Monsoon–ocean coupling

Sulochana Gadgil

Centre for Atmospheric and Oceanic Sciences, Indian Institute of Science, Bangalore 560 012, India

The Indian monsoon is maintained by propagation of convective systems of synoptic (lows, depressions, etc.) and planetary scale (tropical convergence zones) from the warm tropical oceans, onto the heated subcontinent. As a result, the monsoon variability on sub-seasonal scales (between wet and dry spells) and on interannual scales (good monsoons and droughts) is linked to variation of the convective systems over the ocean, where variability in turn depends on the sea surface temperature through complex relationship. The nature of this relationship and the mechanisms suggested for understanding it are discussed. It is suggested that for further insights into monsoon–ocean coupling detailed observations particularly over the Indian seas with new satellites and special observational experiments are essential.

OVER the last two decades the important role of tropical oceans in the variability of monsoon rainfall over the Indian subcontinent has been clearly demonstrated in studies carried out in India and abroad. In this paper I discuss the present understanding of the coupling of the monsoon to the oceans, focusing on the summer monsoon (June–September) season and point out the challenging problems that need to be addressed for further insights into this complex system.

The next section contains a brief background about the Indian monsoon, its variability on intraseasonal and interannual time scales and its link to convection over the ocean. Next, energetic constraints on the occurrence of the basic dynamical system in the tropics, viz. the tropical convergence zone are discussed. Then, the relationship of convection/precipitation over the ocean to the sea surface temperature (SST) is considered. Finally, the problems that need to be addressed for further insight are mentioned.

The Indian monsoon

The monsoonal regions of the world are characterized by a large seasonal variation in rainfall (Figure 1). The large-scale rainfall in the tropics is associated with a planetary scale system – the tropical convergence zone (TCZ). The Indian monsoon is a manifestation of the seasonal variation of the TCZ, which migrates onto the Indian subcontinent in the northern hemispheric summer. A TCZ is characterized by convergence of moist air in the lower troposphere and strong ascent up to the upper troposphere leading to deep convective clouds (Figure 2). The air ascending in the TCZ descends in the surrounding region. Figure 2 shows the Hadley cell in which the circulation is primarily in the meridional (north-south) plane with descent in the region equatorward and poleward of the TCZ. In addition, the air also subsides in the cloud-free regions eastward and westward of the TCZ in what is known as a Walker cell. The intense zonal cloud band associated with a TCZ is a prominent feature of the satellite imagery of the region (Figure 3 *a*). A TCZ can extend over thousands of kilometers in the east-west direction but its latitudinal extent is relatively small, seldom exceeding 10° (Figure 3 *b*).

The large-scale rainfall during the summer monsoon (June–September) over the Indian region is associated with the occurrence of a TCZ over the heated subcontinent. A TCZ occurs over the equatorial Indian Ocean in the pre-monsoon months of April and May. During the onset phase, the TCZ propagates across the peninsula and gets established near its seasonal mean location over the Indo-Gangetic plains (e.g. Figure 3) around the beginning of July. During the peak monsoon months of July and August it fluctuates around this position. Variation of the monsoon rainfall within the season (between active and weak spells) as well as of the seasonal rainfall from year to year (between the good monsoons and droughts) is associated with variation in the intensity and location of the continental TCZ[1,2]. Even after the continental TCZ is established in early July, the TCZ over the equatorial Indian Ocean occurs intermittently throughout the season. At intervals of 2–6 weeks, this oceanic TCZ propagates onto the subcontinent in a manner similar to that of the onset phase (Figure 4). These northward propagations of TCZ across the Bay of Bengal and eastern Arabian Sea play an important role in maintaining the continental TCZ.

It is important to note that a TCZ is characterized by cyclonic vorticity associated with large shear of the zonal (east-west) wind in the north-south direction, at all levels in the lower troposphere above the atmospheric boundary layer. However the vorticity/clouding is seldom of uniform intensity throughout. Generally synoptic scale vortices (monsoon disturbances such as lows and depressions) are embedded in it (e.g. Figure 5 for the circula-

*For correspondence (e-mail: sulo@caos.iisc.ernet.in)

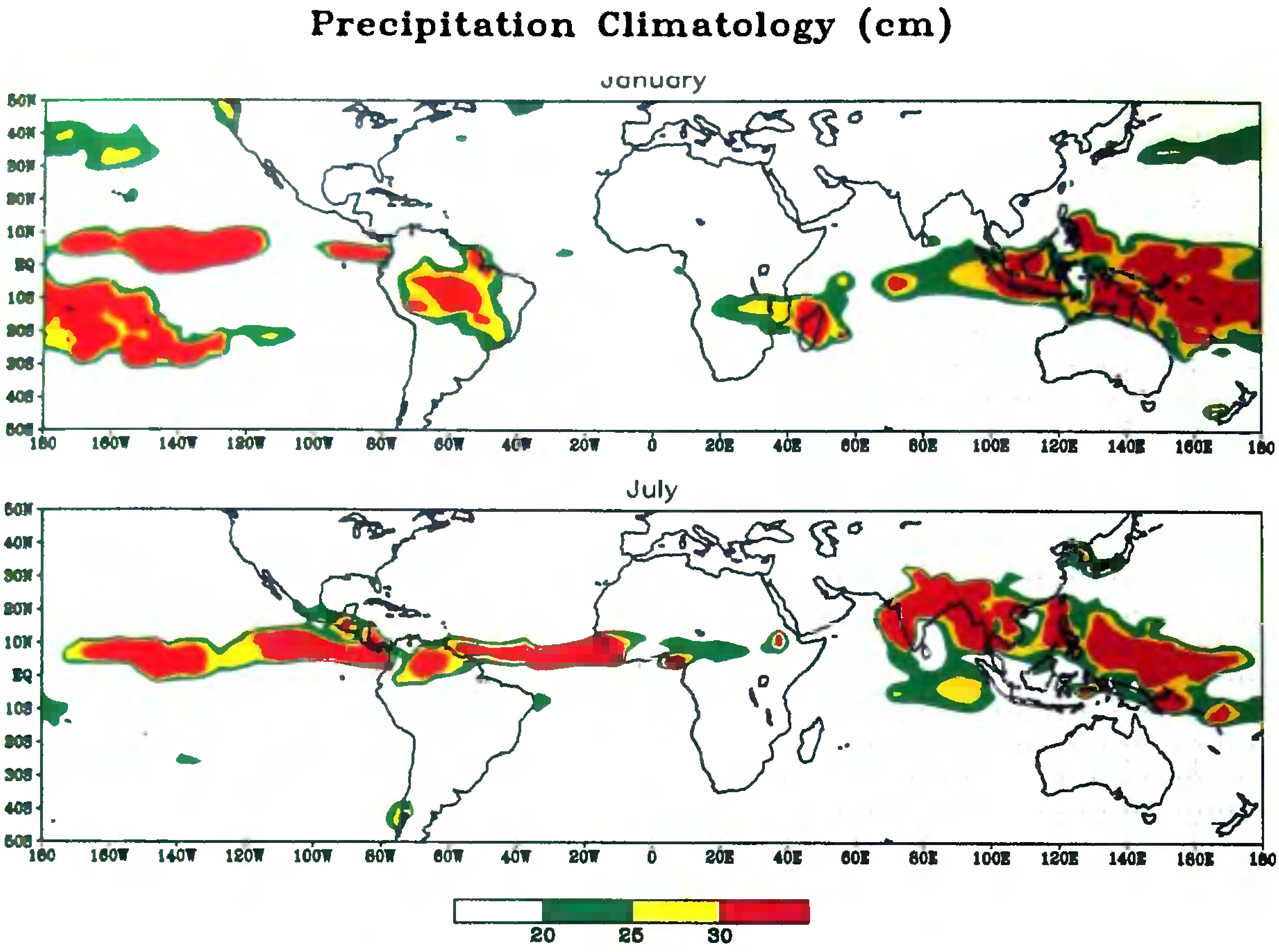


Figure 1. Mean rainfall in cm during January (above) and July (below).

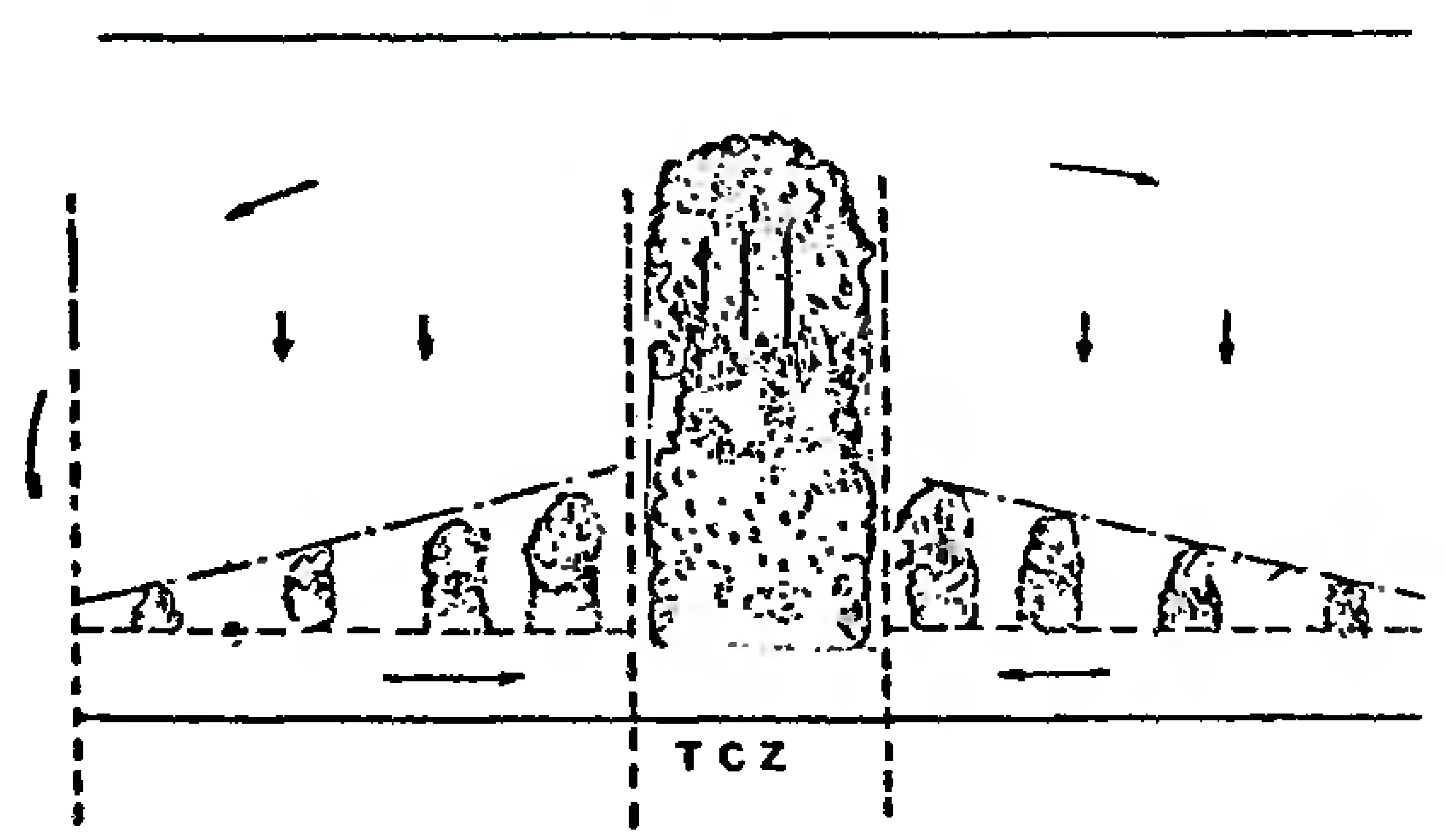
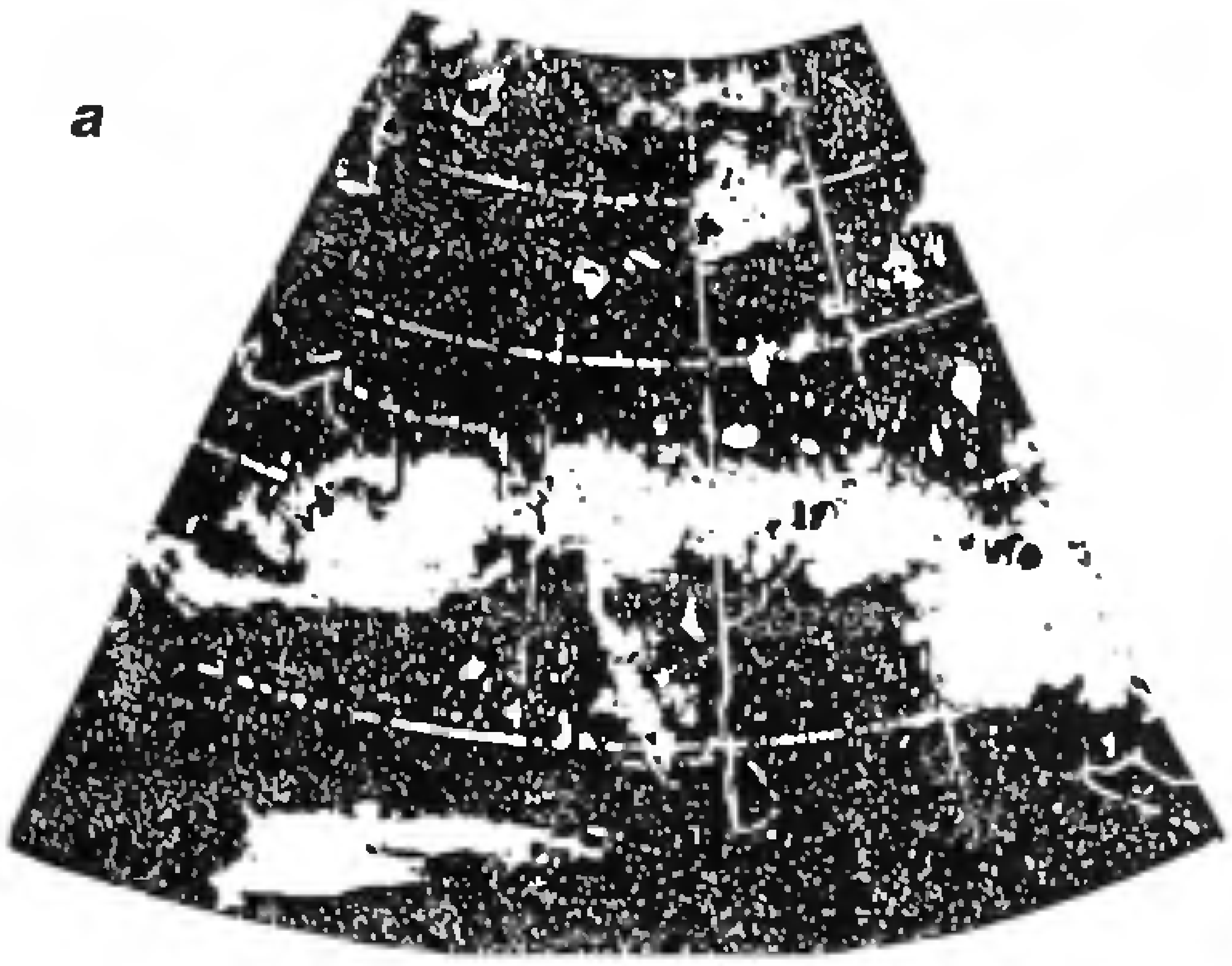


