Variability of the Indian Monsoon in the ECHAM3 Model: Sensitivity to Sea Surface Temperature, Soil Moisture, and the Stratospheric Quasi-Biennial Oscillation

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ABSTRACT

The variability of the monsoon is investigated using a set of 90-day forecasts [MONEG (Tropical Ocean Global Atmosphere Monsoon Numerical Experimentation Group) experiments] and a set of AMIP-type (Atmospheric Model Intercomparison Project) long-term simulations of the atmospheric circulation with the ECHAM3 model. The large-scale aspects of the summer monsoon circulation as represented by differences of dynamical quantities between the two extreme years 1987 and 1988 were reproduced well by the model in both kinds of experiments forced with observed sea surface temperature (SST). At the regional scale the difference of precipitation over India during summer 1987 and 1988 was well reproduced by the model in the 90-day forecasts using interannually varying SSTs; however, similarly good results were achieved in forecasts using climatological SSTs.

The long-term simulations forced with interannually varying SST at the lower boundary of the atmosphere over a period of 14 years, on the other hand, only partly reproduce the observed differences of precipitation over India between 1987 and 1988. For the ensemble mean of five simulations averaged from June to September and for the whole of India an increase from 1987 to 1988 is simulated by the model as observed but with smaller values. The difference in observed precipitation between 1987 and 1988 is of opposite sign for May to that for September. The simulations and observations agree in the manifestation of this sense of opposing variability within a monsoon season for these two years and also for other years. The simulations and observations differ most during July.

The paper concentrates on the question why the interannual variability in the long-term simulations on one hand and the 90-day forecasts and in the observations of precipitation on the other hand differ so strongly during the peak of the monsoon in July. Large-scale dynamics over India are mainly forced by the anomalies of Pacific SST. For the variability of precipitation over India other forcings than the Pacific SST are important as well. Due to enhanced evaporation, warmer SSTs over the northern Indian Ocean lead to increased precipitation over India. Changes in the SST there within the range of uncertainty (0.5 K) can lead to clear impacts.

As a further boundary forcing, the impact of soil moisture is investigated. The use of realistic soil moisture differences between 1987 and 1988 in the MONEG forecasts resulted in improved skill of precipitation forecasts over India. Also the two individual AMIP simulations with realistic precipitation differences over India had more realistic soil moisture differences over east Asia in the beginning of the monsoon season between the two years than those experiments that failed to produce the correct precipitation differences.

The years 1987 and 1988 were quite different with respect to the phase of the stratospheric quasi-biennial oscillation (QBO). As atmospheric circulation models cannot yet reproduce stratospheric QBOs realistically, their impact was tested by nudging observed QBOs into AMIP simulations for July 1987 and 1988. Seven out of eight experiments showed an impact toward a more realistic simulation of precipitation over India; however, during the west phase of the QBO (1987) impacts are very small.

None of these forcings gave a dominant effect. If this finding is confirmed by further experimentation, improvements of practical long-range forecasts may be very difficult as two of these quantities are hardly known with the required accuracy (northern Indian Ocean SSTs and the Eurasian soil moisture) and because models are not yet able to simulate the stratospheric QBO realistically.

This study confirms that El Niño has two direct effects: it reduces the precipitation over India and reduces the surface winds over the Arabian Sea. Due to the latter, the SST of the Arabian Sea can increase as there is less mixing and upwelling in the ocean. Here it is suggested that because of this increased SST there would be more precipitation over India, thus counteracting the expected decrease from the direct El Niño effect.

Sensitivity experiments were carried out with the ECHAM3 model to substantiate this hypothesis. The results may be model-dependent and model deficiencies might influence sensitivities from boundary forcings adversely. Therefore observational data have been investigated as far as possible to seek independent confirmation of the findings obtained through the model simulations.

1. Introduction

A set of model experiments was proposed by the TOGA (Tropical Ocean Global Atmosphere) Monsoon Numerical Experimentation Group (MONEG) in order to assess the skill of complex numerical models in simulating interannual variations in regional climate, in particular, the Northern Hemisphere monsoon circulations in Africa and Asia (WMO 1992). It was proposed to the modeling community to investigate the Northern Hemisphere summer monsoon seasons of 1987 and 1988. In the following we shall discuss aspects of only the Indian monsoon. In 1987

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there was a ''poor'' monsoon over India, whereas in 1988 the Indian monsoon precipitation was abundant. A large number of parameters are assumed to be valuable predictors of the strength of the Indian monsoon. The MONEG program addresses mainly the relations between the variations in the monsoon and the El Niño–Southern Oscillation (ENSO) cycle—that is, sea surface temperature (SST) anomalies in the tropical Pacific Ocean, but also relations with SST anomalies in the Indian Ocean are being investigated. Further, the MONEG group suggested to studying influences from land surface processes on the monsoon variability.

Experiments using the same initial data and the same boundary forcings were carried out by a large number of research groups worldwide (WMO 1992). The present study focuses on the performance of the Hamburg climate model (ECHAM3-T42; Roeckner et al. 1992) to SST anomalies by examining the large-scale response in terms of the wind field at 850 and 200 hPa (section 4a) and the regional response by examining the precipitation field over India (section 4b). As precipitation variability over India is represented differently in the MONEG forecast experiments and in long-term climate simulations using interannually varying SSTs as suggested by the Atmospheric Model Intercomparison Project (AMIP; Gates 1992) but extended to 14 yr, additional experiments were carried out to address the sensitivity of the monsoon variability to the Indian Ocean SST anomalies (section 5a), the sensitivity to soil moisture (section 5b), and the state of the stratospheric quasi-biennial oscillation (QBO; section 5c).

In the following we shall use the term ''simulations'' for long-term runs over 14 yr to indicate that initial data are negligible and the only forcing comes from the surface. We use the term ''forecasts'' for the 90-day or shorter runs to indicate that the initial values may play an important role as well.

The choice of the year 1987 as an example for an El Niño year is somewhat unfortunate because low precipitation amounts over India are mostly found in summers before the peak of El Niño events (growing El Niños), whereas summer 1987 represents the already decaying phase of the El Niño. The connection between the Indian summer monsoon precipitation and the growing phase of El Niño is demonstrated in Fig. 1 in which 16 yr of all-India June–September (JJAS) seasonal mean precipitation from Parthasarathy et al. (1994) are correlated with seasonal mean SST fields (Reynolds 1988) from the extended AMIP dataset before, during, and after the monsoon season. It is obvious that the precipitation variability over India is leading the El Niño–Southern Oscillation (ENSO) variability. It is a sign of data problems which in maps similar to Fig. 1 but that use the GISST2.2 SST prepared by Rayner et al. (1996) the actual values of correlation differ considerably, mainly in terms of a noisier appearance and slightly lower correlations. Also, using the data for the 1903–1994 period again results in considerable changes in the amount and patterns, especially a higher correlation for the Niño-3 area during June–August (JJA). Common to the correlation maps from different data sources is that there is weak negative correlation between the precipitation of all India for the JJAS season and the SST over the tropical central Pacific during JJA. This correlation is stronger for SSTs of the fall and winter after the monsoon season and does not exist or is positive for the SSTs preceding the summer monsoon. Among others, Webster and Yang (1992) have noticed this lead of the monsoon and discuss the question if the monsoon is the active part and ENSO the passive part in their interaction. It is, however, well known and will be shown below that the large-scale dynamical response of El Niño is reaching into the area of India. If one concluded that the ENSO is the passive part in the interaction with the monsoon, this would imply that a prediction of the variability of the monsoon over India would be very difficult.

Meehl (1994) has suggested a mechanism that could explain a biennial variability of precipitation over India in connection with ENSO. The chain of events in his argument requires that a minimum of precipitation over India is preceded by cold SSTs during December–February (DJF) over the tropical Indian Ocean and vice versa. Such a correlation is seen in Fig. 1a but the signal over the Arabian Sea is much stronger.

There is no general definition of the strength of the Indian monsoon. Therefore there are years (e.g., 1984–1986) that are called strong by a definition based on dynamical quantities (Ju and Slingo 1995) and weak by a definition based on precipitation (Kane 1997). For economical and societal impacts the precipitation amount is probably the best indicator of the strength of the monsoon; large amounts of precipitation mean mostly a good harvest and what is commonly called a good or strong monsoon. Precipitation is, however, a quantity that varies considerably in time and space and therefore area means of precipitation can be obtained only with a large margin of uncertainty. It is also difficult for models to simulate the right amount and patterns of the precipitation; therefore, in the past scientists have preferred to use a more robust (dynamical) quantity for defining the strength of the monsoon [e.g., the monsoon index defined by Webster and Yang (1992)]. There is also a strong belief that the strength of the monsoon depends on the phase of ENSO and that dynamical quantities show such a connection much better than the precipitation—a fact that might have helped the use of the dynamical quantities for defining the strength of the monsoon. We shall associate the strength of the monsoon with precipitation anomalies.

Ropelewski and Halpert (1987) showed that the interannual variability of precipitation over northwestern India is connected with ENSO events, with less precipitation during El Niño and more precipitation during La Niña. When setting up the MONEG experiments, it was therefore anticipated that forecasts forced with inter-

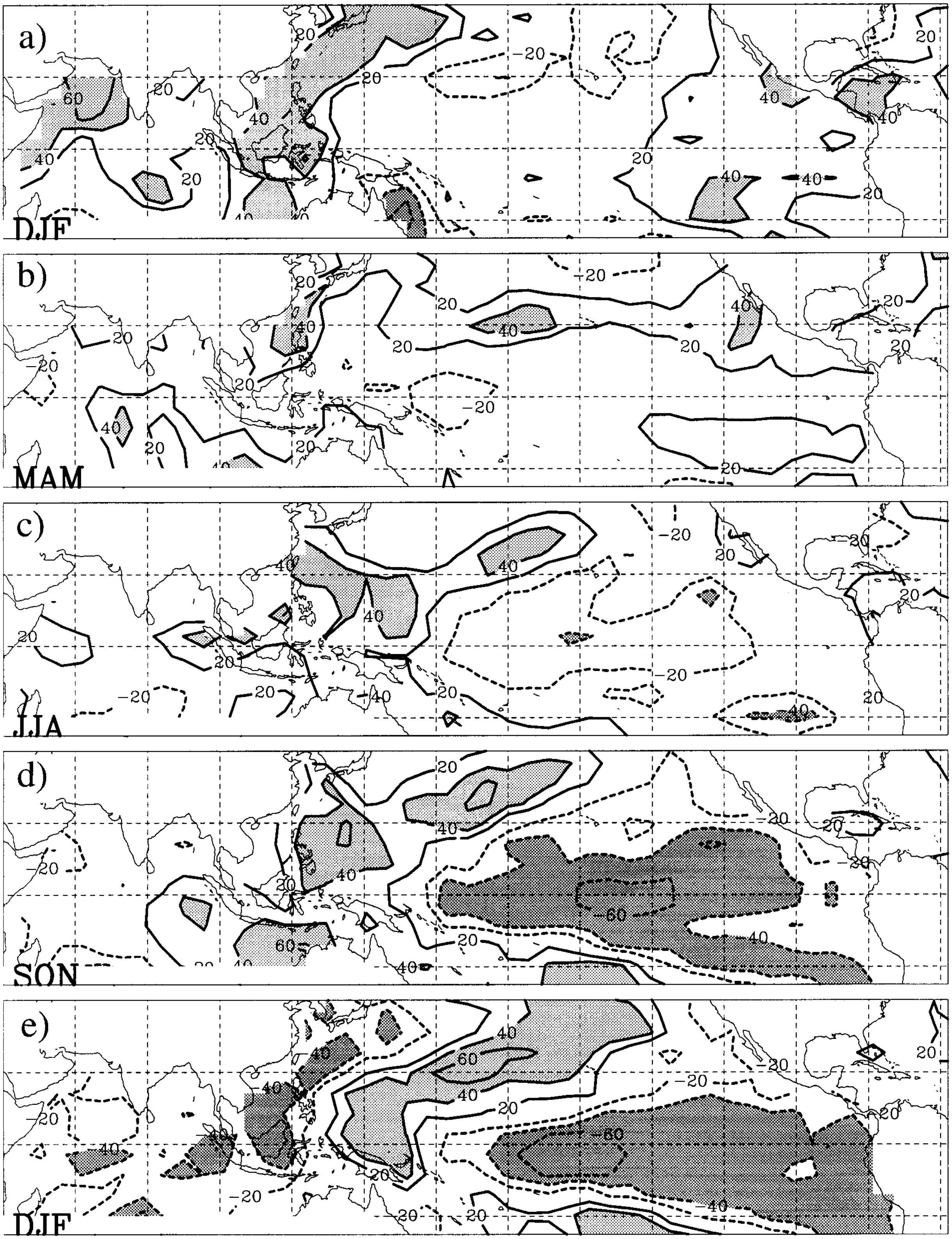

FIG. 1. Correlation between all-India precipitation for the whole summer season from Parthasarathy et al. (1994) with seasonal mean SSTs before, during, and after the summer season. Data from 1979 to 1994 have been used.

annually varying SST boundary conditions would show greater skill in representing the interannual variability of the Indian monsoon than those forced with a climatological mean SST—that is, that they would allow us to distinguish, for example, the ''good'' monsoon year of 1988 from the ''poor'' monsoon year of 1987. However, Ropelewski and Halpert (1987) did not include any event of the 1980s in their statistics.

