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Interannual Variations of Summer Monsoons: Sensitivity to Cloud Radiative Forcing

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ABSTRACT

The sensitivity of the interannual variations of the summer monsoons to imposed cloudiness has been studied with a general circulation model using the initial conditions prepared from the European Centre for Medium-Range Forecasts analyses of 1 May 1987 and 1988. The cloud optical properties in this global model are calculated from prognostically computed cloud liquid water. The model successfully simulates the contrasting behavior of these two successive monsoons. However, when the optical properties of the observed clouds are specified in the model runs, the simulations show some degradation over India and its vicinity. The main cause of this degradation is the reduced land-sea temperature contrast resulting from the radiative effects of the observed clouds imposed in such simulations. It is argued that the high concentration of condensed water content of clouds over the Indian land areas will serve to limit heating of the land, thereby reducing the thermal contrast that gives rise to a weak Somali jet. A countermonsoon circulation is, therefore, simulated in the vector difference field of 850-hPa winds from the model runs with externally specified clouds. This countermonsoon circulation is associated with an equatorial heat source that is the response of the model to the radiative effects of the imposed clouds. Indeed, there are at least two clear points that can be made: 1) the cloud-SST patterns, together, affect the interannual variability; and 2) with both clouds and SST imposed, the model simulation is less sensitive to initial conditions. Additionally, the study emphasizes the importance of dynamically consistent clouds developing in response to the dynamical, thermal, and moist state of the atmosphere during model integrations.

1. Introduction

The “vagaries of monsoon” are well known to the people of the Indian subcontinent. In some years the monsoon is good, bringing more than normal rainfall to this region, while for some other years, the seasonal total of the rainfall remains well below normal, due to a bad monsoon. In order to understand the intriguing behavior of the monsoon phenomenon, several researchers have conducted both observational and modeling studies. From observational studies on rainfall, pressure, and wind fields, it is now a well-established fact that the summer monsoon over and around India exhibits variations both on intraseasonal and interannual time-scales (Madden and Julian 1994). Modeling studies on monsoon variability have started ever since Charney and Shukla (1981) pointed out that a large part of low-latitude variability is due to boundary anomalies in such quantities such as sea surface temperature (SST), albedo, and soil moisture. Later, Shukla (1981) also argued that this low-frequency (seasonal) variability should be predictable from atmospheric general circulation models (GCMs). The slow evolution of these boundary forcings thus makes the prospect of atmo-

spheric predictions achievable on seasonal timescales (Palmer et al. 1990).

The contrast monsoon years of 1987 and 1988 provide an excellent opportunity for modelers who are interested in studying the interannual variations of Northern Hemisphere summer monsoons with the help of atmospheric GCMs. Seasonal integrations of these models are required in order to assess how well they simulate the low-frequency variability of monsoon circulation. The GCMs are also used to test any hypothesis made on the interannual variability of the Indian monsoon (Chen and Yen 1994). Observational studies are utilized in evaluating the performance of atmospheric models in simulating such modes of oscillations. Regarding the summer monsoons of 1987 and 1988, the evolution of several important parameters, which are believed to be responsible for these two successive but contrasting monsoons, have been studied in detail by Krishnamurti et al. (1989, 1990). Their studies reveal that during 1987 the lower-tropospheric streamfunction anomaly showed pronounced countermonsoon circulation that was absent in 1988. Since moisture supply is important for good performance of monsoons, Krishnamurti et al. (1990) have pointed out that the warm SST anomaly over the Bay of Bengal contributed to the supply of moisture during the active monsoon season of 1988. It may be remarked that the seasonal (June–September) difference field of SST for 1988 and 1987 shows weak negative

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four months (May–August) of 1987 and 1988 have been utilized to derive the pertinent parameters. The important features of ISCCP C1 datasets are 1) they represent a detailed description of the global distribution of cloud amount with a resolution of 280 km, 2) they provide variations of cloud amount on timescales from 3 h to several years, and 3) they provide a relatively uniform coverage of the whole globe. An interesting comparison of ISCCP and other cloud amounts has been reported recently by Rossow et al. (1993).

The diabatic heat sources and sinks play an important role in the shaping of the tropical climate. It is known that latent heat release in clouds constitutes an important heat source in the Tropics. The dynamics of the tropical atmosphere is also controlled both by local and remotely situated heat sources in the low latitudes. Gill (1980) elucidated the response of the tropical atmosphere to diabatic heat sources using a simple analytical model. He pointed out that a symmetric heat source over the equator produces, in the lower troposphere, easterly flow (Kelvin response) to the east and strong westerly inflow (Rossby response) to the west of the source region. But when the heating is displaced north of the equator, flow similar to the monsoon circulation of July is produced. This important study has thus emphasized the role of heat sources in the tropical atmosphere and particularly its response to their location and distribution about the equator. Further studies with the help of an axisymmetric model (Lindzen and Hou 1988; Hou and Lindzen 1992) showed that the strength of the Hadley circulation depends on the location of heat sources. Moreover, even small changes in the distribution of tropical heating can profoundly change the intensity of the Hadley circulation and the baroclinicity in the extratropics. These studies, without feedback between forcing and circulation, nevertheless show the sensitivity of the zonally averaged low-latitude circulation to displacement of heating from the equator. In a recent study, Zhang and Krishnamurti (1996) obtained the generalized Gill's solutions in the entire tropical belt to describe the response of the tropical circulation to the global tropical heat sources and sinks retrieved from satellite-based field of outgoing longwave radiation. These solutions exhibit most of the climatological features of the wintertime and the summertime (including the Asian monsoon) circulations of the lower troposphere. The Asian summer monsoon arises due the land-sea temperature contrast, which is influenced by external conditions and internal feedbacks (Meehl 1994a,b). The strength of the monsoon increases (decreases) as this contrast increases (decreases). It has been illustrated and suggested by Meehl (1994a) that external conditions involving land albedos (Charney et al. 1977) could reduce the land-sea temperature contrast leading to a weak monsoon, while internal feedback could increase it leading to a good monsoon. Moreover, this temperature contrast introduces considerable asymmetry in important meteorological parameters.

