Modeling Interannual Variations of Summer Monsoons

T. N. Palmer, Č. Branković, P. Viterbo, and M. J. Miller

European Centre for Medium-Range Weather Forecasts, Reading, United Kingdom

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ABSTRACT

Results from a set of 90-day integrations, made with a T42 version of the ECMWF model and forced with a variety of specified sea surface temperature (SST) datasets, are discussed. Most of the integrations started from data for 1 June 1987 and 1 June 1988. During the summer of 1987, both the Indian and African monsoons were weak, in contrast with the summer of 1988 when both monsoons were much stronger. With observed SSTs, the model is able to simulate the interannual variations in the global-scale velocity potential and streamfunction fields on seasonal time scales. On a regional basis, rainfall over the Sahel and, to a lesser extent, India showed the correct sense of interannual variation, though in absolute terms the model appears to have an overall dry bias in these areas.

Additional integrations were made to study the impact of the observed SST anomalies in individual oceans. Much of the interannual variation in both Indian and African rainfall can be accounted for by the remote effect of the tropical Pacific SST anomalies only. By comparison with the effect of the Pacific, interannual variability in Indian Ocean, tropical Atlantic Ocean, or extratropical SSTs had a relatively modest influence on tropical large-scale flow or rainfall in the areas studied.

Integrations run with identical SSTs but different initial conditions indicated that for large-scale circulation diagnostics, the impact of anomalous ocean forcing dominated the possible impact of variations in initial conditions. In terms of local rainfall amounts, on the other hand, the impact of initial conditions is comparable with that of SST anomaly over parts of India and Southeast Asia, less so over the Sahel. While this may suggest that a nonnegligible fraction of the variance of month-to-seasonal mean rainfall on the regional scale in the tropics may not be dynamically predictable, it is also quite possible that the disparity in the apparent predictability of rainfall and circulation anomalies is a reflection of model systematic error.

1. Introduction

A major component of the Tropical Ocean/Global Atmosphere (TOGA) program is focused toward the development of a seasonal time-scale prediction capability using global coupled ocean–atmosphere general circulation models. In order to test the potential of such models for seasonal forecasting, it is necessary to assess how well they can simulate interannual variations in regional climate when the observed sea surface temperatures (SSTs) for a given season are prescribed (i.e., assuming a “perfect” ocean model environment). Some aspects of this question have already been addressed by model intercomparison studies of the large-scale atmospheric response to (northern) wintertime El Niño SST anomalies (WMO 1986, 1988). However, of particular practical importance in this respect are the interannual variations in summer monsoon rains, since failure of these rains almost invariably has devastating consequences on crop production for the countries affected. A rationale for expecting significant seasonal predictability of monsoon rains and circulations has been given by Charney and Shukla (1981) and Shukla (1981) and is discussed further below.

In addition, seasonal integrations are relevant to the needs of shorter time-scale weather prediction. Long integrations of numerical weather prediction (NWP) models allow an assessment of systematic biases in the simulation of atmospheric low-frequency variability. The ability to simulate the correct climatological frequency of blocking events, for example, is of paramount importance in medium-range NWP.

In this paper, we present some results from a set of seasonal integrations made with the T42 version of the ECMWF NWP model over the northern summers of 1987 and 1988 with specified SST. These years were of particular interest for their contrasting behavior in monsoon regions (Krishnamurti et al. 1989, 1990). In particular, 1987 was a severe drought year over both India and the Sahel, while 1988 was an above-average monsoon year for India, and rains over the Sahel were close to the climatological mean. During these two years, the El Niño/Southern Oscillation (ENSO) cycle swung from a warm to a cold phase. We shall focus our discussion on the ability of the ECMWF model to simulate both global and regional seasonal and monthly mean monsoon circulations and associated rainfalls.
during the northern summer. Additional integrations designed to study the influence of SST anomalies in the individual oceans and the sensitivity of month and seasonal mean fields to initial conditions are also described.

2. Experimental design

All integrations considered here are 90 days in duration and were initialized from ECMWF operational archived data from 1 June 1987 or 1 June 1988 (except for one integration that started on 2 June 1988). The integrations were made with a T42L19 version of the (cycle 34) ECMWF model (Simmons et al. 1988) without cumulus momentum transport but including a revision to the bulk parameterization of sensible and latent heat flux. The principal effect of this revision, described in a companion paper (Miller et al. 1991), was to increase surface evaporation significantly in regions of weak grid-box–mean wind. As shown in Miller et al., the new treatment of the fluxes had a substantial impact on the model’s sensitivity to SST anomalies in the warm pool regions.

