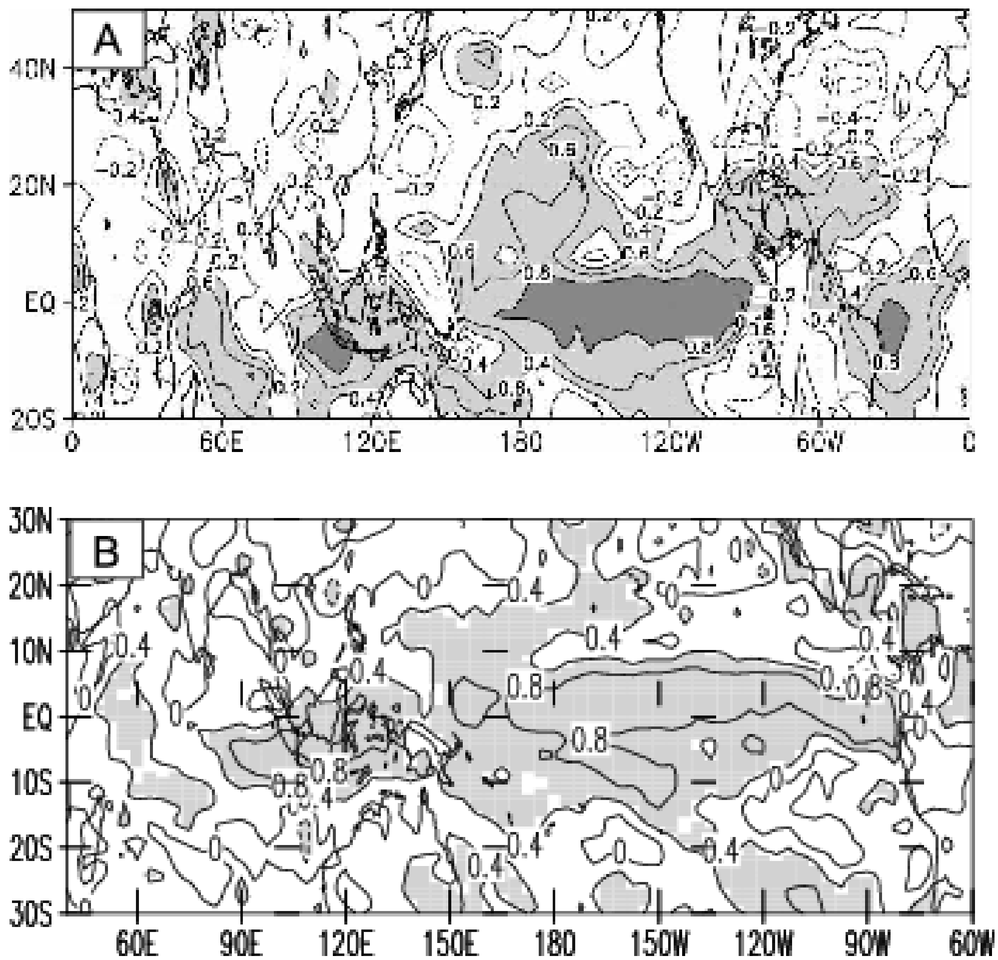

Figure 11: (A) Correlations between predictions of JJA precipitation for 21 years and observations where each prediction is a mean of a 10 member ensemble of predictions (from Kang et al. 2004, Copyright American Meteorological Society), (B) Correlations between ensemble mean (from 5 member) simulation of JJAS precipitation by LMD GCM and observation for a 20 year period (Xavier et al. 2004).