The atmosphere can be heated in a variety of ways

involving several complicated processes. When one considers only the dry atmosphere, most of the solar radiation will be absorbed at the earth surface, and the air parcels after receiving this energy in the form of sensible heat will become unstable and move vertically up (updrafts) in order to cool down. Thus, adiabatic ascent during daytime, when ground is hot, will act as a heat sink (cooling), while nighttime descent will act as a heat source for maintaining the temperature of lower layers as the ground cools. The diurnal variations are thus caused by vast differences of thermal effects of the surface. Chinese researchers (Yeh 1982) have noted important diurnal variations over the western Tibetan Plateau. Earlier studies (Flohn 1960; Koteswaram 1958) have also highlighted its importance as an elevated heat source for the establishment and maintenance of the summer monsoon. The interannual variability of heat sources has been observed to be large in the regions where the heating is strong (Pal et al. 1993).

However, the situation is different for the moist atmosphere. When radiatively active clouds are absent, the situation is initially same as that for the dry atmosphere. But if there is a moisture supply from the surroundings, moisture convergence into this region will give rise to clouding after parcels reach the lifting condensation level. In the unstable layers of the atmosphere the buoyancy of parcels will be maintained and deep convection can develop if moisture convergence continues in the region. Two things happen at this stage. First, due to precipitation over this region, the ground cools down, which will reduce the thermal contrast with the surrounding region. Second, the incoming solar radiation will be reflected back to space at the cloud tops, diminishing the supply of energy into the surface-atmospheric (or the ocean-atmosphere) column. This might result in an overall cooling (clouds heat the atmosphere and cool the surface) of the column, which could initiate more stable lapse rate conditions. In the course of time, the atmospheric column will again turn cloud free. Thus, clouds possess a strong autoregulatory mechanism.

Another scenario can also be constructed on the role of clouds during the monsoon period. Yasunari (1979) has noted two dominant modes, of around 40 and 15 days, in the summer monsoonal fluctuations after analyzing satellite cloudiness. In a later study, Sikka and Gadgil (1980) have shown two preferred locations of maximum cloud zones (MCZ): one over India (north of 15°N) and another (secondary) over the equatorial Indian Ocean. They have also found that the cloud bands have a strong tendency to move northward in all phases of a monsoon and could persist for a few days at a particular latitude. When a monsoon is active, warmer air ascends over monsoon land areas and relatively colder air descends over oceanic areas, which goes with an east–west circulation (Krishnamurti 1971). Like latent heat release, radiative forcing associated with clouds is another important diabatic process that drives

the circulation of the atmosphere. The clouds are generally categorized into three types: low, medium, and high (the ISCCP clouds are also classified into these categories). In fact, these clouds of different height categories affect radiation in different ways. The radiation can also affect clouds significantly. If one considers the presence of only one cloud type and its associated radiative effect in a surface–atmospheric column, then low clouds cool the column above, high clouds warm the column below, while middle clouds warm the column below and cool the column above the level where they are present (Paltridge and Platt 1976). Low clouds in practice radiate at the temperature of the underlying earth's surface, which results in strong cooling of overlying atmospheric layers. Thus, interaction of long-wave radiation with clouds may alter the thermal stratification of cloud layers through cloud-top cooling and cloud-base warming mechanism. Also, the presence of clouds may induce differential cooling between clear and cloudy regions. Hence, the longwave radiative process may increase convergence into the cloud system, which may, in turn, enhance precipitation at the surface in the cloudy regions.

With this background knowledge, it is now possible to perceive the variability induced by clouds in the monsoon rainfall due to their radiative effects only. Let us consider a situation during the active phase of the monsoon when the MCZ is located over India and the Indian Ocean region is initially cloud free. Then, if low clouds appear in the equatorial zone, they shall cool the column above and sinking motion of cold air over the adjacent cloud-covered ocean will increase. It could, in turn, strengthen the rising motion over India and, possibly, increase the quantum of rainfall there, too. In a similar situation, if high clouds cover the Indian Ocean, the quantum of monsoon rainfall over India will decrease in view of the arguments presented in the previous paragraph. But if convection grows over the equator, while the monsoon is still active over India (i.e., active convection at both the preferred locations of MCZ), a zone of downward motion will appear somewhere in between the equator and the position of MCZ over India. With the establishment of this zone of descending motion, the monsoon trough (a region of low pressure) has to adjust its position over the monsoon land areas. If it shifts northward then it will mark the beginning of break monsoon over India (Raghavan 1973; Dakshinamurti and Keshavamurti 1976). Thus, it is clearly expected from the above discussion that large variability can be introduced by the radiative effects of clouds. Moreover, it depends on cloud type and the location of cloud bands in the meridional direction. In other words, clouds can significantly influence the temperature contrast in the monsoon region affecting the external conditions and internal feedbacks discussed earlier by Meehl (1994a,b) in an attempt to explain the varying performance of monsoon from one year to another.

The design of the experiments for this study requires

the radiative forcing of externally imposed clouds when SST forcing is also present. But, for specifying the observed clouds in the model, their optical properties—which directly influence their radiative effects—must be known *a priori*. We have used the ISCCP C1 datasets for this purpose. The ISCCP clouds have been taken because the data are as close to observations as possible and the optical properties of different categories of clouds are fairly accurate (Rossow et al. 1993). To our knowledge, no such study has been reported so far that has used ISCCP clouds (or the observed clouds) to examine the interannual variability of monsoons. This study therefore constitutes a first attempt in this direction using the ISCCP data in a more complete atmospheric general circulation model. It is important to point out here that the purpose of these experiments is not to express any judgment on the impact of ISCCP clouds on the model simulations of monsoon variability. It is rather to stress the importance of time-dependent cloudiness that is consistent with the model dynamics and to examine how a model responds if this condition is not satisfied. The study may, therefore, provide a new insight into the response of the model when the radiative feedbacks of clouds are considered together with the SST data.