The integrations were made with a number of specified SST datasets updated from 5-day mean files. The experiments are summarized in Table 1.

In this study, climatological SSTs were taken from the Reynolds and Roberts (1987) dataset (see also Reynolds 1988). Observed SSTs were taken from the National Meteorological Center SST analyses used for operational ECMWF forecasts. Differences between the observed and climatological SST for the individual months of June, July, and August of both years are shown in Fig. 1. It can be seen that a number of features are common to both years, notably the relative warming in the equatorial and South Atlantic and Indian oceans. In addition, the extratropical North Pacific is cooler than climatology. This pattern is fairly consistent with patterns of interdecadal variability over the last three decades studied from other analyses of SST (e.g., Folland et al. 1986, 1990).

In addition, however, these maps reveal clear evidence of interannual variability, primarily associated with ENSO, which swung from a warm to a cold phase between these two years, with SST anomalies in the equatorial eastern Pacific evolving from more than +2 K in 1987 to more than −2 K in 1988. Note that the cold anomalies in 1988 were restricted to a few degrees of latitude from the equator, while the warm anomalies in 1987 had a much greater latitudinal extent. For reference, Fig. 2 shows the monthly mean differences between SST in 1988 and 1987. The effects of El Niño are dominant over the whole season in the tropical Pacific Ocean. There is a tendency for the Indian Ocean SSTs to be cooler in 1988 than 1987 (up to 1 K locally in June). This may be associated with the effect of the relatively strong atmospheric monsoonal flow in 1988 on the ocean (Shukla 1987).

For the “individual ocean” experiments described in Table 1, the boundary between the Indian and Pacific oceans was taken at 120°E. Where SST anomalies were included in the “tropics” only, they were taken from 30°S to 30°N. Where SST anomalies were included in the “extratropics” only, they were taken poleward of 30° lat.

3. Global diagnostics

a. Integrations with global SST anomalies

For verification purposes we show in Fig. 3 maps of seasonal mean (JJA) 200-mb velocity potential (left-hand column) and streamfunction anomalies (right-hand column) for 1987 and 1988 from ECMWF 12Z analyses. The “climate” used to construct these anomalies is a mean of ECMWF analyses for five summers from 1985 to 1989; analyses of tropical divergent flow, taken from years earlier than this, suffer from serious deficiencies and were therefore not included [for example, see the discussion in Sardeshmukh and Hoskins (1987)]. The reference climate fields are shown in the bottom panels of Figs. 3e,f.

Figure 3 shows clear evidence of significant large-scale atmospheric interannual variability. Over the eastern Pacific, consistent with the differences in SST, there is anomalous large-scale 200-mb divergence in 1987 (Fig. 3a) and anomalous convergence in 1988 (Fig. 3e). In 1987 the anomalous divergence extends across the Atlantic, with anomalous convergence over Indonesia and Southeast Asia. In 1988 the anomalous Pacific convergence extends across the whole Pacific, with relatively strong anomalous divergence centered.
Fig. 1. Sea surface temperature anomalies for (a) June 1987, (b) July 1987, (c) August 1987, (d) June 1988, (e) July 1988, and (f) August 1988.

Contour interval 1 K, with ±0.5 K contours. Coarse stipple > +0.5 K and fine stipple < -0.5 K.
over the western Indian Ocean. The anomalous rotational wind is predominantly westerly across the tropics in 1987 (Fig. 3b) and easterly, particularly over the Atlantic, in 1988 (Fig. 3d).

In Fig. 4 the 90-day mean 200-mb 1200 UTC velocity potential and streamfunction for the differences (87O–87C) and (88O–88C) between experiments with observed and climatological SSTs are shown. For reference, in the bottom panels in Figs. 4e,f the full fields averaged for 87C and 88C are shown.

For 87O–87C, the simulated pattern of large-scale divergence (Fig. 4a) corresponds reasonably to Fig. 3a, with anomalous convergence over the Pacific and the Atlantic. However, the anomalous divergence is positioned farther west than in Fig. 3a. Moreover, the amplitude of the response is much larger than in Fig. 3a. The pattern of 200-mb streamfunction for 87O–87C (Fig. 4b) is broadly in agreement with the observed anomalies. In particular, in both model and analysis (Fig. 3b) there are strong westerly anomalies over much of the tropics during 1987, excluding the eastern Pacific, though again the model response is too strong in amplitude.