Figure 2. Schematic of the meridional circulation and clouds associated with a tropical convergence zone.

tion associated with the cloud band in Figure 3 *a*). A vast majority of these disturbances are generated over the warm oceans surrounding the Indian subcontinent. The northward propagations of the TCZ often involve propagation of disturbances over the Arabian Sea and the Bay, linked with less intense clouding over the intervening land surface. Also, during July and August a large number of disturbances are generated over the head Bay and move westward along the seasonal mean location of the continental TCZ (Figure 6).

Thus the continental TCZ is maintained primarily by genesis of disturbances over the equatorial Indian Ocean, Bay of Bengal and the eastern Arabian Sea. Variability of the continental TCZ and hence the monsoon is therefore, linked to the variability of convection over these warm oceans. For example, during breaks (i.e. dry spells) of the monsoon, the TCZ over the equatorial Indian Ocean is anomalously active (Figure 7 after Gadgil *et al.*[3]). The first study of intraseasonal variation over the Indian longitudes with satellite imagery has suggested that there is a competition between the continental TCZ and the TCZ over the equatorial Indian Ocean, with weak/active spells of the continental TCZ occurring simultaneously with active/weak spells of the oceanic TCZ[1]. Such a competition is an obvious consequence of the two TCZs occurring over the same longitudinal belt with each forcing descent of air equatorward and poleward of its location.



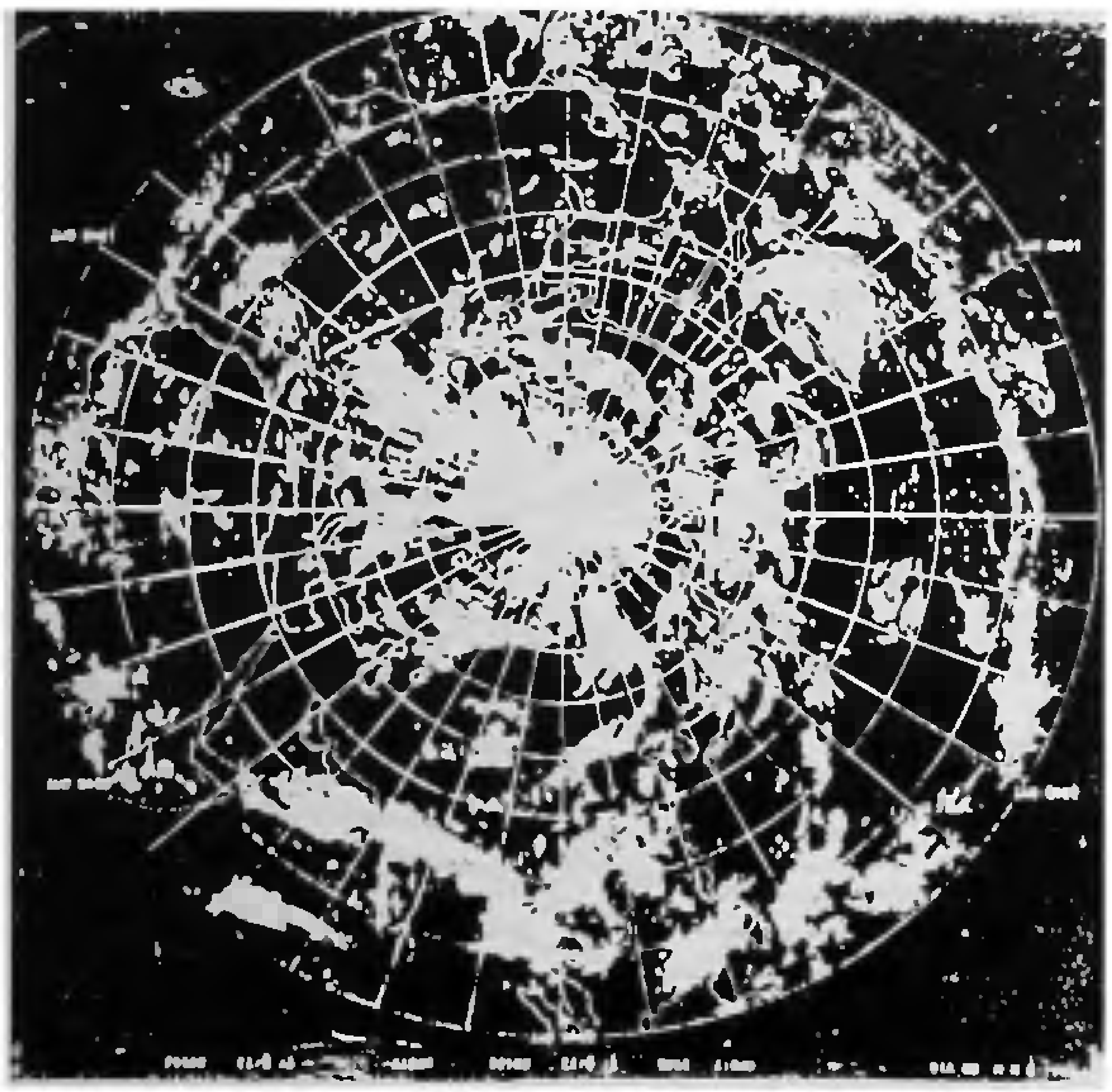


Figure 3. Satellite imagery (visible) of *a*, the Indian sector; and *b*, the northern hemisphere for 8 July 1973.

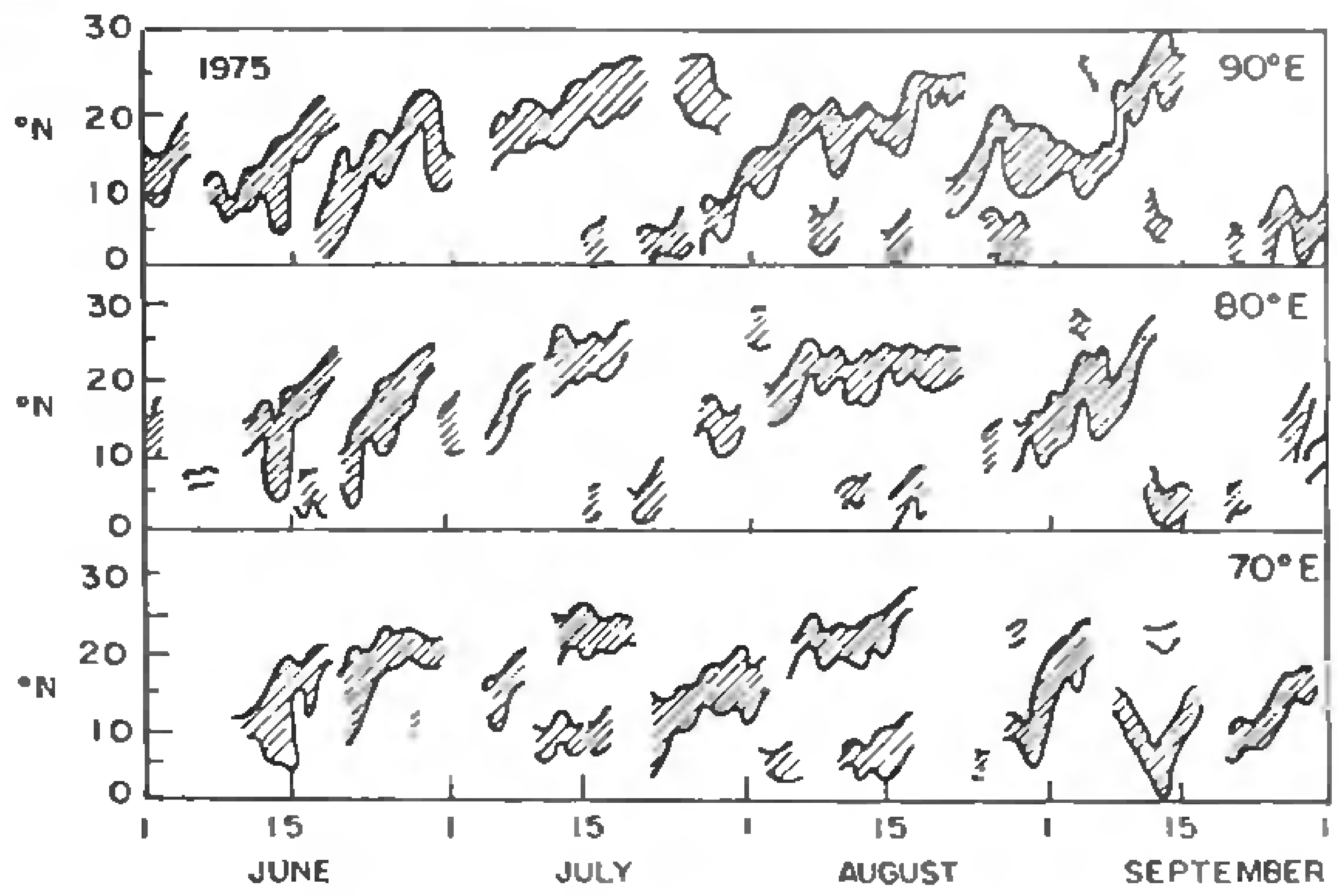


Figure 4. Variation of the location and latitudinal extent of the cloud band of the type in Figure 3 at 70, 80, and 90°E during June–September 1975.

The monsoon is not only linked to the variation over the oceans surrounding the subcontinent but also to that over the Pacific Ocean. This is not surprising, since the planetary scale TCZ often extends continuously from the Indian region all the way across the Pacific (Figure 3 *b*). The strong correspondence between droughts of the Indian monsoon and El Niño over the Pacific (Figure 8) is well established[4,5]. During an El Niño, a TCZ extends as a coherent zone of convection across the Pacific. In contrast, in the climatological average, there is a gap between TCZs over the western and eastern Pacific, with hardly any convection over the Central Pacific (Figure 9). Notice that the TCZ over the equatorial Indian Ocean is more active during the El Niño of 1987 which was also a poor monsoon. This is not surprising, since, as first pointed out by Krishnamurti and Bhalme[6], poor monsoon seasons generally have long dry spells/breaks (Figure 10) within the season, and breaks are associated with an anomalously active oceanic TCZ (e.g. Figure 7).

To summarize, what has been established so far is (i) that the continental TCZ is maintained by genesis of organized convection (in synoptic scale disturbances and/or the TCZ) over the oceans and subsequent propagation onto the continent, and (ii) there are strong links between the variability of the continental TCZ (and hence the monsoon) and the variability of the TCZ over the oceans surrounding the subcontinent and the Pacific Ocean. However, while the links have been clearly established, the cause–effect relationships are not clear. For example, while the probability of deficit monsoon rainfall during an El Niño is large (Figure 8), there is no consensus yet on whether the El Niño triggers the monsoon deficit or vice versa. The variation in the timing of different El Niño episodes *vis-à-vis* the seasons makes the problem more complex. Similarly, whereas it is clear that the TCZ over the equatorial Indian Ocean is active during dry spells or breaks of monsoon, and also during poor monsoon seasons, whether the anomalous activity of the oceanic TCZ is an effect or cause of drought or break monsoon is not clear.

