The effect of a sea surface temperature anomaly on a prediction of the onset of the south-west monsoon over India

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SUMMARY

At the time of the onset of the monsoon in June 1979 the sea surface temperatures in the eastern Arabian Sea were higher than normal. A prediction experiment designed to reveal the effects of this anomaly on the onset is described. This involves two 8-day forecasts made using the same model and the same initial conditions. They differ only in using different sea surface temperatures in the eastern Arabian Sea. A better prediction of the onset is obtained by using the anomalously high temperatures rather than the normal climatological values. In particular, the development of a tropical storm over the Arabian Sea, the strengthening of the Somali jet, and the northward movement of the rainfall over India are all better predicted. The mechanism of the onset in the model is discussed and two important feedback loops are identified, one involving moisture-flux convergence and the other involving the surface fluxes of sensible and latent heat. The most important factor in both loops is the release of latent heat over the Arabian Sea. A linear model is used to show that the onset may be thought of as a response to this heating. The roles of barotropic and baroclinic processes in the onset are also discussed.

1. INTRODUCTION

Seetaramayya and Master (1984) pointed out that the surface temperatures in the eastern Arabian Sea were higher than normal at the time of the onset of the south-west monsoon in 1979. Their paper also contained the suggestion that the development of a tropical storm in this region was aided by the warm anomaly. Many research groups have attempted numerical predictions of the onset in 1979, the year of the First GARP Global Weather Experiment (FGGE), with varying degrees of success (Krishnamurti et al. 1983a). In particular, most attempts did not predict the formation of the tropical storm (also known as the onset vortex). Notable exceptions were the experiments by Krishnamurti et al. (1984) which used an enhanced (envelope) orography, and the results reported by Kershaw (1985a) who used a modified parametrization of convection and radiation. However, even these relatively successful predictions failed to simulate the strength of the circulation of the vortex. As far as can be seen from published results, all of these earlier prediction experiments used climatological estimates of sea surface temperatures, rather than observed values. If the sea surface temperature anomaly did aid the development of the vortex, then it is not surprising that numerical predictions made without the anomaly were unsuccessful. So, in order to test the hypothesis that the anomaly was important, two predictions were made, one with and one without the warm anomaly. Preliminary results from this experiment have been reported, briefly, in Kershaw (1985b). This account contains a more detailed description and a fuller discussion of the results.

2. EXPERIMENTAL DETAILS

A global model of the atmospheric circulation was used for these experiments. It has a horizontal resolution of 2° of latitude and 3° of longitude. It has 11 layers, and the vertical coordinate is sigma (pressure normalized by surface pressure). It contains parametrizations of physical processes, including convection, radiation, large-scale precipitation, and turbulent mixing in the boundary layer. A complete description may be found in Slingo (1985). The model is not normally used for numerical weather
prediction. Its main use is in the study of the general circulation and climate change (e.g. Cunnington and Rowntree 1986).

The initial state of the atmosphere was identical in the two predictions. Both started from an analysis for 12 GMT on 11 June 1979. The (level IIIb) analysis was made by the European Centre for Medium Range Weather Forecasts (ECMWF) using observations specially collected for the FGGE (Lorenc 1981). The same analysis was used for the original study of this case reported in Krishnamurti et al. (1983a). Horizontal and vertical interpolation was carried out from the grid of the European Centre's model to the grid of the model used for the predictions. No special initialization procedure was used after the analysis or the interpolation.

The sea surface temperatures were kept fixed for the eight days of each prediction. Over most of the ocean both predictions used the same surface temperatures: long-term climatological averages appropriate for the time of year. The control experiment also used these climatological temperatures for the eastern Arabian Sea (Fig. 1(a)). The anomaly experiment used temperatures based on the analysis of Sectaramayya and Master (1984) for this region (Fig. 1(b)). Figure 1(c) shows the difference between the surface temperatures used in the experiments. The maximum warm anomaly was 1.7 K. Notice also that the anomaly experiment has a much stronger gradient in surface temperature across the Arabian Sea.