For 88O–88C, both the amplitude and the simulated pattern of large-scale divergence (Fig. 4c) is broadly similar to that analyzed, though there is a less pro-
Fig. 3. June–August 200-mb ECMWF analysis anomalies: (a) velocity potential for 1987, contour interval $1 \times 10^4$ m$^2$ s$^{-1}$; (b) streamfunction for 1987, contour interval $3 \times 10^4$ m$^2$ s$^{-1}$. (c) As in (a) but for 1988; (d) as in (b) but for 1988. June–August 200-mb ECMWF climate (3-yr) mean: (e) velocity potential, contour interval $2 \times 10^4$ m$^2$ s$^{-1}$; (f) streamfunction, contour interval $20 \times 10^4$ m$^2$ s$^{-1}$.

nounced minimum over the western Indian Ocean than in Fig. 3c. The streamfunction difference for 880–88C (Fig. 4d) correctly shows anomalous westerlies over the equatorial Pacific and anomalous easterlies over south America and the tropical Atlantic.

The full field streamfunction (Fig. 4f) compares well

Fig. 4. Days 1–90 200-mb differences: (a) 870–87C velocity potential, contour interval $1 \times 10^4$ m$^2$ s$^{-1}$; (b) 870–87C streamfunction, contour interval $3 \times 10^4$ m$^2$ s$^{-1}$. (c) As in (a) but for 880–88C; (d) as in (b) but for 880–88C. Days 1–90 200-mb average of 87C and 88C: (e) velocity potential, contour interval $2 \times 10^4$ m$^2$ s$^{-1}$; (f) streamfunction, contour interval $20 \times 10^4$ m$^2$ s$^{-1}$. 
with the analyzed climatological full field (Fig. 3f), and both maps show clearly the upper-level Tibetan high, associated with the Asian monsoonal flow. The full field velocity potential in both model and analysis (Figs. 4c and 3e) similarly compares well over Asia, though the model appears to have developed erroneous divergence over Central America.

A quantitative comparison of Figs. 3 and 4 is hampered because of ambiguities in the calculation of atmospheric and SST climatologies. Hence, in order to verify the integrations 87O and 88O in a way that is independent of climate, Figs. 5a,b show the analyzed differences in 200-mb velocity potential and streamfunction between JJA 1988 and 1987, and Figs. 5c,d show the difference in these fields between 88O and 87O. It can be seen that while the overall simulated pattern is quite realistic, the model response, particularly in terms of the velocity potential, is again somewhat strong. On the basis of these results, it is possible that the model is too responsive to the warm El Niño SST anomalies of 1987; on the other hand, the analyses themselves were made with versions of the ECMWF model that predate the one used for the integrations here (in particular, excluding the flux revision described in Miller et al. 1991), introducing further uncertainty in the verification process.

It can be remarked that studies such as these would benefit from consistent SST and atmospheric climatological datasets and would lend support to the requirement for atmospheric reanalysis (Bengtsson and Shukla 1988).

b. Impact of initial conditions

As emphasized by Charney and Shukla (1981), seasonal predictability of the monsoons arises in part because the evolution of the large-scale tropical atmosphere is relatively insensitive to atmospheric initial conditions, in comparison with its dependence on boundary conditions—SST in particular. This can be seen in Figs. 5e,f, which show the difference (88O*–88O) in 90-day mean 200-mb velocity potential and streamfunction; integrations are with identical SST but initial conditions one day apart. Compared with the differences shown in Figs. 5c,d, it can be seen, particularly in the tropics, that the response to different SST overwhelms the response in the large-scale flow to differences in initial conditions.

As a further test of the sensitivity of the tropical simulations to initial conditions, velocity potential and streamfunction differences between integrations 88C and 87C, initialized 1 year apart and run with identical SSTs, have been calculated (but are not shown). Again, the response is very small compared to that shown in Figs. 5a–d.

