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Interactions among deep convection, sea surface temperature and radiation in the Asian monsoon region

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सार - 1983-1990 तक की अवधि के उपग्रह से प्राप्त विभिन्न आँकड़ों का उपयोग करते हुए गहन संवहन, समुद्र सतह तापमान एशियाई मानसून क्षेत्र में और विकिरण के बीच होने वाली जलवायविक अन्योन्य क्रिया की जॉच की गई है। भूमध्यरेखीय हिन्द महासागर के पूर्वी क्षेत्र और पश्चिमी प्रशांत महासागर के समीपवर्ती तथा इंडोनेशिया के क्षेत्रों में गहन संवहन की वार्षिक औसत आवृत्ति अधिकतम है। मानसून वर्षा ऋतु (जून - सितंबर) के दौरान गहन संवहन की अधिकतम आवृत्ति बंगाल की खाड़ी के क्षेत्रों में पाई जाती है।

हिंद महासागर में गहन संवहन की आवृत्ति तथा समुद्र सतह तापमान की विभिन्नताओं के बीच संबंध कमजोर पाए गए हैं। 1987 के एल निनो के दौरान उष्ण समुद्र सतह तापमान के बावजूद, अधिकांश उष्णकटिबंधीय हिन्द महासागर में गहन संवहन की क्रिया मंद पड़ गई थी। प्रशांत महासागर के बेसिन की तुलना में हिन्द महासागर के बेसिन में गहन संवहन की आवृत्ति और समुद्र सतह तापमान के बीच अन्तःवार्षिक विभिन्नता की पद्धति भिन्न भिन्न ढंग की है। एशियाई मानसून क्षेत्र में गहन संवहन मेघ विकिरण को बहुत अधिक प्रभावित करने के परिणामस्वरूप नेगेटिव विकीर्णीय वलीयन मेंघों का बड़ा घेरा बनता है। गहन संवहन आवृति (FDC) में परिवर्तन से वायुमंडल में सतह शार्टवेव मेघ विकिरण (SWCRF) और दीर्घतरंग मेघ विकिरण वलीयन (LWCRF) में 70 प्रतिशत से भी अधिक विभिन्नता का पता चलता है।

अन्तःवार्षिक स्केल पर हिन्द महासागर के समुद्र सतह तापमान के गहन संवहनों में कदाचित अनिवार्य रूप से वृद्धि नहीं हो सकती है। तथापि एशियाई मानसून क्षेत्र में गहन संवहन मेघ अन्तःवार्षिक विभिन्नता सतह वायुमंडल प्रणाली के विकिरण को काफी प्रभावित करती है। उपग्रह से प्राप्त प्रेक्षणों से यह पता चलता है कि गहन संवहन मेघों के घेरे में हुई कमी से अवशोषित सौर विकिरण में हुई वृद्धि के परिणामस्वरूप हिन्द महासागर में समुंद्र सतह ताप अधिक उष्ण हो जाता है।

ABSTRACT. The climatic interactions among deep convection, sea surface temperature and radiation in the Asian monsoon region have been examined using various satellite-derived data sets of the period 1983-90. Annual average Frequency of Deep Convection (FDC) is maximum over the equatorial east Indian ocean and adjoining west Pacific and Indonesian region. Maximum FDC zone shifts to Bay of Bengal during the monsoon (June-September) season.

There is a weak relationship between the variations in FDC and SST in the Indian ocean. Deep convective activity was suppressed over most of the tropical Indian ocean during El Niño of 1987 in spite of warmer SSTs. The pattern of interannual variation between FDC and SST behaves differently in the Indian ocean basin as compared to the Pacific ocean basin. Deep convective clouds interact with radiation very effectively in the Asian monsoon region to cause large net negative cloud radiative forcing. Variation in FDC explains more than 70% of the variation in surface shortwave cloud radiative forcing (SWCRF) and long wave cloud radiative forcing (LWCRF) in the atmosphere.

On inter-annual scale, warmer SSTs may not necessarily increase deep convection in the Indian ocean. However, the inter-annual variation of deep convective clouds influences significantly the radiative budget of the surface-atmosphere system in the Asian monsoon region. The satellite observations suggest that warmer SSTs in the Indian ocean might have resulted from an increase in the absorbed solar radiation at the surface due to a reduction in deep convective cloud cover.

Key words – Cloud radiative forcing, Deep convective cloud, Asian monsoon region, Sea surface temperature, Cloud-climate interaction.

1. Introduction

Cloud -climate interaction is an area of significant uncertainty in projecting future climate change caused by anthropogenic changes. Variation of clouds associated with climate change can either amplify or decrease the direct radiative forcing due to increase in greenhouse gases (Mitchell *et al.* 1989, Wetherald and Manabe 1988). A recent study by Cess *et al.* (1990) showed that there are still large disagreements in the longwave and shortwave components of cloud feedbacks in GCM simulations.

Clouds exert a strong influence on radiative transfer within the earth's atmosphere. The Cloud Radiative Forcing (CRF) affects circulations of the atmosphere and ocean by altering surface energy fluxes and atmospheric heating rates. Many studies (Slingo and Slingo 1988, Randall *et al.* 1989, Sherwood *et al.* 1994) have confirmed that tropical circulations are substantially altered if CRF is neglected. CRF at the surface is a major component of surface energy fluxes and contributes directly to ocean circulations (Chen *et al.* 1994).

Global distributions of cloud radiative forcing at the top of the atmosphere (TOA) can be derived from satellite measurements(Ramanathan 1987, Ramanathan *et al.* 1989). Previous studies (Kiehl 1994, Weare 1995, Pai and Rajeevan 1998, Rajeevan and Srinivasan 2000) suggest that both Short-wave(SW) and Long-wave(LW) cloud radiative forcing is caused by high clouds(clouds with top below 440 hPa) which include the deep convective clouds. Kiehl and Ramanathan (1990) and Kiehl (1994) showed that over the deep convective regions in the west Pacific, there is a near cancellation between SW and LW cloud radiative forcing. Rajeevan and Srinivasan (2000) showed that conclusion of near cancellation between and Long-wave Cloud radiative forcing (LWCRF) and Short-wave cloud radiative forcing(SWCRF) has been invalid in the Asian monsoon region during June-September due to the presence of optically thick high clouds, with amounts exceeding 50%. It was also shown that Asian monsoon region is unique in the tropical region for having large amounts of optically thick high and deep convective clouds thus causing large negative net cloud forcing.

Another aspect of special interest is the relationship between deep convection and sea surface temperatures (Ramanathan and Collins 1991, Fu *et al.* 1996, Hartmann and Michelson 1993). The variation of tropical convection over oceans with SST was examined by Gadgil *et al.* (1984),Graham and Barnett (1987) and Waliser *et al.* (1993). They have suggested a highly non-linear relationship between SST and tropical convection. Bhat *et al.* (1996) suggested a strong link between the frequency of tropical convection and Convective Available Potential Energy (CAPE). Zhang *et al.* (1996) studied the global as well as regional aspects of the relationship between cloud radiative forcing and SSTs. They showed that warmer tropical oceans as a whole are associated with less long wave greenhouse effect of clouds and less cloud reflection of solar radiation to the space.