The main aim of this study is twofold. On the one hand, it investigates how a GCM responds when observed clouds are specified externally over the Asian summer monsoon region and simulates the interannual variability of monsoon, while on the other hand, it examines the sensitivity of rainfall over India to initial conditions. A series of sensitivity experiments (120-day simulation runs) were performed with the atmospheric global model of the Laboratoire de Météorologie Dynamique (LMD GCM). The total number of such numerical experiments was eight; the first four experiments were performed with initial conditions of 1 May 1987, and the other four with that of 1 May 1988. In the next section, the important features of the LMD GCM and the derivation of cloud optical properties from ISCCP C1 data are briefly described. Next, the results of this study are presented with the help of 90-day mean (June–August) fields of precipitation, streamfunction, velocity potential, and circulation, and their differences from various experiments performed for this study. Although monsoon rainfall during the month of September contributes appreciably, in some years, to the seasonal total rainfall amounts in this region, it has not been considered here as the length of each simulation run is restricted to just 120 days. Finally, the main findings of this study are summarized in the last section.

2. Model and experimental design

We have used the LMD GCM for carrying out the sensitivity experiments to initial conditions, the SST forcing, and the radiative forcing of clouds. This section is devoted to the general description of the model that

has evolved over a period of time and to the description of the methodology to incorporate ISCCP clouds into this model. The first version of this model has been described by Sadourny and Laval (1984). This is a finite-difference general circulation model that uses Arakawa's C-type grid for discretization of operators, and the grid points are regularly spaced in longitude and sine of latitude. One of the important characteristics of this model is that it has enhanced meridional resolution in the Tropics resulting from the area-conserving meshes. The horizontal resolution adopted for this study has 96 points in longitude against 72 points in sine of latitude. The model uses normalized pressure ($\sigma = p/p_s$) as its vertical coordinate with 15 layers. The time integration scheme is a combination of explicit schemes (the basic time step of 15 min is split into one Euler-backward step followed by four leapfrog steps). Fourier filtering is applied to the longitudinal derivatives near the poles to ensure uniform and longer time steps. Lateral diffusion is modeled by a mixed bi-Laplacian in which the inner part acts on sigma coordinate and the outer parts on pressure coordinate.

The physics of the model is fairly complete except that the diurnal cycle is not included into this version. Diurnal cycle must be included while studying clouds and solar radiation sensitivities with climate models. But here we do not resolve anyway the diurnal cycle of the clouds over the monsoon area due to the absence of the Indian geostationary satellite (INSAT) data. Moreover, deep convection dominates the active monsoon periods and diurnal cycle may affect only low clouds, which are generally present over the adjacent ocean. The diurnal variation in convection over deep tropical oceans is again relatively weak. It is, therefore, expected that no significant error is introduced by neglecting the diurnal cycle in the model simulations. Physical parameterizations include solar radiation, longwave radiation, large-scale condensation, adjustment for small-scale convection, cumulus convection, and a prognostic equation for cloud water.

At the surface, eddy fluxes are calculated using a bulk method with drag coefficients varying with vertical gradient properties. The description of all the parameterizations of physical processes included in this model may be found in Le Treut et al. (1994). The hydrological exchanges between vegetation and atmosphere are computed using a scheme, called *sechiba*, that allows eight types of vegetation cover within a grid box (Ducoudré et al. 1993; Laval et al. 1996).

This sensitivity study uses primarily the prescribed cloud fields, which are chosen to be as realistic as possible. Hence, the observed cloud fields of 1987 and 1988 from the ISCCP data have been utilized. It is necessary to allocate the ISCCP cloud fractions to various model layers, but this cannot be done without some hypothesis. The cloud radiative properties in the model are calculated from the path of LWC. It is also necessary to derive this fundamental quantity from the ISCCP data. The

scheme to obtain cloud cover and LWC is first presented here for the sake of completeness. It may be recalled that ISCCP C1 analysis provides information of 132 variables, such as cloud amount, cloud-top temperature, albedos, cloud optical depths, among others, every 3 h over a spatial resolution of 280 km². However, validation studies are required for these analyses. One such comparison of C1 cloud amounts with those derived from Advanced Very High Resolution Radiometer (AVHRR) images has been done by Weare (1992). The mean C1 cloudiness was found to be only 6% higher than that of AVHRR. The global patterns of cloud optical thickness with temperature have been reported by Tselioudis et al. (1992).

The ISCCP C1 analyses for May–August for the years 1987 and 1988 were utilized over a region between 30°S–30°N and 30°–120°E. This region practically covers the entire area of influence of the Asian summer monsoon. The ISCCP C1 data, however, do not contain information from the Indian geostationary satellite positioned in this region.