Despite the relative insensitivity of these global diagnostics to initial conditions, it will be shown in section 4c that, over parts of the Indian Ocean, Southeast

![Figure 5](image-url)

**Fig. 5.** June–August 200-mb ECMWF analysis difference 1988–1987: (a) velocity potential; (b) streamfunction. Days 1–90 200-mb difference 88O–87O; (c) velocity potential; (d) streamfunction. (e) As in (c) but for 88O*–88O; (f) as in (d) but for 88O*–88O. Contour interval: $1 \times 10^6$ m$^2$ s$^{-1}$ in (a), (c), and (e); $3 \times 10^6$ m$^2$ s$^{-1}$ in (b), (d), and (f).
Asia, and the western Pacific, regional values of July
- August rainfall differences are somewhat sensitive
to the initial conditions.

c. Integrations with regional SST anomalies

To explore further whether the global response to
the full SST anomalies, as shown in Figs. 4 and 5, can
be understood in terms of the responses to the individu-
als oceans, Figs. 6 and 7 show 90-day mean 200-
mb velocity potential and streamfunction anomalies
for the individual ocean experiments.

With tropical Pacific SST anomalies only, Fig. 6a
(87OP–87C) shows a large-scale response in velocity
potential very similar indeed to the full response in
Fig. 4a. The amplitude of this pattern is even larger
than with the global SST anomalies. With tropical In-
dian Ocean anomalies only, Fig. 6c (87OI–87C) shows
a more localized response, predominantly comprising
anomalous divergence over the Indian Ocean. The ef-
effect of the tropical Atlantic Ocean SST anomalies (Fig.
6e; 87OA–87C) is felt mainly over the Atlantic
(anomalous divergence) and the eastern Pacific
(anomalous convergence). Finally, Fig. 6g shows the
velocity potential response (87OE–87C) to the ex-
tropical SST anomalies. It can be seen that, compared
with any of the tropical SST anomalies, the response
in the tropics to extratropical SST anomalies is weak.

The streamfunction response to the 1987 tropical
Pacific anomalies is shown in Fig. 6b. Between the two
centers of anomalous velocity potential over the Pacific
and Indian ocean are anomalous equatorial easterlies,
with strong anticyclonic anomalies in both hemispheres
at about 120°E. To the east, over tropical South Amer-
ica, Atlantic Ocean, and Africa, there are anomalous
westerlies. This overall pattern qualitatively resembles
Gill’s (1980) linear model response to imposed tropical
heating anomalies. The equatorial streamfunction
anomalies associated with both Indian Ocean (Fig. 6d)
and Atlantic Ocean (Fig. 6f) SST anomalies for 1987
can be similarly viewed as weaker examples of this kind
of linear model response, relative to the appropriate
region of maximum 200-mb anomalous divergence.

The effect of the extratropical SST anomalies on the
tropical streamfunction field is again relatively weak
(Fig. 6h).

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**Fig. 6.** Days 1–90 200-mb differences: (a) 87OP–87C velocity potential, contour interval: $1 \times 10^6$ m$^2$ s$^{-1}$; (b) 87OP–87C streamfunction, contour interval: $3 \times 10^6$ m$^2$ s$^{-1}$; (c) as in (a) but for 87OI–87C; (d) as in (b) but for 87OI–87C; (e) as in (a) but for 87OA–87C; (f) as in (b) but for 87OA–87C; (g) as in (a) but for 87OE–87C; (h) as in (b) but for 87OE–87C.
It can be seen from Fig. 6 that in terms of these large-scale atmospheric flow diagnostics, the effects of the Indian Ocean and the Pacific SST anomalies are partially canceling over the Indian Ocean. Similarly, the Atlantic SST anomalies tend to partially cancel the effect of the Pacific anomalies over the Pacific itself.

The upper-air results for 1988 are consistent with those for 1987 (but not shown, for reasons of space). The response in velocity potential and streamfunction to the Pacific SST anomalies in the La Niña year 1988 (88OP–88C) is broadly opposite to the response for 1987 shown in Figs. 6a and 6e, though the magnitude of the response is weaker in 88OP. The difference in the magnitude of the response is possibly associated with the fact that the spatial extent of the anomalous equatorial cold pool was much smaller in the summer of 1988 than the spatial extent of the anomalous warm pool in 1987 (cf. Fig. 1). As noted above, the 1988 SST anomalies in the tropical Atlantic and Indian oceans are approximately the same as for 1987. Consistent with this, the patterns of difference fields (88OI–88C) and (88OA–88C) are approximately the same as (87OI–87C) and (87OA–87C), respectively, both in terms of velocity potential and streamfunction. The response to extratropical SST anomalies (88OE–88C) is again relatively small.