Extensive experimentation by, for example, Palmer and Anderson (1994) showed that multiple integrations are needed to extract the correct signals from boundary forcings in the atmosphere. The real atmosphere can, however, provide only one realization and this study will help to understand what processes might disturb the ENSO signal in the precipitation over India.

Of special interest is the question to what extent the SST anomalies over the northern Indian Ocean (Arabian Sea and Bay of Bengal) are influencing the monsoon. Shukla (1975) performed simulation experiments in which he used relatively large SST anomalies and found clear impacts. Shukla and Misra (1977) established a weak positive correlation between Indian precipitation and the SST over the Arabian Sea also from observational data. Ju and Slingo (1995) found evidence that the atmosphere is forcing the SST over the Arabian Sea. For a better understanding of the ocean–atmosphere interaction, we have designed sensitivity experiments with our atmospheric model (see section 5a).

The work related to the impact of snow cover and soil moisture over Eurasia on the precipitation over India is reviewed and discussed, for example, by Barnett et al. (1989). It is suggested that enhanced snow depth and as a consequence increased soil moisture over Siberia or Tibet will lead to delayed and reduced monsoon precipitation over India. We shall address this question by discussing results from 90-day forecast experiments and by studying sensitivity experiments related to soil moisture in section 5b.

It is also well known that there is a strong biennial variability in the precipitation over India (Mukherjee et al. 1985). Meehl (1994) suggests a chain of events with interactions between the atmosphere and the ocean as well as between the atmosphere and the soil, which can explain such a variability (see section 4a.1). Gray et al. (1992) suggest a connection between the precipitation over India with the stratospheric QBO via the tropical convection. Observational studies by Mukherjee et al. (1985) found a positive correlation between the 30-hPa mean zonal wind at Balboa and the precipitation in several Indian regions of 30%–42%. We found that the observed 50-hPa zonal wind over Singapore and the observed all-India precipitation for JJAS are positively correlated by 32%. We shall discuss this further in section 5c in connection with sensitivity experiments with our atmospheric model. Further investigations were carried out by Giorgetta (1996).

When discussing the correlation between precipitation over India and the SST in Fig. 1 the problem of uncertainties in analyses was mentioned and this is a problem throughout this investigation. Whenever possible we shall validate model results against several observational datasets to indicate how much differences between model fields and analyses might be due to uncertainties of the analyses. Operational analyses and the reanalyses produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) in Reading, United Kingdom, and the National Centers for Environmental Prediction (NCEP) in Washington, D.C., are available. We use these data for comparisons with upper-air model results but also in respect to precipitation, although precipitation fields from these analyses are purely model generated. For precipitation we also use the all-India seasonal averages by Parthasarathy et al. (1994) and gridded values by Singh et al. (1992), GPCP (Rudolf et al. 1992), Hulme (1994), and Schemm et al. (1992). Precipitation amounts over the oceans used in this study are estimates using microwave observations from satellite (microwave satellite unit; Spencer 1993) and estimates using infrared observations from satellites[Geosynchronous Operational Environmental Satellite (GEOS) Precipitation Index (GPI); Janowiak and Arkin 1991].

Below we shall see that on the one hand there is a large margin of uncertainty in the absolute values of precipitation, but on the other hand all estimates have a very similar interannual variability although they were derived using quite different methods. As this study focuses mainly on the interannual variability, it seems to be less critical which of the different analyses are used for validating the model results except in section 3.

TABLE 1. Long-term simulations.

5	AMIP simulations with interannually varying SST 1979–92
1	AMIP type simulation with interannually varying SST only over the tropical Pacific 1979–92

2. Design of the experiments

In this study the ECHAM3-T42 model (Roeckner et al. 1992) is used. The monsoon variability was studied by exploiting the MONEG experiments (WMO 1992), which consist of a series of 90-day forecasts starting from ECMWF operational analyses of 1–3 June 1987 and 1988, and by exploiting five 14-yr simulations forced by global SSTs, which vary from year to year (extended AMIP integrations). The five realizations start from different Januaries of a control run. The global SSTs are referred to as the COLA/CAC AMIP SST and Sea-Ice Dataset (Reynolds 1988). SSTs from the monthly mean global fields were interpolated to the time step of the model. The climatological SST used in our experiments is the mean from 1979 to 1988 of these monthly means. Table 1 gives an overview of the long-term simulations.

All forecasts in the set of MONEG experiments are

TABLE 2. Boundary forcings in the ensembles of 90-day forecasts (MONEG), all starting on 1, 2, and 3 June 1987 and 1988.

Forcing	Conditions
var. SST everywhere	Interactive soil moisture calculation
var. SST only over the tropical Pacific	Interactive soil moisture calculation
var. SST only over the tropical Indian Ocean	Interactive soil moisture calculation
var. SST only over the tropical Atlantic	Interactive soil moisture calculation
clim. SST everywhere	Interactive soil moisture calculation
clim. SST everywhere	Climatological soil moisture
clim. SST everywhere	ECMWF analysis soil moisture

integrated over 90 days and are initialized from ECMWF analyses at T42 resolution for 1–3 June 1987 and 1988, respectively. Following the suggestions of MONEG, the experimental design is the same as employed by Palmer et al. (1992) in a study using the ECMWF model: in one set of experiments the boundary conditions are taken from climatological global SSTs and in another from global SSTs for the individual years of 1987 and 1988, which are the same as in the AMIP simulations. In addition several regional SST experiments were performed in order to study the contribution from individual ocean basins (see Table 2).

In another set of experiments the sensitivity to soil moisture is studied (section 5b). Here the soil moisture fields were globally prescribed from climatology (Mintz and Serafini 1992), from global operational ECMWF analyses or computed interactively by the model. Table 2 gives an overview of 90-day forecasts (MONEG). Further experiments are described in the following sections.

The precipitation will be displayed in units of mm month^{-1} throughout the paper, also for seasonal means, to ease comparison. Observational data have been scaled accordingly.

3. Climate of the model

The various generations of the Hamburg climate model have been known to represent the monthly averaged parameters associated with the monsoon with reasonable skill even at the low resolution (T21) employed, for example, by Barnett et al. (1989). The quality of the circulation and precipitation of the present model was documented by Roeckner et al. (1992), and the hydrological cycle of the ECHAM3 model and its dependence on the horizontal resolution is discussed by Arpe et al. (1994). Bengtsson et al. (1996) have shown that the ECHAM3 model is able to reproduce the interannual variability of the atmosphere due to forcings from SST anomalies realistically. The long-term simulations used by Bengtsson et al. (1996) are the same as those used in this study. Below, only some aspects that are relevant for the Indian summer monsoon are mentioned.

Compared to the 15-yr mean ECMWF reanalysis, the velocity of the tropical easterly jet at 200 hPa is well represented by the ECHAM3-T42 model in long-term means with an underestimation of the strength by up to 5 m s^{-1} (Fig. 2). In long-term means the Somali jet at 850 hPa is slightly weaker compared to the analysis and there is a southward displacement of the wind pattern in the Arabian Sea. In the years of 1987 and 1988 the model jet streams are weaker than is analyzed in a similar manner as in the long-term means. The simulation of too weak jet streams in the monsoon area is common to other models as well, for example, a similar weakening can be found in the ECMWF model (Sperber and Palmer 1996) and an even stronger weakening in the COLA model (Fennessy and Shukla 1992).

There is a considerable spread between the available precipitation climatologies for this region and season. In Fig. 2 (lower panels) a comparison of the 14-yr mean model simulation with a 9-yr mean of the analysis by GPCP (Rudolf et al. 1992) is presented. Roeckner et al. (1992) have shown that the T42 resolution improves the distribution of maximum and minimum precipitation in the Indian region, as compared to the lower-resolution versions. However, the total amount of precipitation (Fig. 2) is underestimated mainly due to too-weak precipitation during July. In the simulations the precipitation over India starts generally too early. Due to this the soil is wet already in May. This reduces the heating of the soil during the normal time of monsoon onset and may be the cause for a too-weak monsoon precipitation in the later months.

4. Impacts from SST anomalies in AMIP and MONEG experiments

Following the guidelines of the MONEG experiments, results will be displayed mainly in terms of the difference fields for 1988 and 1987. WMO (1992) and Krishnamurti et al. (1990) discuss the anomalies that were observed over India during these two years, the summer monsoon being classified as good in 1988 and as poor in 1987. Figure 3 shows the differences in the boundary forcing between the two years. The SST differences using the AMIP dataset (Fig. 3a) show the main patterns of an El Niño with opposite sign—that is, large negative values in the tropical central and eastern Pacific. This is accompanied by differences of opposite sign over the western and subtropical Pacific as well as by differences of the same sign but much smaller amplitudes over the Indian Ocean. A similar plot but using GISST2.2 (Global Ice and Sea Surface Temperature)