As a consequence, the analysis for this region relies entirely on the coverage by the passage of polar-orbiting satellites. The time resolution of the cloudiness is poorer. A composite field is first prepared for each day using the ISCCP data available at 0600, 0900, and 1200 UTC. These hours generally correspond to daytime hours over the monsoon region. From the composite data of each day, cumulated histograms of 5-day cloud statistics are produced that serve to determine mean cloud properties and are then interpolated to the model grid. Note that the technique of histogram partitioning used here allows the actual definition of averaged cloud fields, which is less meaningful when performed by direct averaging of these quantities. The histogram partitioning used for ISCCP divides the cloudiness by optical depth (five levels) and cloud-top pressure (seven levels). Each cloudy pixel thus belongs to one of the 35 categories depending on the optical depth and the top pressure of the cloud it represents (Fig 1). From the ISCCP classification, it is also possible that at a particular pressure level, clouds with different optical depths could be present. Since cloud cover is a key parameter in the model, it must be appropriately defined because any error in cloud cover will affect both shortwave and longwave radiative transfer calculations. For the experiments presented here, an effective cloud cover, C_f , has been first computed using cloud fractions at a given interval of cloud optical depth. Thus,

$$C_f = \sum_{i=1}^5 C_i, \quad (1)$$

where C_i are cloud fractions at given intervals τ_i of cloud optical depth. The effective cloud optical depth is then computed as a weighted mean given by

$$\tau_f = \sum_{i=1}^5 C_i \tau_i / C_f. \quad (2)$$

TABLE 1. List of numerical experiments.

Experiment	Initial conditions	SST	Clouds
I87-S87-CM	1 May 1987	1987	Model clouds
I87-S88-CM	1 May 1987	1988	Model clouds
I87-S87-C87	1 May 1987	1987	ISCCP clouds 1987
I87-S88-C88	1 May 1987	1988	ISCCP clouds 1988
I88-S87-CM	1 May 1988	1987	Model clouds
I88-S88-CM	1 May 1988	1988	Model clouds
I88-S87-C87	1 May 1988	1987	ISCCP clouds 1987
I88-S88-C88	1 May 1988	1988	ISCCP clouds 1988

for the period from May to August. These experiments (both control and sensitivity runs), of 120-day duration each, are summarized in Table 1. The first column in this table contains the abbreviations used for referring to a numerical experiment in the text. Thus, the abbreviation I87-S87-CM refers to the simulation that has been carried out with the initial conditions of 1 May 1987 (I87), sea surface temperatures that have been specified for 1987 (S87), and model clouds (CM) that are internally computed using the prognostic equation for cloud water. An experiment with above initial conditions and SST, but using the ISCCP clouds of 1987, say, will be referred to as I87-S87-C87 in the text. The other simulations can also be identified easily in this manner.

For each set of initial conditions, two integrations were forced by SST alone where cloud parameters were internally calculated by the model. The rest of the model runs were forced with a coherent combination of SST and ISCCP clouds for 1987 and 1988 (i.e., only the combinations S87-C87 and S88-C88 of SST and clouds are considered in these simulations). Mean fields were then computed by considering the last 90 days July–August (JJA) in each integration. However, the important features of this sensitivity study are discussed here by using such 90-day average fields of streamfunction, velocity potential, precipitation, and winds. The mean fields of streamfunction and velocity potential, as well as their differences, of 1987 and 1988, are presented over the whole globe, while precipitation fields and 850-hPa circulation are presented over a region 20°S–40°N, 20°W–130°E. The average fields on visual inspection also demonstrate the clear differences in precipitation, velocity potential, and circulation of the model integrations performed with realistic sea surface temperatures for the two monsoons. The model thus captures the interannual features of monsoon when forced with SST alone. This is in agreement with the conclusions of Palmer et al. (1992). However, the distinct features of various runs are better illustrated with the help of difference fields; hence, the various model runs have been compared in the light of differences of JJA mean fields of precipitation and other meteorological parameters of interest.

a. Difference fields: June–August

In this section, various simulation experiments are compared in order to discuss mainly the variability of a monsoon and its sensitivity to radiative forcing, and the impact of initial conditions. The comparison between I87-S87-CM and I88-S88-CM will show how the model simulates the interannual variations of a monsoon when model produces its own clouds as the integrations progress, whereas a comparison of I87-S87-C87 and I88-S88-C88 will show how the cloud–SST patterns affect the interannual variability when the clouds are externally prescribed and need not be associated with any precipitation. The sensitivity of model simulations to initial conditions, with SST and model generated or imposed clouds in the monsoon region, is investigated from the whole ensemble (shown in Table 1) of eight numerical experiments performed for this study.

1) PRECIPITATION

The 90-day mean precipitation for the four experiments performed with I87 and forced with SST and clouds of 1987 are shown in Fig. 3. It is evident from Fig. 3a that in experiment I87-S87-CM (i.e., initial conditions of 1 May 1987, SSTs of 1987, and model clouds), precipitation is absent in the central part and is mainly concentrated in the eastern part and the west coast of India. But precipitation increases, as shown by Fig. 3c, when the ISCCP clouds (I87-S87-C87) are specified. However, when the integration is forced by SSTs of 1988 (I87-S88-CM), precipitation increases over India with model clouds (Fig. 3b). As expected, when the ISCCP clouds of 1988 are imposed (I87-S88-C88), the radiative forcing has further increased the precipitation (Fig. 3d) over India. Thus, the initial conditions I87 are not as important as the SST and clouds. The model simulations, I88-S87-CM and I88-S88-CM, show increased precipitation over central India in comparison to that from the simulations I87-S87-CM and I87-S88-CM. Similar experiments were performed with initial conditions I88. On comparison, it appears that the initial signal in I88 was very strong relative to that in I87. Some spectacular changes have been noted in the rainfall over India in the two experiments I88-S87-C87 and I88-S88-C88. Surprisingly, with I88, the precipitation decreases (figures not shown) over parts of central India with S88 and C88. In the equatorial Indian ocean, a strong belt of precipitation has been simulated in some of the experiments where 1988 SSTs have been specified. Why does the model simulation degrade, with initial conditions I88, when observed clouds C88 are imposed together with SSTs of 1988? This may, however, be answered by the weaker monsoon circulation arising from the suppressed land–sea contrast in temperature. This point will be elaborated further in the next section. The distribution of precipitation in all the experiments can be compared with the climatology (JJA) of Jaeger