The present study is designed to examine (*i*) the spatial and temporal variation of Frequency of Deep Convection (FDC) in the Asian monsoon region, (*ii*) the local and remote (teleconnection) response of the deep convective clouds to sea surface temperatures (*iii*) the interaction between the deep convective clouds and radiation in the Asian monsoon region and (*iv*) the inter annual variability of FDC and its relationship with SST and its role on modulating the surface and atmospheric energy budget.

2. Data sets and methodology

2.1. *Frequency of Deep Convection (FDC)*

In this study, observations of high optically thick cloud cover derived from the International Satellite Cloud Climatology Project (ISCCP) was used as a proxy of deep convection. Deep convective cloud is diagnosed by using the ISCCP C2 monthly mean cloud data set (Rossow and Schiffer 1991). Deep convective clouds are defined as those cloudy pixels for which the optical thickness is greater than 22.4 and the cloud top pressure is lower than 440 hPa. The optical thickness threshold is approximately equivalent to an albedo of 0.7 (Fu *et al.* 1990). FDC represents the number of pixels classified as deep convective clouds in a $2.5^{\circ} \times 2.5^{\circ}$ latitude-longitude grid cell which is at a scale comparable to a deep convective cluster and can be interpreted as an index of fractional amount of deep convective clouds. The features of FDC in the ISCCP data set are consistent with other satellite estimates of convective activity (Waliser *et al.* 1993).

2.2. *Radiative Fluxes and SST*

The SW and LW radiative fluxes at TOA have been taken from the Earth Radiation Budget Experiment (ERBE) (Barkstrom 1984). The data include the fluxes under both all sky as well as clear sky conditions gridded at a $2.5^{\circ} \times 2.5^{\circ}$ latitude X longitude resolution. In the ERBE data, there are few grid boxes with missing clear sky fluxes. These missing clear sky fluxes are estimated by filling with the mean of the clear sky fluxes at the nearest grid boxes (Zhang *et al.* 1996). Since the clear sky fluxes over the oceans are relatively uniform on small

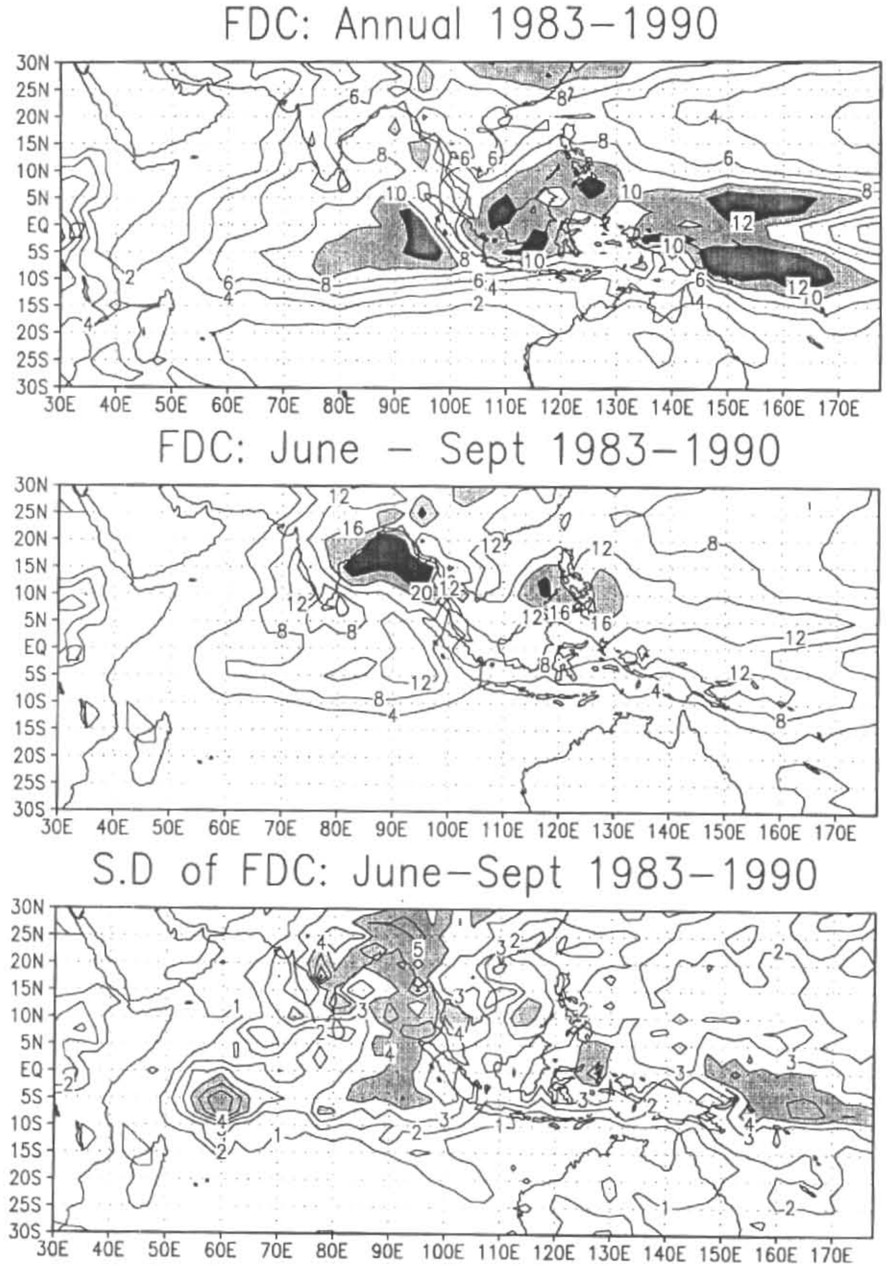


Fig.1. Geographical maps of (a) the annual average frequency of deep convection (FDC) (top), (b) average FDC during summer monsoon (June-September) (middle) and (c) standard deviation of FDC during the summer monsoon (bottom). Period : 1983-90. Contour interval 2%(top), 4%(middle) and 1%(bottom)

scales, the errors resulting from the filling are expected to be small.

The surface radiative fluxes cannot be measured by satellites. However they can be realistically determined using a radiative transfer model using the observed atmospheric parameters and cloud fields and observed TOA radiative fluxes measured by the satellites. The surface SW radiative flux data was taken from the Version 1.1 Surface Radiation Budget (SRB) SW products for the period March 1985 through December 1988. Inputs to this product are from the ISCCP and ERBE. It uses two