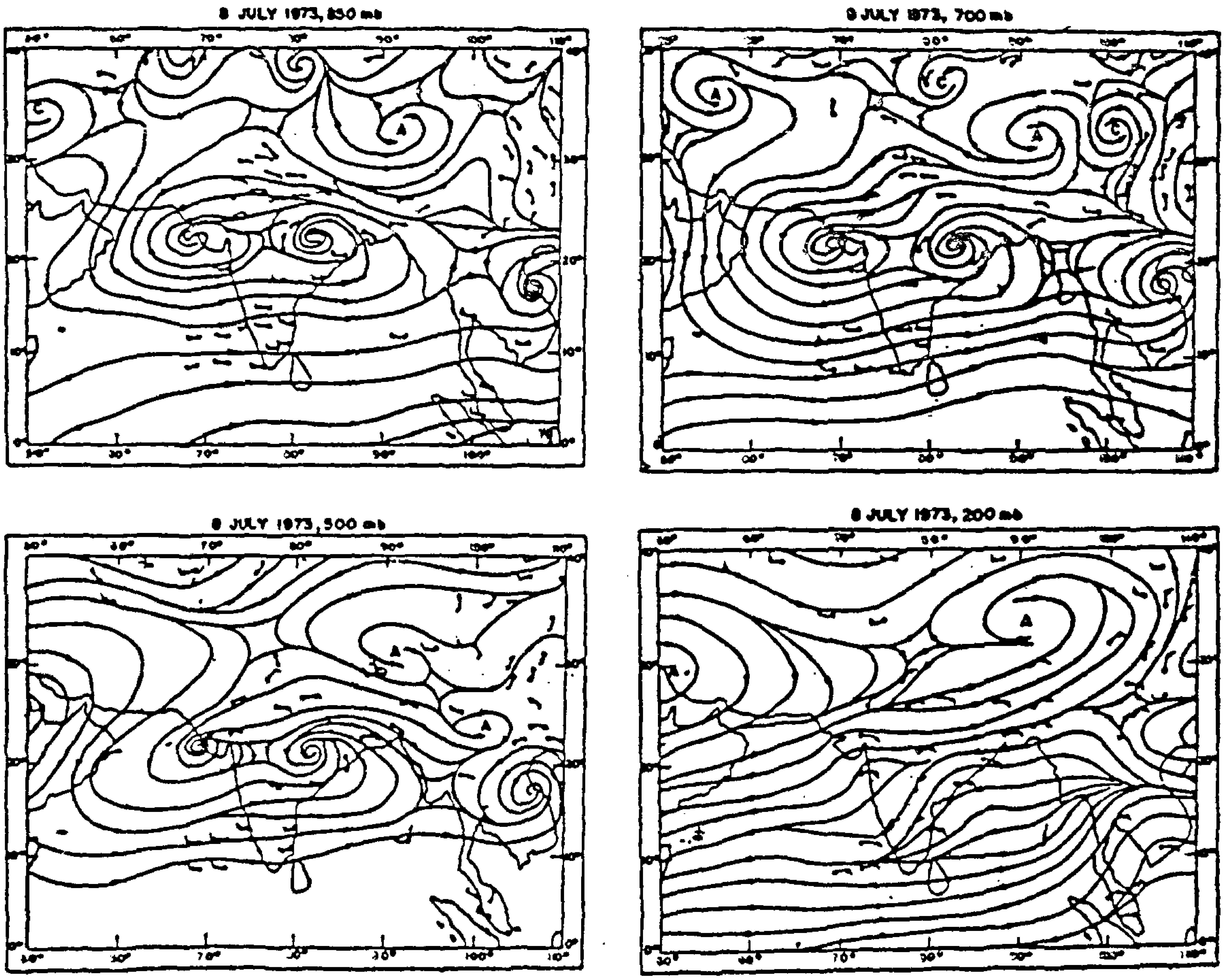
Understanding the complex interactions between the continental TCZ over the Indian region and the TCZs over the Indian and Pacific Oceans is undoubtedly one of the most challenging problems in monsoon meteorology today. The first step in addressing this problem is to gain an insight into the factors that determine the occurrence of a TCZ over the oceans. We shall address this in the next section.

Tropical convergence zones

As pointed out earlier, the monsoon is a manifestation of the seasonal variation of the large-scale rainfall in the tropics associated with seasonal migration of the TCZ. This seasonal migration of TCZ occurs in response to the seasonal variation of the incident radiation. The latitude at which the sun is overhead varies between the Tropic of Capricorn during the winter solstice to the Tropic of Cancer during the summer solstice. Charney[7] suggested that TCZs can occur only over regions which are radiative sources, i.e. over those regions where the net radiation received by the earth–atmosphere system is positive. In contrast, deserts occur over regions that are radiative sinks.

In fact, there is a correspondence between the seasonal variation of the zonal average of the net radiation received by the earth–atmosphere system and that of the precipitation (Figure 11). However, the latitudinal range of variation of precipitation is much smaller. Also, there is considerable variation in the east-west direction because of the presence of continents and oceans. Thus in the northern hemispheric summer, whereas the region around 20°N over the Indian subcontinent receives considerable rainfall, the conditions over the African longitudes are markedly different, the Sahara desert being a prominent feature (Figure 1). Using the first direct observations of the total radiation obtained by the Earth Radiation Budget Satellite, Srinivasan and Smith[8] showed that organized deep convection is indeed restricted to regions with positive net radiation, while deserts occur over radiative sinks. Figure 12 shows that regions with deep convection (low outgoing longwave radiation (OLR)) occur only within regions of positive net radiation. However, there are large regions with positive net radiation which do not have deep convection. This is clearly brought out in Figure 13 (after Srinivasan[9]) which shows that all grids with OLR < 225 wm^{-2} (i.e. with deep clouds) are characterized by positive net radiation, for grids with large positive net radiation, OLR ranges from less than 200 wm^{-2} up to 270 wm^{-2} (i.e. cloud-free conditions).

Neelin and Held[10], using a simple two layer model, showed that convergence of moisture in the lower troposphere (which is an essential feature of a TCZ) can occur only in regions in which there is a net convergence of



Circulation pattern at different levels on 8 July 1973.

Figure 5. Circulation patterns on 8 July 1973 at 850, 700, 500 and 200 mb levels.

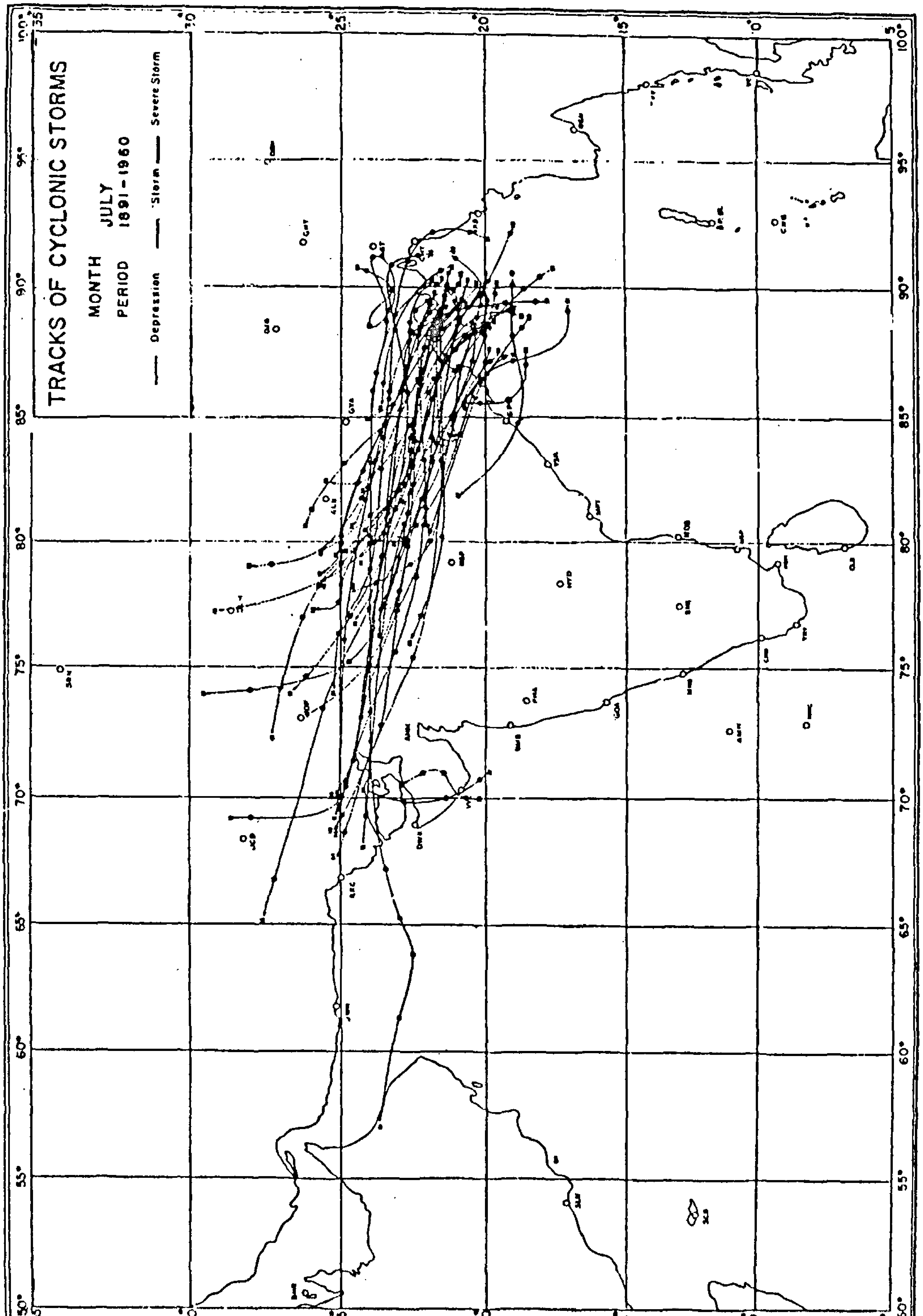


Figure 6. Tracks of cyclonic disturbances in July during 1891–1960; India Met. Dept.

energy (including radiative, latent and sensible heat fluxes) in the troposphere. Srinivasan[9] showed that regions with deep convection are embedded only within regions with positive energy convergence.

Thus the condition on net radiation/energy convergence is a necessary but not sufficient condition for the occurrence of a TCZ. Obviously, other parameters are also important for determining where tropical convergence zones should occur. It is well known that one of the most important oceanic variables for the occurrence of organized convection over the oceans is the sea surface temperature (SST). Next we consider the relationship between organized convection and SST.

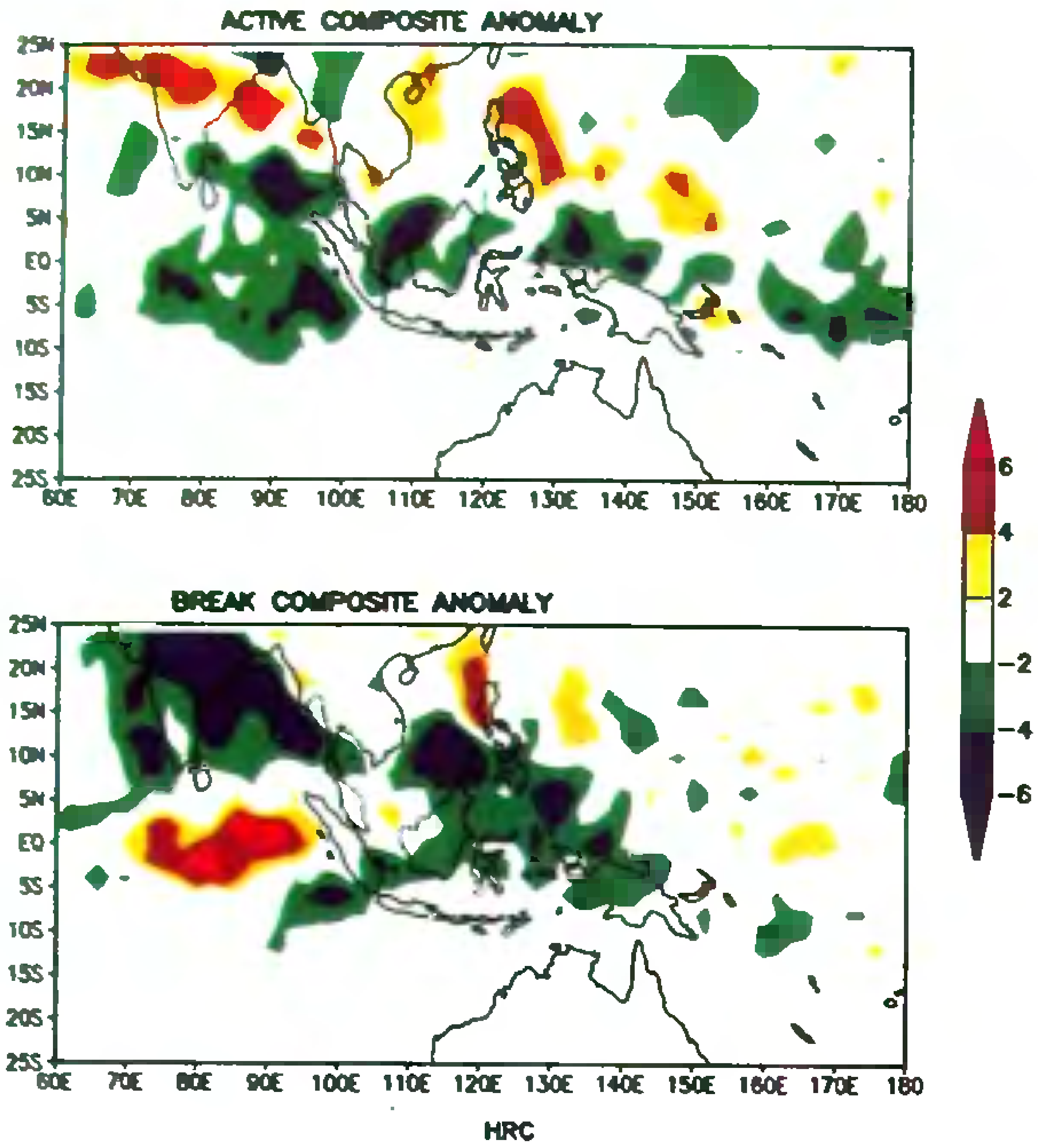


Figure 7. Composite HRC anomaly patterns for active spells and breaks of the Indian monsoon (after Gadgil *et al.*[3]).

Convection over oceans and SST

Observations about the relationship between organized convection over tropical oceans and SST and the hypotheses suggested for the observed relationships are considered in this section.