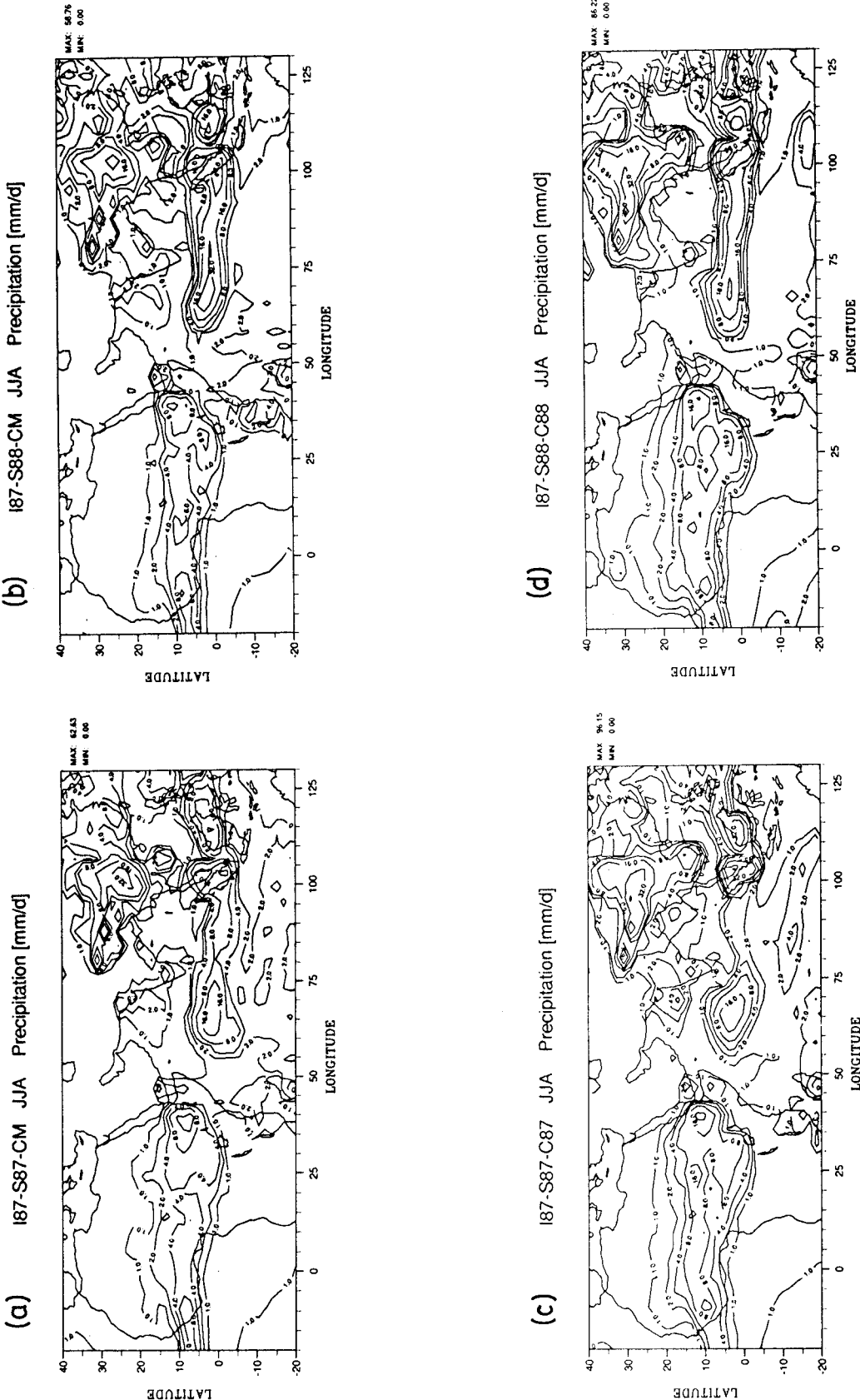


FIG. 3. Simulated June–August mean precipitation with 1 May 1987 initial conditions for (a) SST 1987, (b) SST 1988, (c) SST 1987 and observed clouds 1987, and (d) SST 1988 and observed clouds 1988. Contour intervals (mm day^{-1}): 1, 2, 4, 8, 16, ...