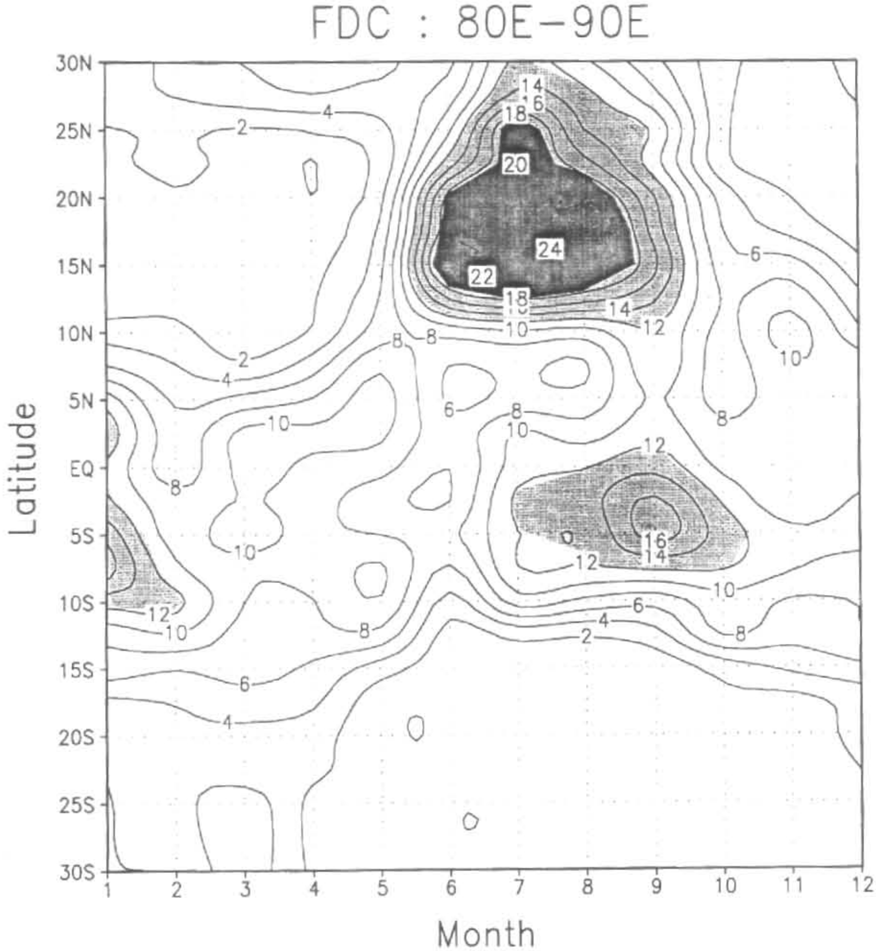


Fig.2. Latitude-month variation of FDC averaged between the longitudes 80°E-90°E. Period : 1983-90. Contour interval 2%

methods known as Pinker algorithm and Staylor algorithm to estimate surface SW radiative fluxes (Whitlock *et al.* 1995). In this study, the Pinker algorithm was used. In this method, the all sky and clear sky upward and downward fluxes at TOA and surface were computed from the mean cloudy and clear radiances and from cloud fractions of the ISCCP C1 data, using the shortwave radiation budget algorithm (SASRAB) developed by Pinker and Laszlo (1992). The differences in atmospheric absorption when compared to high resolution computations are about 2% and 7% for the clear and cloudy cases respectively. The monthly mean LW fluxes at the surface were taken from the NCEP reanalysis data (Kalnay *et al.* 1996). These radiative fluxes are estimated using the radiative transfer model of the NCEP numerical model.

The SW cloud radiative forcing (SWCRF) and LW cloud radiative forcing (LWCRF) at the TOA are calculated as SWCRF= S (A_{clr} - A) and LWCRF= F_{clr}–F, where the A_{clr} is the clear sky albedo and F_{clr} is the clear sky long wave radiative flux at TOA. S is the incoming solar radiation at the top of the atmosphere. Surface CRF is also defined as the difference between surface radiative fluxes under all-sky and clear sky conditions.

The TOA (T), surface (S) and atmospheric (A) CRF are related as

$$CRF(A) = CRF\ (T) - CRF(S)$$

Positive CRF parameters indicate a warming of the system and negative values indicate a cooling.

The SST data was taken from the monthly analysis of the National Centre for Environmental Prediction (NCEP), USA, which is derived from ship, buoy and satellite measurements (Reynolds 1988). This data set provides monthly mean SST on a 2° × 2° spatial grid. This

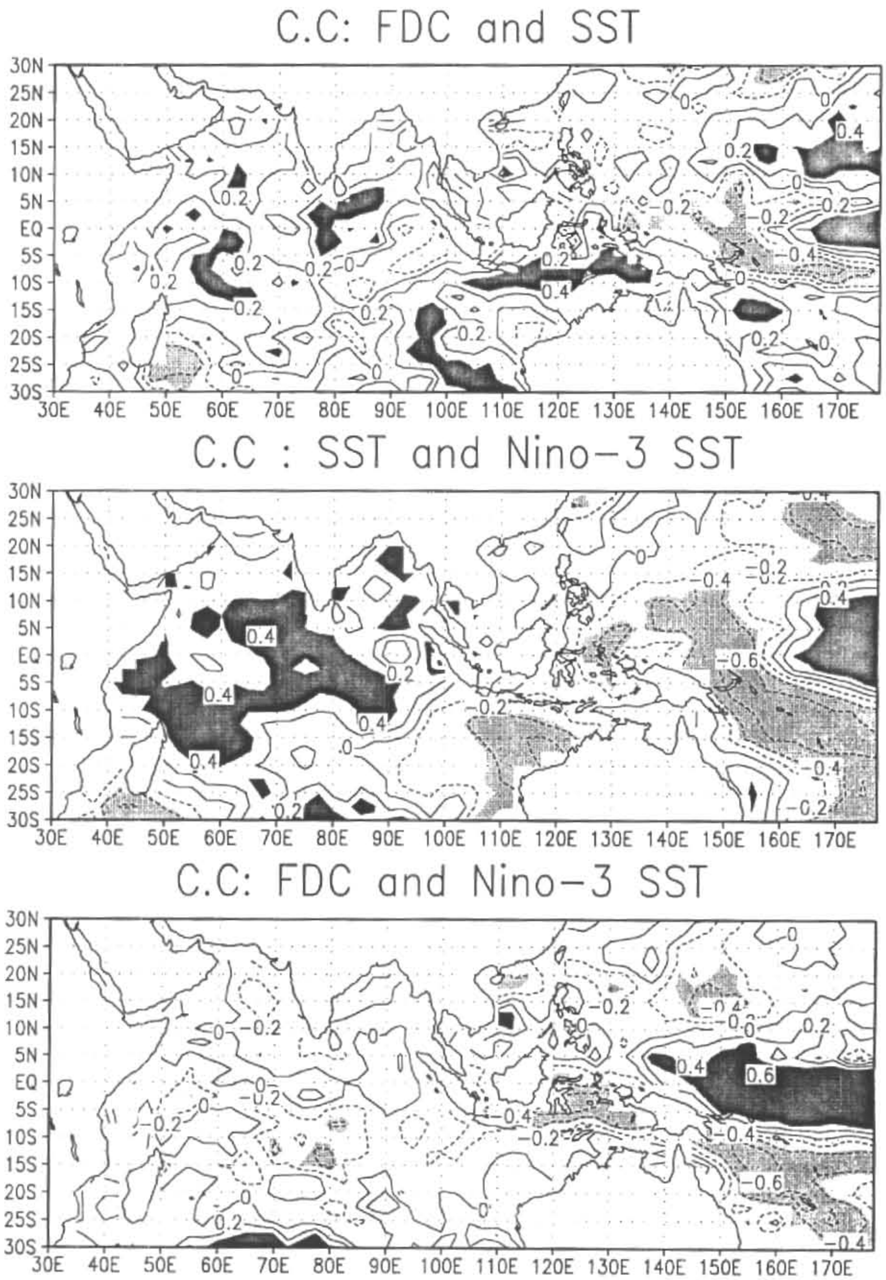


Fig.3. Geographical maps of correlation coefficients between (a) local SST and FDC anomalies (top), (b) SST anomalies and Nino-3 SST (middle) and (c) FDC anomalies and Nino-3 SST(bottom). Period: June-September, 1983-90. Contour interval : 0.2 . Negative (positive) CCs are indicated by dotted (continous) lines. Negative (positive) CCs exceeding 0.35 are shaded light(dark)

data is available in the GEDEX CD-ROM. This data set was then interpolated into $2.5^{\circ} \times 2.5^{\circ}$ grids of ERBE and ISCCP data.