3. RESULTS

During the onset of the south-west monsoon the atmospheric circulation undergoes a considerable change. In the lower troposphere a westerly jet becomes established over the Arabian Sea, and in the upper troposphere an easterly jet forms. The rainfall increases in south-west India and moves northwards along the west coast of the peninsula. In June 1979 these changes occurred more rapidly and somewhat later than is usual (Pearce and Mohanty 1984). Figure 2(a) shows the analysed wind field at the 850 mb level, at 12 GMT on 11 June. At this stage the Somali jet had not yet fully developed. It had a maximum speed of 14 m s⁻¹ a few degrees to the east of the Somali Peninsula. The main branch of the monsoon flow curved south-eastwards from the jet towards another area of strong winds south of Sri Lanka. By 12 GMT on 15 June the Somali jet had strengthened and extended eastwards (Fig. 2(b)). The onset vortex had formed on its northern flank, close to the Indian coast, and south-westerly winds were affecting the southern peninsula. By 12 GMT on 19 June the vortex had strengthened and had moved first north then west. It lay just a few degrees to the east of the Arabian Peninsula (Fig. 2(c)). The jet had strengthened further and moved northwards, and the strong westerly winds were affecting the entire west coast of the peninsula.

In the control experiment the jet did not extend eastwards quickly enough. Neither did it attain sufficient strength. By day 4 of the prediction (12 GMT 15 June, not shown) a trough had formed on the north-eastern edge of the jet, but no vortex had formed. By day 8 (12 GMT 19 June) a weak vortex had formed but it remained stationary (Fig. 3(a)). The jet had attained a maximum speed of only 24 m s⁻¹, compared with the analysed speed of 29 m s⁻¹. The strong westerly winds were still only affecting the southern part of the peninsula. Thus the control prediction failed to capture the full intensity of the changes that occurred during the onset of the monsoon.

The anomaly experiment produced a much better prediction of these changes. By day 4 (not shown) the jet had extended further eastwards and the vortex had already formed. By day 8 (Fig. 3(b)) the vortex had strengthened and moved north-westwards. The jet had attained a maximum speed of 29 m s⁻¹, identical to the observed maximum
speed, and the entire west coast of the peninsula was under the influence of the strong winds. The prediction is not perfect, but it is very good for an 8-day forecast. Note, however, the following errors: the area of the jet is too extensive; the centre of the vortex is not far enough north or west; the orientation of the jet is westerly in the prediction.
but south-westerly in the analysis; the westerly flow into the Bay of Bengal is too weak. Nevertheless, in several respects this is the best prediction for this case study published so far. In particular, it is a very good forecast of the formation of the onset vortex and the increase in kinetic energy over the Arabian Sea (4°S to 20°N, 50°E to 70°E; Fig. 4).
Figure 3. Wind vectors and isotachs at 850mb for day 8, 12 GMT 19 June 1979; from (a) control forecast, (b) anomaly forecast. Units m s$^{-1}$, contour interval 5 m s$^{-1}$.

Figure 4. Kinetic energy per unit mass at 850mb, averaged over Arabian Sea. Full line—ECMWF FGGE IIIb analysis; dashed line—control forecast; crosses—anomaly forecast.
The kinetic energy increased more in the anomaly experiment than in the control experiment, and the total increase in the anomaly experiment was similar to the observed increase. However, over the last four days of the experiment the predicted increase actually exceeded the observed increase.

An increase in kinetic energy also occurred at 200 mb. The analysis on 11 June (Fig. 5(a)) shows a south-easterly flow over the Arabian Sea. A westerly flow is evident over northern India; this is a southern branch of the subtropical jet. By 15 June (not shown) the easterly jet had strengthened and acquired a northerly, cross-equatorial, component. By 19 June the easterly jet had strengthened further and the jet core was over southern India (Fig. 5(b)). (This northward movement of the jet core happened between 18 and 19 June. On 18 June the jet core was still over the equator and was stronger than on 19 June.) Over northern India the subtropical jet had receded northwards leaving a weak easterly flow. The westerly jet itself was strongly anticyclonic. Both experiments (see Figs. 6(a) and (b)) predicted the strengthening of the easterly jet and the increase in cross-equatorial flow. Neither experiment predicted the replacement of

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**Figure 5.** Wind vectors and isotachs at 200 mb from ECMWF FGGE IIb analysis. (a) 12 GMT 11 June 1979; (b) 12 GMT 19 June. Units m s\(^{-1}\), contour interval 10 m s\(^{-1}\).
the weak westerlies by weak easterlies over northern India. The anomaly experiment was superior in that it predicted a stronger easterly jet, more like the observed jet (although for 19 June the prediction was a little too strong).