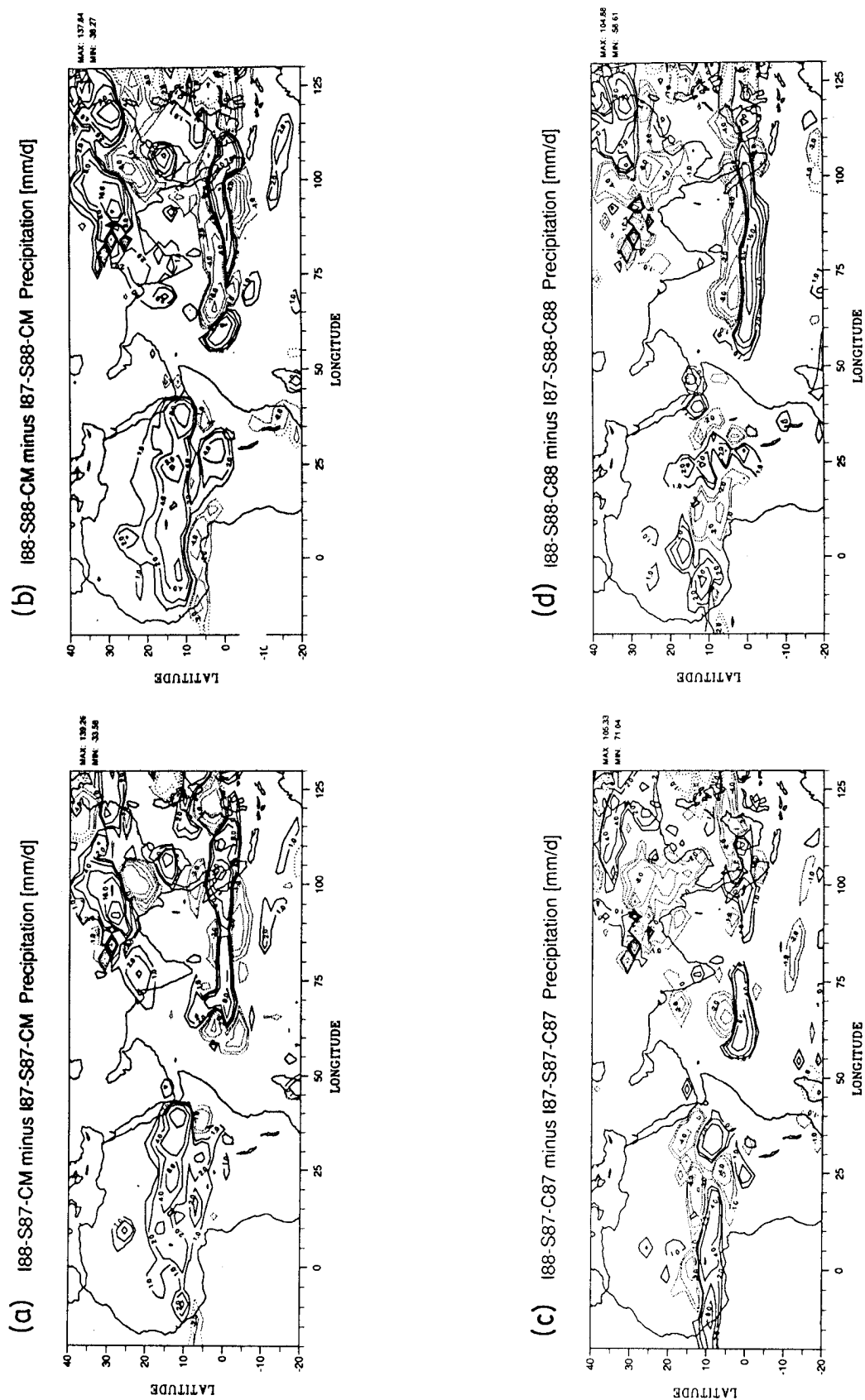


FIG. 4. Precipitation differences for coherent forcing but different initial conditions: (a) SST 1987, (b) SST 1988, (c) SST 1987 and observed clouds 1987, and (d) SST 1988 and observed clouds 1988. Contour intervals (mm day^{-1}): $-16, -8, -4, -2, -1, 1, 2, 4, 8, 16$.

divergence over Central America (Fig. 9a) has vanished in the 200-hPa velocity potential differences (Fig. 9b) computed from simulations I88-S88-C88 and 87-S87-C87. Thus, under the radiative forcing of prescribed clouds, the velocity potential difference field forms a wavenumber 1 structure showing divergence over monsoon region and convergence over the Pacific Ocean; nevertheless, the gradients of the isopleths in Fig. 9b have weakened relative to those in Fig. 9a. It may therefore be inferred from the above discussion that the response of the model to externally imposed cloudiness (over the monsoon region) deteriorates in the large-scale flow, relative to its response to dynamically consistent clouds.