Convection over oceans and SST – Observations

Half a century ago, Palmen[11] 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[12] showed that the variation of convection over the tropical Pacific could be attributed to variations of SST. Systematic investigation of variation of convection and its relationship with SST became possible only after the availability of satellite data. Studies on the variation of tropical convection over oceans with SST based on different measures of organized deep convection such as cloudiness intensity[13], OLR[14,15], and the frequency of highly reflective clouds[16] (HRC), all suggest a similar and highly nonlinear relationship. In the first such study by Gadgil *et al.*[13], Sadler data on daily cloudiness intensity were used to identify for each month, the number of days with deep convection (cloudiness intensity ≥ 6) and hence the total cloudiness intensity of such days. The

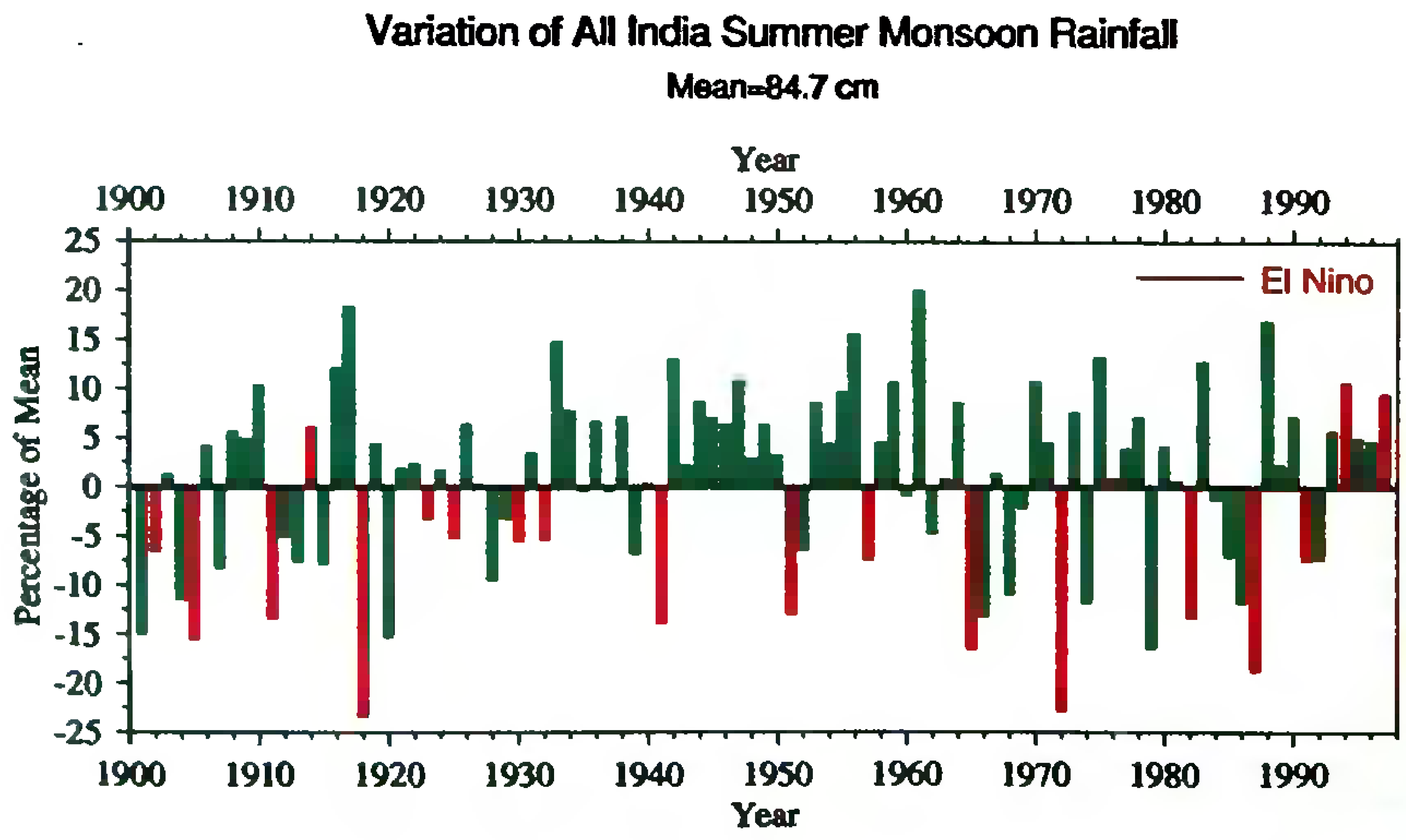


Figure 8. Variation of the departure of the all-India monsoon rainfall from average; the El Niño years are indicated.

relationship of this cloudiness intensity to the monthly SST for the Indian Ocean region compiled by Joseph and Pillai[17] for each of the 5° boxes during June–September 1966–72 is depicted in Figure 14. It is seen that over cold oceans, (below the Palmen threshold) no organized convection occurs (Figure 14 *a*). The propensity for convection increases with SST and around 27.5° the mode shifts with the maximum frequency implying occurrence of organized deep convection (Figure 14 *b*). However, for SST above this threshold there is a large variation of cloudiness. From this, Gadgil *et al.*[13] concluded that there is a threshold value of SST of about 27.5°C above which the propensity for organized deep convection is high. They further noted that SST being above the threshold is a necessary but not sufficient condition for organized deep convection. This is not surprising, since tropical convection is also known to depend on other factors such as large-scale convergence.

In fact, SST appears to play an important role in determining the variability in convection when SST variations are around the threshold as for the central Pacific; and not when SST is maintained well above the threshold as for the Indian Ocean (Figure 15). Thus the relationship between convection over warm oceans (with SST above the threshold) to oceanic conditions is more complex. Hence, although it is clear that variability of the monsoon on intraseasonal and interannual scales is linked to the convection over oceans, correlations between monsoon rainfall over India and SST of the Indian seas are poor[18,19]. In fact the impact of atmosphere on the ocean comes out as the stronger signal with SST in the seasons following a good monsoon being lower[17,20].

Graham and Barnett[14] showed that the nonlinear relationship between SST and convection first discovered by analysis of data over the Indian Ocean[13] is a basic feature of organized deep convection over the tropics by analysis of a better data set on convection, viz. OLR. They also showed that the mean convection (mean OLR) increases (decreases) rapidly in a range of about 1°C around the threshold. This is clearly seen in the observed variation of

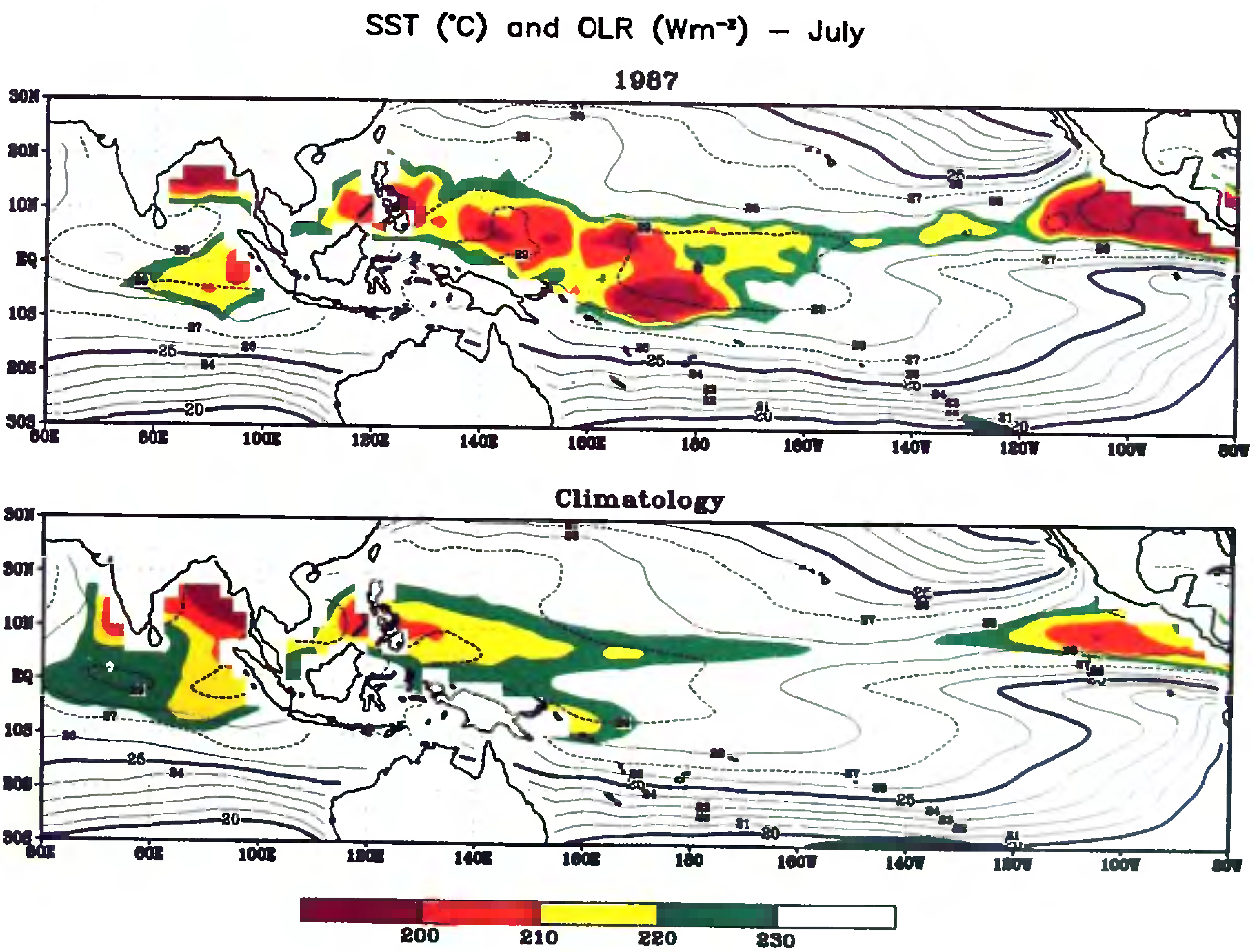


Figure 9. OLR pattern during July 1987 (El Niño) (above) and the average July OLR.

the mean HRC with SST, for different tropical oceans shown in Figure 16 *a*. An idealized representation of this variation is shown in Figure 16 *b*. It is seen from Figure 16 *a* that although the threshold value of SST varies a little with the region, being somewhat lower for the Atlantic, slopes of HRC with SST in the critical range for the different regions are almost identical. Graham and Barnett[14] attributed the variation in convection for SST above the threshold to the variation in the surface divergence.

In the last decade, the quality and quantity of atmospheric data has increased enormously with the availability of estimates of precipitation over the ocean using microwave sounding unit (MSU) and NCEP reanalysis (Kalnay *et al.*[21]) for various dynamical quantities of interest on a global scale. A detailed study by Bony *et al.*[22] has clearly demonstrated that the variation of precipitation with SST is maximum around the threshold of 28°C (Figure 17), when the entire tropical region is considered. Analysis of the important dynamical fields such as vertical velocity at 500 hpa, and upper level divergence, from the NCEP reanalysis data set by Bony *et al.*[23] and Lau *et al.*[24] has suggested that relationship between the zones with ascent of air and regions with high SST is much stronger than hitherto supposed. Variation of the percentage of region between 20°N and 20°S with upward velocity of different magnitudes with SST is shown in Figure 18. It is seen that 90% of the region with SST between 29 and 30°C is characterized by ascent. For SST above 30°C, the fraction of the ascending zone decreases to about 75%, indicating cloud-free conditions over a quarter of the warmest oceans. If only the regions with strong ascent (deep convective clouds) are considered, then the fraction over waters with SST above 29°C reduces to about 60%. The question of where the belt of intense convection and precipitation is located over the warm ocean will be addressed later in the article.

SST and convection – Mechanisms

The first possibility suggested is the nonlinearity between the water holding capacity of air (saturation water vapour mixing ratio) and its temperature (Clausius–Claperon relation). However, it has been shown that this nonlinearity is not adequate to explain the observed nonlinearity of the SST–convection relationship[10,25]. Deep convection in the tropics can be maintained only when there is a positive cloud buoyancy over a large depth of the atmosphere. Buoyancy of the cloud depends not only on the properties of air near the surface of the ocean but also on those of the upper air through which it rises. Hence, we expect a