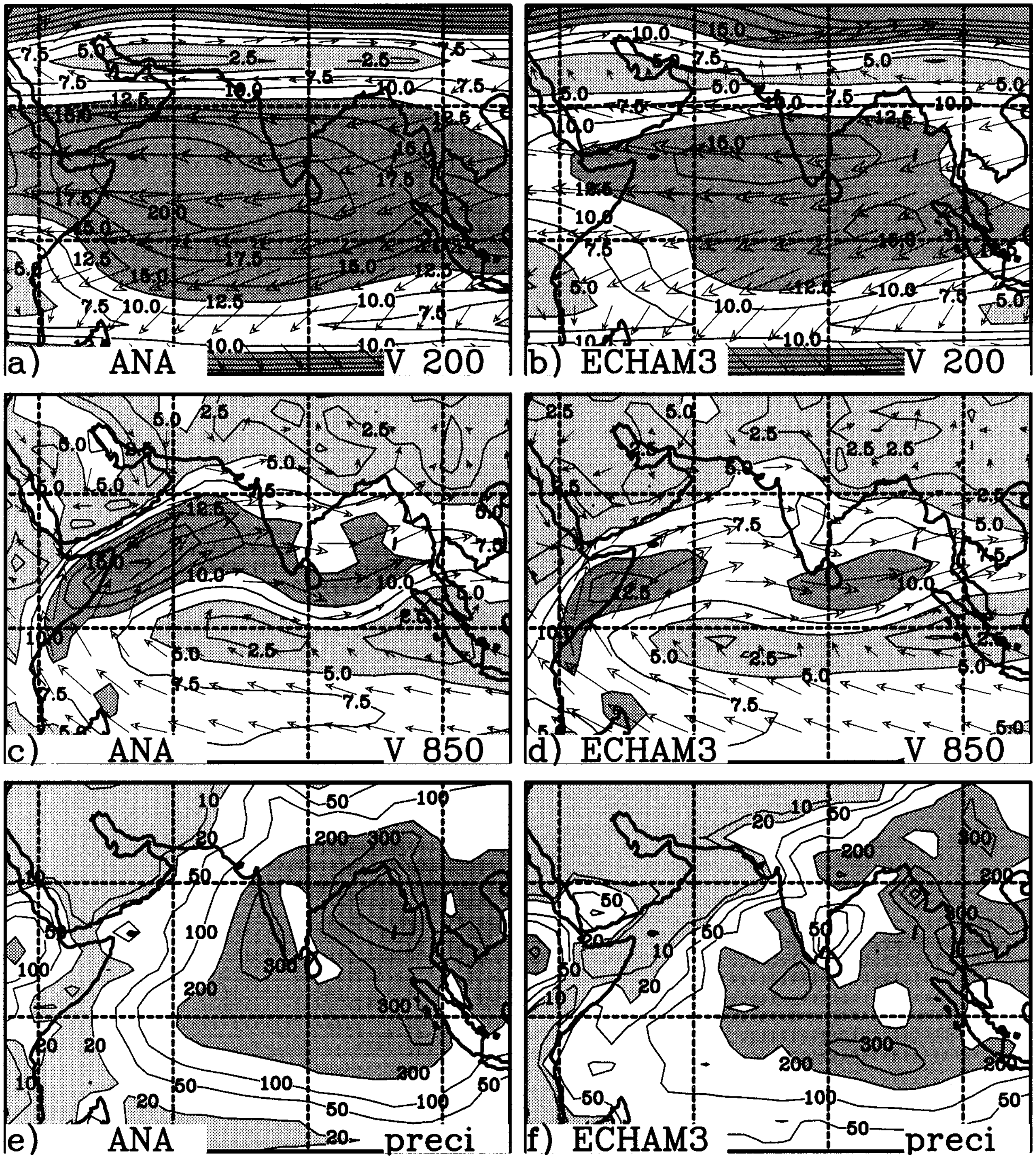

FIG. 2. Long-term mean wind at (upper panels) 200 hPa and (middle panels) 850 hPa, and (lower panels) long-term mean precipitation during JJA as (left panels) analyzed or (right panels) simulated with the ECHAM3 model. Contour interval of the wind fields is 2.5 m s^{-1}, heavy shading for winds >12.5 m s^{-1} at 200 hPa and >10 m s^{-1} at 850 hPa, light shading for winds <5 m s^{-1}. The precipitation fields have contours at 10, 20, 50, 100, 200, 300, 400, and 500 mm $month^{-1}$, heavy shading for precipitation >200 mm $month^{-1}$, and light shading for precipitation <20 mm $month^{-1}$.

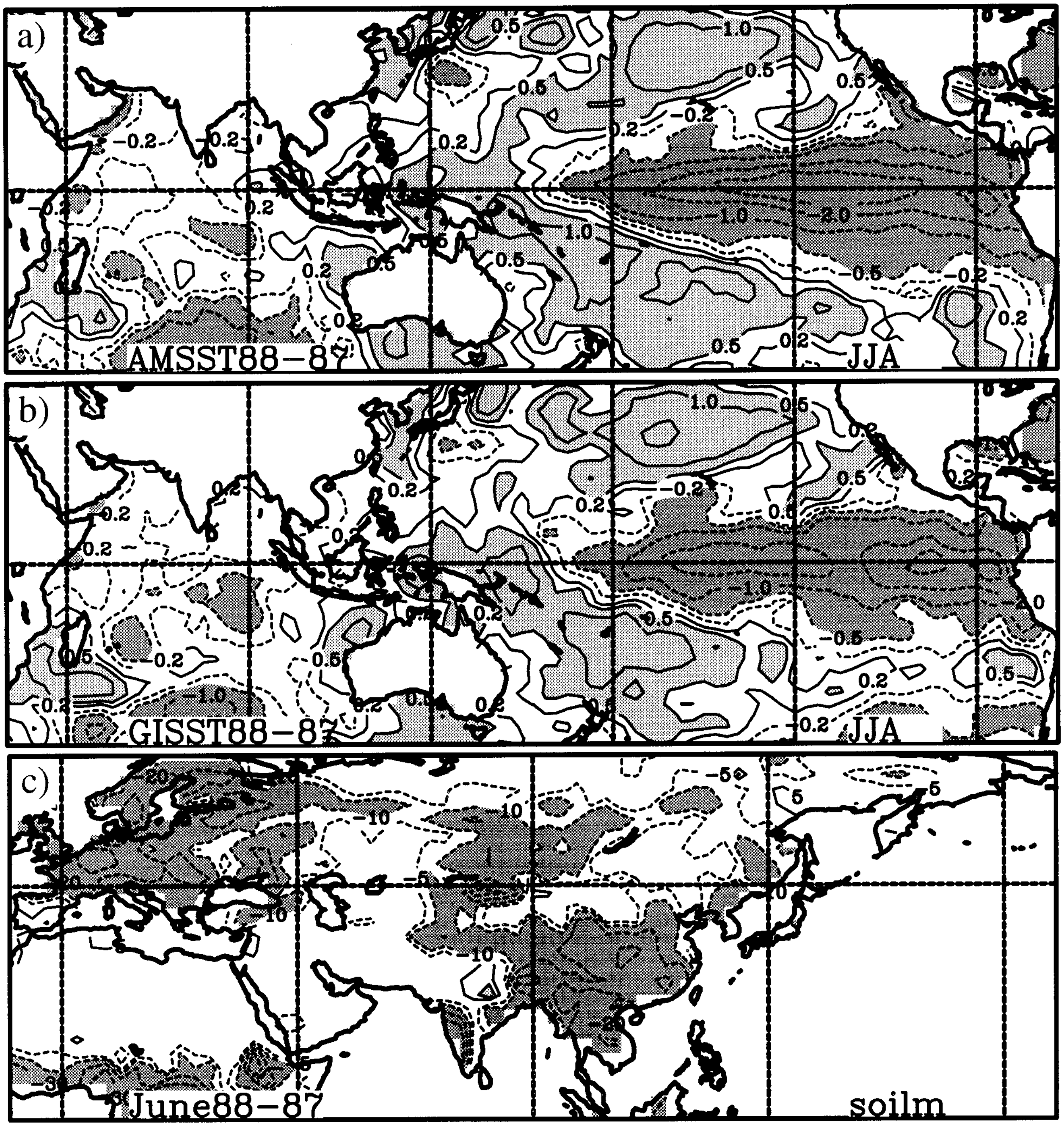

FIG. 3. JJA 1988 minus 1987 differences of SST (a) in the AMIP dataset and (b) in the GISST2.2 dataset. Contours at ±0.2, 0.5, 1, 2, 3 K, negative contours are dashed, shading for differences >0.5 K. (c) 1988 minus 1987 differences of soil moisture in the ECMWF analyses for June. Contours in relative numbers.

data in Fig. 3b (Rayner et al. 1996) shows a very similar pattern but the actual values are quite different in some areas; for example, at 135°W at the equator the AMIP dataset has a difference between 1988 and 1987 of −3.9 K, whereas the GISST2.2 has only a difference of −2.3K and at the northern Bay of Bengal (22°N, 90°E, two grid points) the difference is −0.2 K for the AMIP data and +0.4 K for the GISST data. The latter difference between the two datasets is small but we shall see below that small differences in this area may be important for the simulation of precipitation over India. In Fig. 9 it is shown that the GISST data show generally smaller SST amplitudes in the Niño-3 area (150°–90°W, 5°S–5°N) than the AMIP data.

The soil moisture differences between 1988 and 1987 are shown in Fig. 3c. There is less soil moisture for

1988 than 1987 for almost all of Eurasia. Figure 3c shows June values but JJA means are very similar. As might be expected, the soil in the forecasts is warmer in 1988 than 1987 all over Asia (not shown).

In section 4a we shall discuss the large-scale circulation aspects of the impacts of these forcings on the monsoon in terms of anomalies and differences of the velocity potential at 200 hPa and by showing the time series of the Webster–Yang index for the full integration length of the AMIP simulations. The regional aspect will be investigated in terms of the precipitation variability over India in section 4b.

a. Dynamical large-scale response

1) Simulations with interannually varying SST (extended AMIP)

The velocity potential provides an insight into the large-scale aspect of the impacts of the SST on the atmospheric circulation. In Figs. 4a and 4b the differences of the velocity potential between 1988 and 1987 are given for the ECMWF analysis (a for the reanalysis, b for the operational analysis) and in Fig. 4c for the ensemble mean of five long-term simulations with varying SST (AMIP). Taking the uncertainty of the analysis of this quantity as indicated by the difference between the reanalysis and the operational analysis (Figs. 4a and 4b) into account, we find that the simulated differences are in good agreement with the analyzed ones.

The major differences in velocity potential between the two years of 1987 and 1988 occur in the tropical Indian Ocean and central Pacific. The mean anomalies of the 200-hPa velocity potential for JJA in both years (a minimum over the central tropical Pacific and a maximum over the Indian Ocean during 1987 and vice versa for 1988) are found in each of the individual AMIP experiments (not shown). The AMIP simulations have minima of the velocity potential anomaly over the central Pacific and maxima over the Indian Ocean for all the summers of years with growing El Niño (1982, 1986, and 1991) and extremes of the opposite sign during all La Niña years (1984, 1988, and 1989). El Niño and La Niña years are simply defined here by large positive or negative SST anomalies over the Niño-3 area. The range of variation within the ensemble of experiments is very low compared to the interannual variability (not shown). This is a clear indication of the strong impact of the SST on the tropical velocity potential.

A time series of the Webster–Yang index (U_{850hPa}–U_{200hPa}, 40°–110°E, 0°–20°N) for JJA means in Fig. 5 shows that all individual AMIP experiments show the strong increase from 1987 to 1988 similar to the one in the ECMWF reanalysis (ERA), except that the ERA values are considerably larger throughout, this is due to the stronger winds at 850 and 200 hPa as shown in section 3. Two out of three years with decaying El Niño, 1983 and 1987, have clear minima. This index indicates that the atmospheric circulation over India is dominated by the Pacific SST, as the run using interannually varying SST only over the tropical Pacific (dotted line, Pac) is very similar to the ones using interannually varying SST over the whole globe. For 1992, a year with decaying El Niño, one finds a minimum only in the ERA data and in the Pacific run, whereas the AMIP runs have a minimum already one year earlier. The values for only July are very similar to the ones shown in Fig. 5 for JJA and are therefore not shown.

Table 3 gives an overview of the Webster–Yang index during July in the individual experiments and analyses. There is a large variability between the individual AMIP experiments but nevertheless the values for 1988 are clearly larger than those for 1987. The AMIP-type simulations using varying SST only for the tropical Pacific give a similarly large difference. This indicates that it is mainly the tropical Pacific SST that is forcing this measure of monsoon strength. Table 3 contains for comparison also values from the reanalyses by ECMWF and by NCEP. The difference between both years is in both analyses nearly the same but the absolute values differ considerably indicating the large uncertainty in the analyses in this part of the world.

2) 90-day forecasts

The difference between 1988 and 1987 in the velocity potential at 200 hPa in the forecast experiments is displayed in Figs. 4e–h. The velocity potential difference in the forecasts with interannually varying SSTs everywhere (Fig. 4e) is simulated in good agreement with the analyses in a similar way as discussed for the AMIP simulations above. The model response given by the ECHAM3-T42 model is consistent with the model response shown in Fennessy and Shukla (1992) for the COLA model, both in magnitude and position. In the forecasts using global climatological mean SSTs, the differences between the two years are much smaller in magnitude and also of smaller horizontal extent (Fig. 4f).

In addition, the contributions to the monsoon variability originating from the SSTs of individual ocean basins are examined. In 1988 and 1987 the situation is such that the influence from the Pacific Ocean SSTs is largest (Fig. 4g). The difference in the velocity potential is slightly larger in the case with interannually varying SSTs only over the Pacific than in the one with globally varying SSTs. This is due to a counteracting influence from the Indian Ocean SST, which causes positive velocity potential anomalies to the west of the Indian subcontinent (Fig. 4h). Also the AMIP simulation using only Pacific SST anomalies (Fig. 4d) gives a slightly stronger response than the run with global SSTs (Fig. 4c). The influence of SST anomalies from the Atlantic Ocean and the extratropics is negligible (not shown).