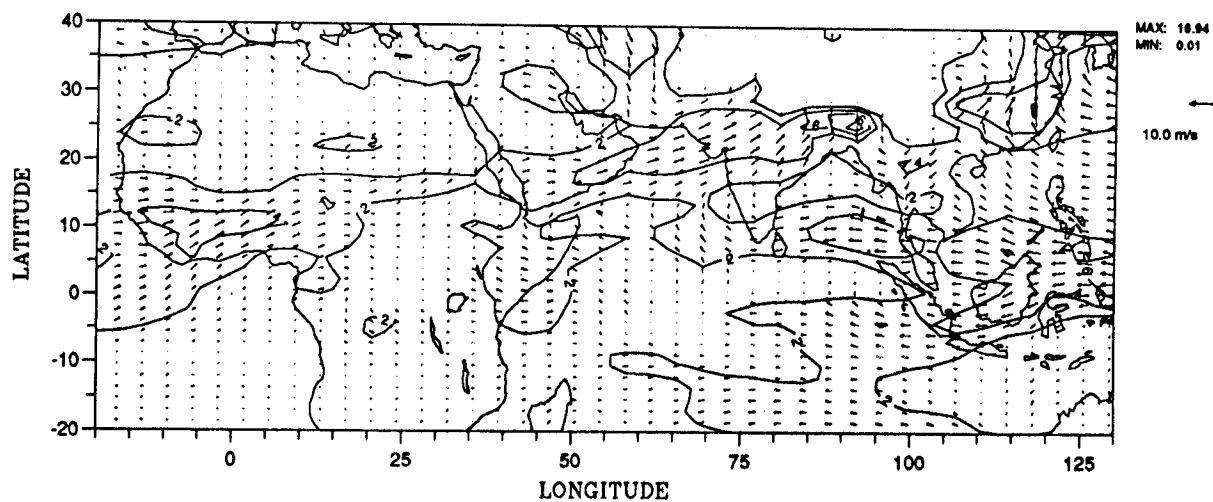
3) 850-hPa WIND

The streamfunction and velocity potential fields have clearly shown that radiative forcing of clouds has an important bearing on the monsoon circulation. We now examine its influence on the low-level circulation of this region. The onset of monsoon occurs with the establishment of the low-level cross-equatorial flow and the Somali jet over the Arabian sea. The Somali jet brings much-needed moisture over the Indian peninsula; therefore, its intensity directly affects the quantity of monsoon rainfall received by this region. The 850-hPa mean wind fields for the JJA period for the two simulations performed with the 1987 initial conditions (I87) have been shown in Fig. 10. The model has well reproduced the cross-equatorial flow and the Somali jet that characterize the monsoon circulation both with interactively computed clouds and with prescribed clouds. But the intensity of this low-level jet is weaker in I87-S87-CM than that simulated in I87-S87-C87. The cloud radiative properties of imposed clouds have thus induced an enhancement of the Somali jet in the latter simulation. This increase is possibly related to the thermal contrast between the Indian Ocean and the continent (the Indian landmass). It appears that the changes in the cloud radiative properties, besides increasing the convective activity over the monsoon region, have also increased the surface heating over India in the simulation I87-S87-C87 relative to I87-S87-CM, accentuating the thermal difference between the land and ocean. Such a pronounced meridional thermal gradient should be consistent with a relatively intense low-level jet over the Arabian sea in the experiment I87-S87-C87. In more specific terms, Fig. 10b shows wind speeds in excess of 10 m s^{-1} that occur over the entire Arabian sea region in association with strong easterlies (wind speeds exceeding 5 m s^{-1}) over the southern Indian Ocean and a broad cross-equatorial flow. Also note that the model has unrealistically simulated a strong southwesterly flow (wind speeds exceeding 15 m s^{-1}) at 850 hPa over northwest India. However, the aforementioned features of the low-level circulation are weaker (especially the Somali jet) in simulation I87-S87-CM (Fig. 10a) in comparison to

those simulated in I87-S87-C87 (Fig. 10b). In another set of simulations (figures not shown) performed with the 1988 initial conditions (I88), the strength of the Somali jet (speed exceeding 10 m s^{-1} over the Arabian Sea) in I88-S88-CM is rather weak compared to the maximum velocity ($>15 \text{ m s}^{-1}$) at the core of the observed jet in the 1988 JJA 850-hPa ECMWF analysis (WCRP 1992, p. 1.24). Nevertheless, the overall curvature of the cross-equatorial flow has been captured by the LMD model. However, when the ISCCP clouds are imposed together with SST in the experiment I88-S88-C88, slight weakening (speed less than 10 m s^{-1}) of the low-level jet has been noted over some parts of the Arabian Sea. It may therefore be inferred from the above discussion that the model response to externally imposed clouds together with SST has produced contrasting results: a better monsoon in the sensitivity run of 1987 (I87-S87-C87) than that of 1988 (I88-S88-C88). This point is further examined with the help of vector difference of the velocity fields from these two runs.

The vector wind difference field at 850 hPa, shown in Fig. 11, perhaps presents the most remarkable result of this sensitivity study. The interannual changes (1988–1987) in the ECMWF analysis (WCRP 1992, p. 2.4) show stronger equatorial easterlies over the Indian Ocean and an enhanced westerly inflow into western Africa. As can be seen from Fig. 11a, the wind differences show a stronger easterly flow north of 5°N, when the cloud radiative properties are internally calculated in the model. However, the difference flow is much weaker over the equator in comparison to the stronger easterly flow noted earlier in the ECMWF analysis differences. One can also notice stronger southwesterlies in Fig. 11a over the Arabian Sea and India. Though the observed and simulated difference fields disagree in the equatorial Indian Ocean, the vector wind differences at 850 hPa from the simulations I88-S88-CM and I87-S87-CM resemble those of the ECMWF 1988–87 analysis in the Arabian Sea. This shows that model has successfully simulated an intense southwest monsoon in 1988 in comparison to 1987 (WCRP 1992, p. 2.4). The stronger westerly inflow into West Africa can also be noted in Fig. 11a. But when ISCCP clouds are imposed over the monsoon region, the 850-hPa circulation shows dramatic changes. The difference field, shown in Fig. 11b, illustrates the interannual changes in the wind field resulting from the specification of observed clouds in simulations I88-S88-C88 and I87-S87-C87. Indeed, this figure shows that model has simulated a stronger monsoon in the sensitivity run I87-S87-C87 relative to I88-S88-C88. This is in contravention to reality, hence some explanation is warranted. We know that the monsoon arises as a result of land–sea temperature contrast. Therefore, a weaker Somali jet in I88-S88-C88 relative to I87-S87-C87 is consistent with the weaker thermal contrast caused by the imposition of observed cloud radiative properties, while the reverse is true in the simulation I87-S87-C87. The precise mechanism of this

