The Indian Monsoon

4. Links to Cloud Systems over the Tropical Oceans

Sulochana Gadgil

In this article, I discuss the links between the variation of the monsoon and the variation of convection over the Indian and Pacific Oceans, on subseasonal scales of a few days and on an interannual scale. The relationship of the variation of the convection over the oceans to the sea surface temperature is elucidated and the implications for the Pacific and Indian Oceans considered. The role played by the El Nino and Southern Oscillation over the Pacific and the Equatorial Indian Ocean Oscillation in determining the year-to-year variation of the monsoon is discussed.

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[1] Part 3. Physics of the monsoon, *Resonance*, Vol 12, No.5, pp.4–20, 2007.

[2] Part 1. Variations in space and time (*Figure* 7), *Resonance*, Vol 11, No.8, pp.8–21, 2006.

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1. Introduction

We have seen that the monsoon is a manifestation of the seasonal migration of a planetary scale rain belt in response to the seasonal variation of the latitude at which the sun is overhead (*Figure* 1) [1]. This rain belt is associated with a planetary scale system in the tropics known as the inter-tropical convergence zone (ITCZ). The most important characteristic of the ITCZ is the strong convergence of the moist air in the lower troposphere, which leads to an ascent up to the upper troposphere and to the deep convection (clouds). The ITCZ appears as an east-west cloud band in the satellite imagery of the tropics (*Figure* 2). The large-scale rainfall over the Indian monsoon zone[2] (*Figure* 3) during the summer monsoon season is associated with the ITCZ. This ITCZ is called the continental ITCZ (CTCZ) to distinguish it from the more common ITCZ seen over the tropical oceans.

With the advent of meteorological satellites in the 70s, it became possible to literally see the planetary-scale cloud bands associated with the ITCZ on a day-to-day basis. Two important features of the variation of the ITCZ, over the Indian longitudes within the

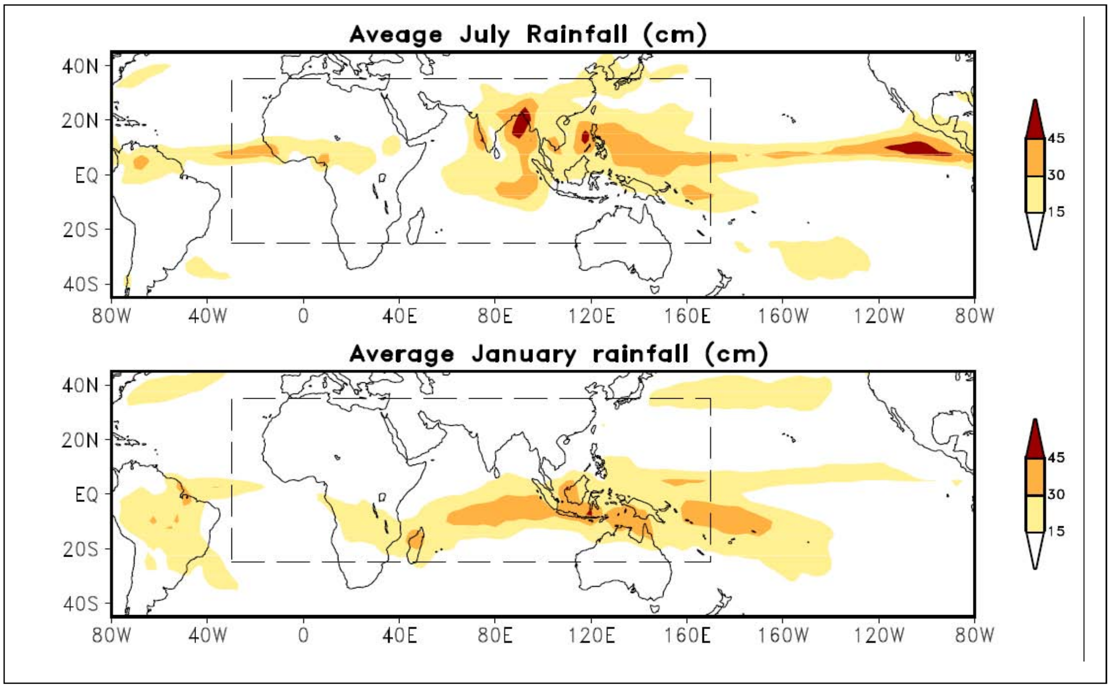

Figure 1. Average rainfall during July and January (cm).

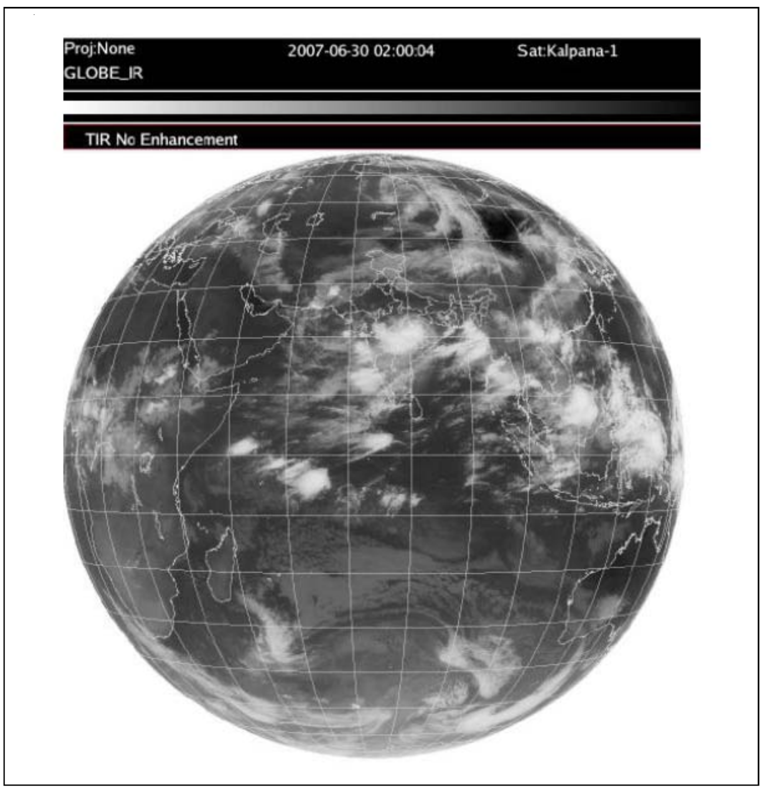

Figure 2. INSAT picture of the cloud systems on 30 June, 2007.