The anomaly experiment also produced a better prediction of precipitation than the control (see Figs. 7(a) and (b)). By the end of the forecast period the anomaly experiment had significant amounts (>20 mm d\(^{-1}\)) over much of the west coast of the peninsula, whereas in the control experiment the precipitation was still restricted to the southern part of the peninsula. The predicted amounts in the anomaly experiment were comparable with those observed over the west coast of India at the time. (Krishnamurti et al. (1983b) give an analysis of daily rainfall during June 1979 using conventional and satellite data.) It is not clear whether the increased precipitation over the Arabian Sea in the anomaly experiment is correct. The predicted maxima are greater than those estimated by Krishnamurti et al. However, their estimates may be too low. A more recent and more detailed analysis by Martin and Howland (1985) shows rainfall maxima exceeding 300 mm d\(^{-1}\) during the onset. If this later analysis is correct, then the anomaly experiment does have a more realistic distribution of precipitation over the Arabian Sea than the control.
These results support the initial hypothesis that the sea surface temperature anomaly was instrumental in the development of the onset vortex. Moreover, the use of more accurate sea surface temperatures improved the prediction of other aspects of the onset of the monsoon. The strengthening of the Somali jet and the easterly jet, and the northward movement of the rainfall were all predicted more accurately in the anomaly experiment than in the control. This is a very good example of the beneficial impact that the use of observed (rather than climatological) sea surface temperatures can have on numerical weather prediction for the tropics.

4. MECHANISM OF THE ONSET

(a) Role of surface fluxes

Whilst the result of the experiment is interesting in itself, so is the way in which the sea surface temperature anomaly produces the change in the prediction. The results increase our understanding of the mechanism of the onset in the model. The most obvious effect of the anomaly is to increase the fluxes of heat and moisture into the atmosphere from the sea. There are three components: sensible heat, net radiation, and latent heat. Time series of the differences in these fluxes averaged over the area of the anomaly (4°N to 20°N, 60°E to 76°E) are shown in Fig. 8. The major changes in the prediction are probably caused, indirectly, by these increased fluxes. Several points about the fluxes
are noteworthy. Firstly, the increase in latent heat flux is much bigger than the other components, but its energy can only be released when precipitation occurs. Secondly, the increases in sensible and latent heat fluxes remain steady for the first few days but increase dramatically thereafter; this suggests that a feedback is occurring. Lastly, the increase in radiative heat flux remains small and steady throughout; hence radiative effects are probably not important. The increased fluxes warm and moisten the atmosphere. However, these changes in the temperature and moisture structure are too small to produce the large changes in the prediction directly. They are also restricted to the boundary layer, whereas the circulation changes affect the entire troposphere.