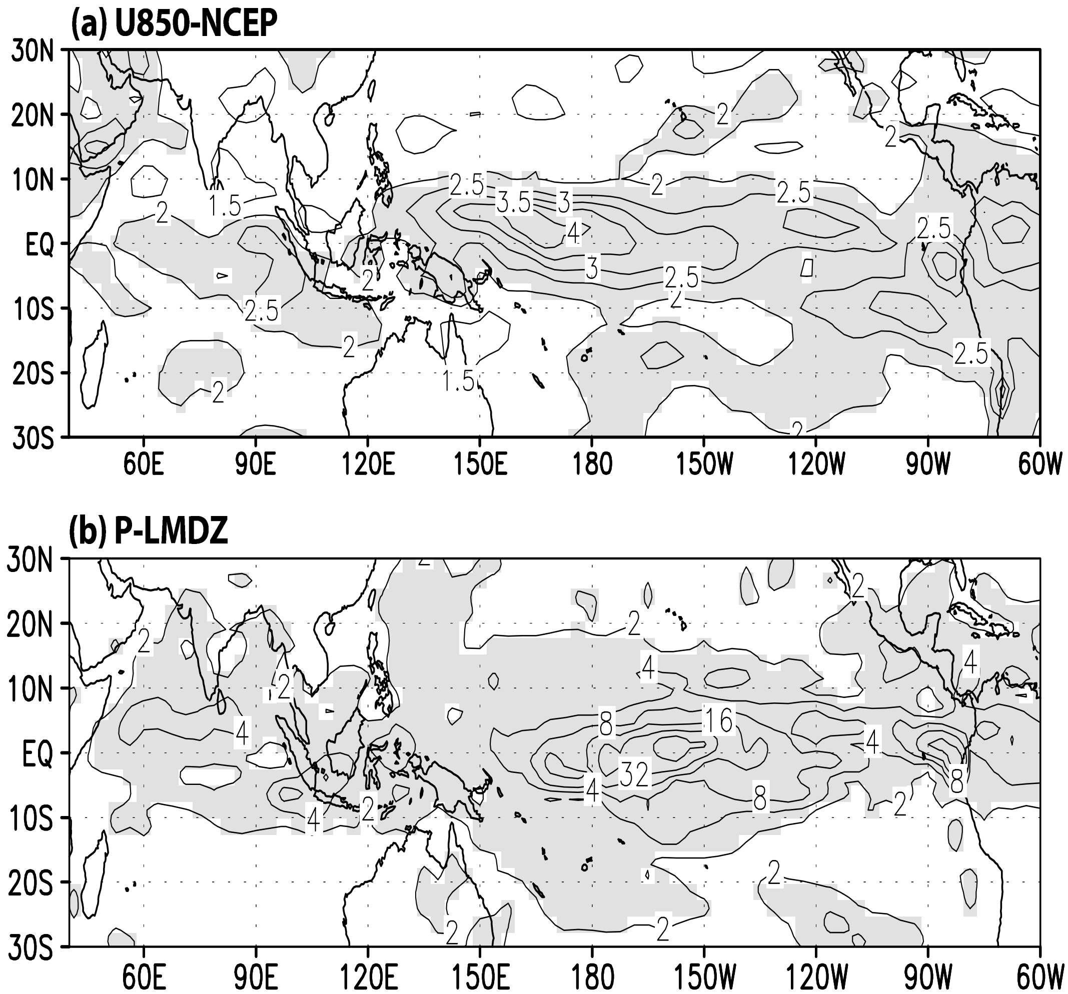

Figure 12: Estimate of predictability for (a) zonal winds at 850 hPa and (b) precipitation during northern summer (JJAS). (a) Ratio (F) between 'total' interannual variability of monthly means during June-September and 'internal' variability of zonal winds at 850 hPa based on daily NCEP/NCAR reanalysis data between 1979 and 2002. (b) Ratio (F) between 'total' interannual variability of JJAS seasonal mean precipitation from a 5 member ensemble of 20 years simulation of LMD GCM and estimate of 'internal' variability of the seasonal mean.