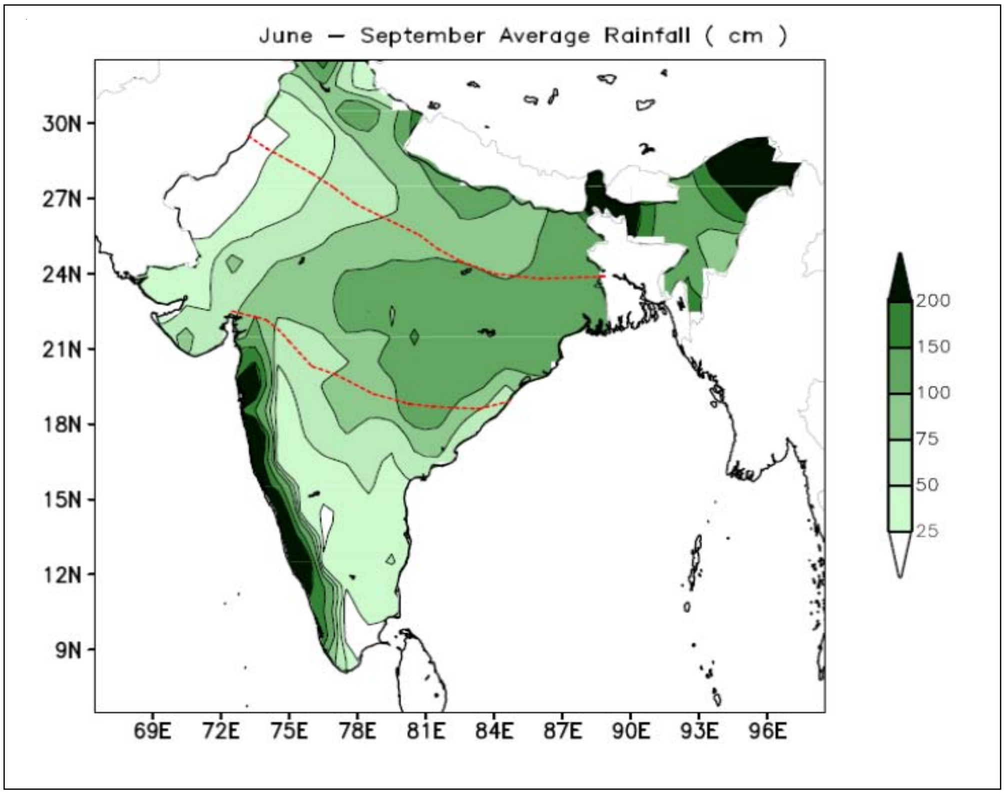

Figure **3. The average June –September rainfall (cm) over the Indian region. The boundaries of the monsoon zone, delineating the region of the large-scale monsoon rainfall, are shown as red dashed lines.**

summer monsoon season (June–September), were discovered in the first study of the daily satellite imagery [1]. Firstly, it was shown that even after the ITCZ gets established over the Indian monsoon zone in July, an east-west cloud band associated with a tropical convergence zone appears intermittently over the equatorial Indian Ocean throughout the summer monsoon season. Several times in the monsoon season, this cloud band moves northward onto the Indian region. These northward propagations of the cloud band occur at intervals varying from 2 to 6 weeks. In fact, the most prominent feature of the subseasonal variation of the cloud band/rain belt associated with the ITCZ over the Indian longitudes, is this series of northward propagations. This is seen in the daily variation in the latitudinal location of the rain belt at 90°E (longitude of Kolkatta) obtained from the satellite-derived rainfall data (*Figure* 4) for the summer monsoon of 1986 [2].

Most of the cloud systems that give us rainfall (such as lows,

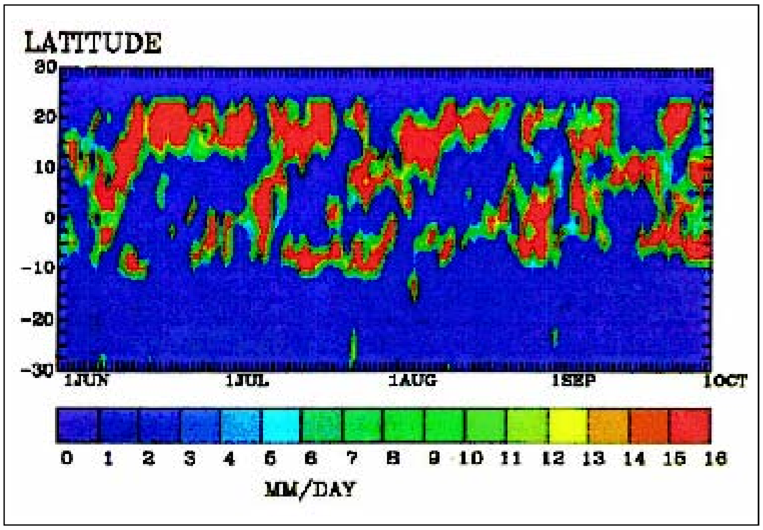

Figure 4. Variation of the satellite derived rainfall along 90˚E during June–Sepember 1986.

depressions, etc.,)[3] are generated over the oceans around us i.e., the Arabian Sea, the Bay of Bengal and the equatorial Indian Ocean. For example, the infrared imagery for August 16, 2006 (*Figure* 5) shows a cloud system over the head Bay as well as another which has moved onto the western parts of the Indian monsoon zone. These systems over the Bay of Bengal are often a result of the westward propagation of synoptic-scale low pressure

[3] Part 2. How do we get rain? *Resonance*, Vol. 11, No 11, pp.8–21, 2006.

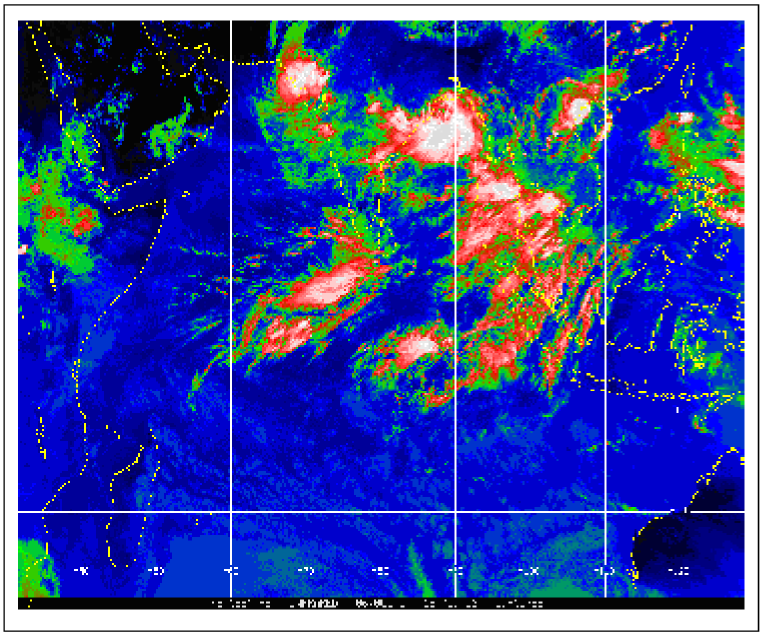

Figure 5. Infrared imagery on 16 August, 2006. Note the cloud system over the head of Bay of Bengal and another which has moved to the western part of the monsoon zone.