Table 3 (Webster–Yang index) shows a similar picture—that is, strongest impacts from the forecasts with

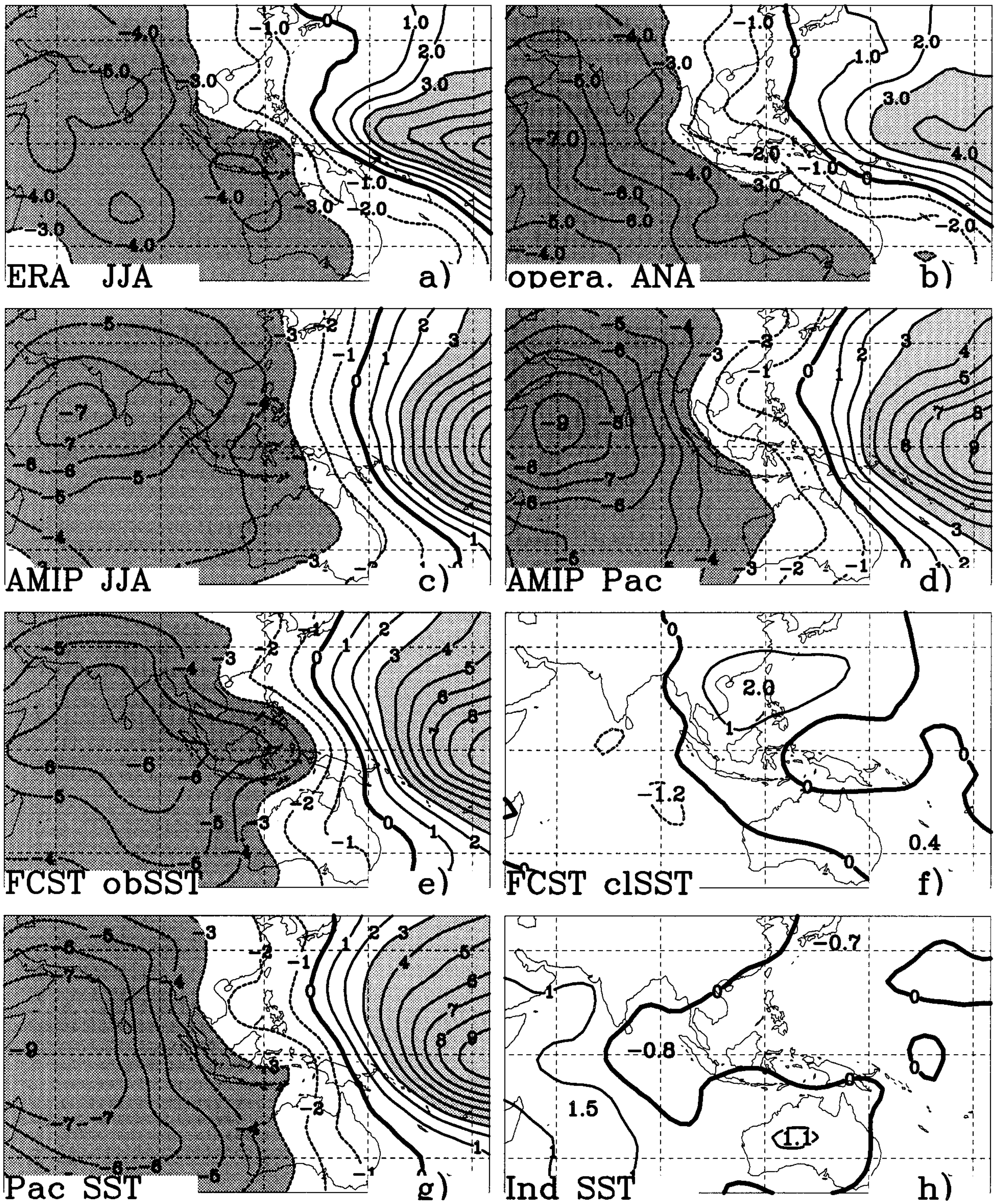

FIG. 4. Differences of velocity potential at 200 hPa during JJA, 1988 minus 1987. Contour interval: 10^{-6} m^2 s^{-1}, negative contours are dashed, shading for absolute values >3. 10^{-6} m^2 s^{-1}: (a) ECMWF reanalysis, (b) ECMWF operational analysis, (c) AMIP simulation, (d) AMIP-type simulation but interannually varying SSTs only over the Pacific, (e) forecasts using interannually varying SSTs, (f) forecasts using climatological SSTs, (g) forecasts using interannually varying SSTs only over the Pacific, and (h) forecasts using interannually varying SSTs only over the Indian Ocean. All forecasts and simulations are using interactive soil moisture calculations.

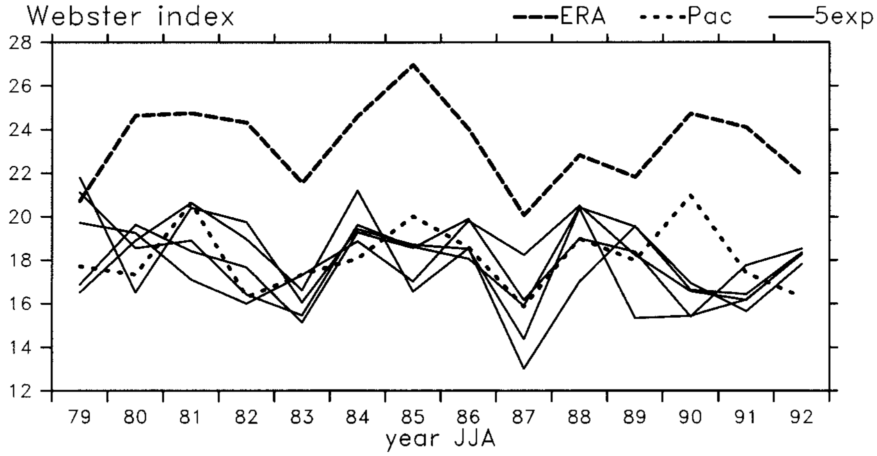

FIG. 5. Webster–Yang index for JJA 1979–92. AMIP simulation (5exp) are compared with the simulation with interannually varying SSTs only over the Pacific (Pac) and with ECMWF reanalysis (ERA).

interannually varying SST only over the tropical Pacific and almost an opposite difference between the two years for the runs with interannually varying SST over the Indian Ocean only. The forecasts using global climatological SST show slightly larger values for 1988 than for 1987 indicating that the Webster–Yang index is influenced not only by the Pacific SST but also by the initial atmospheric data in the forecasts, which may contain already the influence from the Pacific SST.

b. Precipitation (regional aspects)

1) SIMULATIONS WITH INTERANNUALLY VARYING SST (EXTENDED AMIP)

In Fig. 6 the differences of precipitation during JJA between 1988 and 1987 are shown for an area covering India and the western Pacific. For comparison the GPCP precipitation analysis and the estimates from ERA are shown in Figs. 6a and 6b. In the AMIP simulations (Fig. 6c) many observed aspects of the variability of precipitation, particularly over the tropical Pacific, are reproduced. In the Pacific region the impact from the SST anomalies is very strong directly over the SST anomalies. For India, however, the AMIP simulations reproduce the observed precipitation differences between 1987 and 1988 only partly, although the large-scale response discussed above (velocity potential and Webster–Yang index) is represented realistically. This agrees with findings by Ju and Slingo (1995) in their Fig. 21 that the variability of the Webster–Yang index hardly agrees with the variability of precipitation over India. It is interesting to note that the AMIP run using only Pacific SST (Fig. 6d) shows a more realistic response than the run using interannually varying SST everywhere, but we do not know if this is a significant result as two of the five AMIP simulations also show the observed response.

In Fig. 7 the interannual variability of precipitation in the individual extended AMIP simulations and in the observations is presented. In Fig. 7a averages of precipitation for the tropical Pacific region, known as Niño-3 are shown as seasonal means from March–May (MAM) 1979 to September–November (SON) 1992. It includes the five realizations of our extended AMIP simulations and two estimates of the truth. The estimates from MSU observations (Spencer 1993) show clearly higher precipitation amounts than the estimate from satellite observations of cloud-top temperatures (GPI; Janowiak and Arkin 1991) or the model simulations, but the interannual variability is very similar in all curves. The model reproduces the true variability of precipitation in the Niño-3 region quite realistically while the spread between the different integrations is small. The

TABLE 3. Webster–Yang index during July.

Exp	1987	1988
AMIP mean	17.7	22.7
AMIP1	21.0	24.3
AMIP2	18.2	24.4
AMIP3	14.0	18.9
AMIP4	20.0	23.0
AMIP5	15.1	23.1
Pacific–SST run	16.8	21.7
ECMWF	22.8	25.2
NCEP	26.0	28.1
MONEG obsSST1	21.1	25.4
MONEG obsSST2	17.3	23.3
MONEG obsSST3	21.9	21.7
MONEG cliSST1	21.1	21.3
MONEG cliSST2	21.1	22.7
MONEG cliSST3	20.2	23.3
MONEG PacSST1	17.8	25.4
MONEG PacSST2	21.3	22.7
MONEG PacSST3	17.8	23.8
MONEG IndSST1	22.0	22.4
MONEG IndSST2	25.4	22.7
MONEG IndSST3	23.7	23.9

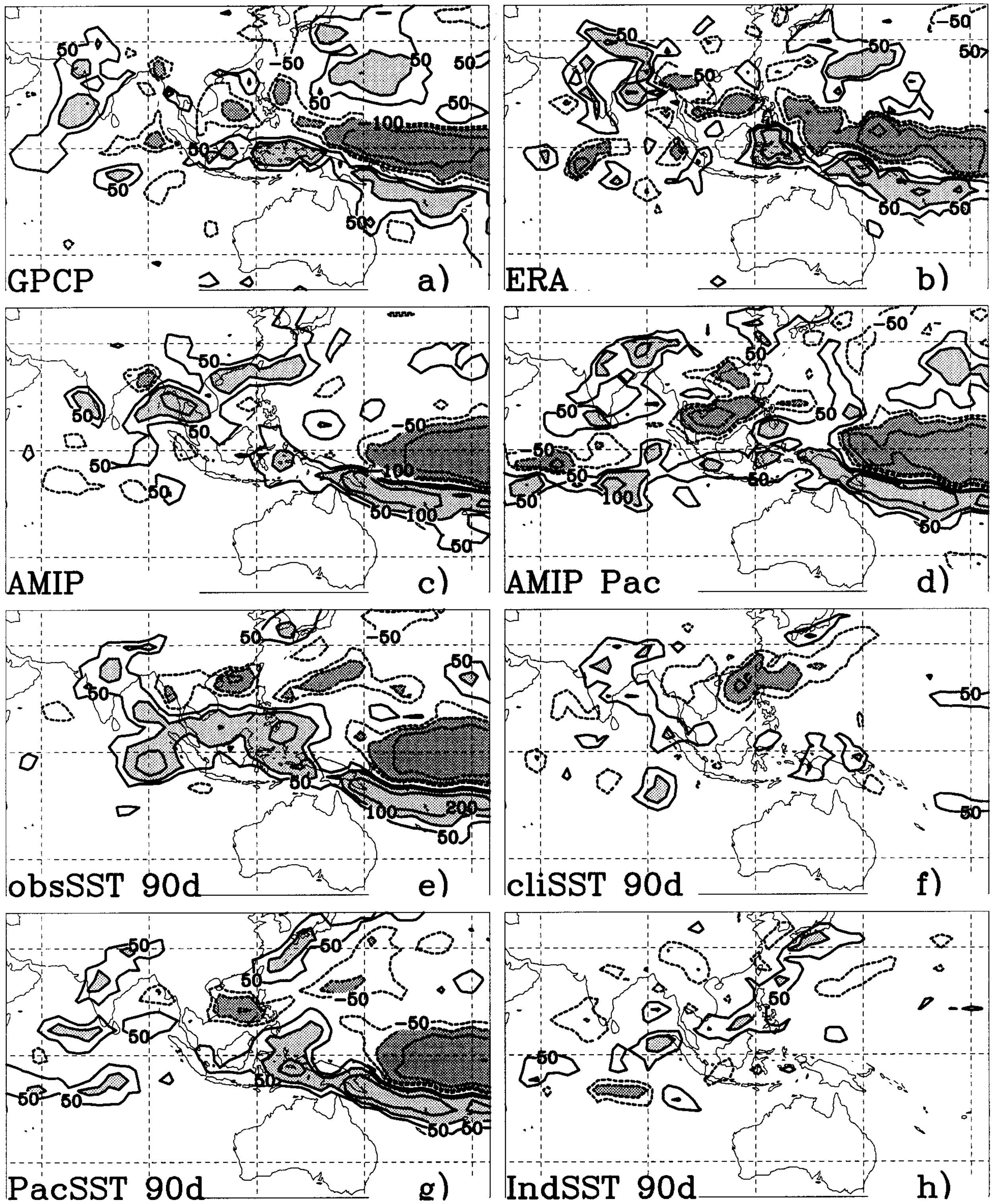

FIG. 6. JJA 1988 minus 1987 differences of precipitation. Contours at ±50, 100, 200, 500 mm month⁻¹. Shading for absolute values >50 mm month⁻¹: (a) GPCP analysis (Rudolf et al. 1992); (b) ECMWF reanalysis; (c)–(h) as in Fig. 4.