The increased fluxes increase the precipitation, and it is probably the increased release of latent heat which causes the circulation changes. Figure 9 shows the time series
of the difference in the rate of precipitation \((P)\) averaged over the same area as the fluxes. The difference in the rate of evaporation \((E, \text{ proportional to latent heat flux})\) is plotted on the same graph for comparison. The increase in precipitation is largely convective and it is promoted by the warming and moistening of the atmosphere over the anomaly. Note that the precipitation increase exceeds the evaporation increase, except on day 1. Initially, all the moisture necessary to sustain the increase in precipitation is provided by increased evaporation. The total moisture content of the region is increasing and the moisture-flux convergence is positive too. The difference in evaporation stays near its initial value during the first three days of the experiment. During this time the precipitation difference is increasing rapidly (by 400%). So although the initial increase in precipitation can be explained by the increase in evaporation alone, the subsequent increase must involve some other mechanism. In fact the precipitation increase is sustained by increased moisture-flux convergence. The difference between the precipitation and evaporation curves in Fig. 9 is a measure of this flux convergence (neglecting the small increase in moisture content). Associated with the initial rise in precipitation there is an increase in latent heat release. Some of this warms the atmosphere but most is compensated by increased adiabatic cooling due to ascent. The increased ascent is associated with convergence of mass at low levels and divergence of mass at high levels. Associated with the mass convergence is convergence of moisture flux. This enhances the moisture supply and promotes increasing precipitation, completing the feedback loop. The same mechanism is operating from days 5 to 7; the moisture-flux convergence \((P - E)\) continues to grow through this period. However, another mechanism is acting to increase the precipitation. From day 4 onwards the surface fluxes (and in particular the evaporation) are increasing because the wind speed is increasing. So part of the increase in \(P\) is being forced by the increase in \(E\). It remains only to show that the precipitation is causing the increase in wind speed to complete another feedback loop.

(b) Role of precipitation

The simple model developed by Gill (1980) is a useful aid to understanding the dynamical effects of precipitation. It assumes a simple baroclinic structure with maximum heating and vertical velocity, and one reversal of winds, in mid-troposphere. An example is shown in Fig. 10. The contour encloses the imposed heating region and the arrows show the equilibrium response of the low-level wind field. (Note that the upper-level response is obtained by reversing the arrows.) With the particular value of the frictional dissipation parameter used in this example, equilibrium is achieved in less than one day. The solution is obtained analytically using equations which have been linearized about a basic state at rest. It shows that the circulation responds to the heating by generating a strong westerly flow into the heating zone and a cyclonic circulation to the north-west. Weak easterlies are induced to the south and east. The forcing and the response in the numerical experiment show the same pattern. Figure 11 shows the difference fields at day 5 (12 GMT 16 June) for the precipitation and the wind at 850 mb. The similarity between Figs. 11 and 10 suggests that the circulation change is induced by the heating associated with the precipitation. The difference fields at day 8 (12 GMT 19 June) are shown in Fig. 12. By this time the cyclonic anomaly has started to move northwards and a prominent maximum has appeared in the precipitation field in the south-west quadrant of the depression. Even after eight days the circulation anomaly is largely restricted to the Arabian Sea, and it still looks like Fig. 10, even though its magnitude is such that nonlinear effects might be expected to be important. The circulation anomalies are similar in amplitude to those predicted by the linear model. However, the response of the linear model is sensitive to the precise value of the frictional dissipation parameter used, so this
Figure 10. Equilibrium response of low-level wind (shown by arrows) to a region of heating (enclosed by contour) north of equator \((y = 0)\) in a linear model. One unit of length is approximately 10° latitude. (After Gill (1980), courtesy of P. J. Phillips.)

Figure 11. Difference charts, anomaly minus control forecast at 12 GMT 16 June 1979 (day 5). (a) Rate of precipitation averaged over preceding day. Units \(\text{mm d}^{-1}\), contour interval \(20 \text{mm d}^{-1}\). (b) Vector wind at 850 mb. Contours show magnitude of difference. Units \(\text{m s}^{-1}\), contour interval 5 \(\text{m s}^{-1}\).
may be fortuitous. Nevertheless, the linear model gives the link which completes the second feedback loop identified above; the latent heat release generates the circulation changes and the increased winds generate the increases in the fluxes. Figure 13 summarizes the two feedback loops, loop 1 involving the moisture-flux convergence and loop 2 involving the surface fluxes. The diagnostics presented so far show that these two mechanisms are important; other mechanisms could also be operating. In particular, instabilities might be aiding the formation of the onset vortex. This problem is discussed in subsections 4(d) and 4(e).