As discussed, the difference between observed and climatological SSTs for the Indian and Atlantic oceans in particular, may principally reflect interdecadal rather than interannual variability. In order to study the impact of interannual variations in regional SST, Fig. 7 shows the 200-mb velocity potential and streamfunction differences between 88OP–87OP, 88OI–87OI, 88OA–87OA, and 88OE–87OE. The results clearly show the predominance of the remote effect of El Niño. By comparison, the impact of the other ocean SST differences are quite negligible. Moreover, it is difficult to ascribe any statistical significance to the non-Pacific response. For example, the velocity potential difference over the western Indian Ocean in Fig. 7c (88OI–87OI) changes sign between the second and third months of the integrations. In particular, therefore, we would assert that the remote effect of El Niño is overwhelmingly more important in determining interannual variability.
of the Asian monsoon than the more local effect of Indian Ocean SST anomalies between 1987 and 1988.

4. Regional diagnostics

a. Integrations with global SST anomalies

In this subsection we discuss simulation of regional rainfall and low-level flow for the simulations 87O and 88O, focusing on the Sahelian and Asian monsoons. Successful simulations of interdecadal fluctuations in rainfall over the Sahel, using the U.K. Meteorological Office model, have been presented by Folland et al. (1986, 1991). Palmer et al. (1990) found that the observed and predicted interannual variations of monthly monsoon rainfall for the Sahel agreed relatively well, although not as well for the south Asian region. In this section we concentrate on averages over July and August; however, some full field rainfall maps for July and August are separately shown. In addition to the interannual variability, these illustrate how the model successfully captures aspects of the variation of the annual cycle of rainfall within the summer season, particularly over the Sahel.

1) The Sahel

In Figs. 8a,b the ECMWF-analyzed moisture flux for July-August 1987 and 1988 for level 17 of the model (about 970 mb) are shown. Values greater than 50 g kg\(^{-1}\) m s\(^{-1}\) are shown stippled. There are a number of characteristics of the moisture flux that are consistent with earlier studies by Lamb (1983). The most striking differences lie to the east of Lake Chad, where in 1988, relatively strong fluxes extended across the African continent almost to the Red Sea. These differences are captured well by the model simulations (87O and 88O) as shown in Figs. 8c,d for days 31-90, which correspond to July-August average. Note, for example, the relatively strong moisture flux across to Sudan in 88O.

Rainfall maps associated with the intertropical convergence zone (ITCZ) over Africa for July and August in 87O and 88O are shown in Fig. 9. The relative dryness in 87O is apparent for both months. In parts of the Sahel around 15°N, there is essentially no July rainfall in 87O, while in 88O, values generally lie between 1 and 2 mm day\(^{-1}\). In August of both years, the ITCZ has moved north, and rainfall values over the Sahel have increased. However, as for July, values are noticeably weaker in 1987 than in 1988, which have rates up to 4 mm day\(^{-1}\). Note also the “tongue” of rainfall over Ethiopia and the Red Sea in 88O, which is absent in 87O.

In order to verify these rainfall amounts, Fig. 10 depicts monthly mean station values in the Sahel from WMO CLIMAT reports (obtained from the U.K. Meteorological Office) for July and August 1987 (figures above station) and 1988 (figures below station). In July (Fig. 10a), rainfall rates are generally higher in 1988 than in 1987, except perhaps for stations near the Atlantic coast. In the latitude band around 15°N to the east of the Greenwich meridian, values lie around 2 mm day\(^{-1}\) in 1987 and 4 mm day\(^{-1}\) in 1988. In August (Fig. 10b) the difference between the relatively wet year in 1988 and the relatively dry year of 1987 can be clearly seen in most of the station values. At 15°N values in 1988 are up to about 8 mm day\(^{-1}\), yet they are only about 3 mm day\(^{-1}\) in 1987. While observed rainfall rates over Ethiopia and surrounding areas are not shown, the tongue of relatively strong rain in Fig. 9d does appear to be realistic. The flooding of the Nile during the summer of 1988 was well documented.

Compared with these station values, the model appears to have an overall dry bias. Consistent with this, as discussed later, this version of the model significantly underestimates transient variability over the Sahel on the time scale of easterly waves and squall lines with which rainfall is known to be associated. On the other hand, the principal sense of interannual variation is captured well by the model simulations.