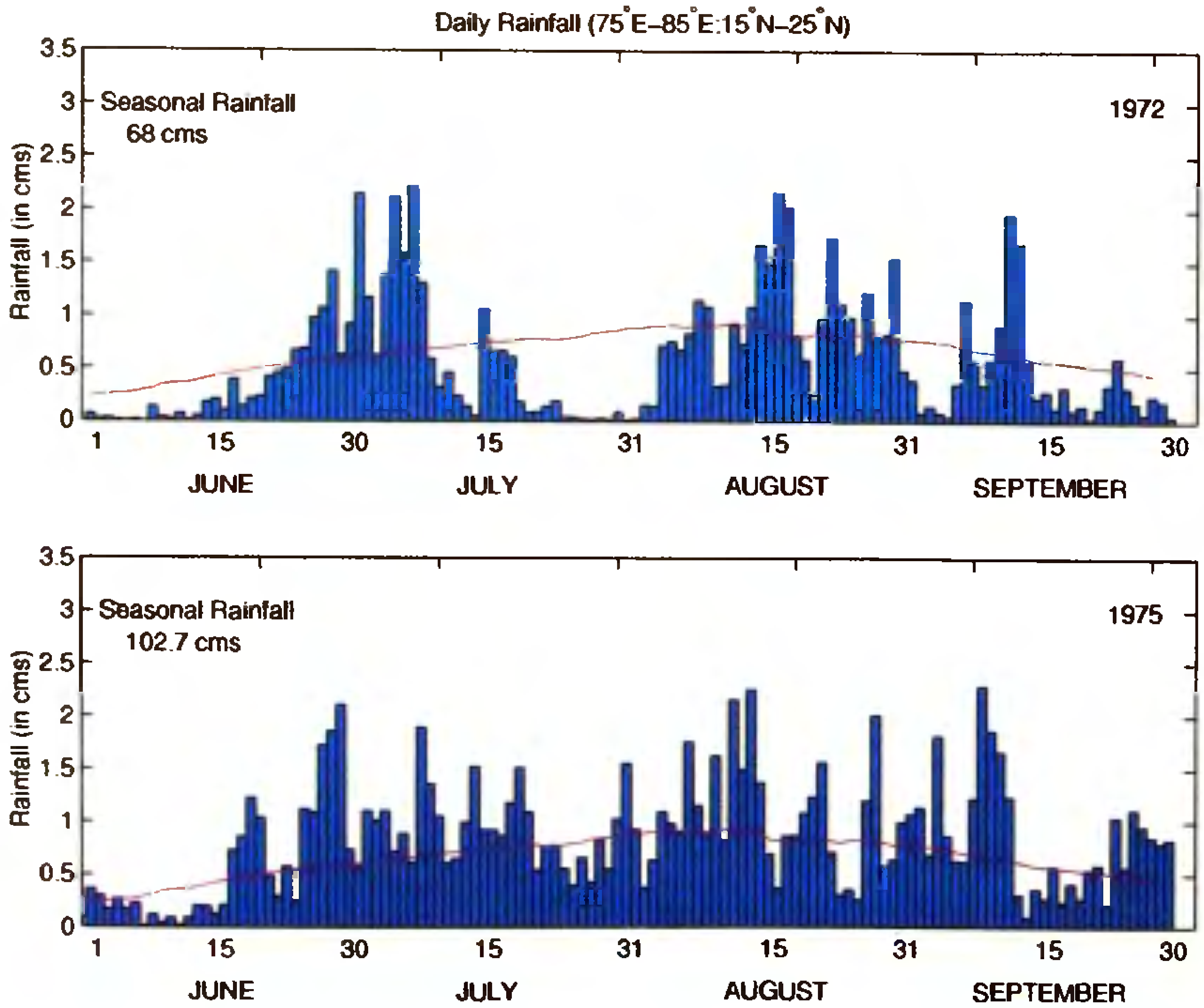


Figure 10. Daily variation of rainfall over central India during the poor monsoon season of 1972 (with all-India monsoon rainfall of 65 cm, i.e. 23% deficit) and the good monsoon season of 1975 (with all-India monsoon rainfall of 96 cm, i.e. 13% excess).

parameter which incorporates the effects of surface and upper air to be better correlated with deep convection than one that includes only the properties of surface air. Bhat *et al.*[26] suggested that the convective available potential energy (CAPE), which determines whether the surface air will ascend is one such parameter. They demonstrated that CAPE (determined from monthly mean profiles) represents the work potential of the atmospheric heat engine (of the type in Figure 2) with ascent in the region of deep clouds and descent in the cloud-free region and that the frequency of deep convection is highly correlated with CAPE (Figure 19 *a*). Thus the relationship between CAPE and SST is strikingly similar to that between frequency of convection and SST (Figure 19 *b*). Hence they concluded that the relationship between convection and SST could be understood in terms of variation of CAPE.

Srinivasan[9] suggested that another important parameter for determining organized convection over tropical oceans and land is the moist static energy of the lower troposphere. He showed that the relationship between convection and moist static energy is nonlinear and similar to that of convection with SST. There is a threshold of about 330 kJ/kg, below which there is hardly any convection;

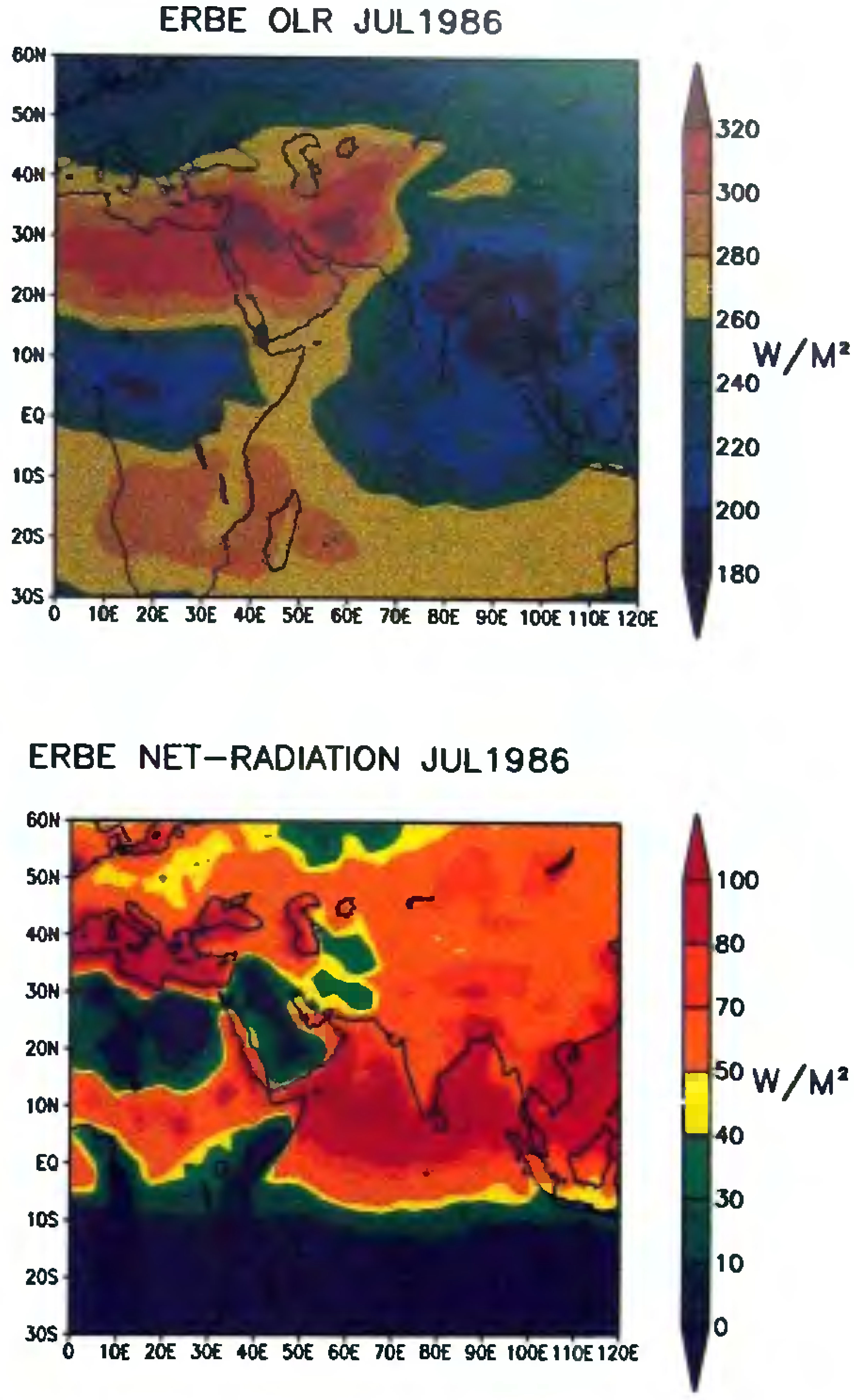


Figure 12. OLR and net radiation from ERBE for July 1986.

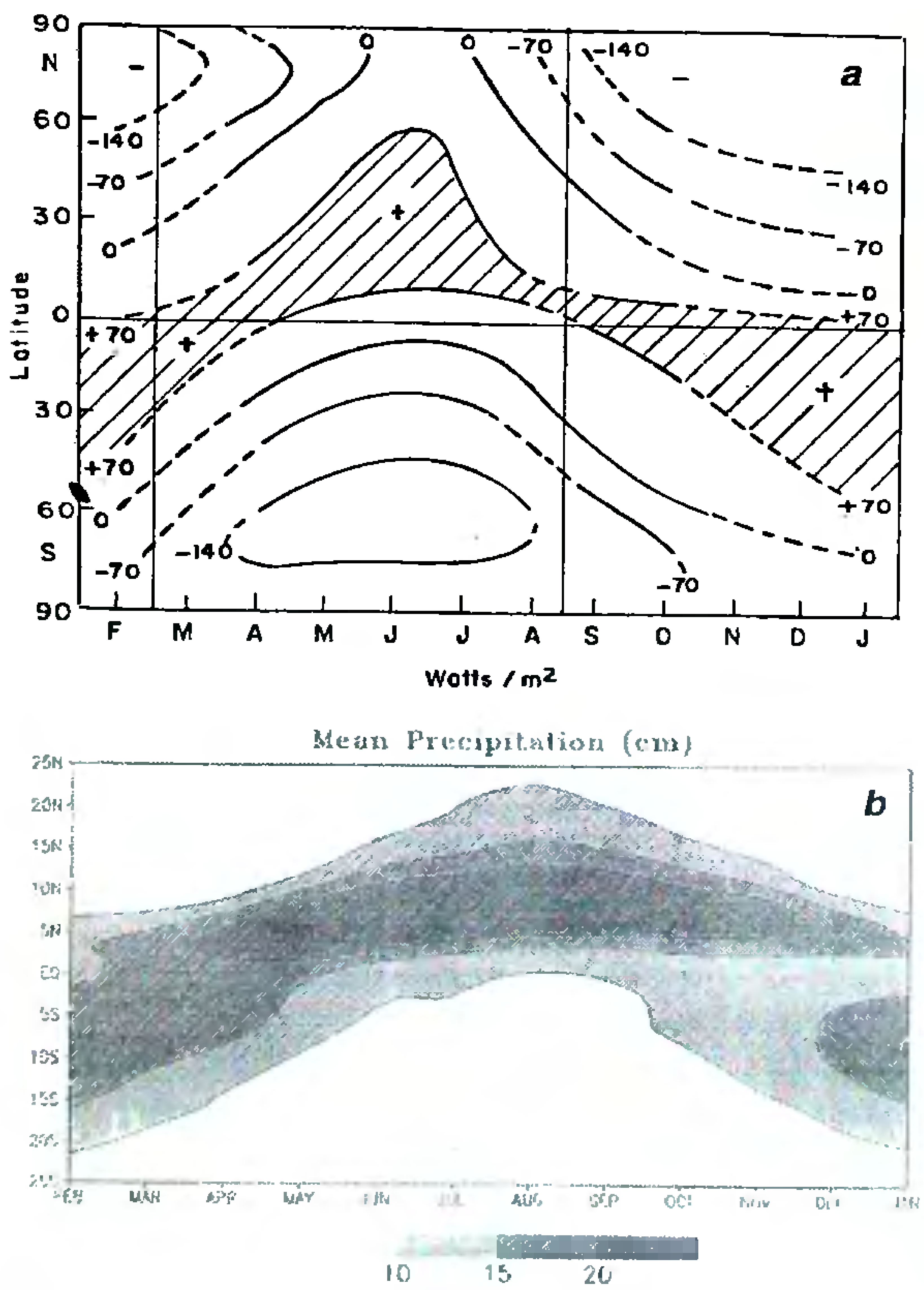


Figure 11. ***a***, Seasonal variation of the zonal average of net radiation; and ***b***, precipitation with latitude.

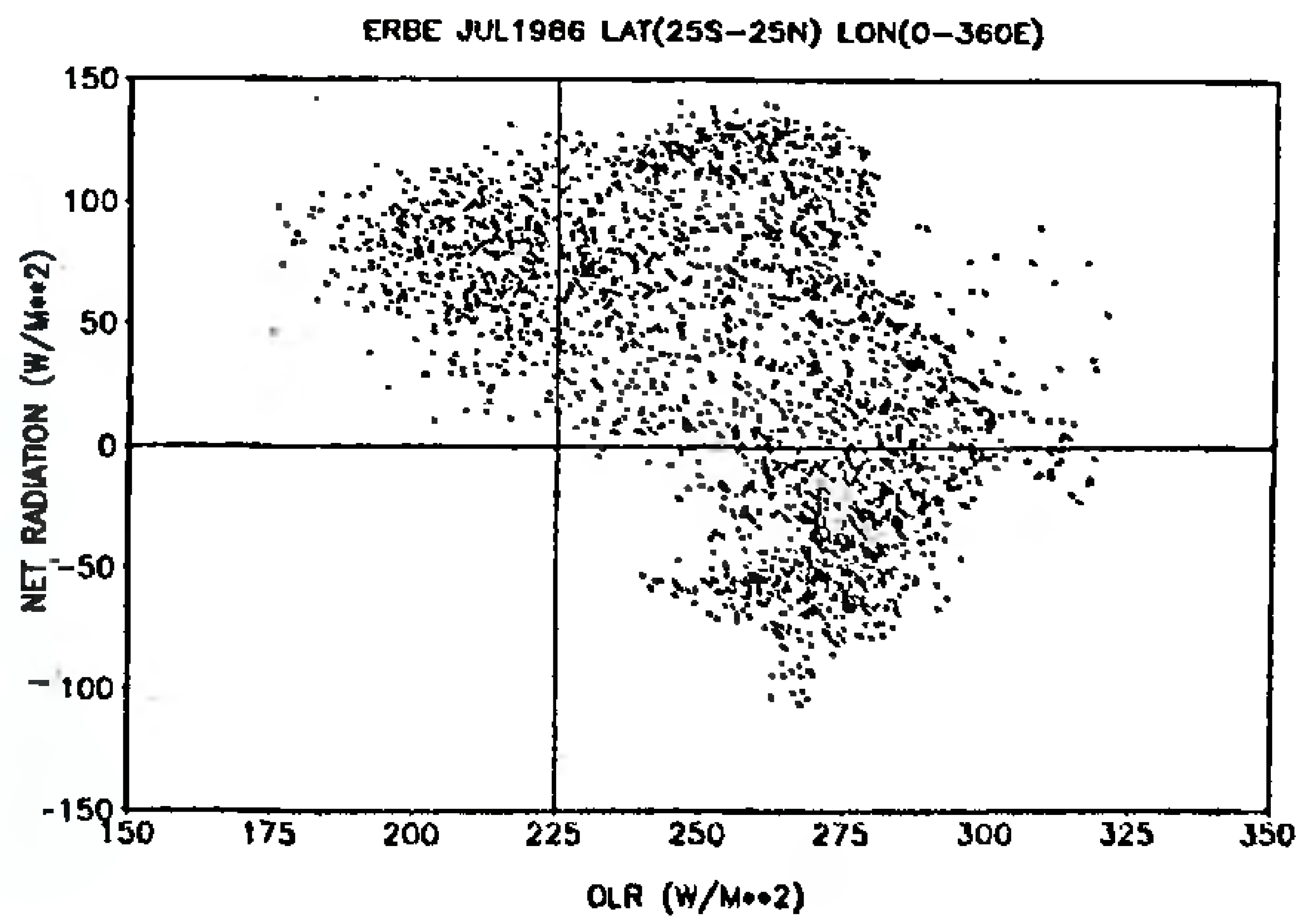


Figure 13. Variation of net radiation with precipitation over the tropic for July 1986.

convection increases rapidly with moist static energy between 330 and 340 kJ/kG. A related parameter is a simple measure of the vertical moist stability of the atmosphere, viz. the difference between the moist static energy of the lower and upper layers of the troposphere. Srinivasan[9] showed that deep convection occurs in regions with low values of the vertical moist stability, i.e. when the atmospheric profiles are near neutral stability. He also showed that low values of vertical moist stability is also a necessary, but not sufficient condition for deep convection/precipitation. Thus, while the regions with high precipitation are characterized by a low value of stability, regions with low value of stability have precipitation ranging from zero to high values. He further demonstrated that there is a close correspondence between regions characterized by low values of vertical moist stability and those with high values of CAPE.