Figure 12. Difference charts, anomaly minus control forecast at 12 GMT 19 June 1979 (day 8). (a) Rate of precipitation averaged over preceding day. Units mm d\(^{-1}\), contour interval 20 mm d\(^{-1}\). (b) Vector wind at 850 mb. Contours show magnitude of difference. Units m s\(^{-1}\), contour interval 5 m s\(^{-1}\).

Figure 13. Mechanism of onset in model, showing feedback loops 1 and 2 identified in text.
(c) Nonlinear effects

A more detailed comparison of the predicted anomalies with the results from the linear model reveals some interesting differences. At 850 mb, the linear model represents the effect of precipitation in the prediction model quite well. However, the nonlinear horizontal advection terms at 850 mb are just as large as the other terms in the momentum equation. The fact that they do not appear to alter the nature of the response is in agreement with the findings of Gill and Philips (1986). The vertical structure of the response in the prediction model is similar to that assumed in the linear model; the circulation difference at 200 mb (Fig. 14) shows anticyclonic flow above the low-level cyclone. Note, however, that the high-level flow is not simply a reversal of the low-level flow. The high-level circulation anomaly extends into the southern hemisphere. The strongest easterly flow is to the south of the precipitation anomaly, rather than immedi-

![Figure 14](image_url)

Figure 14. Difference (anomaly – control forecast) in wind vector at 200 mb at 12 GMT 19 June 1979 (day 8). Contours show magnitude of difference. Units m s⁻¹, contour interval 5 m s⁻¹.

ately above it. Also the horizontal scale of the anomaly is bigger, extending from the Red Sea to the Bay of Bengal. These factors suggest that nonlinear effects might be more important at this level. To some extent the advective effects of the north-easterly flow in the control experiment explain this response. A similar response is seen in Gill’s model when a non-zero mean flow is imposed (Phlips and Gill 1987). The high-level flow is strongly divergent, whereas the low-level flow is convergent; moreover, the maxima are located differently. At 850 mb the major anomalous convergence centre is around 10°N 70°E (Fig. 15(a)), in the south-east quadrant of the low-level circulation anomaly (Fig. 12(b)). At 200 mb the major centre of anomalous divergence is north-west of this (Fig. 15(b)), associated with the maximum rainfall anomaly in the south-west quadrant (Figs. 12(a) and (b)). As the divergence field may be regarded as forcing the vorticity at each level, this difference in location may explain the greater westward extent of the circulation anomaly at 200 mb. It does not explain the greater southward extent.

Further insight into the upper-level response is provided by the non-divergent component of the circulation anomaly at 200 mb (Fig. 16). This shows two anticyclonic centres, one over the Arabian Peninsula and the other to the south-east. The feature over Arabia agrees with the expected response of the linear model to the heating in the
Figure 15. Difference (anomaly – control forecast) in divergence at (a) 850 mb; (b) 200 mb; at 12 GMT 19 June 1979 (day 8). Units $10^{-5}$ s$^{-1}$, contour interval $10^{-5}$ s$^{-1}$, zero contour omitted.

Figure 16. Difference (anomaly – control forecast) in non-divergent wind vector at 200 mb at 12 GMT 19 June 1979 (day 8). Contours show magnitude of difference. Units m s$^{-1}$, contour interval 5 m s$^{-1}$. 
south-west quadrant of the depression. This is generated only in the last few days of the experiment. The southernmost high is generated on day 1 and moves south-eastwards as the integration progresses. This initial circulation anomaly at 200 mb is well predicted by the linear model; but after a few days the detailed development diverges from the linear prediction. So the linear model explains the circulation anomaly at low levels better than that at high levels. Sardeshmukh and Hoskins (1985) came to the same conclusion in their study of the 1982–83 El Niño. Incorporating a non-zero mean wind in the linear model helps to explain some of the effects observed. However, there are still some details which are not explained. Thus the experiment is a useful demonstration of the potential and limitations of the linear model.