Figure 6. Satellite imagery of the Indian sector (above) and the northern hemisphere (below) on 8 July 1973. Note that the cloud band over the Indian monsoon zone extends right across the Pacific.

systems from the West Pacific which intensify over the Bay of Bengal. Satellite imagery also shows that on several days during the summer monsoon, the cloud band over India is a part of a very large-scale band stretching eastward up to and beyond central Pacific (*Figure* 6). Thus, we expect the variability of the monsoon rainfall to be also linked to the variability of the rainfall over the equatorial Pacific.

The northward propagations of the cloud bands from the equatorial Indian Ocean also play an important role in maintaining the cloud band associated with the CTCZ over the Indian monsoon zone. However, the contribution of the convergence zone over the equatorial Indian Ocean to the CTCZ is not entirely positive. There is also competition between these two, with active phases of one coinciding with the weak phases of the other [1].

The latent heat of condensation is a major source of energy for cloud systems.

Since the cloud systems generated over the oceans around the subcontinent as well as the Pacific make a major contribution to the rainfall we receive during the monsoon season, the variability of the monsoon on the scale of a few days (characterizing the synoptic scale systems viz., lows and depressions) as well as of about two weeks (characterizing the active spells and weak spells of the monsoon) is linked to the variability of the cloud systems (convection, henceforth) over the surrounding ocean. In fact, the prediction of rainfall a few days ahead involves the prediction of genesis, intensification and propagations of these systems over the Bay of Bengal and the Arabian Sea.

On the scale of the season also, the variation of the Indian summer monsoon rainfall (ISMR) is linked to the variation of convection over the Indian and Pacific Oceans. In this article I discuss first what determines the variability of deep convection (clouding) over the tropical oceans and then what we understand about the nature of the links between ISMR and convection over the Indian and Pacific oceans.

2. What Determines the Variability of Deep Convection over Tropical Oceans?

It has been known for a long time that the sea surface temperature (SST) plays an important role in determining the variability of convection over tropical oceans. In 1948, Palmen [3] suggested that SST has to be above a threshold of 26.5 °C for tropical disturbances to intensify to tropical cyclones or hurricanes. In a pioneering study, Bjerknes [4] showed that the variation of deep convection/rainfall over the central equatorial Pacific could be attributed to variations of SST.

Why does the SST have an impact on the variation of cloud systems over the tropical oceans? We know that for clouds to form, the water vapour in the air has to condense to form liquid water. The capacity of air to hold water vapour decreases with decreasing temperature. (That is why droplets appear on the outside of a glass which contains very cold water). The temperature of the air at the surface of the oceans is close to the SST, since the air at the surface is in contact with the sea. So we expect the quantity of water vapour in the surface air to increase with SST. How much water vapour the surface air has is important, because the latent heat of condensation which is released when water vapor condenses to form liquid water in the clouds, is a major source of energy for cloud systems. The important role played by water vapour in the atmosphere in the energetics of these cloud systems becomes evident when we consider the fate of intense tropical systems, such as hurricanes, when they come over land. These systems are generated and intensify over warm tropical oceans as they supply water vapour continuously to the surface air. Over land, the supply of water vapour to the surface air ceases and the hurricanes start dissipating. Hence it is not surprising that SST is important in determining whether cloud systems form over the oceans.

The systematic investigation of the variation of convection and its relationship with SST became possible only after the availability of satellite data. In the first such study, data on subjectively determined daily cloudiness intensity from satellite images were used to identify the total cloudiness intensity for each month for each 500 km $\times$ 500 km grid over the equatorial Indian Ocean, Bay of Bengal and Arabian Sea [5]. Analysis of these monthly data on cloudiness intensity and monthly SST obtained from ship reports showed that the relationship between SST and convection is highly nonlinear. There is a threshold of about 27.5 °C of SST above which the propensity of convection is high and below which convection tends to get suppressed. Subsequently, quantitative data on satellite-derived outgoing long-wave radiation (OLR) became available. As the height of the radiating surface (cloud top) increases, the temperature decreases and hence the OLR also decreases. Thus, very low values of OLR characterize regions with deep clouds which have tops high in the troposphere. OLR has been extensively used as a measure of moist convection in the tropics. Graham and Barnett's [6] analysis of the OLR data set showed that this nonlinear relationship, first discovered by the analysis of cloudiness data over the Indian Ocean, is a basic feature of the organized deep convection over the tropics. Studies on the variation of tropical convection over oceans with SST (based on the frequency of deep clouds identified as highly reflective clouds from satellite data), also suggest a similar and highly nonlinear relationship. Here I consider the relationship of OLR with SST.