Further studies are required to understand the relationship between these two parameters which are important for organized convection. However, it is clear that the impact of SST on the tropical atmosphere is via the changes in the vertical moist stability/CAPE. Observations of the vertical structure of the monsoonal atmosphere over the Indian seas during different phases of convection planned under the Indian Climate Research Programme[27] will provide further insight into the role of the stability.

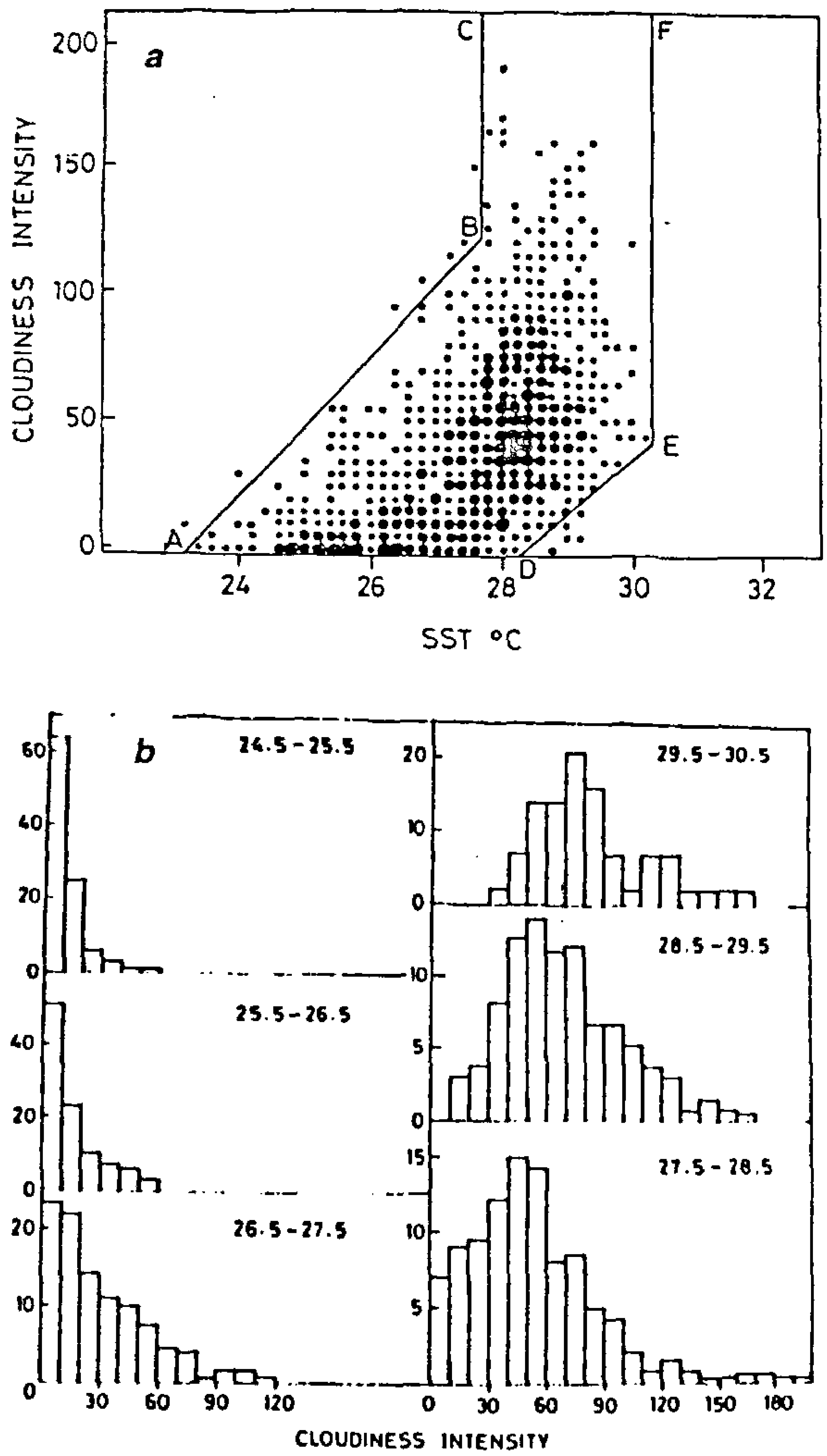


Figure 14. *a*, Two-dimensional histogram of the occurrence of different levels of cloudiness intensity at different SSTs (after Gadgil *et al.*[13]). *b*, Percentage of different levels of cloudiness intensity for different SSTs (after Gadgil. *et al.*[13]).

Axis of the TCZ and SST

The next question concerns the location of the maximum convection/precipitation or the axis of the TCZ over the warm oceans. When the SST is specified as a boundary condition, Charney's[28] simple model of the Hadley cell suggested that TCZ would be located over the region with maximum SST (when the maximum SST is not at the equator); this is a consequence of convergence in the Ekman layer at the latitude of maximum SST and the associated ascent of moist air and triggering of deep convection (as in Figure 2) in the conditionally unstable atmosphere. Studies by Manabe *et al.*[29,30] with an atmospheric general circulation model and Schneider and Lindzen[31] with a simpler axisymmetric model also showed that the TCZ was located over such an SST maximum. However, Hastenrath[32] pointed out that the maximum convection cannot be tied to the SST maximum because of the sharp decrease in insolation in association with the TCZ. The problem of understanding the relationship of the location of the axis of the monthly mean TCZ *vis-à-vis* the SST distribution is complex because even if the ascent and convection initially start at the latitude of maximum SST, the impact of the TCZ on SST will change the SST distribution and hence the location of the TCZ. In general, monthly distributions reflect the integral over the space–time variations of TCZ as well as SST. Only in special regions the interactions between the TCZ and the ocean lead to a relatively steady location of the TCZ over a sharp SST maximum[33,34].

Figure 20 shows the latitudinal variation of the average convection/precipitation, and SST for zonal averages across sectors of Atlantic and Indian Oceans for the winter and summer of 1987. It is seen that whereas over the Atlantic the axis of the rainbelt (associated with the TCZ) and SST maximum are co-located, over the equatorial Indian Ocean the axis of the rainbelt appears to be displaced by a few degrees from the SST maximum. However, it is important to note that the difference is not larger than the latitudinal extent of TCZ.

It is important to document variations of convection and SST during active and weak spells of the TCZ over regions such as the equatorial Indian Ocean which are characterized with SSTs higher than the threshold throughout the season. Only by understanding the feedback between the TCZ and the underlying oceans on synoptic and intraseasonal scales can further insight be gained about this important problem.

Challenging problems

On the subseasonal scale, the most important problem involving coupling of the ocean and the monsoon is understanding the mechanisms for (i) genesis of the TCZ and the synoptic scale systems (monsoon disturbances) over oceans, and (ii) their propagation across the ocean and land.

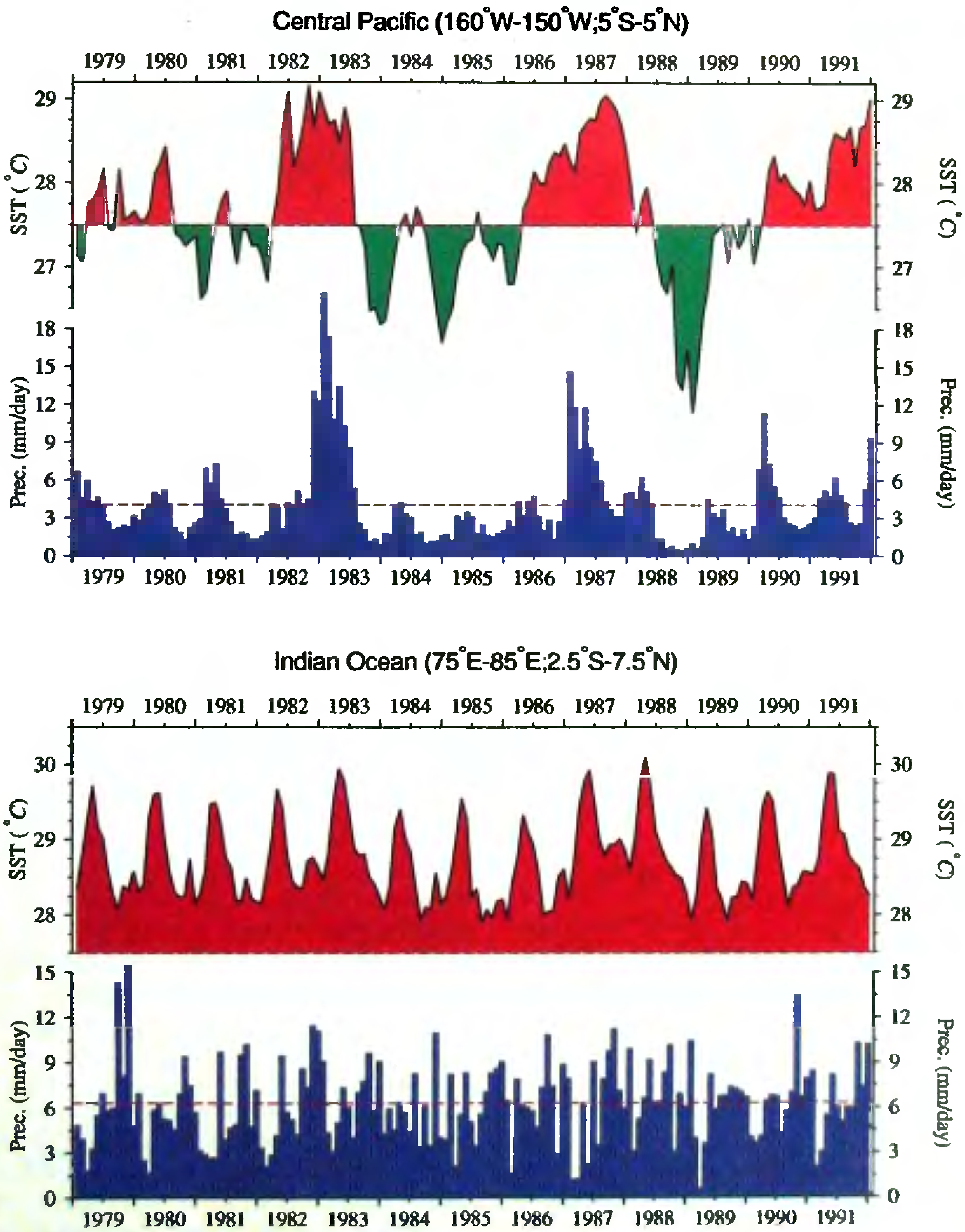


Figure 15. Variation of rainfall derived from MSU and SST over central Pacific (*above*) and equatorial Indian Ocean (*below*).

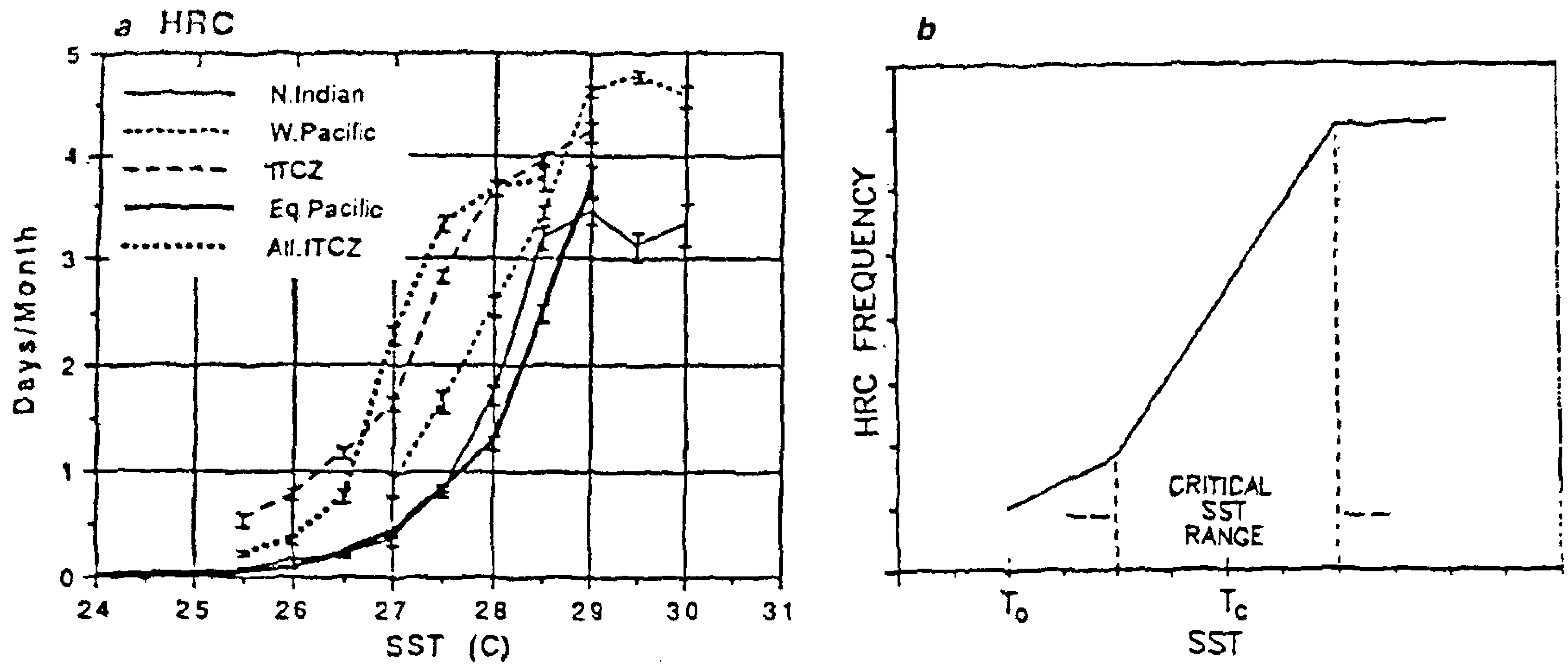


Figure 16. *a*, Variation of mean HRC (frequency of highly reflective clouds) with SST for different tropical oceans (after Waliser *et al.*[16]). *b*, An idealized representation of variation of convection with SST.