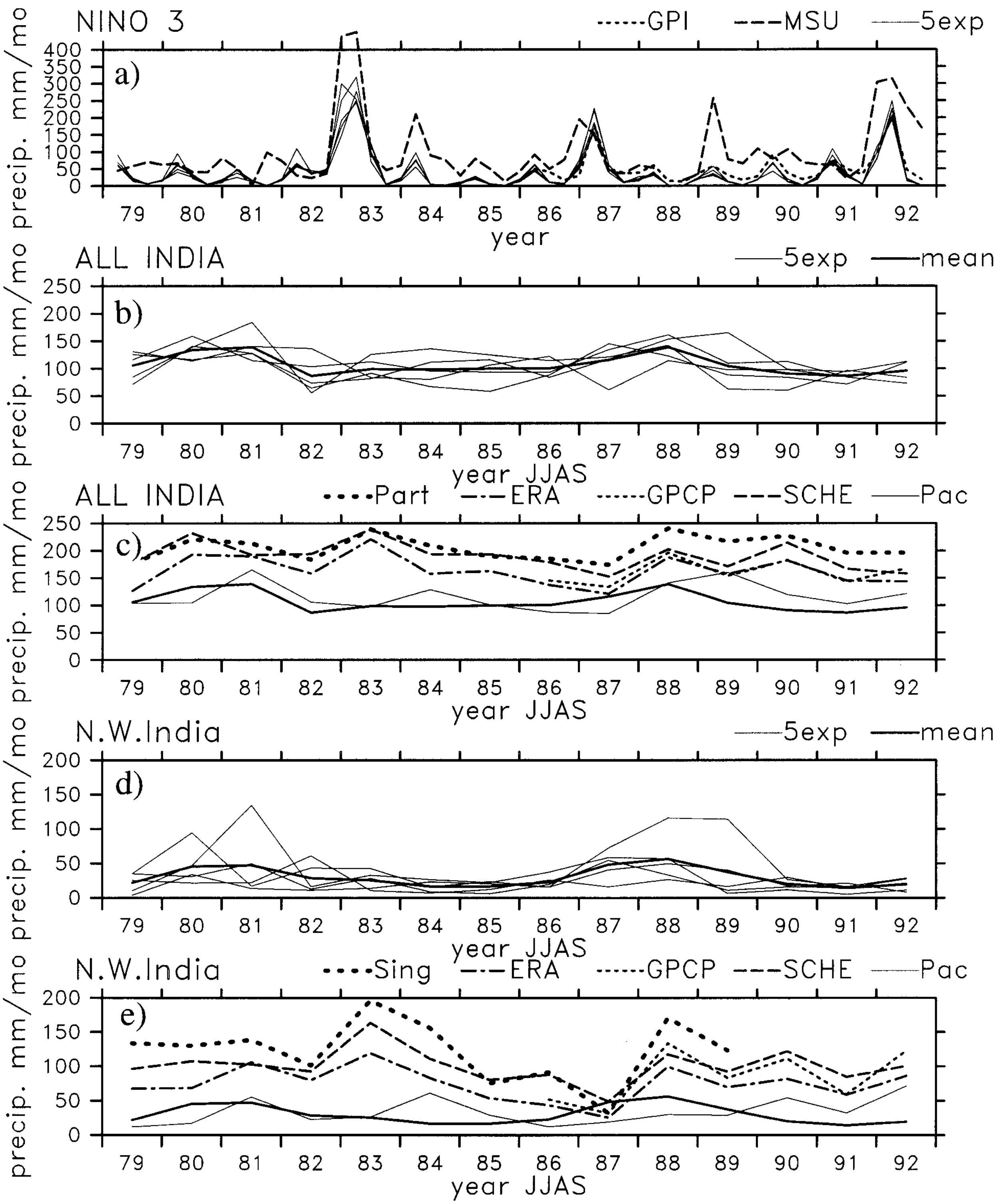

FIG. 7. Interannual variability of precipitation for area averages: Niño-3 (150°–90°W, 5°S–5°N), all India, and northwestern India from 1979 to 1992. Seasonal means in the top panel (MAM 1979–SON 1992) and JJAS means in the lower panels. Thin lines: individual AMIP simulations (5 exp) or simulation with interannually varying SSTs only over the Pacific (Pac); heavy solid line: ensemble means of AMIP simulations; dashed lines: estimates by Spencer (1993) for oceans and by Schemm et al. (1992) for land; light dotted lines: estimates by GPI for oceans and by GPCP for land; heavy dotted lines: observational data by Parthasarathy et al. (1994) or Singh et al. (1992).

exact match between the simulations and the GPI estimates is remarkable.

Figures 7b and 7c show a similar plot for the region ''all India'' as defined by Parthasarathy et al. (1994) for JJAS means. Here the five experiments show much less consistency than for the Niño-3 area. All experiments except one show an increase from 1987 to 1988. The precipitation is generally low in the analyses as well as in the simulation during growing (1982, 1986, 1991) and decaying (1983, 1987, 1992) El Niño years except in 1983 in the analyses. The simulation using interannually varying SSTs over the Pacific only (thin solid line in Fig. 7c, Pac) shows the best representation of low precipitation during El Niño years. It also shows a strong increase from 1987 to 1988.

Figure 6 indicates that the strongest difference between 1987 and 1988 in the analyses as well as most forecasts is in northwestern India and therefore we shall concentrate on this area in the following. Northwestern India is defined here as the Indian state bounded by 20°–30°N, 69°–75°E, which corresponds to the boxes 3, 6, 7, 16, 17, 26, 27 of Singh et al. (1992). It covers five grid points in a T42 model. The AMIP simulations disagree very strongly with the analyses in this area (Figs. 7d and 7e). All the datasets representing observations have a dramatic increase of JJA precipitation from 1987 to 1988, in the data of Singh et al. (1992) from 25 to 170 mm month^{-1}, whereas only a modest increase of about 10% is found in the simulations.

The interannual variability of precipitation for northwestern India is shown in Fig. 8 for the single months of May, July, and September instead of JJAS means, which were presented in Figs. 7d and 7e. In May the five individual extended AMIP simulations show a consistent variability with maximum precipitation in 1982–83, 1987, and 1990–91, though less consistent and with smaller amplitudes for 1990–91. The analyses show maxima in the same years but the maxima are weaker in the analyses especially in the reanalysis data by ECMWF than in the model simulations. The 1987–88 change in all curves is opposite to that observed for the whole summer monsoon season.

During September (Fig. 8c) the extended AMIP simulations show quite a large variability among themselves. A consistent feature is a steep increase from 1986–87 to 1988 as observed (one simulation has only a weak increase). The steep increase from 1987 to 1988 is found in the observations already during July, whereas this feature emerges only during August (not shown) in the simulations. For September a variability of the precipitation can be seen that is connected with ENSO, that is, low precipitation during years with growing El Niños, clearer in the observations than in the simulations.

The precipitation of the five AMIP experiments during July (Fig. 8b) deviate most strongly from each other. A very weak consistency leads to a variability in the ensemble mean (not shown) that has relatively high precipitation during 1980, 1981, 1986, 1987, and 1988 and low values during 1979, 1983, 1984, 1985, 1990, 1991, and 1992—that is, a variability that is not connected directly to ENSO events. This applies for northwestern India as well as for all India (not shown). Composites of SST anomalies of those two sets of years (not shown) reveal that the precipitation correlates positively with the SSTs over the northern Indian Ocean. Using the mean of the five AMIP simulations over 14 yr for July, we find a correlation of SST anomalies over the Indian Ocean north of 22°N with precipitation for all India of 45% and for northwestern India of 61%, whereas it is −30% and −3%, respectively, for the correlation between the precipitation in those two Indian areas and the Niño-3 SST.

Note that the run with varying SST only over the Pacific in Fig. 8 has a variability very similar to the one of the AMIP simulations for May and September, but it is quite different for July. In July the Pacific run has a variability that is significantly different from the AMIP simulations and shows relative minima in years with growing or decaying El Niños except in 1992. This suggests that there are other SST anomalies than those over the tropical Pacific that influence the precipitation over India for this month.

For the observed precipitation over India the correlations with the northern Indian Ocean SST are much lower and with the Niño-3 SST much higher than in our AMIP runs. On the one hand there may be an oversensitivity to SST anomalies over the northern Indian Ocean in the ECHAM3 model, but on the other hand we have seen above in the discussion of Fig. 3 that there is an uncertainty in the SST analyses. Widespread differences between the SST analyses by AMIP or GISST of 1 K in seasonal means are numerous. In some areas the differences of the SSTs produced by the two analysis methods are of the same magnitude as their interannual variability—for example, over the northern Indian Ocean. The difference between AMIP and GISST values for the Arabian Sea amounts to 0.5 K in several months (see Fig. 9d) although both analyses are derived from virtually the same observational data. We do not know if the real atmosphere was forced by an SST similar to the analysis by AMIP or by GISST or by an SST that differs even more because both analyses may have accepted erroneous observational data. Therefore differences in the variability of the precipitation between the real atmosphere and the simulations may partly be due to uncertainties in the analyzed SSTs.

The correlation of the observed July all-India precipitation data by Singh et al. (1992) for the period given in Fig. 8b with the Niño-3 SSTs yields an impressive value of −62%. The choice of a different observational precipitation dataset or a different time period generally lead to lower values (Table 4). The precipitation data by Singh et al. (1992) are probably superior to the ones by Hulme (1994) because of the higher resolution and because more observational data have been incorporat-

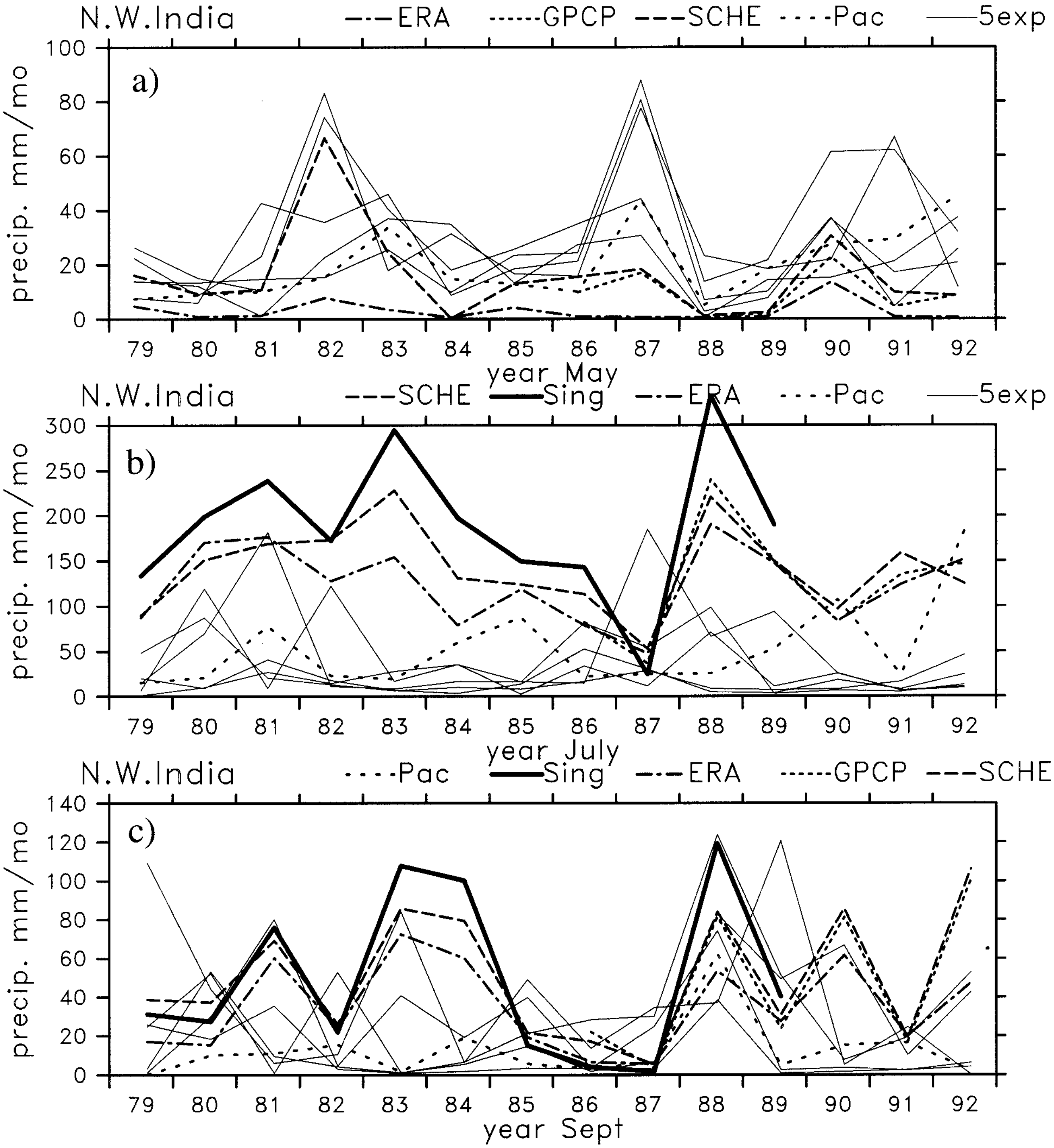

FIG. 8. Variability of precipitation for northwestern India from 1979 to 1992 for three selected months: (top panel) Monthly means of May, (middle) July, (bottom) September. Thin lines: individual AMIP simulations; light dotted lines: simulation with interannually varying SSTs only over the Pacific (Pac); dashed lines: estimates by Schemm et al. (1992); dense dotted lines: GPCP; heavy solid line: observational data by Singh et al. (1992); dash–dotted line: ERA.