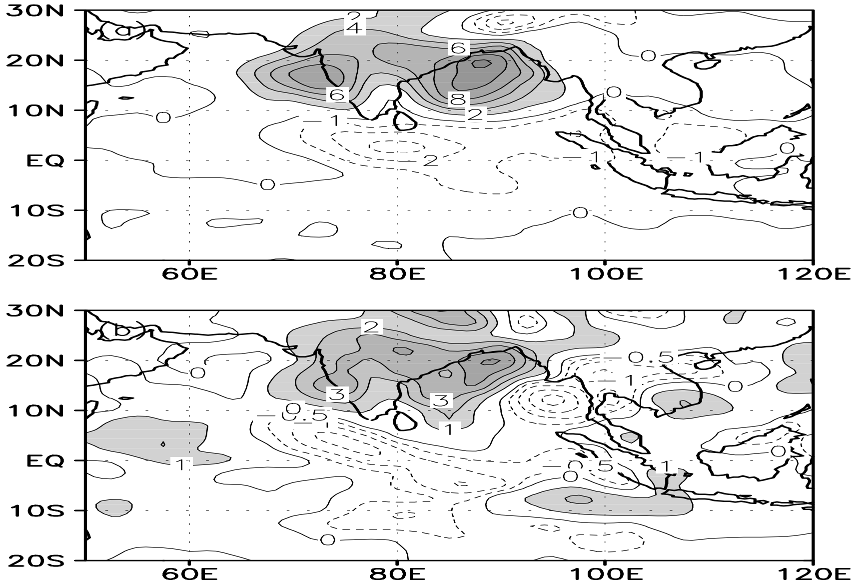

Figure 13: (a) Spatial pattern of dominant ISO mode in precipitation ($mmday^{-1}$). Composite of all 'active' minus 'break' conditions 10-90 day filtered precipitation between June 1 and September 30 for the period between 1979 and 2002. (b) Spatial pattern of dominant IAV of Asian monsoon. Composite of 'strong' minus 'weak' JJAS precipitation based on six 'strong' and four 'weak' years. See text for definition of ' active'/'break' conditions. The strong (weak) monsoon is selected from normalized time series of JJAS precipitation averaged over the area between 70°E-90°E, 10°N-30°N being greater than 1.25 (< -1.25) standard deviation.

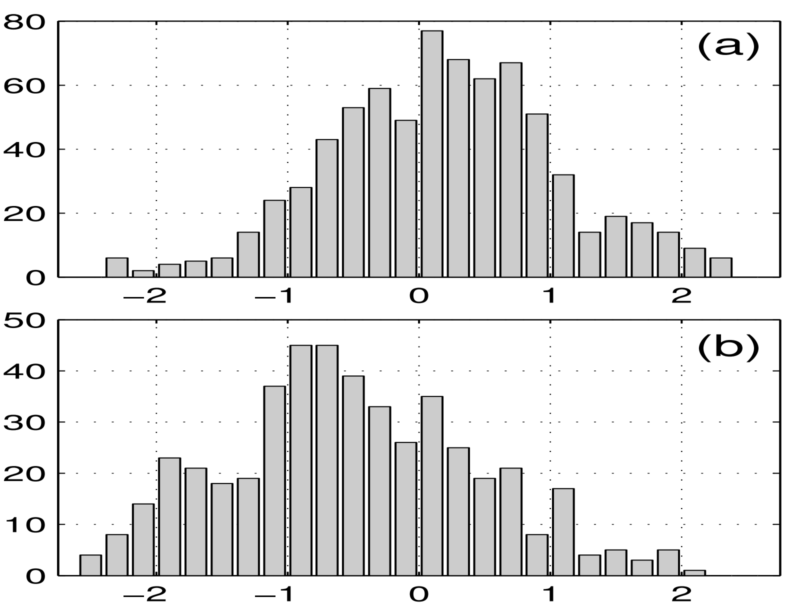

Figure 14: Frequency distribution of 10-90 day filtered precipitation anomalies averaged over 70°E-90°E, 10°N-30°N for (a) six strong monsoon years between June 1 and September 30 normalized by its own standard deviation. (b) same as (a) but for 4 weak monsoon years from NCEP/NCAR reanalysis.