2) India

Analyzed and simulated July-August mean 850-mb winds for 1987 and 1988 over the Indian Ocean are shown in Fig. 11. The analyzed low-level wind maximum off the Somali coast is noticeably stronger in 1988 (Figs. 11a,b). On the other hand, over India, the Bay of Bengal, and parts of southeast Asia, the low-level flow is stronger in 1987, suggesting stronger wind convergence over India and the Arabian Sea in 1988. In the integrations 87O and 88O, there is clearly a systematic westerly bias in the winds of more than 5 m s\(^{-1}\) in places. Nevertheless, the analyzed interannual differences over India, the Bay of Bengal, and southeast Asia have been captured, with noticeable stronger winds in 87O. While there appears to be stronger low-level convergence over India in 88O, the stronger analyzed Somali wind maximum in 88O has not been successfully simulated.

Figure 12 shows monthly mean rainfall maps from 87O and 88O for July and August. In 87O, much of the interior of India receives less than 1 mm day\(^{-1}\) in July (Fig. 12a). For 88O, in these areas, the comparable rates are up to 4 mm day\(^{-1}\) (Fig. 12b). The west coast maximum is also stronger in 88O than in 87O. These differences are consistent with the 90-day mean 200-mb velocity potential maps discussed above. Figures 12c,d show rainfall for August from 87O and 88O. Rainfall in the interior of India has increased from July values in both 87O and 88O; however, overall values for 88O are still noticeably weaker than in 88O. On the other hand, it is interesting to note that the rainfall maximum on the west coast is now somewhat weaker in 88O than in 87O.

Observed rainfall values over India are given in Fig.
Fig. 8. July–August mean model level 17 moisture flux (kg m⁻² s⁻¹) over the Sahel. (a) ECMWF analysis for 1987, (b) ECMWF analysis for 1988, (c) 870, and (d) 880. Shaded areas for values greater than 50 kg m⁻² s⁻¹.
13 for July and August 1987 and 1988. These values are subdivisional monthly means as reported by Das et al. (1988, 1989). The relative drought in July 1987 is manifested throughout India, with less rainfall in 1987 than 1988 in almost all subdivisions. In this sense, the simulated rainfall differences between 87O and 88O are reasonable. However, it can be seen that, as in the Sahel, the simulated rainfall rates are considerably smaller than those observed, particularly in the Indian interior. For August, the difference in observed rainfall in the two years is less marked, though values in the Indian interior are still somewhat lower in 1987. Interestingly, rainfall in the west coast subdivisions are not significantly higher in 1988 and are noticeably lower for one subdivision. In this sense, the simulated rainfall differences between 87O and 88O are not unrealistic for August, though, as in July, absolute values are still too small. As for the Sahel, we note that these rainfall deficiencies are consistent with the model's underestimation of transient variability over India on the time scale of monsoon depressions.

For reference (see section 3c), the mean rainfall difference for days 31–90 between 88O and 87O over Africa and south Asia is shown in Fig. 14. The extensive
Fig. 11. July-August mean 850-mb wind (m s⁻¹) over the Indian Ocean and surrounding regions: (a) ECMWF analysis for 1987, (b) ECMWF analysis for 1988. (c) 870, and (d) 880. Contours every 5 m s⁻¹, stippled if greater than 10 m s⁻¹.
region of positive values over the Sahel and the southern Indian Ocean is consistent within the individual months comprising this 60-day average. Differences over India, while generally positive, are less consistent.

b. Influence of initial conditions

In the discussion in section 4a comparing observed and simulated rainfall values, one might question the meaningfulness of a detailed comparison of values at the station or subdivisional level. In particular, if these values are not intrinsically predictable on these scales, such detailed comparison is not warranted. In section 3b it was noted that the impact of interannual differences in SST, particularly over the Pacific, on seasonal mean velocity potential and streamfunction dominated the impact of initial conditions. To assess whether this conclusion still holds for regional estimates of rainfall, Fig. 15 depicts rainfall differences for days 31–90 of 88O–88O; integrations are with identical SST fields but started one day apart.