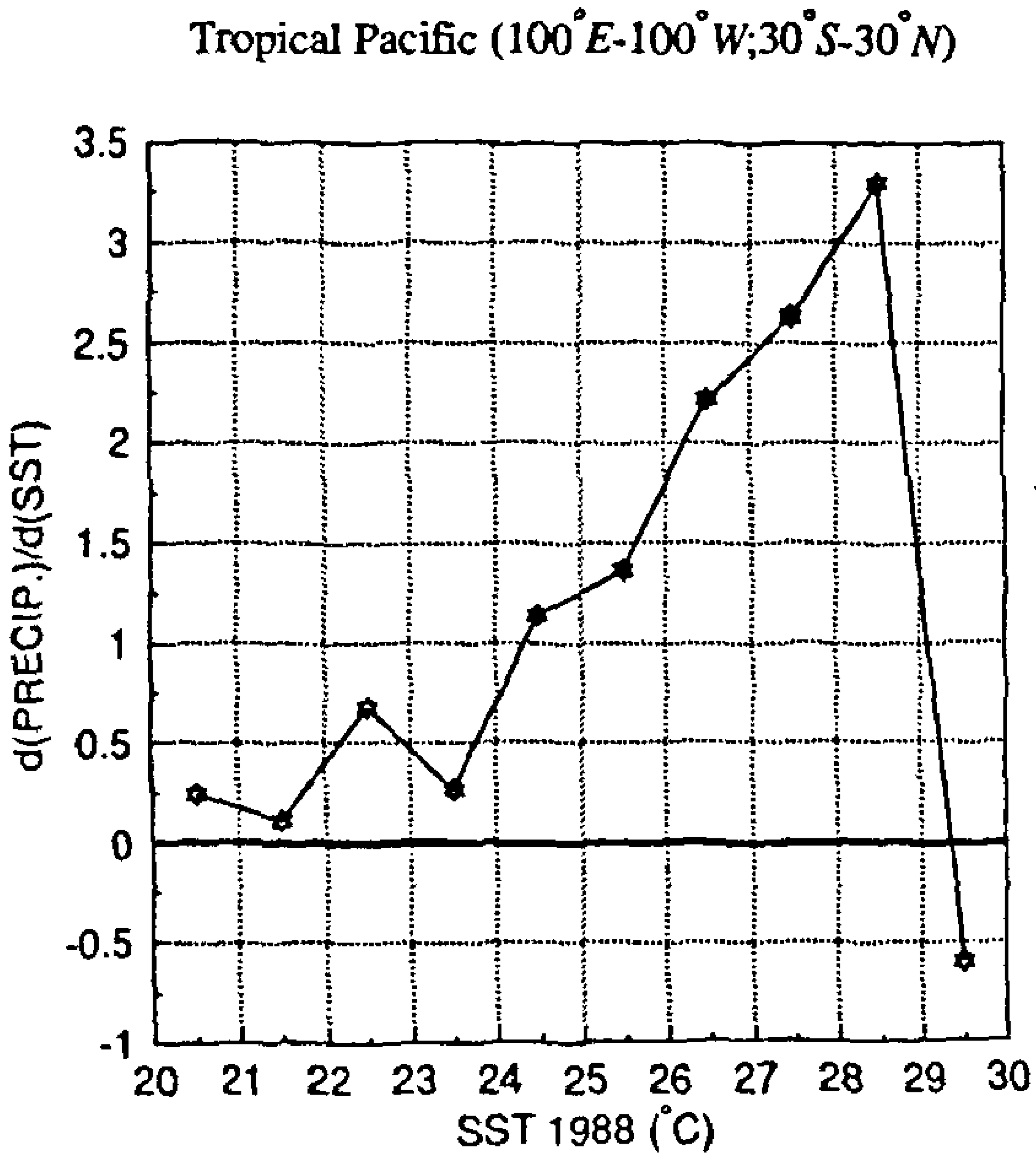


Figure 17. Variation of rate of change of precipitation with SST vs derived SST from data for 1987 and 1988 (after Bony *et al.*[22]).

It is believed that monsoon disturbances are generated with the formation of a low pressure area due to the instability of the wind (associated with the shear in horizontal and vertical direction, i.e. barotropic and baroclinic instability) which triggers growth of cumulus clouds in the conditionally unstable tropical atmosphere. Warming of the atmosphere due to the release of latent heat of condensation of water vapour in the clouds leads to an intensification of the low pressure system. While some positive

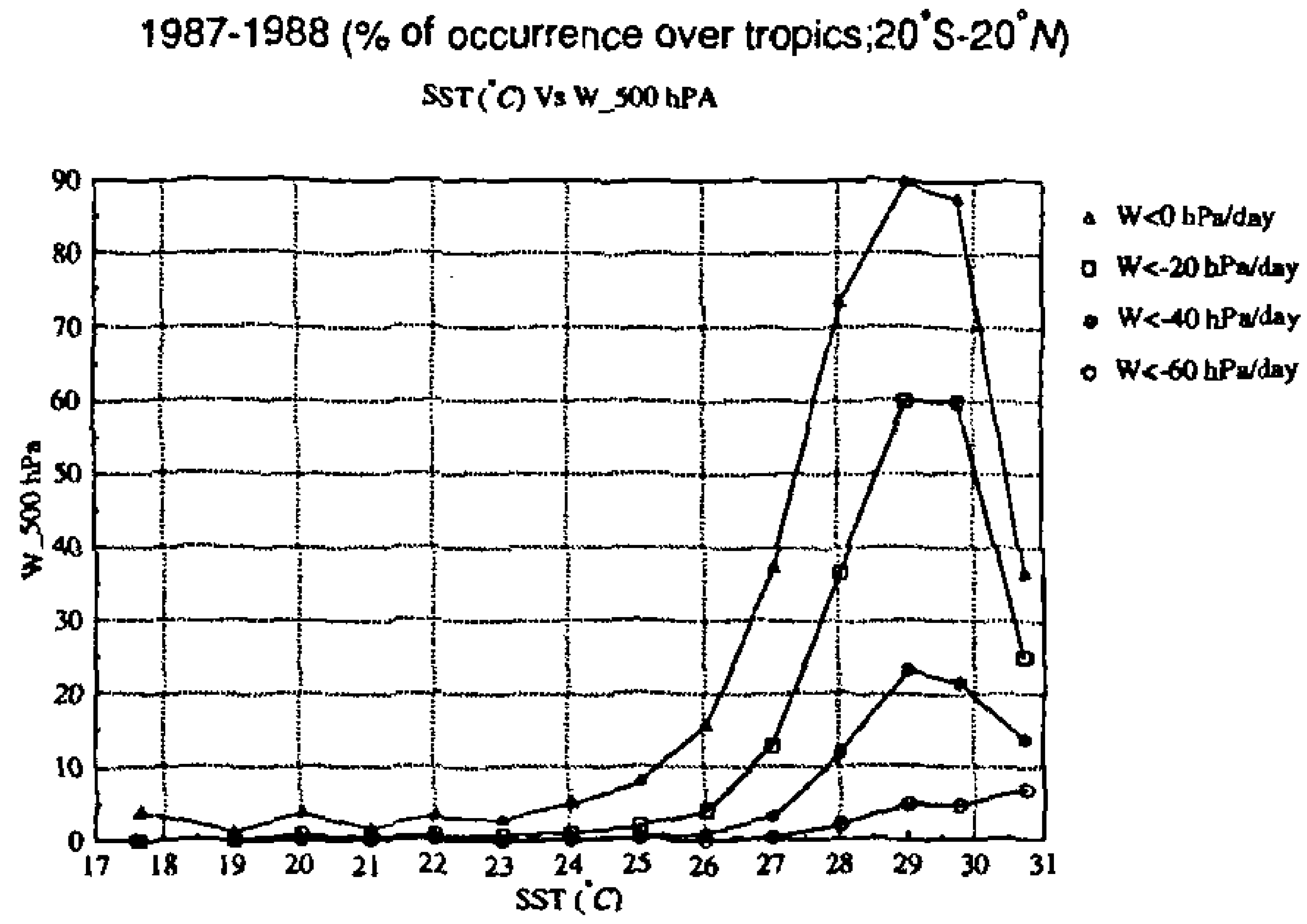


Figure 18. Variation of fraction of the region with upward velocity of different magnitude with SST (after Bony *et al.*[23]).

feedbacks such as this, as well as negative feedbacks between clouds and radiation are known, detailed observations for all the critical parameters during genesis and intensification of these systems are needed to test the hypotheses proposed. Similarly, modelling studies have suggested that propagations of the TCZ occur towards the region of higher atmospheric moist instability[35–37]. These hypotheses also have to be tested. The cause–effect relationship in the competition between continental and oceanic TCZ has to be determined. Hence the role of oceanic conditions (such as SST) in determining the intensity of oceanic TCZ in different phases of the monsoon on the intraseasonal scales (active spells and breaks) has to be elucidated. This should provide insights into the role of the equatorial Indian Ocean in interannual variation of the monsoon as well.

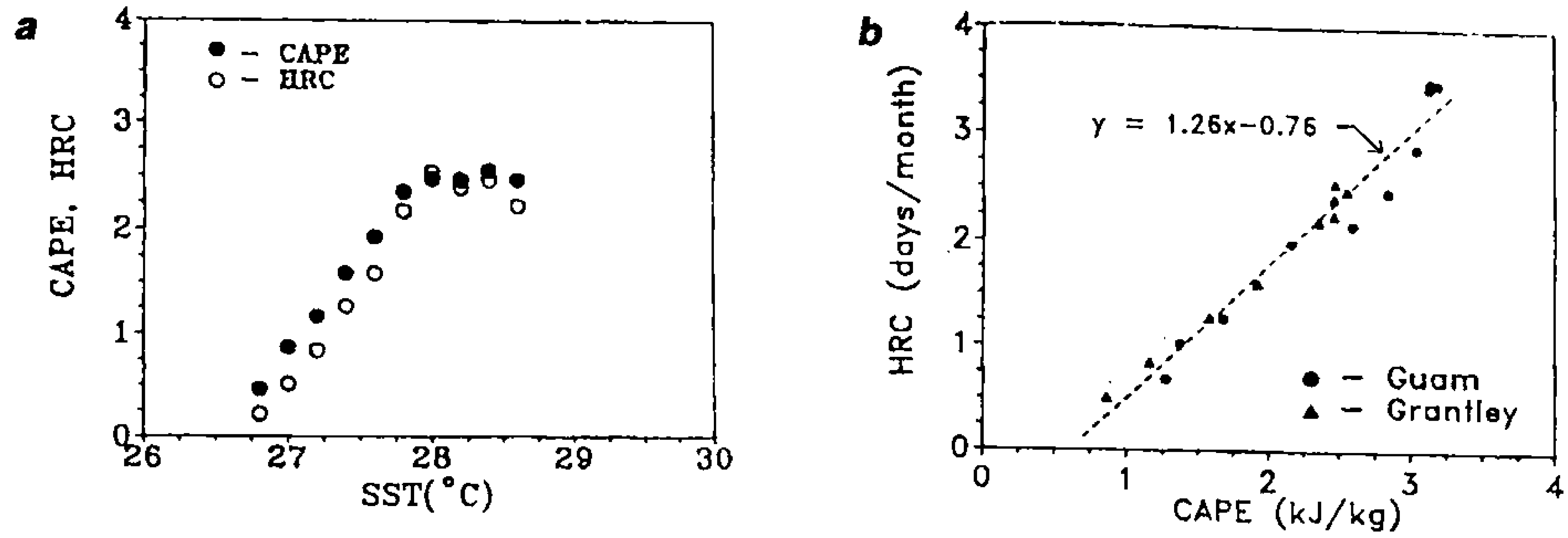


Figure 19. *a*, Variation of mean CAPE (kJ/kg) and HRC (day/month) with SST for Grantley Adams AP, assuming an air–sea temperature difference of –1°C. Data period is January 1979–December 1986; *b*, Scatter plot of HRC vs CAPE. Values of HRC ≥ 0.5 day/month are only shown.