FDC data are available from June 1983 to December 1990, whereas the ERBE and surface radiative budget data are available from 1985 to 1988.

3. Results and discussion

3.1. *Spatial and seasonal variations of FDC*

The spatial pattern of annual frequency of deep convection is shown in Fig.1 along with the FDC during the monsoon period (June-September) and its standard

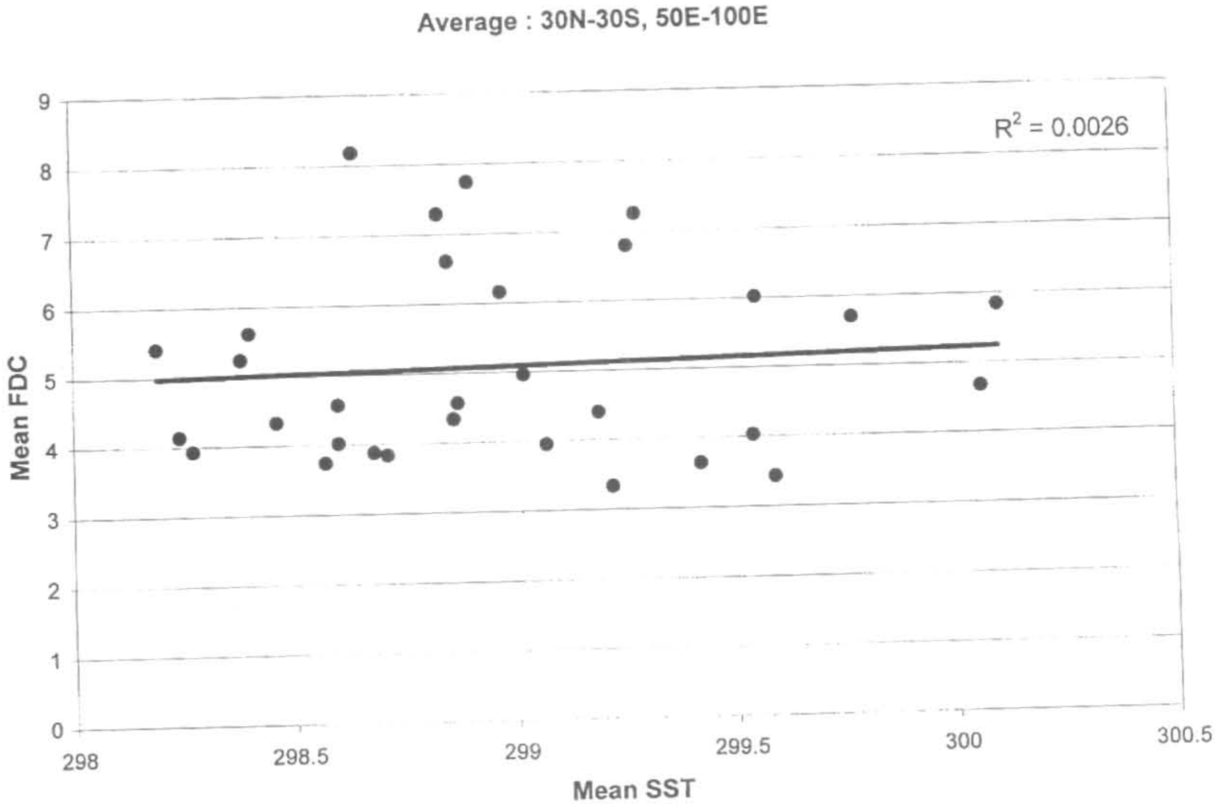


Fig.4. Scatter between monthly and area averaged (30° S-30° N, 50° E –100° E) SST and FDC . Period: June-Sept, 1983-90

deviation. Annual FDC is maximum (exceeding 12%) close to equator over the east Indian ocean and adjoining west pacific and Indonesian region. During the summer monsoon period of June-September, maximum FDC (exceeding 20%) is observed over Bay of Bengal. Another smaller area of maximum is observed over the west Pacific ocean. Over the Indian region, FDC decreases rapidly towards the west. Maximum standard deviation exceeding 5% is observed over Bay of Bengal and the equatorial Indian ocean. The latitude-month variation of FDC averaged between the longitudes 80°E and 90°E is shown in Fig.2, which clearly shows the northward progression of maximum FDC from the equatorial region to Bay of Bengal from March to July. Maximum FDC exceeding 20% is observed over north Bay of Bengal from June to August. The zone of maximum FDC shifts to the equatorial region once the summer monsoon is withdrawn from the Indian region.

3.2. *Sea surface temperature and deep convection*

The local and remote correlations between SST and deep convection during the monsoon period, June–September were examined to explore the relative influence of local SST versus large scale circulation. Local correlation coefficients calculated between monthly mean SST anomalies and FDC anomalies, are shown in Fig.3a. Positive (negative) correlations suggest that warmer SSTs increase (decrease) deep convective clouds. Significant positive correlations between SST and FDC are observed over the central equatorial Pacific, east Indian ocean and over small regions in the Arabian Sea and south Bay of Bengal. However over Indonesia and adjoining east Indian ocean, variations in convection tend to be out of phase with variations in SST. The weak correlations exist between SST and FDC over most of Arabian Sea and Bay of Bengal, indicating a weak relationship between FDC and SST over these regions.