Perhaps surprisingly in view of the earlier results, the differences in Fig. 15 are by no means negligible. Compared with interannual differences in Fig. 14, the differences are smaller and somewhat incoherent over the Sahel, though are more comparable both in scale and magnitude over south Asia and the Indian Ocean. Indeed, one might expect the Indian Ocean and west Pacific to be regions where local rainfall predictability is smallest, since they are areas where internal intra-seasonal variability is largest.

From these results it appears that month-to-seasonal time-scale rainfall is more predictable over the Sahel than over India. This is consistent with the conclusions of Palmer et al. (1990), who studied 30-day mean rainfall predictions over a set of 4 years.

Considerable caution is required, however, before deducing that the predictability of month-to-seasonal time-scale rainfall is relatively small over south Asia, compared with the predictability of the large-scale tropical circulation discussed in section 3. As mentioned above, analysis of this version of the ECMWF model reveals a significant underestimation of tropical transient activity on the time scale of monsoon depressions and easterly waves (not shown here). In practice, much of the observed tropical rainfall is associated with such transients. In reality, these transients are dynamically coupled to the planetary-scale flow in the deep
tropical troposphere. An underestimation of such transients in the model may therefore imply that simulated rainfall variability is not sufficiently well coupled to simulated variability in planetary-scale flow. In such a situation, the relatively small apparent predictability of rainfall compared with 200-mb flow may merely reflect the inadequacy of such coupling and may therefore be a consequence of systematic deficiencies in the model.

It is possible that the reason for such a weak coupling is associated with the choice of closure in the model's mass-flux convection scheme. Preliminary studies of

Fig. 13. Observed rainfall over India (mm day$^{-1}$) based on subdivisional means (Das et al. 1988, 1989): (a) July 1987, (b) July 1988, (c) August 1987, and (d) August 1988.

Fig. 14. Rainfall differences averaged between days 31–90 for 880–870. Coarse stippling for values greater than 2 mm day$^{-1}$, fine stippling for values less than $-2$ mm day$^{-1}$. Contours at 2, 4, 8, and 16 mm day$^{-1}$. 
seasonal model integrations with a different convection scheme, that is, with the adjustment scheme described in Betts and Miller (1986), have revealed a significant increase in simulated tropical transients on time scales of 5 days or less. The mass-flux scheme has a low-level moisture convergence closure and hence is only weakly coupled to mid- and upper-tropospheric variability. The adjustment scheme, on the other hand, is more closely coupled to this variability through its use of reference profiles throughout the entire convective troposphere.

c. Integrations with regional SST anomalies

In Figs. 16a–c, July – August mean rainfall difference fields over Africa and south Asia are shown between “individual ocean” experiments 88OP–87OP, 88OI–87OI, and 88OA–87OA, respectively, for the last 60 days of the experiments. The SST difference field between the two years was shown in Fig. 2, and the seasonal mean velocity potential and streamfunction difference fields between these experiments were shown in Fig. 7.

The 88OP–87OP difference (Fig. 16a) shows a relatively large-scale impact of tropical Pacific SST variability on monsoon rainfall. Over most of the Sahel, East Africa, India, and the south Indian Ocean, rainfall is enhanced. On the other hand, there is some decrease in rain over the coastline of Burma. These differences are similar to those shown in Fig. 14 for the full global observed SST fields and are consistent within the individual months comprising the 60-day mean.

For 88OI–87OI and 88OA–87OA (Figs. 16b,c), the response is generally weaker and less coherent. For the Indian Ocean, these results are consistent with those of Dube et al. (1990), who concluded that interannual variability of SST in the Arabian Sea can be largely thought of as the passive response to variability in the monsoon system. Similarly, the impact of interannual variability in Atlantic SST appears to be less important in accounting for interannual variations in Sahelian rain than the impact of El Niño. At first sight, this may seem inconsistent with earlier studies of the impact of SST on Sahelian rainfall (e.g., Folland et al. 1986; Palmer 1986), in which the importance of the Atlantic SST anomalies was emphasized. However, these studies were primarily focused on the interdecadal variability of Sahelian rainfall where coherent large-scale variability in Atlantic SST was evident.