(d) Role of barotropic processes

The discussion so far has implied that the formation of the onset vortex is forced by latent heat release over the Arabian Sea. However, previous research has suggested that barotropic instability was important in the formation of this vortex (Krishnamurti et al. 1981). Does the same mechanism operate in the numerical model? To investigate this the diagnostic

$$E = (v'^2 - u'^2, -u'v')$$

is used. $u$ and $v$ are the westerly and southerly components of the wind, and primes denote departures from time means (overbars). Hoskins et al. (1983) give details of the applications of $E$ and the approximations involved in using it. $E$ enables the interaction between the time-mean flow and the transient eddies to be quantified. For example, convergence of $E$ implies easterly forcing of the time-mean flow by the eddies. The quantity $\overline{u} \nabla \cdot E$ represents the exchange from the kinetic energy of the eddies to kinetic energy of the mean flow. (More precisely, it represents the local contribution to this barotropic exchange.) $\overline{E}$ and the zonal component of the mean flow for the ECMWF analyses are shown in Fig. 17(a). The averaging period is the eight days of this study. Note that $E$ has a large magnitude on the northern flank of the jet and that the vectors point south-west and converge where $\overline{u}$ is positive. In Fig. 17(b) $\overline{u} \nabla \cdot E$ is strongly negative when averaged over the Arabian Sea, implying that the transient eddies grew at the expense of the mean flow.

Krishnamurti et al. (1981) showed that the necessary condition for barotropic instability was satisfied by the longitudinally averaged flow in this region at this time. They also diagnosed that energy was transferred from the zonal flow to the vortex. At first sight this E-vector analysis confirms that diagnosis. However, there are several difficulties over diagnosing barotropic processes near an extending jet. The basic flow is zonally varying, so the theory of instability of a zonally uniform flow is of doubtful relevance. Although the E-vector approach eliminates this problem it raises another. It is difficult to define eddies in a way which isolates them from the jet. The use here of deviations from the time mean is only partially successful. Because the jet is extending, $u'^2$ is large in the jet exit, so $\overline{E}$ has a substantial component towards the west. It would have even if no vortex formed. Thus the barotropic energy conversion diagnosed here is, in part, an artefact of the definition of eddies. Furthermore, the existence of barotropic energy conversion does not necessarily imply that barotropic instability is present. Nevertheless, the E-vector analysis presented here is consistent with the growth of the onset vortex by barotropic processes.

The same diagnostics are shown for the control prediction in Fig. 18. The arrows are drawn to the same scale as in Fig. 17 but they are much smaller. The energy conversion term in Fig. 18(b) is only weakly negative, indicating that barotropic processes are
Figure 17. (a) $\mathbf{E}$ (units m/s$^{-2}$) and isotachs of $\mathbf{u}$ (units m/s$^{-1}$, contour interval 5 m/s$^{-1}$); (b) $\mathbf{u} \cdot \mathbf{E}$ (units $10^{-3}$ m$^2$/s$^{-2}$, contour interval 0.5 x $10^{-3}$ m$^2$/s$^{-2}$); averaged over 8 days from 12 GMT 12 June 1979 to 12 GMT 19 June, calculated from ECMWF FGGE IIIb analyses at 850 mb.

not very effective in the control experiment. This is consistent with the development of the onset vortex being underestimated in that prediction. For the anomaly prediction (Fig. 19) the $\mathbf{E}$ vectors are more like those diagnosed from the analysis (Fig. 17), and the energy conversion term is more negative. These results are consistent with barotropic processes being important in the growth of the vortex in the model, but further research is needed to verify this.

(e) Role of baroclinic instability

The relative importance of barotropic and baroclinic processes in the formation of monsoon depressions has attracted much attention (Moorthi and Arakawa 1985; Lindzen et al. 1983; Das 1986). It is of interest, therefore, to ask whether baroclinic instability is also important in the formation of the onset vortex. To shed some light on this a Q-vector analysis has been performed. Q is defined as

$$\left( - \frac{g}{\theta_0} \frac{\partial V}{\partial \theta_0} \cdot \nabla \theta, - \frac{g}{\theta_0} \frac{\partial V}{\partial \theta_0} \cdot \nabla \theta \right)$$
Figure 18. (a) $\overline{E}$ (units $m^2s^{-2}$) and isotachs of $\overline{u}$ (units $m s^{-1}$, contour interval $5 m s^{-1}$); (b) $\overline{u} \overline{\nabla \cdot E}$ (units $10^{-4} m^3 s^{-3}$, contour interval $0.5 \times 10^{-4} m^3 s^{-3}$); averaged over 8 days from 12 GMT 12 June 1979 to 12 GMT 19 June, calculated from control forecast at 850 mb.