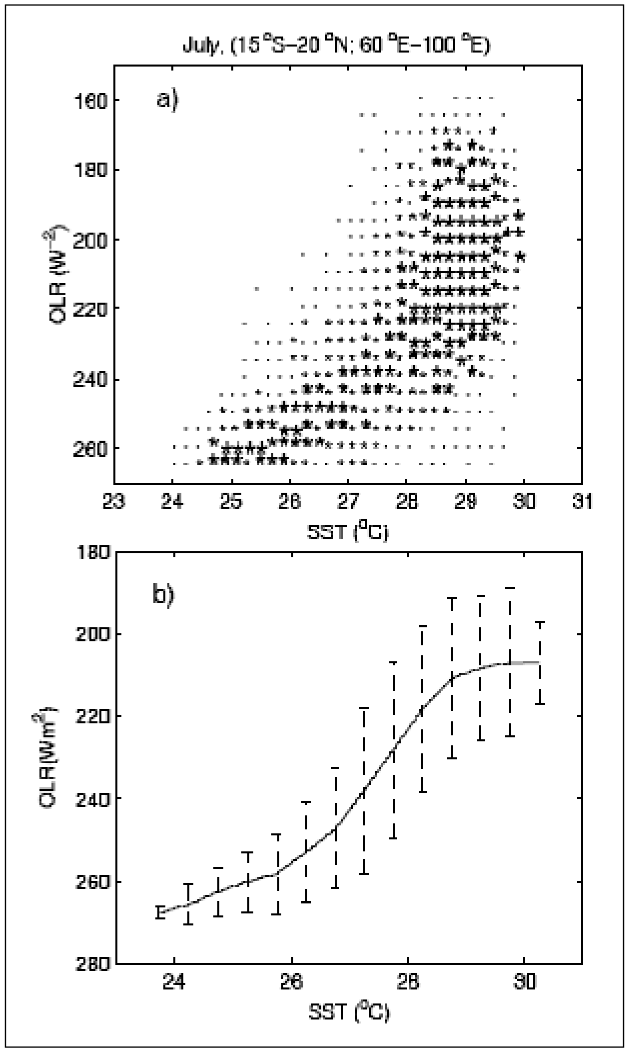

Figure 7. For July (1982–1998) over the Indian Ocean (15°S–20°N, 60°E–100°E).
(a) *Variation of OLR and SST.*
(b) *Variation of the mean and standard deviation of OLR with SST.*

The nature of the relationship of OLR with SST is seen from the scatter plot in which each grid point in the domain, for the month of July of each year, is shown at the location appropriate for the values of OLR and SST (*Figure* 7a). In this plot, the number of grid points within bins of SST and OLR are shown with symbols which become bigger and darker as the number of points in the grid increases. The mean OLR versus SST is plotted in *Figure* 7b along with the standard deviation at each SST. The frequency distributions of points in different ranges of OLR for specific SST ranges are shown in *Figure* 8. The values of monthly OLR below 240 Wm^{-2} are generally associated with deep convection. It is seen that over cold oceans (below the Palmen threshold), there are hardly any points with OLR < 240 Wm^{-2} suggesting that convection organized on the scale of the grid (250 × 250 km) is very rare for such low SST. The value

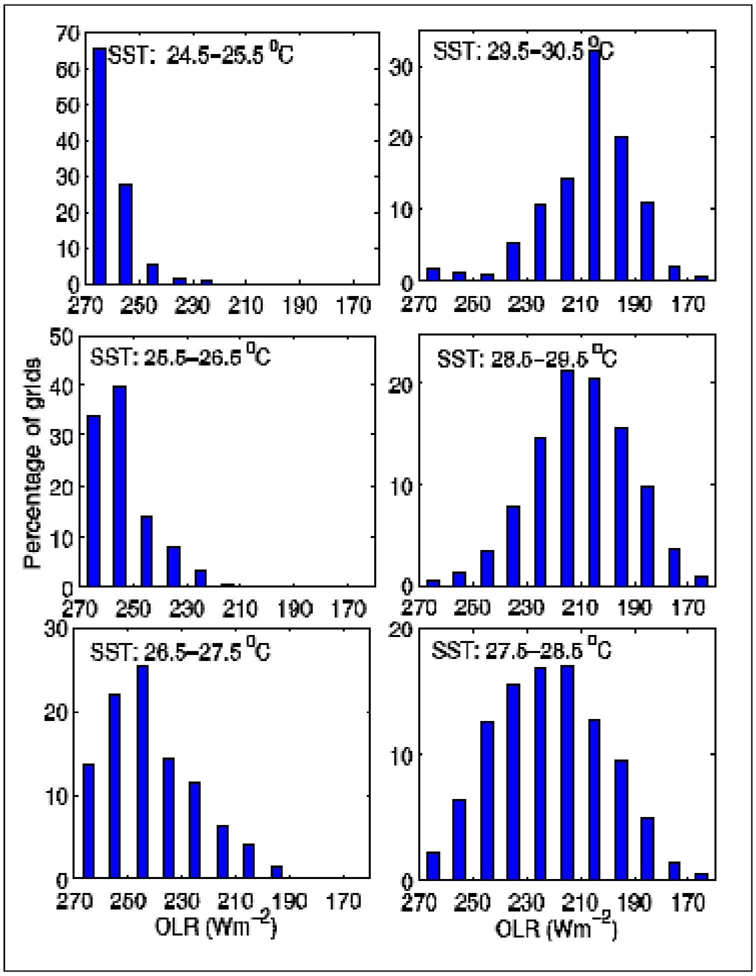

Figure 8. For July (1982–1998) over the Indian Ocean, percentage of grids in different OLR ranges for specified SST ranges.

of OLR with maximum chance of occurrence, i.e., the mode, shifts towards lower values of OLR as SST increases but remains higher than 240 Wm^{-2} for SST < 27.5 °C. Around 27.5 °C the mode shifts to values lower than 240 Wm^{-2}, implying occurrence of organized deep convection (*Figure* 8). Thus there is a threshold value of SST of about 27.5 °C, above which the propensity for deep convection, organized on the spatial scale of hundreds of kilometers, is high.

However, it is important to note that for the SST above this threshold, there is a large variation of OLR ranging from values well below 180 Wm^{-2} indicating the presence of deep convection to large OLR values indicating the absence of convection. This is not surprising, because generally even when SST and hence the specific humidity (i.e. quantity of water vapor in air) of the surface air is high, the air is generally not saturated. In fact, the relative humidity of surface air over the tropical oceans is typically about 80%. For condensation to occur and clouds to form, the air has to become saturated. If the surface air is forced to ascend adiabatically (i.e., without supplying any energy), it expands and cools and the relative humidity increases. If the ascent takes place up to a level at which the air gets saturated, then condensation can occur. Thus for the genesis of deep clouds, the dynamics has to be favourable as well [5].

Hence, SST being above the threshold is a necessary but not a sufficient condition for deep convection organised on a large scale (few hundreds of kms).

Whether the variability of convection over the oceans has a strong link to the SST variability depends on whether the SST variation is around the threshold value. The mean SST of the central Pacific is close to the threshold of 27.5 °C. So when the SST is higher than the average, the region becomes favourable for convection. The strong link between the SST and the convection/rainfall over the central Pacific, first demonstrated by Bjerknes [4], is evident by the variation in the rainfall and SST as depicted in *Figure* 9. It is seen that wet spells coincide with periods when the SST is above the threshold. However, over the Indian Ocean, the SST is always above the threshold, warmth of the ocean surface or surface air is never a limiting resource. Hence, this relationship between rainfall and SST is not seen over the Indian Ocean.