SST anomalies over the central and east Pacific ocean associated with the ENSO events are responsible for large scale circulation anomalies especially over the tropics. To understand the response of the variations in FDC over Asian monsoon region to the SST anomalies over the east Pacific ocean, remote correlations between Nino-3 SST [the mean anomalous SSTs of the

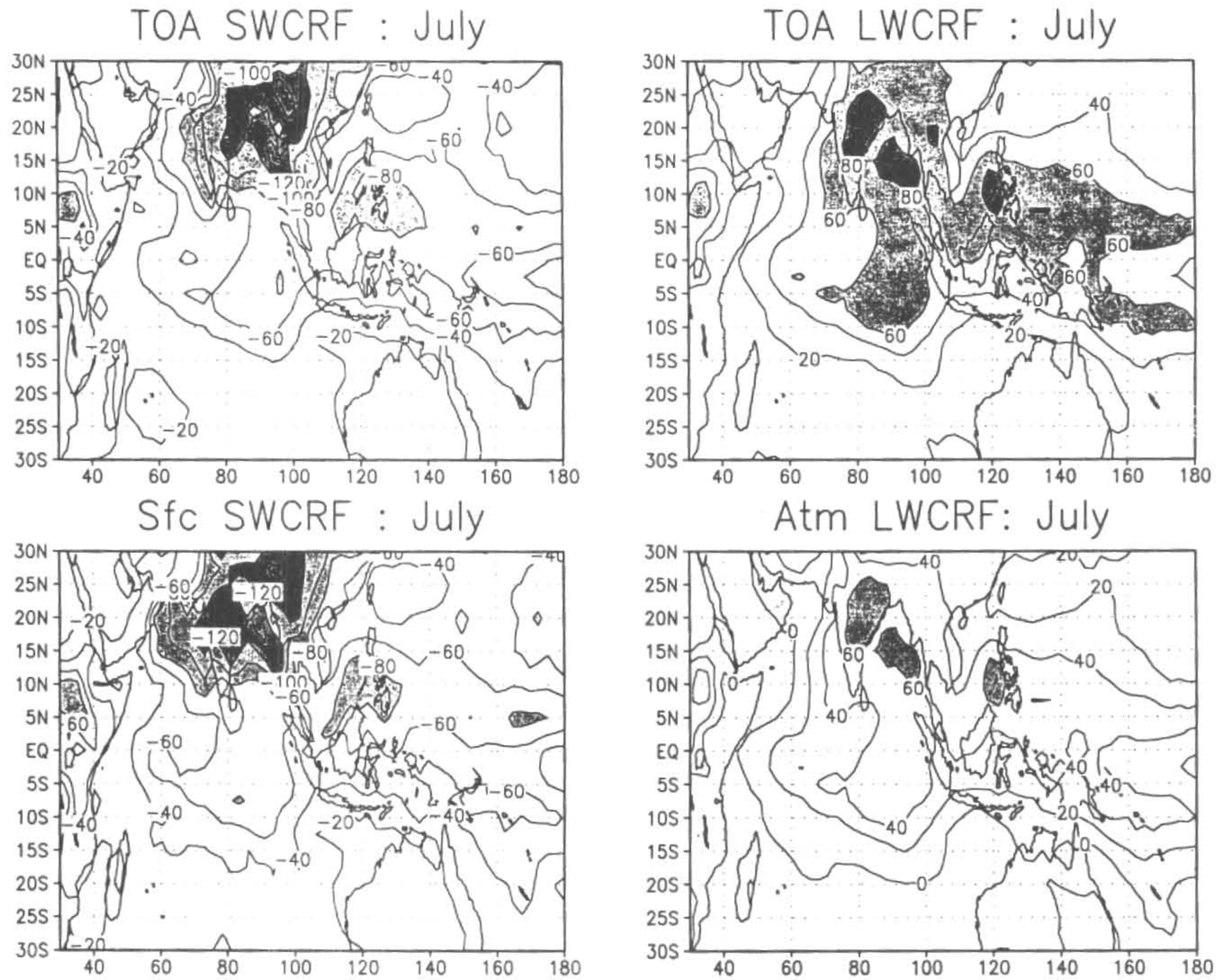
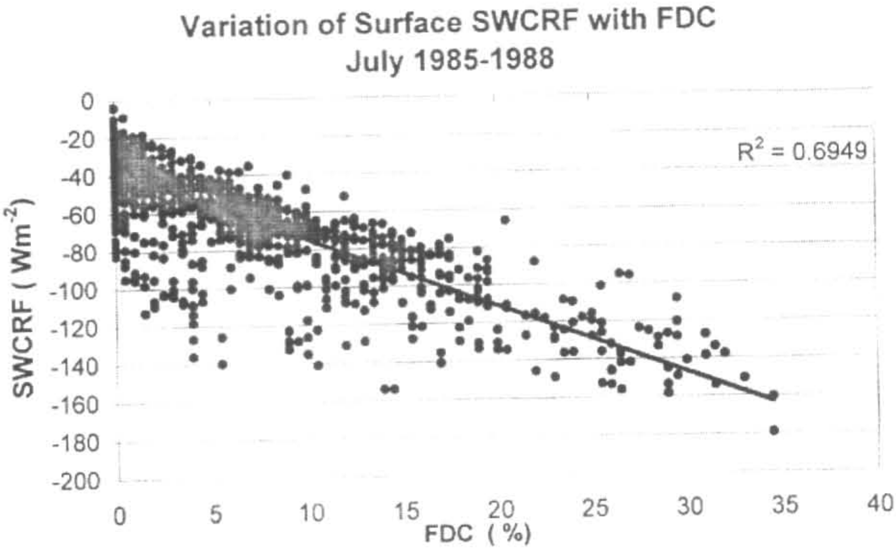


Fig.5. Geographical maps of (a) SWCRF at TOA, (b) LWCRF at TOA, (c)SWCRF at the surface and (d) Atmosphere LWCRF. Period: July, 1985-88. Contour interval: 20 Wm^{-2}

Nino-3 area (5°S- 5°N, 150°W – 90° W)] and FDC were computed for the monsoon period, 1983-90 and are shown in Fig. 3 (c). In the same diagram (Fig 3.b), also shown are the remote correlations between monthly mean SST anomalies and Nino-3 SST. Significant positive correlations between monthly mean SSTs and Nino-3 SST are found over equatorial central Pacific and most of the Indian ocean. Over the Indonesian region and west Pacific, negative correlations are found. The positive correlations over the Indian ocean are consistent with the findings of Meehl (1987) and Godfrey (1994). They reported that SST in the Indian ocean warm up in a composite El Nino event. The reasons for this warming are not clear. Hirst and Godfrey (1993) suggested that the open passage at the southern tip of the Indonesia makes warm ocean currents cross the Indian ocean and these warm the SSTs during an El Nino event. Another suggestion was that of increased surface insolation due to decrease in cloud cover. This aspect will be further examined later in this study. However it is found that increased convective activity over the central tropical Pacific is associated with reduced convection over tropical Indian ocean, despite the fact that this region is associated with positive SST anomalies. Over the west Pacific Ocean also, convection was reduced. However this reduced convection was rather associated with negative SST anomalies.