(a) Vent a 850hPa [m/s], 188-S88-CM minus 187-S87-CM



(b) Vent a 850hPa [m/s], 188-S88-C88 minus 187-S87-C87

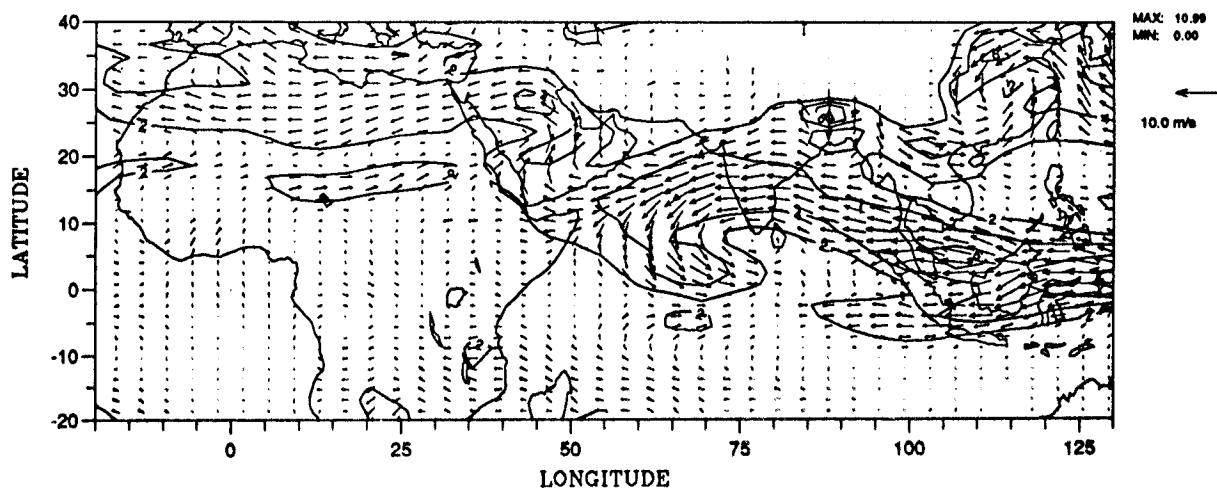


FIG. 11. Same as Fig. 5 but for 850-hPa winds. Contour interval: 2 m s^{-1} .

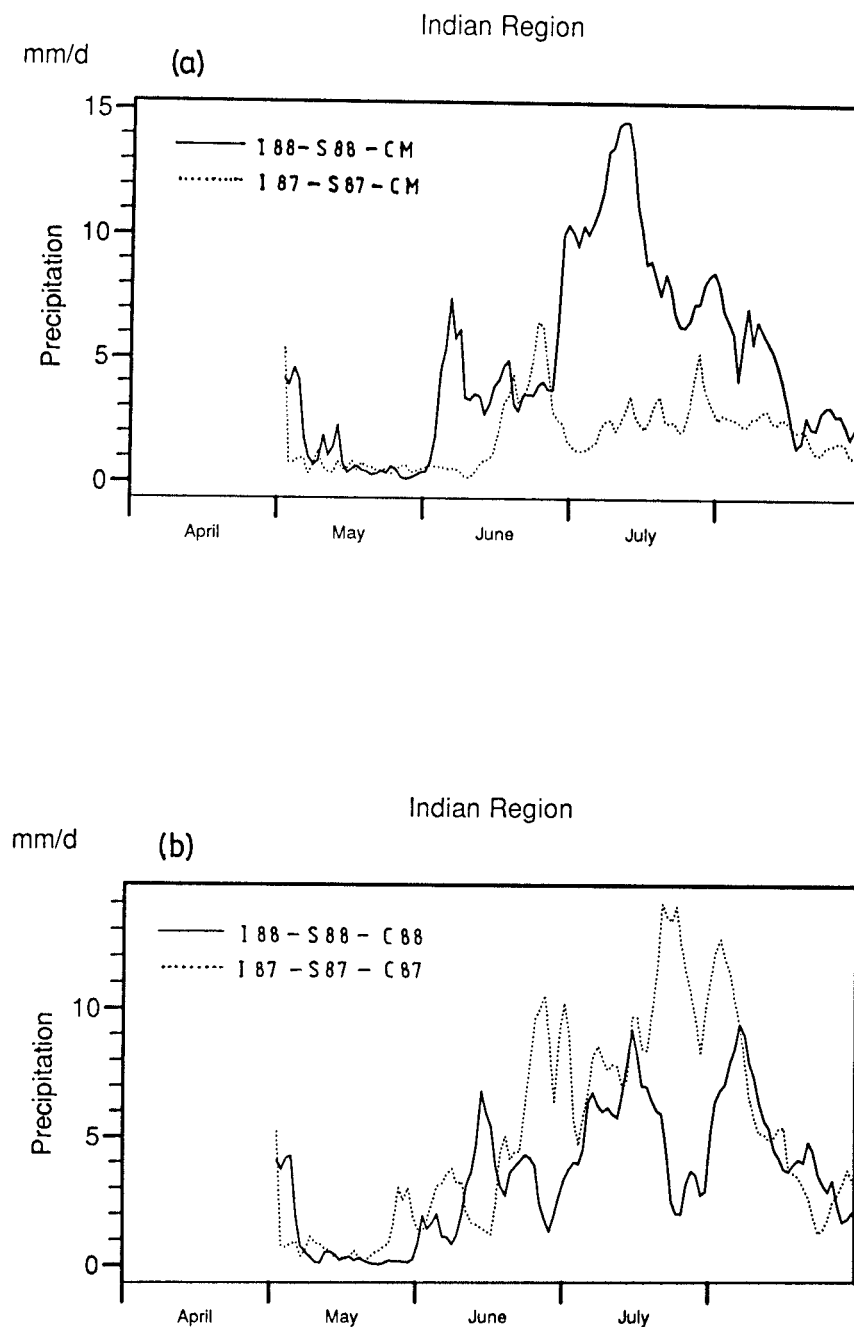


FIG. 12. Evolution of area-averaged precipitation over India: (a) I88-S88-CM, initial conditions 1 May 1988 and SST 1988 (thick line); I87-S87-CM, initial conditions 1 May 1987, and SST 1987 (dotted line). (b) I88-S88-C88: initial conditions 1 May 1988, SST 1988, and observed clouds 1988 (thick line); I87-S87-C87, initial conditions 1 May 1987, SST 1987, and observed clouds 1987 (dotted line). Units: mm day⁻¹.

ferences show some degradation in model-simulated rainfall. The reduced precipitation in the simulation I88-S88-C88 over central India in comparison to the corresponding model run for 1987 gives prominence to the variability introduced by cloud forcing. The precipitation differences are positive over northwest India and negative over central India. The negative precipitation

differences in this region may be attributed to the strong countermonsoon circulation simulated by the model in 1988 relative to the 1987 simulation. Moreover, in the equatorial Indian Ocean there are easterlies in the region east to 80°E and a westerly inflow to the west of this meridian. The structure of this lower-atmospheric flow closely resembles the combined response of a linear

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