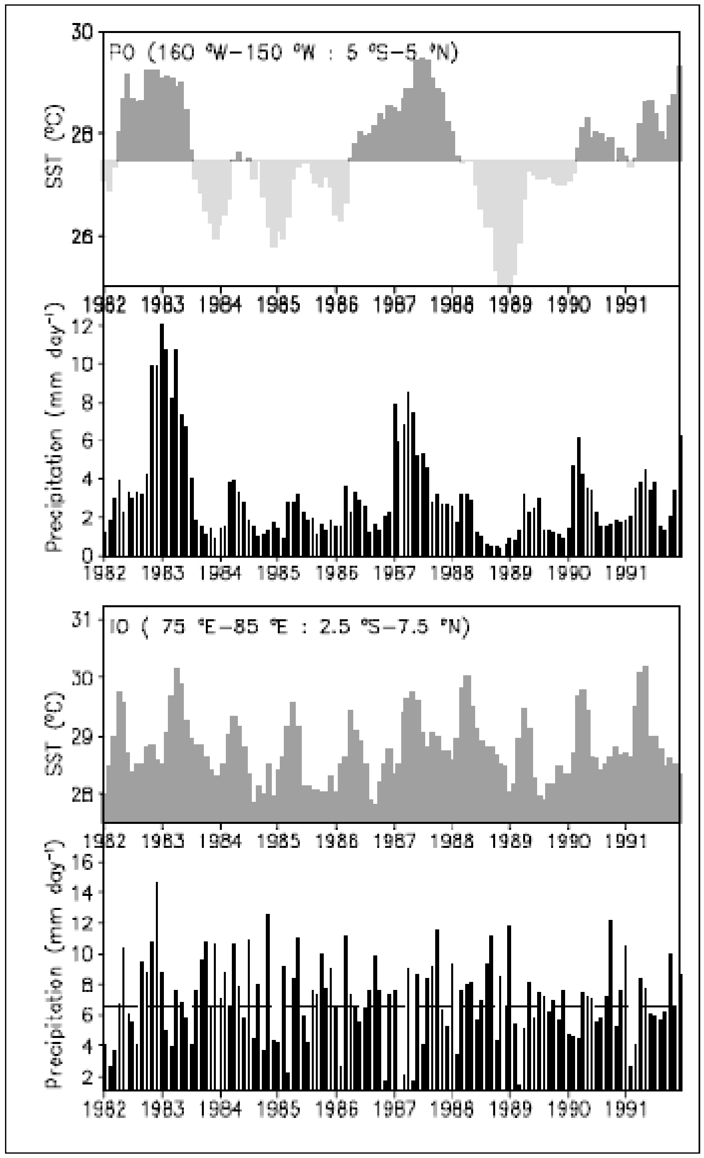

Figure 9. Variation of rainfall derived from MSU and SST over central Pacific (*above*) and equatorial Indian Ocean (*below*).

3. Links with El Nino and Southern Oscillation

El Nino and the Southern Oscillation (ENSO), is the dominant signal of interannual variation of the coupled atmosphere–ocean system over the Pacific. ENSO comprises an oscillation between

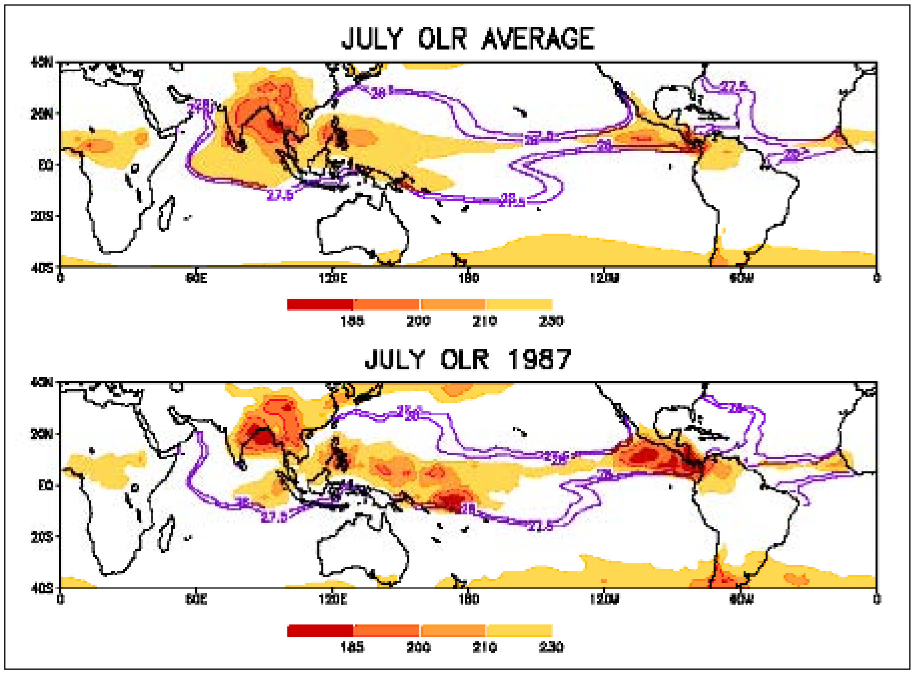

Figure 10. Regions with OLR < 230 Wm^{-2} and SST > 27.5 ˚C for July average (above) and July 1987 (El Nino) below.

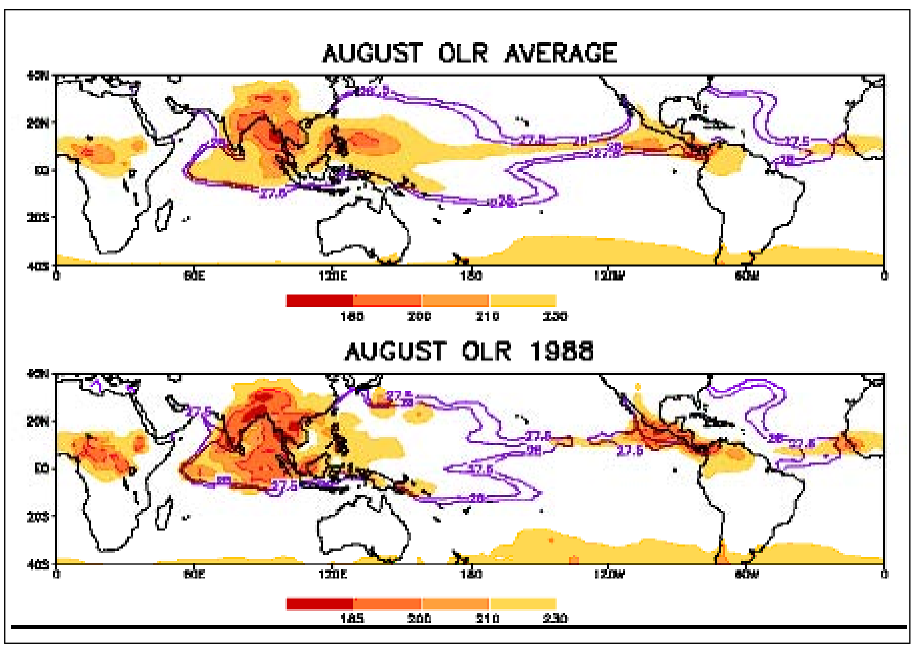

Figure 11. Regions with OLR < 230 Wm^{-2} and SST > 27.5 ˚C for August average (above) and August 1988 (La Nina) below.