It is also useful to examine the relationship between SST and FDC on larger spatial scales. Fig.4. depicts a scatter plot of the area averaged (30° S – 30° N, 50° E – 100°E) SST *versus* corresponding area mean FDC. It can



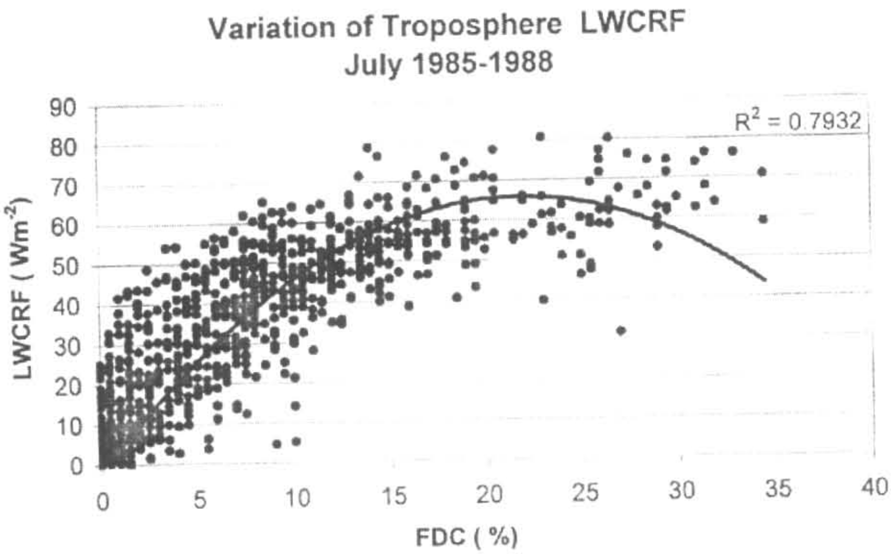


Fig.6. Scatter between FDC (%) and SWCRF at the surface (top) and LWCRF in the atmosphere (bottom). Period : July 1985-88

be seen that temporal variations in the mean FDC are uncorrelated with changes in the mean SST, which suggests that warmer SST may not necessarily increase deep convection over the Indian ocean. In this connection, the suggestion of Betts (1990) is relevant. He suggested that a warmer climate resulting from increased CO_2 would not necessarily lead to a greater convection. Hence the pattern of inter-annual variation between FDC and SST seems to behave differently in the Indian ocean basin as compared to the Pacific ocean basin.

3.3. *Deep convection and cloud radiative forcing*

The CRF at the top of the atmosphere (TOA) represents the overall radiative effect of clouds on the surface-atmosphere system. CRF also can be defined for the surface and atmosphere separately, where the effects of clouds are actually registered on the energetics of the system. Among the atmospheric CRF parameters, long wave cloud radiative forcing (LWCRF) is more important. In the tropics, it warms the troposphere, enhancing large

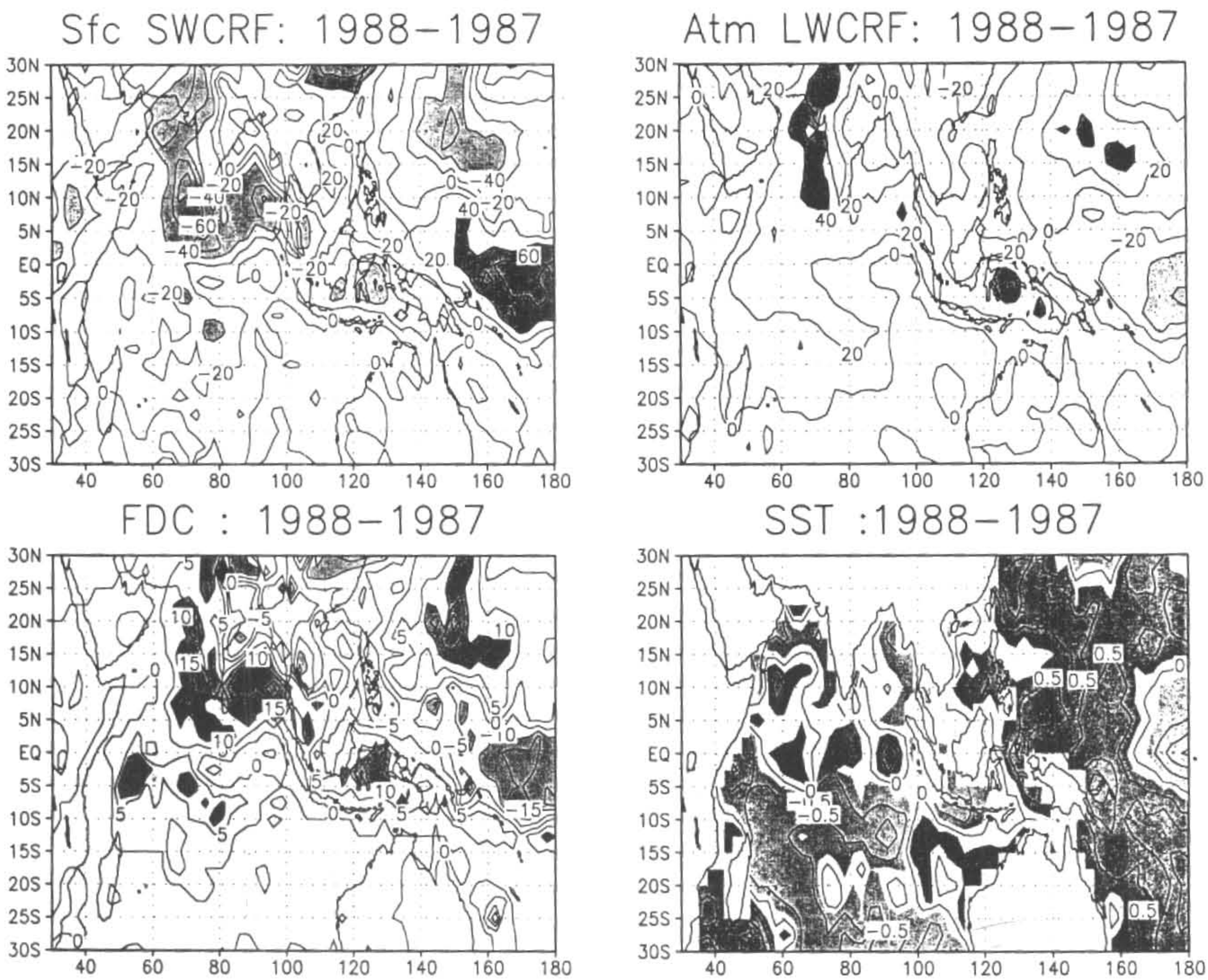


Fig.7. Geographic maps of the differences between July 1988 and July 1987 in (a) SWCRF at the surface, (b) LWCRF in the atmosphere, (c) FDC and (d) SST

scale convection (Randall *et al.* 1989) and it cools the troposphere outside the tropics, thus strengthening the Hadley circulation. In the deep convective regions, the atmospheric LWCRF may be comparable in magnitude to the latent heat released during cloud formation (Randall *et al.* 1989). In contrast, the short-wave cloud radiative forcing (SWCRF) on atmospheric circulation is generally quite small because clouds do not significantly affect the overall absorption of SW radiation in the atmosphere (Harshvardhan *et al.* 1990).