ed, but the Hulme data cover a longer time span. It is obvious from the values in Table 4 that the highest correlation values result from the 1987–88 event, which was shown above to be untypical for the monsoon–ENSO relation. Without this event the simulations and observations obtain similar correlation coefficients. The variability for June and August is not shown in Fig. 8 for brevity and because they have a variability that reflects the transition between the 3 months shown.

In Fig. 7c one can see a strong component of a quasi-biennial oscillation in the analyses. Meehl (1994) has suggested a mechanism that could explain such a vari-

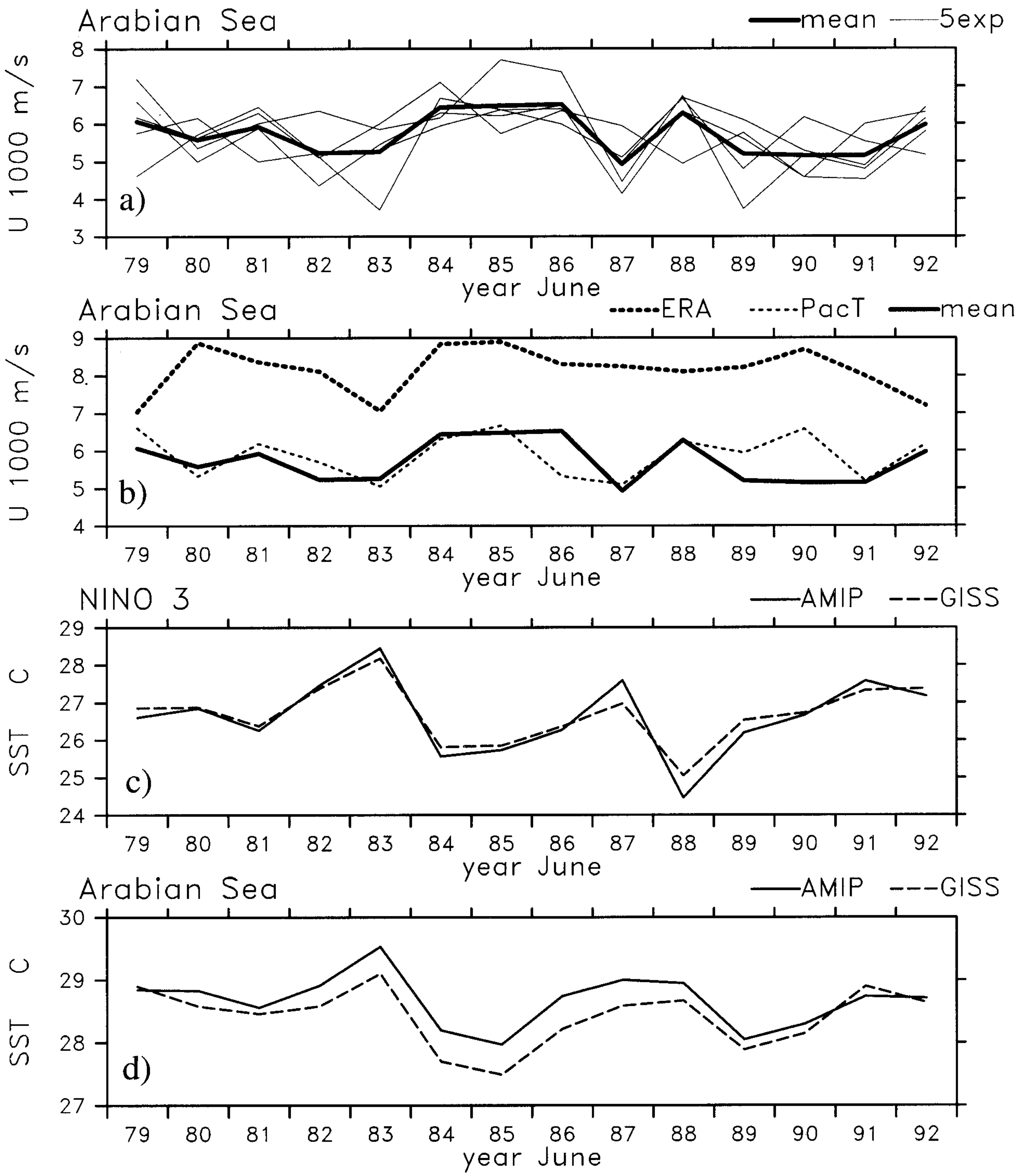

FIG. 9. Interannual variability for June [(a) and (b)] of zonal 1000-hPa wind over the Arabian Sea (45°–77°E, 5°–20°N), (c) of SST over the Niño-3 area (150°–90°W, 5°S–5°N), and (d) of SST over the Arabian Sea. In (a) and (b) thin lines: five individual AMIP experiments; heavy line: their ensemble mean; light dotted line: simulations using interannually varying SST over the tropical Pacific; heavy dotted line: ECMWF reanalysis (ERA). (c) and (d) SSTs from the AMIP dataset are compared with GISST2.2 analysis.

TABLE 4. Correlation between all-India and northwestern Indian precipitation in July and the SST in the Niño-3 area at the same month using different precipitation datasets (Hulme 1994; Singh et al. 1992; and results from the present simulations) and different time periods.

Dataset	All India	Northwestern India
Singh 1979–89	−62%	−46%
Singh 1971–89	−51%	−39%
Singh 1971–86	−25%	−3%
Hulme 1979–89	−51%	−6%
Hulme 1971–89	−41%	−3%
Hulme 1971–86	−25%	+27%
Hulme 1913–92	−22%	−13%
mean of simulations 1979–92	−30%	−3%

ability. The chain of events in his argumentation requires that reduced JJA precipitation over India is preceded by cold SSTs during the preceding DJF over the tropical Indian Ocean and vice versa. In the SSTs of AMIP as well as GISST2.2 (not shown but an indication of this is given in Fig. 1a), we find such precursors for the precipitation. There is a positive correlation of 51% between the DJF SST over the tropical Indian Ocean (50°–95°E, 10°S–10°N) and the all-India precipitation JJAS means by Parthasarathy et al. (1994) when using the data for 1980–92. The correlation is much lower (38%) when extending the period by 2 yr to 1994 and it is even lower for the period 1929–83 (15%). The latter may be due to a long-term trend in the GISST2.2 data. Using the other estimates of precipitation of Fig. 7c, the correlations are within the range of 15%–51%, for example, 27% for the Schemm et al. (1992) dataset for the period 1979–92.

Despite the uncertainties just mentioned, one finds clearly positive correlations between the DJF SST over the tropical Indian Ocean and the all-India precipitation of the following summer when using observational data. In the AMIP simulations this correlation is positive at 38% for the 1980–92 period. Therefore the mechanism suggested by Meehl (1994) may be present also in the AMIP simulations. Figure 1a suggests that the SST of the Arabian Sea may be more important than the SST from the tropical Indian Ocean. Meehl (1994) suggested further that increased summer monsoon precipitation over India is connected with cooler SSTs during the same season. This connection is not supported by Fig. 1. Such a connection can, however, be expected, if one measures the strength of the monsoon with dynamical quantities because a strong monsoon circulation enhances upwelling and mixing and as a consequence cools the Arabian Sea. This process and the relationship with SST anomalies over the Pacific will be discussed next.

Heckley and Gill (1984) have shown with a simple model that a heating source at the equator would induce a surface wind circulation with westerlies in the Tropics to the west of the heat source and easterlies to the east. During El Niño the main equatorial heat source that is placed over Indonesia is weakened and this reduces the westerlies over the Indian Ocean. Ju and Slingo (1995) have demonstrated this weakening of the westerlies over the Arabian Sea in composite maps in analyses as well as model simulations during El Niño years. In Fig. 9a the 1000-hPa zonal wind averaged for the Arabian Sea (50°–70°E, 5°–20°N, sea only) is shown for all months of June in the individual extended AMIP simulations as well as their means. There is a consistent variability with low values during El Niño events especially in their decaying phase. In Fig. 9b we see that this agrees reasonably well with the analysis by ECMWF though there is a clear bias as already discussed in section 3. During El Niño events the zonal wind is low and this is expected to lead to warmer SSTs over the Arabian Sea due to reduced upwelling and mixing (Weare 1979). The positive correlation between the SSTs of the Arabian Sea and the Niño-3 area and the negative correlations between the two SST time series and the zonal wind over the Arabian Sea are documented in Fig. 9. However, the correlation is not perfect; for example, 1987 and 1988 show a positive SST anomaly over the Arabian Sea while the wind anomalies and the ENSO phase are opposite in the two years. This may be due to uncertainties of the SST analysis discussed above and is shown in Figs. 9c and 9d where the SST from the AMIP dataset as well as from the GISST2.2 dataset are displayed. The uncertainties in the SSTs of the Arabian Sea becomes obvious also by referring to the statement of Krishnamurti et al. (1990) that the SSTs in the Arabian Sea in 1987 were below normal, while the presently available datasets show the opposite. However, we believe that it is unlikely that both analysis schemes have accepted erroneous data that could have led to the unexpectedly warm temperatures in 1988.

In Fig. 9b we have also included the experiment that had prescribed interannually varying SST only for the tropical Pacific while using climatological SST everywhere else (PacT, dotted line). It shows the same variability as the simulations with interannually varying SST everywhere; thus confirming that it is mainly the tropical Pacific SST anomalies that are causing the wind anomalies over the Arabian Sea, which then might have influenced the SST anomalies over the Arabian Sea.

In Fig. 9 the June values are shown because this month gives the clearest signal. The May values give an equally good signal. This is in agreement with findings by Ju and Slingo (1995). During later months the signal disappears probably because other factors than the ENSO become more important for the monsoon circulation.

2) 90-DAY FORECASTS OF PRECIPITATION

Considering only the large-scale aspects of the forecast differences of the precipitation between 1988 and 1987, the 90-day forecasts do indeed show the antici-

TABLE 5. July precipitation for all India, values in mm month^{-1}. Forecast experiments with different prescribed SST, all using interactive soil moisture calculations. The five validation datasets come from GPCP, Schemm et al. (1992), Singh et al. (1992), NCEP, and ECMWF reanalyses. Three forecast values refer to forecasts starting on 1, 2, and 3 June.