El Nino, which is characterized by warmer SST over the east and the central Pacific and La Nina, which is characterized by cooler SST over the east and central Pacific [7]. The average patterns of OLR and SST (with regions of low OLR and high SST clearly indicated) for July and the patterns for July 1987 (El Nino) are shown in *Figure* 10. The average patterns of OLR and SST for August and the patterns for August 1988 (La Nina) are shown in *Figure* 11. Consider first the mean July patterns (top of *Figure* 10). As expected from the SST–convection relationship, the region of the low OLR is bounded by the SST contour of 27.5 °C. Note that in the average July pattern, there is a gap between the regions with deep convection over the western and eastern Pacific, with hardly any convection over the Central Pacific. During El Nino (*e.g.,* July 1987 in *Figure* 10), the convection over the central part of the equatorial Pacific is enhanced and the low OLR region (i.e., region with deep convection) extends as a coherent zone of convection across the Pacific. During La Nina, the convection over the central part of the equatorial Pacific is suppressed resulting in a larger gap between the convection over the western and eastern Pacific (*Figure* 11). The patterns of the OLR anomalies for July 1987 (i.e., the difference between the OLR for July 1987 and the average July OLR shown in *Figure* 10) and the pattern of the OLR anomalies for August 1988 (i.e., the difference between the actual OLR for August 1988 and the average August OLR in *Figure* 11) are shown in *Figure* 12. The suppression of convection over the central Pacific during the El Nino of 1987 and the enhancement of convection over the central Pacific during La Nina of August 1988 is clearly seen in *Figure* 12.

A major advance in our understanding of the year-to-year variation of the monsoon rainfall occurred in the 80s, with the discovery of its strong link with ENSO. It was shown that there is an increased propensity of droughts during El Nino and of excess rainfall during the opposite phase, La Nina [8, 9, 10]. During El Nino, convection is suppressed over the equatorial Indian Ocean as well as the Arabian Sea and the Bay of Bengal (*Figure* 12). On

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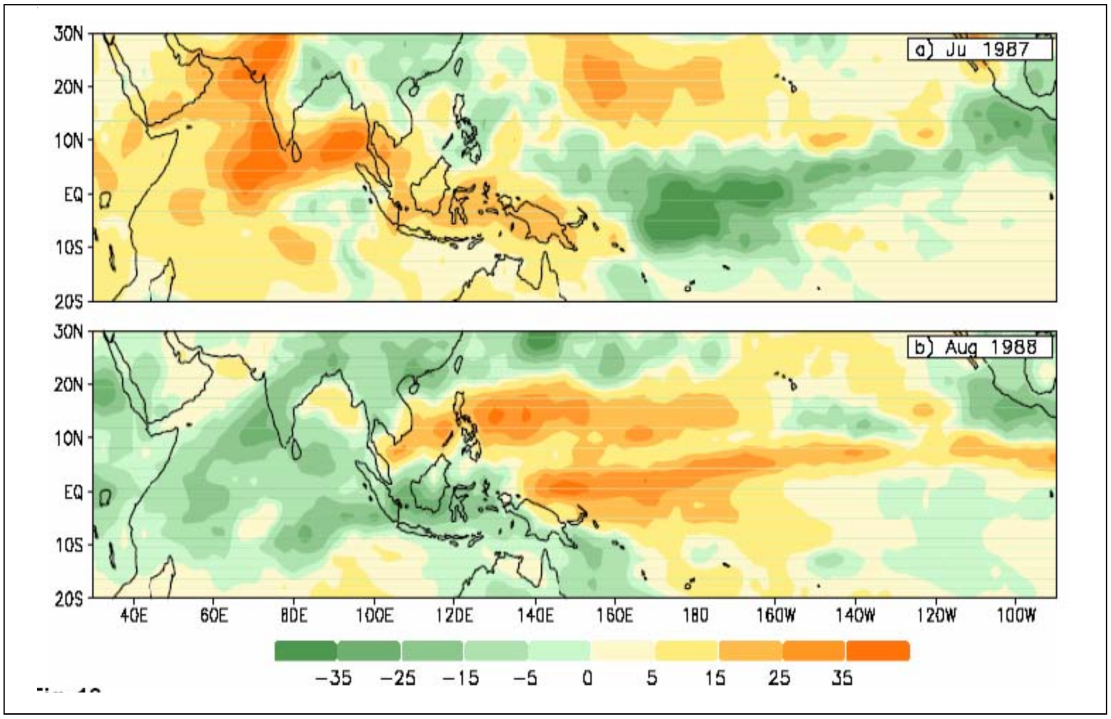

Figure 12. OLR anomaly (difference between actual and monthly average) for July 1987 and August 1988.

the other hand during La Nina, convection is enhanced over these oceanic regions (*Figure* 12). Given that convection over the ocean surrounding the Indian subcontinent is the life-line of the monsoon, it is not surprising that El Nino tends to be associated with droughts and La Nina with excess monsoon rainfall. The last two decades have witnessed phenomenal progress with the elucidation of the nature of ENSO, unraveling of the underlying mechanisms and the improved capability of predictions of the occurrence of El Nino/La Nina with models of coupled atmosphere and ocean. In view of this link between ENSO and the Indian monsoon, it was expected that the capability of prediction of the Indian monsoon will also improve. However, this did not happen. To understand why, we consider in detail the relationship between monsoon and ENSO in the satellite era.