The mean spatial distribution of SWCRF and LWCRF at TOA for the period, July 1985-88 are shown in Fig. 5. Also shown in the diagram are SWCRF at the surface and LWCRF in the atmosphere. Maximum cloud radiative forcing is observed over Bay of Bengal and adjoining area. Over north Bay of Bengal, magnitude of SWCRF at TOA exceeded –120 Wm^{-2} whereas LWCRF at TOA was only of the order of -80 Wm^{-2}, thus causing large net negative cloud radiative forcing of the order of –40 Wm^{-2}. Rajeevan and Srinivasan (2000) suggested that this large negative net forcing in the Asian monsoon region is unique in the tropics. At the surface, magnitude of negative SWCRF exceeds 120 Wm^{-2} over north Bay of Bengal. LWCRF in the troposphere exceeds 60 Wm^{-2} over Bay of Bengal and adjoining land area. It decreases towards south and becomes negative (cooling) in the S.H., causing a large north-south cloud induced heating gradient. Comparison of these spatial patterns of the cloud radiative forcing with the spatial pattern of FDC (Fig.1) shows that regions of large CRF are associated with large FDC. It is further seen that about 80% of the LWCRF of the surface-atmospheric column is manifested in the troposphere while 80-100% of the SWCRF is felt at

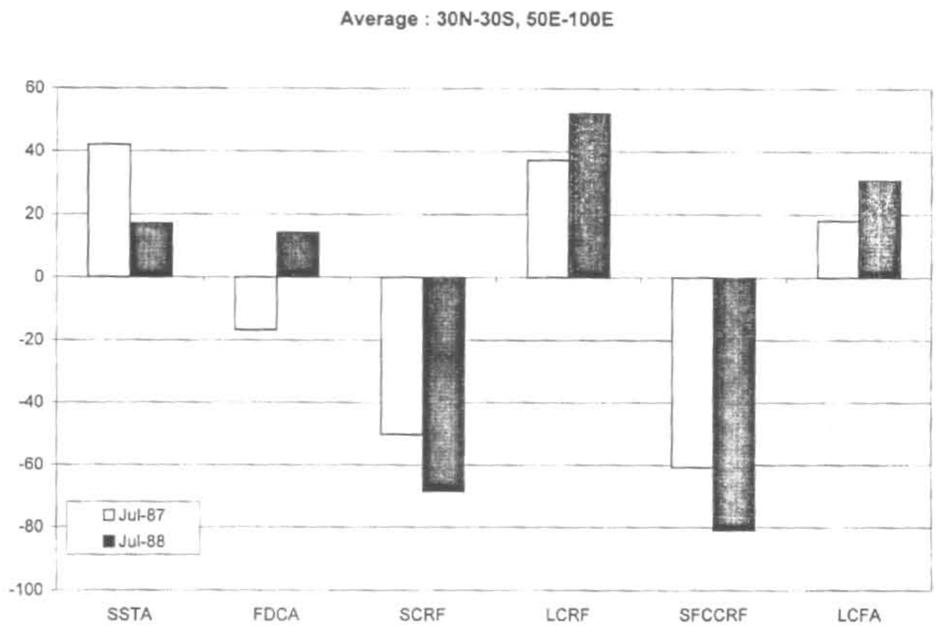


Fig.8. Histograms of area averaged (30° N-30° S, 50° E-100° E) SST anomalies (°C) (multiplied by 100), FDC anomalies (%) (multiplied by 10), SWCRF at TOA (Wm^{-2}), LWCRF at TOA (Wm^{-2}), SWCRF at the surface(Wm^{-2}), LWCRF in the atmosphere (Wm^{-2}) for July 1987 and July 1988

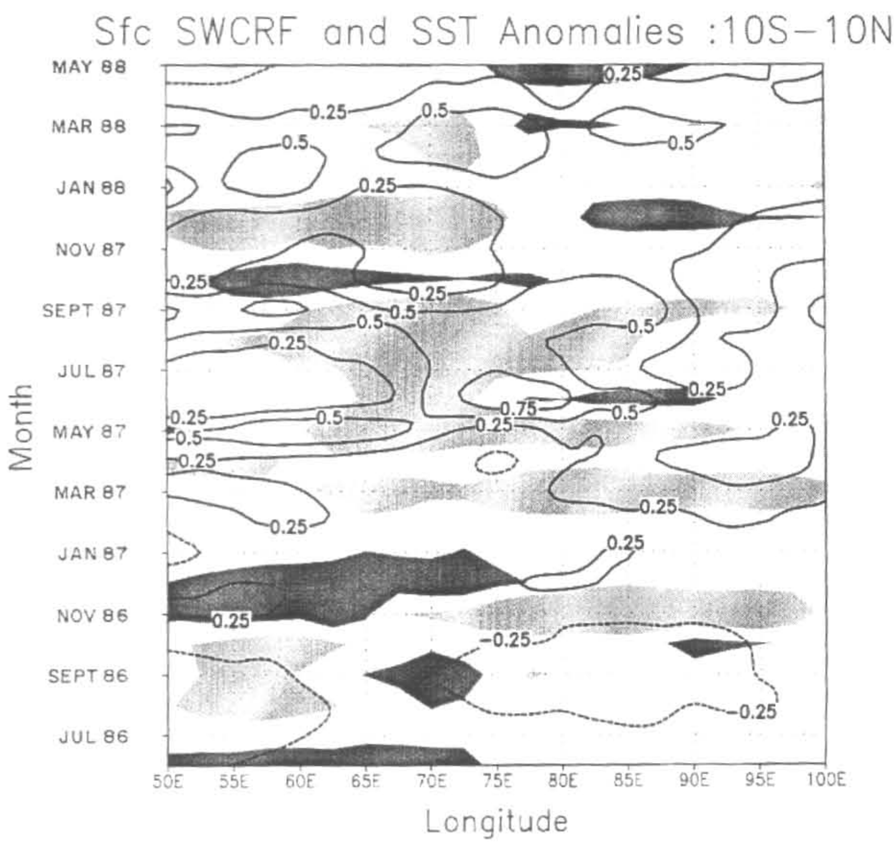


Fig.9. Contemporaneous longitude-month variation of anomalies of surface SWCRF and SST from June 1986 to May 1988. The light(dark) shaded area represents positive(negative) anomalies of surface SWCRF exceeding 10 Wm^{-2} suggesting less (more) cloud forcing. Negative (positive) SST anomalies are represented by dashed(continous) lines. Contour interval : 0.25° C

the surface. Thus the radiative effects of clouds are very similar to that of latent heat release in that they remove heat (solar energy) from the surface and deposit it (as long wave energy) in the troposphere (Ramanathan 1987). Moreover the cloud induced long wave heating of the troposphere can influence the monsoon circulation. The heating induces upward motion in the column and the compensatory subsidence in the surrounding regions. The surface solar cooling effect and the atmospheric long-wave warming effect of clouds will also tend to stabilize the column just as the effect of latent heat release (Ramanathan 1987). Thus in the monsoon deep convective cloud systems, radiative effects of clouds may play an important role in maintaining the vertical thermal structure.

The relationship between the cloud radiative forcing at the TOA and cloud radiative properties have been well documented (Kiehl 1994, Weare 1995 , Pai and Rajeevan 1998, Rajeevan and Srinivasan 2000). In this study, we have examined the relationship of surface SWCRF and atmospheric LWCRF with FDC and the results are shown in Fig.6. There is a linear relationship between surface SWCRF and FDC and about 70% of the variation on the surface SWCRF is explained by the variations in FDC. On the other hand, the atmospheric LWCRF increases almost linearly up to FDC of about 15% and thereafter there is little dependence on FDC is noted. About 79% of the variation in atmospheric LWCRF is explained by the variation in FDC. When FDC is more than 15%, surface SWCRF linearly increases with FDC whereas the atmospheric LWCRF is invariant with FDC. This non-linear response of atmospheric LWCRF causes an asymmetry between the surface SWCRF and atmospheric LWCRF contributing to a overall cooling of the surface-atmosphere system.