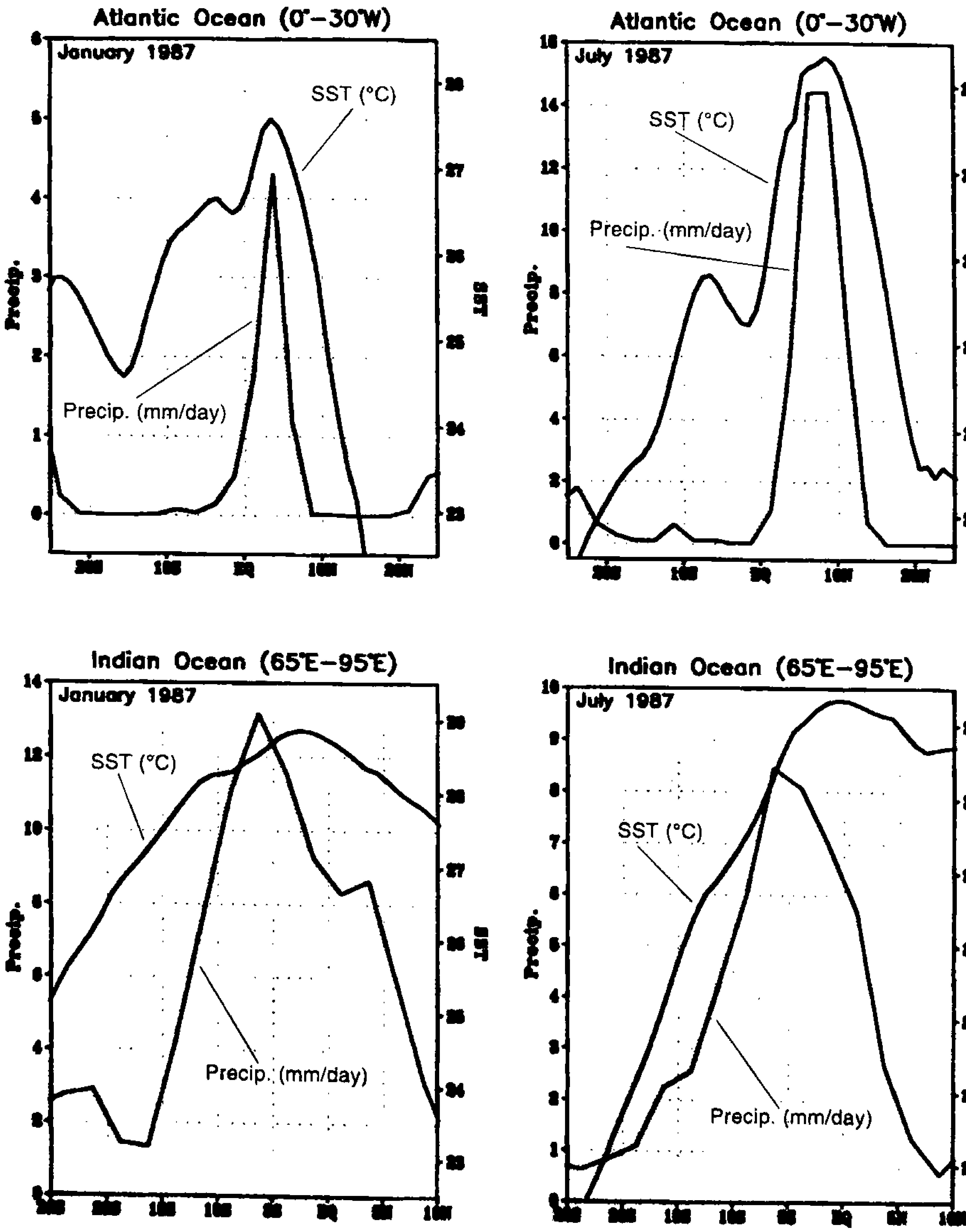


Figure 20. Variation of precipitation and SST over Atlantic and Indian Oceans.

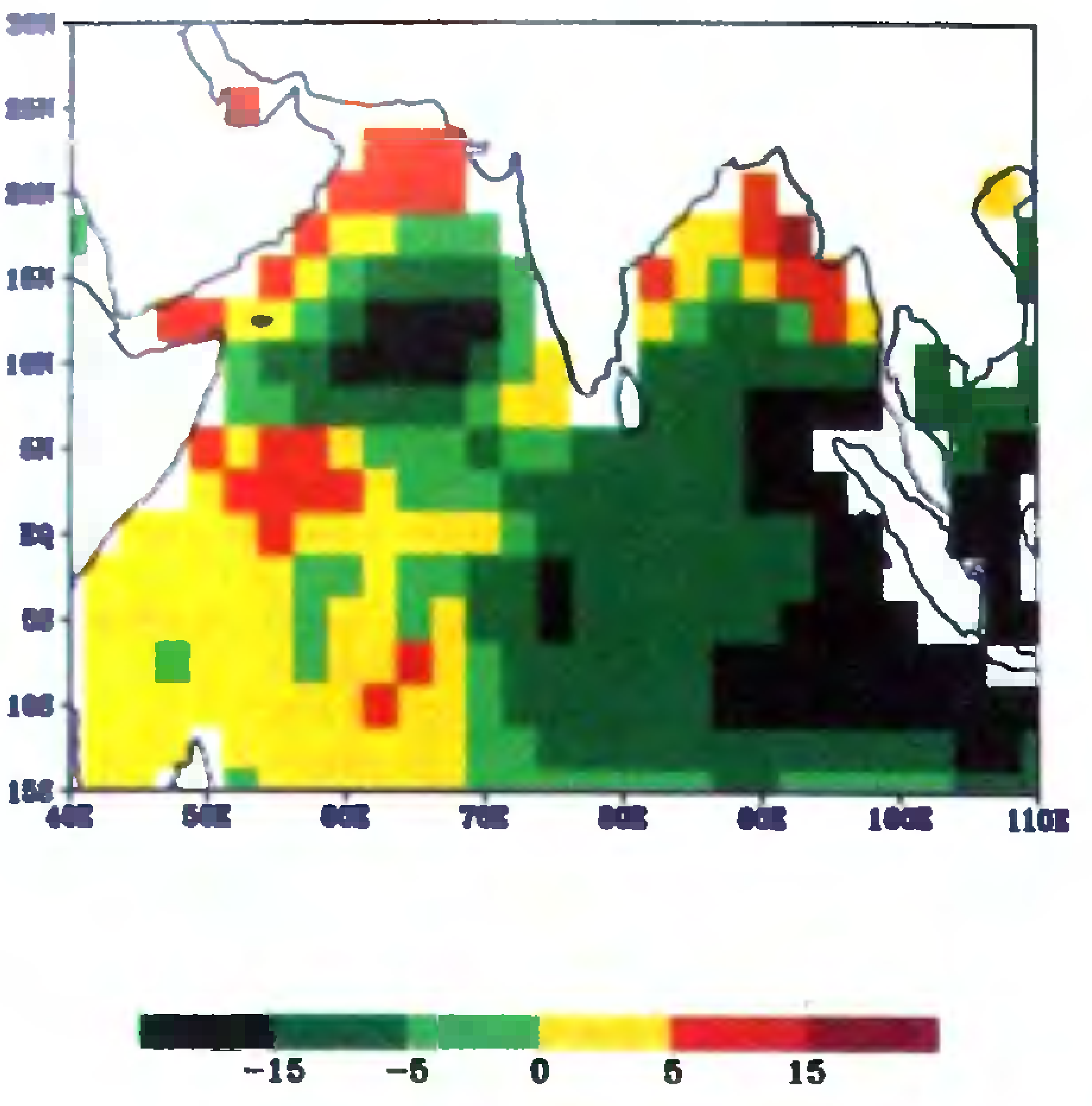


Figure 21. OLR anomaly for June–September 1998.

On the interannual scale there are large variations of the convection/precipitation over the Indian seas. For example, in the last monsoon (June–September 1998) several synoptic systems were generated over the Arabian Sea and the convection/rainfall was anomalously high. However, hardly any systems were generated in the normally fertile part of the Bay of Bengal, viz. the head Bay leading to a deficit rainfall not only over the region but also over the adjoining state of Orissa. The peculiar pattern of convection/precipitation over the Indian seas this year is clearly brought out in the OLR anomaly pattern for June–September (Figure 21).

To understand (and hence eventually predict) the genesis and propagation of organized convection over the Indian seas on the subseasonal scale and variation of convection on the interannual scale, an intensive observational effort is needed with (i) continuous observations of critical parameters with satellites, such as the OCEANSAT to be launched soon and a network of moored buoys (of which a beginning has just been made[38] and (ii) special observational experiments such as BOBMEX and ARMEX with oceanographic ships planned under the Indian Climate Research Programme[27].

A multipronged research programme with special observational experiments, observations, process models as well as studies with general circulation models should lead to deeper insight into the coupling of the monsoon to the ocean within the next few years.

1. Sikka, D. R., and Gadgil, Sulochana, *Mon. Weather Rev.* 1980, **108**, 1840–1853.
2. Gadgil, Sulochana, *Aus. Meterol. Mag.*, 1988, **36**, 193–205.
3. Gadgil, Sulochana, and Joseph, P. V., CAOS Report 98AS12.
4. Sikka, D. R., *Proc. Indian Acad. Sci.* (*Earth Planet. Sci.*), 1980, **89**, 179–195.
5. Rasmusson, E. M. and Carpenter, T. H., *Mon. Weather Rev*, 1983, **111**, 517–528.
6. Krishnamurti, T. N. and Bhalme, H. N., *J. Atmos. Sci.*, 1976, **33**, 1937–1954.
7. Charney, J. G., *Q. J. R. Meterol. Soc.* 1975, **101**, 193–202.
8. Srinivasan, J. and Smith, G. L., *Mon. Weather Rev.*, 1996, **124**, 2089–2099.
9. Srinivasan, J., *J. Indian Inst. Sci.*, 1997, **77**, 237–255.
10. Neelin, J. D. and Held, I. M., *Mon. Weather Rev.*, 1987, **115**, 3–12.
11. Palmen, E., *Geophysica* (*Helsinki*), 1948, **3**, 26–38.
12. Bjerknes, J., *Mon. Weather Rev.*, 1969, **97**, 163–172.
13. Gadgil, Sulochana, Joseph, P. V. and Joshi, N. V., *Nature*, 1984, **312**, 141–143.
14. Graham, N. E. and Barnett, T. P., *Science*, 1987, **238**, 657–659.
15. Zhang, C., *J. Climate*, 1993, **6**, 1898–1913.
16. Waliser, D. A., Graham, N. E. and Gautier, C., *J. Climate*, 1993, **6**, 331–353.
17. Joseph, P. V. and Pillai, P. V., *Mausam*, 1984, **35**, 323–330.
18. Shukla, J. and Mishra, B. M., *Mon. Weather Rev.*, 1977, **105**, 998–1002.
19. Rao, K. G. and Goswami, B. N., *Mon. Weather Rev.*, 1988, **116**, 558–568.
20. Shukla, J., in *Monsoons* (eds Fein, J. S. and Stephens, P. L.) 1987, pp. 399–463.
21. Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J., Zhu, Y., Chelliah, M., Ebisuzaki, W., Higgins, W., Janowiak, J., Mo, K. C., Ropelewski, C., Wang, J., Leetmaa, A., Reynolds, R., Jenne, R and Joseph, D., *Bull. Am. Meteorol. Soc.*, 1996, **77**, 437–471.
22. Bony, S., Sud, Y. C., Lau, K. M., Susskind, J. and Saha, S., *J. Climate*, 1997, **10**, 1441–1462.
23. Bony, S., Lau, K. M. and Sud, Y. C., *J. Climate*, 1997, **10**, 2055–2077.
24. Lau, K. M., Wu, H. T. and Bony, S., *J. Climate*, 1997, **10**, 381–392.
25. Lau, K. M. and Shen, S., *J. Atmos. Sci.*, 1988, **45**, 1781–1797.
26. Bhat, G. S., Srinivasan, J. and Gadgil, Sulochana, *J. Jpn Meterol. Soc.*, 1996, **74**, 155–166.
27. Indian Climate Research Programme, Science Plan, November 1996, Department of Science & Technology, Government of India, New Delhi, 1996.
28. Charney, J., *Planetary Fluid Dynamics* (ed. Morel, P.), D. Reidel Publishing, 1973.
29. Manabe, S., *Mon. Weather Rev.*, 1969, **97**, 775–805.
30. Manabe, S., Hahn, D. G and Holloway, J. L Jr., *J. Atmos. Sci.*, 1974, **31**, 775–805.
31. Schneider, E. K. and Lindzen, R. S., *J. Atmos. Sci.*, 1977, **34**, 263.
32. Hastenrath, S., *Climate Dynamics of the Tropics*, Kluwer Academic Publishers, The Netherlands, 1991, pp. 488.
33. Mitchell, T. P. and Wallace, J. M., *J. Climate*, 1994, **5**, 1140–1156.
34. Philander, S. G. H., Gu, D., Halpern, D., Lambert, G., Lau, N. C., Li, T. and Pacanowski, R. C., *Am. Meteorol. Soc.*, 1996, **9**, 2958–2972.
35. Gadgil, Sulochana. and Srinivasan, J., *Meteorol. Atmos. Phys.*, 1990, **44**, 119–132.
36. Srinivasan, J., Gadgil, Sulochana and Webster, P., *J. Atmos. Phys.*, 1993, **52**, 15–35.
37. Nanjundiah Ravi S., Srinivasan, J. and Gadgil, Sulochana, *J. Meteorol Soc. Jpn.*, 1992, **70**, 529–550.
38. Premkumar, K., Ravichandran, M., Kalsi, S. R., Debasis Sengupta and Gadgil, Sulochana, *Curr. Sci.*, 1999, **78**, 323–330 (this issue).

ACKNOWLEDGEMENTS. I thank Prof. J. Srinivasan, Prof. G. S. Bhat, Dr Satish Shetye and Ms S. Sajani for their valuable inputs and suggestions and ISRO for financial support. This research was supported by the Department of Ocean Development.