We use an index of ENSO based on the SST anomaly (difference between the actual value and the average value) of the central Pacific. The SST anomaly is normalized by the standard deviation. The sign of the index is chosen so that positive value implies

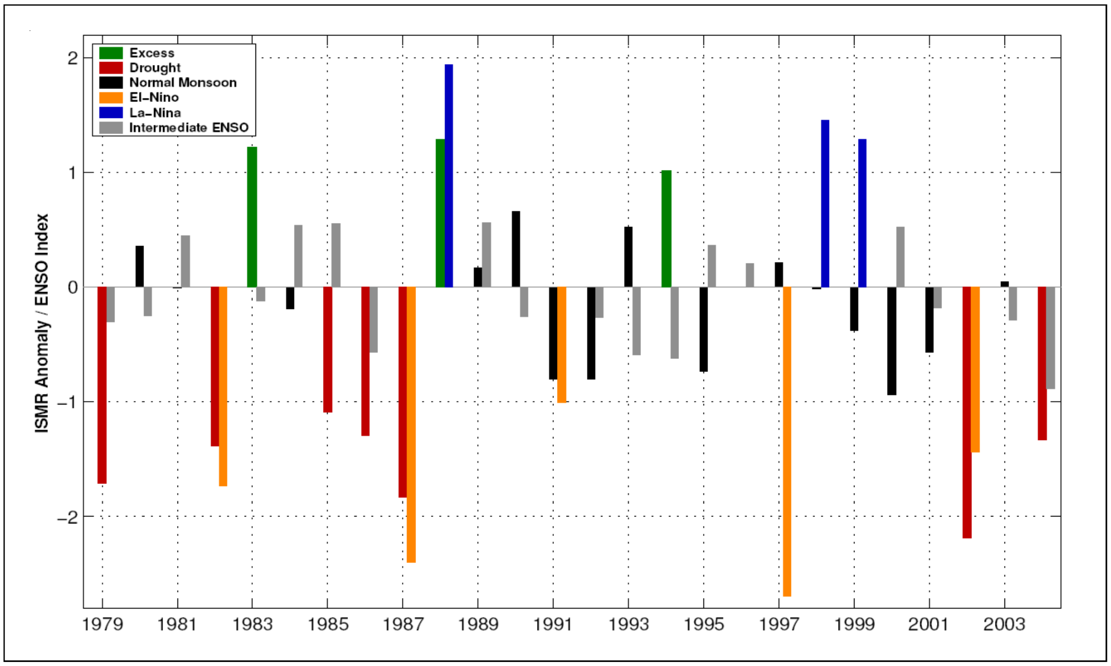

Figure 13. Normalized ISMR anomaly and ENSO index for 1979–2004.

condition favourable for the monsoon. In terms of this index, El Nino/La Nina are characterized by the negative/positive values larger than 1.0 in magnitude respectively.

The variation of the all India summer monsoon rainfall (ISMR) during 1979–2004 is shown in *Figure* 13 along with that of the ENSO index. Droughts/excess monsoon seasons correspond to a deficit/excess in ISMR of magnitude larger than one standard deviation respectively. It is seen from *Figure* 13 that, consistent with the nature of the links of the monsoon with ENSO, the La Nina event of 1988 was associated with large excess of ISMR; and the El Nino events of 1982 and 1987 with droughts. However, there was excess rainfall in 1994 despite an unfavourable ENSO and a drought in 1985 despite a favourable ENSO. Furthermore, for fourteen consecutive years beginning with 1988, there were no droughts despite the occurrence of El Nino. Finally, during the strongest El Nino event of the century in 1997, the ISMR was higher than the long term mean. This leads to the important question: Why is the variation of the ISMR not always what we would expect from the phase of ENSO? The answer lies

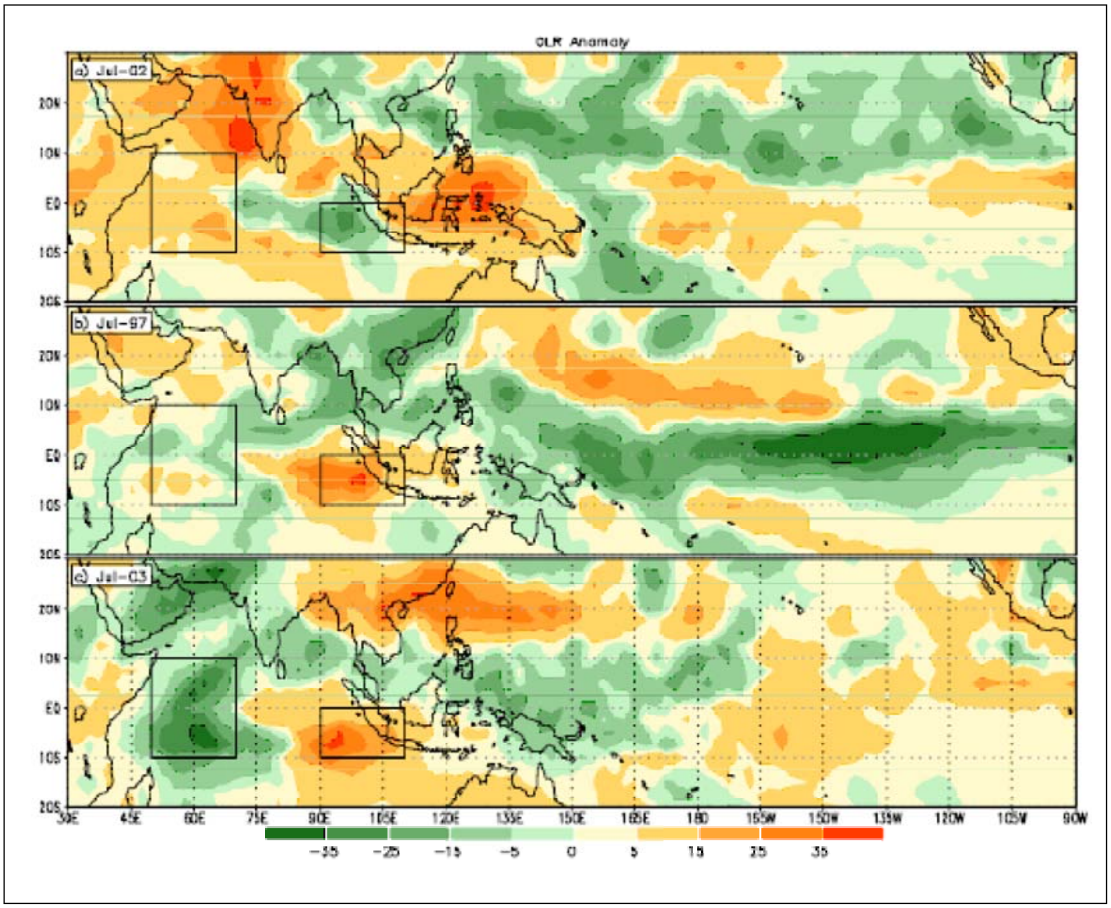

Figure 14. OLR anomaly patterns for July 1997, 2002 and 2003.

in the variation of the convection over the equatorial Indian Ocean.