3.4. *Inter-annual variation of the relationships*

In this section, the inter-annual variation of the relationships among FDC, SST and cloud radiative forcing have been examined. For this purpose, July 1987 and July 1988, two anomalous months, have been considered. Fig.7. shows the spatial distribution of differences in FDC, SST and surface SWCRF and atmospheric LWCRF between 1988 July and 1987 July. SST distribution (Fig. 7d) reveals that most of Indian ocean was warmer in July 1987 associated with the development of the ENSO event. Warm pool area over west Pacific on the other hand was cooler in July 1987. However, deep convection (Fig. 7b) was more active over the Indian ocean in July 1988 compared to July 1987. The differences in FDC exceeded even 15% over eastern parts of Arabian sea and south Bay of Bengal. Thus in spite of warmer SSTs, deep convective activity over most of the tropical Indian ocean was suppressed in June 1987.

Correspondingly, surface SWCRF (Fig. 7a) in July 1988 was more pronounced over the Indian ocean especially along the west coast of India and south Arabian Sea and Bay of Bengal. The differences in SWCRF are of the order of 40-60 Wm^{-2} over Arabian sea and 20 Wm^{-2} over the equatorial Indian ocean. Thus, most of the Indian ocean received excess solar radiation at the surface in July 1987 compared to July 1988. The differences in the atmosphere LWCRF also reveal similar spatial pattern. Over most of the Indian ocean and particularly along the west coast of India , atmosphere LWCRF was more in July 1988 which suggests the eastward shift of the cloud induced LW heating of the atmosphere in July 1987 compared to July 1988. Both the east-west and the north - south gradients in atmosphere LWCRF were stronger in July 1988 suggesting stronger east-west and monsoon circulations. The spatial distribution of differences in the surface SWCRF and atmospheric LWCRF are similar to that of FDC suggesting that the differences in the cloud radiative forcing is caused due to the corresponding changes in FDC.

These aspects are even evident in the area averaged differences. Fig. 8. shows the area averaged (between 30° S-30°N, 50°E - 100°E) anomalies of SST and FDC, SWCRF at TOA, LWCRF at TOA , surface SWCRF and atmospheric LWCRF. Deep convective activity was suppressed in July 1987 over the Indian ocean in spite of warmer temperatures. More FDC in July 1988 caused larger SWCRF and LWCRF at TOA and larger surface SWCRF and atmospheric LWCRF. In July 1988, area averaged surface SWCRF was more by 20 Wm^{-2} compared to July 1987. Thus warmer SSTs in July 1987 over the tropical Indian ocean lead to less warming due to long wave cloud radiative forcing and less cloud reflection of solar radiation to the space.

In addition to the well known dependence on SST, tropical deep convection is also known to depend on factors such as the large scale circulation features(Lau *et al*. 1997, Bony *et al*. 1997). The role of heat fluxes and energy convergence in the troposphere in tropical deep convective zones was addressed by Srinivasan and Smith(1996) and Srinivasan (1997) based on the model proposed by Neelin and Held (1987). These results suggested that the necessary but not sufficient condition for the existence of tropical convergence zone (TCZ) is a positive energy convergence in the troposphere. The Neelin and Held model suggests that TCZ may not exist at the highest SST if the net energy convergence in the troposphere is negative. Based on this model, Srinivasan (1997) explained the relatively smaller occurrence of deep cloud clusters over the warm pool in the west Pacific.

A possibility is that the warmer SSTs in the Indian ocean may actually result from an increase in the absorbed solar radiation at the surface due to a reduction in deep convection related cloud cover. In this way atmospheric circulation anomalies may force SST anomalies rather than *vice versa* (Soden and Fu, 1995). Because the time scale of the ocean mixed layer response to the surface heating change is much faster than one month, the contemporaneous correlation between solar forcing by cloud variation and sea surface temperature changes will be strong, provided changes in other surface heat fluxes are negligible. During the period considered, the year 1987 was an El Nino year and 1988 was a La Nina Year. It has been well documented that in 1987, Indian summer monsoon circulation was weak and India experienced severe drought conditions (Krishnamurti *et al.* 1989). On the other hand,in 1988, Indian summer monsoon was stronger and India received excess rainfall (Krishnamurti *et al.* 1990). These circulation anomalies can influence the variations in deep convective clouds and cause a change in solar cloud forcing. We have seen that in July 1987, Indian ocean on average received 20 Wm^{-2} more solar radiation at the surface compared to July 1988 due to reduced convection which might have helped to cause warmer SSTs. The contemporaneous changes between the surface solar forcing anomalies and SST anomalies during the ENSO episode of 1987 as shown in Fig.9. also reveal the same conclusion. However this aspect needs to be examined with more years of data and case studies before a more concrete conclusion can be drawn.

4. Conclusions

(a) The annual average FDC is maximum over the equatorial east Indian ocean and adjoining west Pacific and Indonesian region. During the summer monsoon (June-September), maximum FDC (exceeding 20%) is observed over Bay of Bengal. The maximum FDC zone exhibits north-south oscillation in association with the advancement/ withdrawal of summer monsoon.

(b) FDC variation in the Indian ocean is weakly correlated with the variation with SST, suggesting that atmospheric circulation patterns play an important role in modulating the FDC variations. In spite of warmer SSTs associated with the 1987 ENSO, deep convection was suppressed over the Indian ocean. Even averaged over a larger spatial scale, temporal variations in the mean FDC are uncorrelated with changes in the mean SST which suggests that warmer SST may not necessarily increase deep convection in the Indian ocean.

(c) Maximum cloud radiative forcing over Bay of Bengal during the summer monsoon season is associated with maximum FDC observed over this region. Variations in FDC explain about 70% and 79% of the variations in surface SWCRF and atmosphere LWCRF respectively. At larger FDC (> 15%) atmosphere LWCRF does not depend on FDC, thus causing large asymmetry between surface SWCRF and atmosphere LWCRF leading to a overall cooling of the surface-atmosphere system.

(d) Interannual variations in FDC however modulate significantly the interannual variations of cloud radiative forcing in the Asian monsoon region. In July 1987, an El Nino year, warmer SSTs, over the tropical Indian ocean lead to less warming due to long wave cloud radiative forcing and less cloud reflection of solar radiation to the space.

(e) There is some observational evidence suggesting that warmer SSTs in the Indian ocean during the 1987 ENSO episode might have resulted from an increase in the absorbed solar radiation at the surface due to a reduction in deep convection related cloud cover.

This study is carried out with available but limited satellite data sets. More observational study is required for better documentation of the interactions among the deep convection, SST and radiation in the Asian monsoon region. The Clouds and the Earth's radiant energy system (CERES) initiated by NASA (Wielicki *et al.* 1996) will provide more useful data sets for the next 15 years so that the features discussed in this study can be verified further. CERES data are now becoming available for research. Also important is to examine whether General Circulation Models can simulate these observed interactions among deep convection, SST and radiation in the Indian ocean which are different from other ocean basins.

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