	Validation					Obs SST			Clim. SST			Pac SST			Ind SST		
1987:	149	173	238	148	141	90	9	24	125	38	62	65	12	32	108	84	116
1988:	266	271	401	258	267	139	110	75	151	230	146	154	66	173	72	104	106

pated increase in skill due to the use of interannually varying SST (Figs. 6e and 6f) as found in section 4a.2 for dynamical quantities. Over the Pacific the forecasts using interannually varying SSTs globally but also the forecasts using interannually varying SSTs only over the tropical Pacific (Figs. 6e and 6g) show differences between 1988 and 1987 that are very similar to the observed ones and to the ones obtained from the AMIP simulations. The main features over the Pacific and Indonesia are caused mainly by the Pacific Ocean anomaly; the other ocean basins, including the Indian Ocean, have only a marginal influence on the precipitation difference over the Pacific (Fig. 6f). Over the Indian subcontinent, however, three forecast sets (with climatological varying SST, interannually varying SST everywhere, and interannually varying SST only over the tropical Pacific, respectively) show the same realistic response in the JJA means, although they differ considerably in the dynamical response.

The differences in the precipitation forecasts between the two years are largest in July. The values are given in Table 5 for an average of all India. Five values for the observations for the two years indicate the range of uncertainty of the true amounts of precipitation. In spite of this uncertainty the signal of almost twice the amount of precipitation in July 1988 compared to July 1987 is very clear. Also for the forecasts three values are given because the ensemble consists of three forecasts starting from 1, 2, and 3 June. They show a considerable variability due to uncertainties in the initial data from the analysis that are amplified by nonlinear interactions during the forecasts.

The ensemble means of all forecast experiment groups except the one with interannually varying SST only over the Indian Ocean reproduce the same signal as the observations—that is, more precipitation in 1988 than 1987. Individual values for 1987 in these groups hardly exceed individual values for 1988 in the same group. The strongest signal is found in the integration using interannually varying SST only over the tropical Pacific, but the forecasts forced with climatological SST reproduce a signal of similar strength. Also the patterns of the means for JJA in Figs. 6f and 6g in the area of India are quite similar. The experiment with interannually varying SST for the whole globe has a smaller signal. From this result together with the slightly larger mean precipitation for 1987 than in 1988 in the forecasts with interannually varying SST only over the Indian Ocean, one can deduce that the SSTs over the Indian Ocean are counteracting the observed precipitation differences between both years.

In Table 6 for the area of northwestern India the differences between 1988 and 1987 are larger than for all India. This applies for the analyses as well as for the forecasts although the forecasts have a large bias (see also Fig. 2). The largest differences are found in the forecasts with climatological SST and with variable SST only over the Pacific. However, the spread from case to case is very large and the results may therefore be statistically not significant to interpret the impacts from different oceans.

3) CONCLUSIONS WITH RESPECT TO PRECIPITATION IN THE SIMULATIONS AND 90-DAY FORECASTS

From the better performance of the forecasts and the simulations with interannually varying SST over the Pacific only compared to the runs with interannually varying SST everywhere, we can deduce that the Pacific SST has a strong impact on the precipitation over India, but this is counteracted by SST anomalies at other places. Especially the SST anomalies of the Indian Ocean may counteract the impacts of the Pacific, as substantiated by the forecast experiments. In section 5a the impact of the SST over the northern Indian Ocean will be further investigated by sensitivity experiments.

From the good performance of the 90-day forecasts with climatological SST we can deduce that it is only partly the Pacific SST that leads to the precipitation differences between the summer monsoons of 1987 and 1988; the information that is contained in the initial data seems to be important as well. The initial data may, however, already contain information from the SSTs through observational data of the atmosphere. We found two features in the initial data that were distinctly different in the two years, which lasted through the whole

TABLE 6. As Table 5 but for northwestern India.

	Validation					Obs SST			Clim. SST			Pac SST			Ind SST		
1987:	36	53	24	87	48	19	0	0	6	0	1	2	0	1	0	0	32
1988:	240	220	333	199	190	25	4	2	19	158	20	22	4	48	8	46	4

TABLE 7. July precipitation for all India, values in mm month^{-1}. Experiments with increased or decreased SST over the Indian Ocean north of 10°N.

Exp	1/87	2/87	3/87	1/88	2/88	3/88
Control	125	38	62	151	230	146
+0.4	113	133	184	168	74	170
−0.4	79	101	25	42	58	78

forecast and which may have had an impact on the precipitation over India—that is, the soil moisture over Eurasia (see section 5b) and the stratospheric QBO (see section 5c). Also, Sperber and Palmer (1996) show important impacts from the initial data in similar experiments; however, they found that the impacts may be sensitive to the model used in such experiments.

5. Results from sensitivity experiments

a. North Indian Ocean SST experiments

Above we have deduced that ENSO events are likely to cause wind anomalies over the Arabian Sea and these would consequently affect SST anomalies over the Arabian Sea. However, interactions with local SST anomalies might cause monsoon anomalies as well and below we shall demonstrate such local effects. One can anticipate that a warmer Arabian Sea leads to enhanced evaporation and consequently to enhanced precipitation over India with a strengthened monsoon circulation. In fact the latent heat flux over the Arabian Sea in the AMIP simulations correlates positively with the SST of the same area. Composites of SST anomalies over the northern Indian Ocean using the years of extreme precipitation over India in the AMIP simulation indicate a positive correlation between precipitation over India and the SST over the northern Arabian Sea or the most northern part of the Indian Ocean. In these experiments the correlation between the SST over the Indian Ocean north of 22°N and the precipitation over all India is 45% [see section 4b(1)].

To investigate the effect of SST anomalies over the Arabian Sea further, we have set up 90-day forecast sensitivity experiments in which we modified the MONEG experiments using climatological SST and interactive soil moisture. A small temperature anomaly was added or subtracted north of 10°N in the Indian Ocean, increasing linearly from 0.0 K at 10°N to 0.4 K at 25°N. Table 7 gives an overview of the results of these experiments. Precipitation amounts for all India for 1987 and 1988 are presented for July as this month provides the clearest impacts. All experiments with increased SST give more precipitation than those with reduced SST. One would expect the control runs to give values somewhere between the two SST anomaly runs but this is true only for half of the forecasts. This sheds some doubts on the fidelity of this finding. A larger sample may be needed to clarify the significance of this signal. The present experiments exhibit, however, the high sensitivity of the ECHAM3 model to small changes of the SST in this area within the margin of uncertainty.

In Fig. 10 we summarize our findings in a schematic diagram: an El Niño has two effects; as a direct effect it causes a shift in the Walker circulation and thereby it reduces the precipitation over India and reduces the surface winds over the Arabian Sea. Due to the latter, as a secondary effect the Arabian Sea can become warmer as there is less mixing and upwelling in the ocean. Because of this increased SST more evaporation over the sea would lead to more precipitation over India, thus counteracting the expected decrease from the direct El Niño effect.

At the same time, a warmer Indian Ocean means a reduced temperature contrast between the ocean and the land, which would weaken the monsoon—that is, coun-

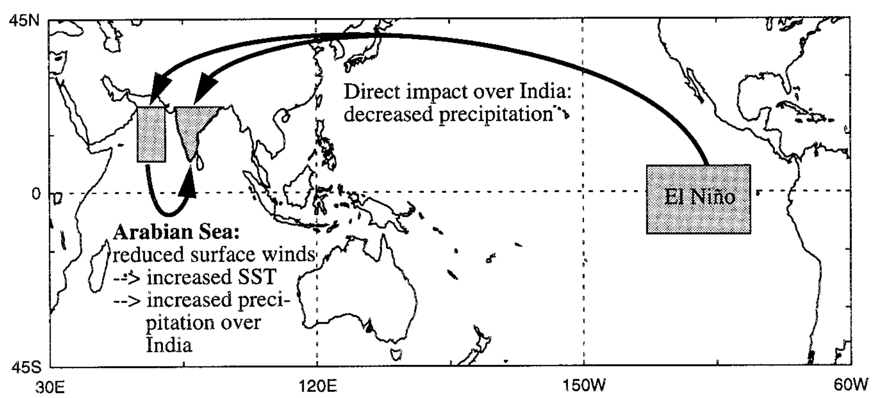

FIG. 10. Schematic diagram of impacts from El Niño on the precipitation over India. Directly it leads to a decrease of precipitation because of a shift in the Walker circulation. The El Niño leads further directly to reduced surface winds over the Arabian Sea. Due to the latter, the SST of the Arabian Sea can increase as there is less mixing and upwelling in the ocean. Because of this increased SST there would be more precipitation over India, this counteracts the decrease from the direct El Niño effect.

TABLE 8. July precipitation for all India, values in mm month^{-1}. Forecasts with interactive soil moisture calculations, prescribed analyzed, or climatological soil moisture. See Table 5 for more details.

	Validation					Interactive			Ana. soil m.			Clim. soil m.		
1987:	149	173	238	148	141	125	38	62	106	127	105	102	116	118
1988:	266	271	401	258	267	151	230	146	181	192	210	137	130	163

teracting the strengthening due to the enhanced evaporation. The effect from evaporation seems to be dominant.

We have seen above that in 1988 the coupling between the Niño-3 SST and the SST over the Arabian Sea was disturbed—that is, the mechanism described in Fig. 10 did not work for this year and warm temperatures over the Arabian Sea were observed in connection with cold temperatures over the Pacific. Both SST anomalies are favorable for enhanced precipitation in 1988 as observed.

b. Impacts of soil moisture on precipitation over India

The importance of soil moisture was already addressed by MONEG and therefore we have carried out the suggested experiments with prescribed soil moisture. In Table 8 the July precipitation for all India in the 90-day forecasts using an interactive soil moisture calculation is compared to those using climatological or analyzed soil moisture, all using climatological SSTs. All the forecasts have larger precipitation in 1988 than 1987, but the ones using interactive soil moisture and varying analyzed prescribed soil moisture show clearly larger differences than the forecasts using climatological soil moisture. This indicates that the different soil moisture anomalies in 1987 and 1988 contained either in the initial data or in the analysis helped to generate different precipitation intensities over India in both years. However, also other information in the initial data contributes toward the observed precipitation difference between the two years. The results for northwestern India in Table 9 give similar differences though with a bias toward too low precipitation.

In Fig. 3 it was shown that throughout Asia the soil is drier in 1988 than in 1987 except for a small area over northern India. We assume that the local effect of the soil moisture over India is less important than the effect from the whole of Asia. This is supported by the AMIP simulation in which the two individual experiments with a realistic precipitation increase from 1987 to 1988 over India have a difference in the soil moisture between 1988 and 1987 in June that is negative for the whole of eastern Asia, while there are also positive differences in the other experiments.

c. Impact of the stratospheric QBO on the precipitation over India

In section 4b it is mentioned that the stratospheric QBO differed considerably between 1987 and 1988. There is an easterly phase at 50 hPa in 1987 and a westerly phase in 1988. Zonal mean values from the ECMWF analysis and observations at Singapore (Naujokat 1986) have the same signal; however, the winds observed at Singapore are more extreme as might be expected. These differences in the QBO were already present in the initial data from which the MONEG forecasts were started and persisted through the 90 days of the forecasts. The west wind phase of 1988, however, was slightly weakened during the forecast (not shown).

The ECHAM model does not generate realistic stratospheric QBOs in climate simulations nor does any other GCM; the tropical stratospheric winds are weakly easterly most of the time. According to Gray et al. (1992) the phase of the QBO may have a direct impact on the convection in the equatorial zone. The stratospheric zonal winds at the equator cause a secondary meridional circulation that leads to a warming at the tropopause level with westerlies and cooling with easterlies. This changes the vertical stability in the upper troposphere and therefore the intensity of the convection. The direct impact on the equatorial convection may also have an impact on the subtropics. The correlation between the zonal wind at 50 hPa over Singapore and the all-India precipitation both for JJAS during the period 1954–90 is 32%, which agrees with investigations by Mukherjee et al. (1985). Gray et al. (1992) regard the wind shear in the lower stratosphere as more important than the winds themselves and indeed using the 70–50-hPa wind difference instead increases the correlation slightly to 36%.

To find out if the precipitation difference over India between the two years might have been caused by the different phases of the QBO contained in the initial fields, we carried out experiments in which we nudged the QBO state taken from observed monthly means of

TABLE 9. Same as Table 8 for northwestern India.

	Validation					Interactive			Ana. soil m.			Clim. soil m.		
1987:	36	53	24	87	48	6	0	1	24	12	9	37	12	18
1988:	240	220	333	199	190	19	158	20	55	161	45	9	56	82

TABLE 10. July precipitation for all India, values in mm month^{-1}. AMIP simulations compared to experiments with nudged QBO. See Table 5 for more details.