4. Links with the Equatorial Indian Ocean Oscillation

In addition to the ENSO mode, in which the convection over the entire equatorial Indian Ocean is suppressed or enhanced, there is a mode in which the anomalies in convection over the eastern and western equatorial Indian Ocean are out of phase. This is seen in the OLR anomaly patterns for July 1997, 2002 and 2003 (*Figure* 14). The OLR anomaly patterns for July 1997 and 2003 have positive anomalies over the eastern equatorial Indian Ocean and negative over the western equatorial Indian Ocean. The OLR anomaly pattern of 2002 over the equatorial Indian Ocean is seen to be of opposite phase to those of 1997 and 2003. It is seen that the convection is enhanced over the Indian region when it is enhanced over the western equatorial Indian Ocean and suppressed over the eastern equatorial Indian Ocean. The anomalies of the sea level pressure and the zonal (east-west) component of the surface wind along the equator are consistent with the OLR anomalies. Thus, when convection is enhanced (negative OLR anomalies) over the western part of the equatorial Indian Ocean and reduced (positive OLR anomalies) over the eastern part, the pressure is reduced over the western part and enhanced over the eastern part. The anomalous pressure gradient (with pressure increasing from west to east) is associated with the easterly (i.e., from the east to the west) anomalies in the zonal wind. On the other hand, when convection is enhanced over the eastern part, there are westerly anomalies of the zonal wind at the equator.

We call the oscillation between these two states as the Equatorial Indian Ocean Oscillation, EQUINOO [11]. The index chosen for EQUINOO is based on the anomaly of the east-west wind over the central Equatorial Indian Ocean. This index, EQWIN, is positive for enhanced convection over the western equatorial Indian Ocean, which is favourable for the monsoon. The ISMR anomaly, ENSO index and EQWIN for all the droughts and excess rainfall seasons during 1979-2004 as well as the special year of 1997 are shown in *Figure* 15. It is seen that during the strong El Nino of 1997, EQUINOO was highly favourable. So in the tug of war between two strong opponents, the monsoon rainfall turned out to be close to the average value. The drought of 1985 occurred despite of a

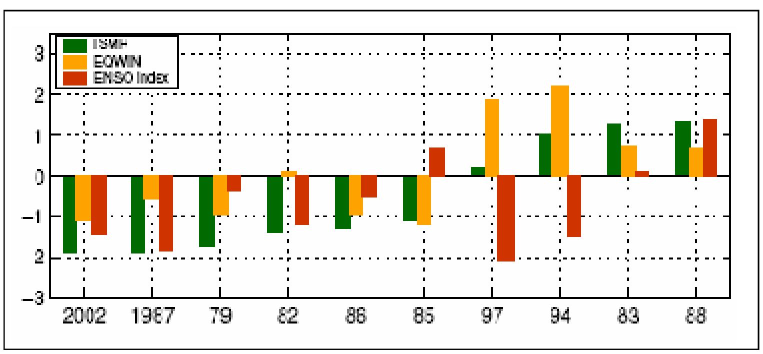

Figure 15. ISMR, EQWIN and ENSO index for all summer monsoon seasons with large deficit or excess during 1979–2004 and the special season of 1997.

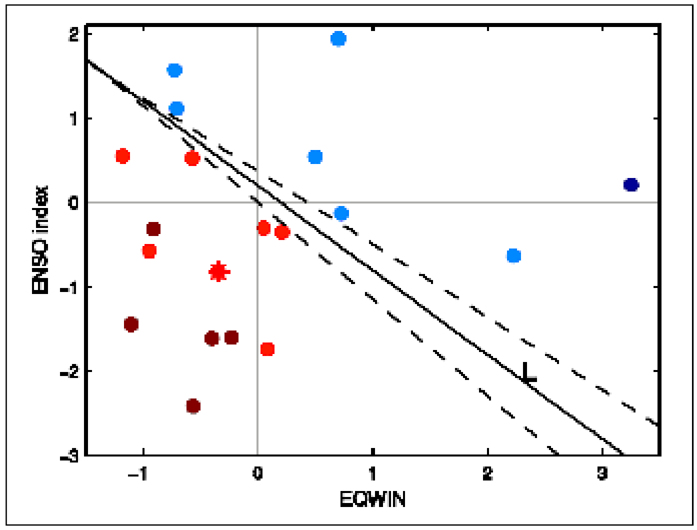

Figure 16. The extremes of ISMR (droughts red and excess seasons blue) during 1958–2004 in the phase plane of EQWIN and ENSO index.

favourable ENSO in association with unfavourable EQUINOO, whereas the excess rainfall season of 1994 was associated with a favourable EQUINOO. Thus we can understand the extremes of ISMR in terms of the favourable/unfavouable phases of one or both of the modes, ENSO and EQUINOO.

The plot of the ISMR extremes in the phase plane of EQWIN and ENSO (*Figure* 16) shows that the droughts (with ISMR deficit from the average larger than one standard deviation) are well separated from the excess rainfall seasons (with excess rainfall larger than one standard deviation). This implies that if the EQWIN and ENSO index for a specific monsoon season is below the line L, it will not be an excess rainfall season and if it is above the line it will not be a drought. Thus if the seasonal values of the ENSO index and EQWIN could be predicted, it will be possible to give a highly reliable prediction about the non-occurrence of droughts.

The year-to-year variation of the Indian summer monsoon rainfall are primarily determined by ENSO over the Pacific and EQUINOO over the Indian Ocean.

5. Concluding Remarks

We have seen that convection over the warm ocean surrounding the subcontinent is the life-line of the monsoon. Most of the systems that give us rain during the monsoon are born over this oceanic region. So prediction of the genesis of cloud systems (such as lows, depressions and large scale convergence zones), and their propagation is essential for prediction of rainfall over the Indian region, on the scale of a few days. We have also seen that the year-to-year variation of the Indian summer monsoon rainfall, and particularly the extremes i.e., droughts and excess rainfall seasons are primarily determined by ENSO over the Pacific and EQUINOO over the Indian Ocean. Prediction of these two modes is, therefore important for prediction of the summer monsoon. In the next article we will consider how predictions of the monsoon are generated and how good they are.

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