In view of the importance of resolving the question of whether Indian Ocean SST anomalies have a significant influence on rainfall in surrounding regions, in Figs. 17a,b, rainfall differences are shown averaged over the last 60 days of the integrations for 87OI–87C and 88OI–88C. As pointed out in the discussion above, the SST anomalies for these two years are indicative of interdecadal rather than interannual variability. As shown in Fig. 6 and discussed in section 3, the velocity potential difference fields indicated anomalous 200-mb divergence over the Indian Ocean. The rainfall differences appear consistent with this and show positive values on the equator between 60° and 70°E. To the east, between 90° and 100°E, there are consistent regions of negative rainfall differences. Note, however, that over India itself, there are no consistent differences between the two years. Again, this suggests a relatively weak causal relationship between Indian Ocean SST and Indian monsoon rainfall.

5. Conclusions

A set of 90-day integrations has been made with a T42 version of the ECMWF operational model from 1 June 1987 and 1 June 1988, forced with both observed and climatological sea surface temperature (SST) datasets. A brief description of the experiments was given in Table 1. The following conclusions can be made.
With observed SSTs the model was able to simulate with reasonable accuracy the observed interannual variations in the global-scale velocity potential and streamfunction fields and in regional low-level monsoonal flow on seasonal time scales. On the other hand, particularly for the integration for 1987, the model response tended to be larger than suggested by the analyses. While this may indicate the need for further adjustment to some of the model's physical parameterizations, particularly with regard to surface flux formulation (cf. Miller et al. 1991), it is likely that the analyzed fields themselves underestimate the strength of the observed atmospheric anomalies.

When observed SSTs were used, simulated rainfall over the Sahel in 1987 was significantly weaker than in 1988, both in the second and third monthly mean.
periods (July and August). Simulated rainfall rates were verified using station data over the Sahel. These verifying values suggested that the model had an overall dry bias over the Sahel. Similar conclusions were found for the Indian subcontinent, though overall agreement between simulations and observations was not as good. A similar dry bias was found for Indian rainfall.

The sensitivity of the simulations to initial conditions was studied by running an additional integration for 1988 from 2 June. While the simulation of large-scale seasonal mean velocity potential and streamfunction in the tropics was relatively insensitive to atmospheric initial conditions compared with the impact of the imposed SST anomalies, July — August regional rainfall was sensitive to initial conditions, particularly over the east Indian Ocean, Southeast Asia, and the west Pacific. Farther to the west, for example, over the Sahel, regional rainfall was less sensitive to initial conditions. While these results may indicate that intrinsic predictability of month to seasonal rainfall over India at the subdivisional scale may be somewhat limited, considerable caution has to be exercised at this stage. In particular, it is possible that the apparent disparity in the predictability of rainfall and circulation may in fact be a consequence of model deficiencies in which simulated rainfall is insufficiently coupled to dynamical instabilities of the large-scale flow. In any case, more significant extraction of the predictable component of the rainfall fields would be achieved using an ensemble of model integrations.

The impact of SST anomalies in the individual oceans was studied. Interannual differences in rainfall over the Sahel and India between 1987 and 1988 were associated almost entirely with the remote effect of the tropical Pacific SST anomalies. The Indian Ocean, somewhat cooler in 1988 than in 1987, did not appear to have a significant impact on the simulated Indian monsoon rains. Moreover, while the Indian Ocean was
found to be warm relative to climatology in both 1987 and 1988 and the impact of these anomalies produced a consistent increase in rain in the model over the Indian Ocean, the impact of these Indian Ocean SST anomalies on rainfall over India itself was not significant. Overall, these results are supported by observational studies suggesting that Indian monsoon rainfall is correlated more strongly with Pacific SST anomalies than with Indian Ocean SST anomalies (Shukla and Paolino 1983). However, it should be stressed that with a study of only two years the conclusion about the influence of individual oceans may not hold in general.

These integrations have been made with the complete ECMWF land surface parameterizations, and variation in land surface variables (e.g., soil moisture) between the two initial datasets are included. In this study, we have deliberately chosen to focus on the impact of SST itself. However, this does not imply that these initial land surface differences are unimportant in accounting for either the observed or simulated differences in monsoon rains. Indeed, comparison of two runs with identical SSTs, with land surface (and atmospheric) initial conditions at one year apart, suggest that the Sahelian rain in particular may be significantly influenced by land surface conditions.

Studies such as these would benefit from consistent SST and atmospheric climatological datasets and lend support to the requirement for atmospheric reanalysis (Bengtsson and Shukla 1988). Moreover, the results of this study justify more extensive experimentation with a higher resolution version of the model. A more extensive investigation of the predictability of month-to-seasonal rainfall over regions of the tropics appears to be warranted.

REFERENCES