	Validation					AMIP (no QBO)				With QBO			
1987:	149	173	238	148	141	127	156	95	187	101	129	83	212
1988:	266	271	401	258	267	71	198	246	118	99	201	247	134

the zonal wind over Singapore (Naujokat 1986) into the equatorial stratosphere of the AMIP simulations. The prescribed zonal wind is assumed to be longitudinally uniform and to follow a Gaussian distribution in the meridional direction (Reed 1965). The relaxation coefficient is set to $N = 1(10 \text{ day})^{-1}$ between 10°N and 10°S, and decays to 0 at 20° lat. For 1987 and 1988 the forecasts for the months of July and August were rerun with this method using initial data from the AMIP simulations valid on 1 July. Table 10 shows the impact of the QBO on the precipitation over the Indian peninsula in these simulations. For practical reasons this nudging was carried out in only four of the five AMIP simulations. In all experiments the incorporation of the QBO signal leads to an increase of precipitation for 1988 though, with quite small differences and in three out of four to a decrease for 1987, in accordance with observations.

Giorgetta (1996) has investigated this further with a more recent version of the ECHAM model. He found the largest impacts of the stratospheric QBO on the precipitation over Indonesia with wave trains toward the northern Pacific but also significant signals over India, similar to those described here—that is, larger impacts during the east phase of the QBO (1987) and only weak impacts during the west phase (1988).

6. Conclusions

We have shown that although there is a large-scale dynamical response of the Pacific SST anomalies on the monsoon circulation in our model results, there is no simple connection between ENSO and precipitation over India. The difference in observed precipitation between 1987 and 1988 is of opposite sign for May to that for September. The simulations and observations agree in the manifestation of this sense of opposing variability within a monsoon season for these two years and also for other years.

During the peak of the monsoon activity (i.e., July) the precipitation over India in the AMIP simulation is not at all correlated with ENSO events. Observed precipitations show a modest negative correlation with Niño-3 SSTs, boosted by the 1987–88 difference. The 1987–88 difference averaged for the whole monsoon season JJAS is reproduced by the model simulations only for very large areal means in four out of five realizations. On the other hand, the 90-day forecasts (MONEG) reproduce the 1987–88 difference of precipitation over India very well. The quality of the forecasts is similar when using climatological or interannually varying SSTs. Therefore other influences than the Pacific SST anomalies seem to be important as well. However, some of the information from the Pacific SST anomalies may be contained in the analyses that are used as initial fields for the forecasts. We investigated the SST anomalies over the northern Indian Ocean, the soil moisture over Eurasia, and the stratospheric QBO as possible forcings for the precipitation variability.

In the sensitivity experiments, warmer SST over the northern Indian Ocean leads to increased precipitation over India due to enhanced evaporation. Stronger monsoon precipitation is mostly connected with a stronger circulation, and increased winds enhance the mixing and upwelling in the Arabian Sea, which leads to cooler SSTs. This seems to be a very delicate negative feedback so that the interannual variation of SST over the Arabian Sea is very low. The interannual variability of the SST has an amplitude as large as the differences between two SST analyses—that is, the interannual variability of the SST lies within the margin of uncertainty.

SST anomalies over the Arabian Sea are positively correlated with SST anomalies over the tropical Pacific (Niño-3 area) via large-scale atmospheric circulation anomalies caused by the ENSO events (anomalies of the Walker circulation). These SST anomalies over the Arabian Sea are positively correlated with the precipitation over India, whereas the SST anomalies of the Niño-3 area are negatively correlated with the precipitation over India. So there are two effects on the Indian precipitation from SST anomalies of the Pacific that counteract each other. There are indications that the direct impact is stronger in the observations and that the indirect impact is stronger in the ECHAM3 simulations.

The use of realistic soil moisture differences between 1987 and 1988 in the MONEG forecasts resulted in improved skill of the precipitation forecasts over India. Also, the two individual AMIP simulations with realistic precipitation differences over India had more realistic soil moisture differences over east Asia in the beginning of the monsoon season between the two years than those experiments that failed to produce the correct precipitation differences.

As atmospheric circulation models cannot yet reproduce stratospheric QBOs realistically, their impact was tested by nudging observed QBOs into AMIP simulations for July 1987 and 1988. Seven out of eight experiments showed an impact toward a more realistic simulation of precipitation over India, however, with very small impacts during the west phase of the QBO (1988) in accordance with experiments by Giorgetta (1996).

In addition to the impacts from SST anomalies over the Pacific, we found impacts from very small SST anomalies over the northern Indian Ocean, from soil moisture anomalies over Eurasia, and from the stratospheric QBO. None of them gave a dominant effect. Although the precipitation over India may be predictable to some extent, in practice the long-range forecasts will be extremely difficult as two of these potential predictors (northern Indian Ocean SSTs and the Eurasian soil moisture) are not known to the required accuracy and because models are not yet able to simulate the stratospheric QBO realistically.

Because of the highly nonlinear response of our model (presumably also of the real atmosphere) to small changes in boundary conditions and to initial data, we intend to carry out further experiments to obtain results with a higher statistical significance. Experimentation with different models would be worthwhile in order to establish how far these results are model dependent.

The signals investigated here are weak; only by a combined study of model simulations and observational data was some confidence in the findings gained.

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REFERENCES

Arpe, K., L. Bengtsson, L. Dümenil, and E. Roeckner, 1994: The hydrological cycle in the ECHAM3 simulations of the atmospheric circulation. *Global Precipitations and Climate Change,* M. Desbois and F. Desalmand, Eds., NATO ASI Series, Vol. 1, NATO, 361–377.

Barnett, T. P., L. Dümenil, U. Schlese, E. Roeckner, and M. Latif, 1989: The effect of Eurasian snow cover on regional and global climate variations. *J. Atmos. Sci.,* **46,** 661–685.

Bengtsson, L., K. Arpe, E. Roeckner, and U. Schulzweida, 1996: Climate predictability experiments with a general circulation model. *Climate Dyn.,* **12,** 261–275.

Fennessy, M. J., and J. Shukla, 1992: Influence of global SST on GCM simulations of the Northern Hemisphere monsoon circulation of 1987 and 1988, simulation of intra-seasonal monsoon variability. WMO Tech. Document WMO/TD 470, World Meteorological Organization, Geneva, Switzerland, 2.37–2.45.

Gates, W. L., 1992: AMIP: The Atmospheric Model Intercomparison Project. *Bull. Amer. Meteor. Soc.,* **73,** 1962–1970.

Giorgetta, M. A., 1996: Der Einfluss der quasi-zweijährigen Oszillation auf die allgemeine Zirkulation: Modellsimultionen mit ECHAM4. Examensarbeit Nr. 40, 146 pp. [Available from Max-Planck-Institut für Meteorologie, Bundesstr. 55, D-20146 Hamburg, Germany.]

Gray, W. M., J. D. Sheaffer, and J. A. Knaff, 1992: Influence of the stratospheric QBO on ENSO variability. *J. Meteor. Soc. Japan,* **70,** 975–995.

Heckley, W. A., and A. E. Gill, 1984: Some simple analytical solutions to the problem of forced equatorial long waves. *Quart. J. Roy. Meteor. Soc.,* **110,** 203–217.

Hulme, M., 1994: Validation of large-scale precipitation fields in General Circulation Models. *Global Precipitations and Climate Change,* M. Desbois and F. Desalmand, Eds., NATO ASI Series, Vol. 1, NATO, 387–406.

Janowiak, J. E., and P. A. Arkin, 1991: Rainfall variations in the Tropics during 1986–1989, as estimated from observations of cloud-top temperature. *J. Geophys. Res.,* **96,** 3359–3373.

Ju, J., and J. Slingo, 1995: The Asian summer monsoon and ENSO. *Quart. J. Roy. Meteor. Soc.,* **121,** 1133–1168.

Kane, R. P., 1997: Relationship of El Niño–Southern Oscillation and Pacific sea surface temperature with rainfall in various regions of the globe. *Mon. Wea. Rev.,* **125,** 1792–1800.

Krishnamurti, T. N., H. S. Bedi, and M. Subramaniam, 1990: The summer monsoon of 1988. *Meteor. Atmos. Phys.,* **42,** 19–37.

Meehl, G. A., 1994: Coupled land-ocean-atmosphere processes and South Asian monsoon variability. *Science,* **266,** 263–267.

Mintz, Y., and Y. V. Serafini, 1992: A global monthly climatology of soil moisture and water balance. *Climate Dyn.,* **8,** 13–27.

Mukherjee, B. K., K. Indira, R. S. Reddy, and Bh. V. Ramana Murty, 1985: Quasi-biennial oscillation in stratospheric zonal wind and Indian summer monsoon. *Mon. Wea. Rev.,* **113,** 1421–1424.

Naujokat, B., 1986: An update of the observed quasi-biennial oscillation of the stratospheric winds over the Tropics. *J. Atmos. Sci.,* **43,** 1873–1877.

Palmer, T. N., and D. L. T. Anderson, 1994: The prospects for seasonal forecasting—A review paper. *Quart. J. Roy. Meteor. Soc.,* **120,** 755–793.

——, C. Brankovic, P. Viterbo, and M. J. Miller, 1992: Modelling interannual variations of summer monsoons. *J. Climate,* **5,** 399–417.

Parthasarathy, B., A. A. Munot, and D. R. Kothawale, 1994: All-India monthly and seasonal rainfall series: 1871–1993. *Theor. Appl. Climatol.,* **49,** 217–224.

Rayner, N. A., E. B. Horton, D. E. Parker, C. K. Folland, and R. B. Hackett, 1996: Version 2.2 of the global sea-ice and sea surface temperature data set, 1903–1994. Climate Research Tech. Note CRTN74, 35 pp.

Reed, R. J., 1965: The present status of the 26-month oscillation. *Bull. Amer. Meteor. Soc.,* **46,** 374–387.

Reynolds, R. W., 1988: A real-time global sea surface temperature analysis. *J. Climate,* **1,** 75–86.

Roeckner, E., and Coauthors, 1992: Simulation of the present-day climate with the ECHAM model: Impact of model physics and resolution. MPI Rep. 93, 171 pp. [Available from Max-Planck-Institut für Meteorologie, Bundesstr. 55, D-20146 Hamburg, Germany.]

Ropelewski, C. E., and M. S. Halpert, 1987: Global and regional scale precipitation and temperature patterns associated with the El Niño/Southern Oscillation. *Mon. Wea. Rev.,* **115,** 1606–1626.

Rudolf, B., H. Hauschild, M. Reiß, and U. Schneider, 1992: The calculation of areal mean precipitation totals on a 2.5 grid by an objective analyses method. *Meteor. Z.,* **1,** 32–50.

Schemm, J., S. Schubert, J. Terry, and S. Bloom, 1992: Estimates of monthly mean soil moisture for 1979–1989. NASA Tech. Memo. 104571, 262 pp.

Shukla, J., 1975: Effect of Arabian sea-surface temperature anomaly on the Indian summer monsoon: A numerical experiment with the GFDL model. *J. Atmos. Sci.,* **32,** 503–511.

——, and B. M. Misra, 1977: Relationships between sea surface temperature and wind speed over the central Arabian Sea, and monsoon rainfall over India. *Mon. Wea. Rev.,* **105,** 998–1002.

Singh, S. V., R. H. Kripalani, and D. R. Sikka, 1992: Interannual variability of the Madden–Julian oscillations in the Indian summer monsoon rainfall. *J. Climate,* **5,** 973–978.

Spencer, R. W., 1993: Global oceanic precipitation from the MSU during 1979–1991 and comparisons to other climatologies. *J. Climate,* **6,** 1301–1326.

Sperber, K. R., and T. N. Palmer, 1996: Interannual tropical rainfall variability in general circulation model simulations associated with the Atmospheric Model Intercomparison Project. *J. Climate,* **9,** 2727–2750.

Weare, B. C., 1979: A statistical study of the relationships between ocean surface temperatures and the Indian monsoon. *J. Atmos. Sci.,* **36,** 2279–2291.

Webster, P., and S. Yang, 1992: Monsoon and ENSO: Selectively interactive systems. *Quart. J. Roy. Meteor. Soc.,* **118,** 877–926.

WMO, 1992: Simulation of interannual and intra-seasonal monsoon variability. WRCP 68, WMO/TD No. 470, 208 pp.