where $V_g$ is the geostrophic wind, $\theta$ is the temperature at the same level, and $\theta_o$ is an arbitrary reference temperature. $Q$ is a convenient diagnostic because it only requires data for a single isobaric level and it only involves horizontal derivatives. It can be used to diagnose vertical velocity; convergence of $Q$ implies ascent and divergence implies descent. (This assumes that the quasi-geostrophic approximation holds and that $f$ is constant. Hoskins et al. (1978) give details.) This diagnosis reveals areas of ascent and descent which are forced dynamically, by the geostrophic flow rather than by the release of latent heat. Figure 20 shows the differences in $Q$ and its divergence between the two predictions at 12 GMT 18 June. This chart is for 500 mb but similar features can be seen at other levels, and at other times. In the region just west of India where the precipitation anomaly for the day beginning 12 GMT 18 June (Fig. 12(a)) has a maximum, $Q$ is divergent, implying descent. But this is the region of maximum mass convergence at 850 mb which we have suggested is the primary forcing region for the development of the onset vortex. It is also a region of positive temperature anomaly (not shown). If baroclinic instability were aiding this development then dynamically forced ascent would be found in this region. As it is not, such instability is probably not important in the development of the vortex.
Figure 19. (a) $\mathbf{E}$ (units $m^3 s^{-2}$) and isotachs of $u$ (units $m s^{-1}$, contour interval $5 m s^{-1}$); (b) $u \nabla \cdot \mathbf{E}$ (units $10^{-3} m^2 s^{-3}$, contour interval $0.5 \times 10^{-3} m^2 s^{-3}$); averaged over 8 days from 12 GMT 12 June 1979 to 12 GMT 19 June, calculated from anomaly forecast at 850 mb.

Figure 20. Difference (anomaly – control forecast) in $Q$ (arrows), $\nabla \cdot Q$ (contours: units $10^{-16} s^{-1} m^{-1}$, contour interval $5 \times 10^{-16} s^{-1} m^{-1}$) at 500 mb at 12 GMT 19 June 1979 (day 8).
However, \( Q \) is convergent near 15\(^\circ\)N 60\(^\circ\)E, and so ascent is being forced dynamically in the south-west quadrant of the circulation anomaly, where a secondary maximum in anomalous precipitation occurs (see Fig. 12). Rainfall maxima are frequently observed in the south-west quadrant of monsoon depressions and Douglas (1985) has suggested that ascent (and hence precipitation) is forced dynamically in that quadrant. The results from this study suggest that this is also true for the onset vortex. The occurrence of this feature might explain the north-westward movement of the vortex, for the linear model suggests that heating in such a region will force cyclonic vorticity to the north-west. This suggestion is, however, purely speculative.

5. **Importance of Sea Surface Temperature Anomalies**

In this case study it was important to specify the correct sea surface temperatures to get a good prediction of developments in the tropics. When else might this be important? Sea surface temperature anomalies can probably contribute to the development of severe tropical storms, for it is well known that these storms develop preferentially over warmer waters. In fact similar feedback mechanisms to the two identified in section 4 are important in their development (Gray 1982). An important factor is the occurrence of strong winds over the anomaly which amplify the response. Monsoon disturbances will therefore be particularly sensitive to sea surface temperature anomalies, because they develop near the low-level jet. The normal gradient of sea surface temperature in the Arabian Sea will also amplify the impact of anomalies in the east, because the prevailing winds flow up the gradient, from cooler to warmer seas. So it seems likely that the onset of the monsoon in other years will be sensitive to such anomalies. In particular, in years when the onset is late, higher than normal temperatures might be expected in the Arabian Sea because of the extra solar heating available before the cloudiness increases at onset. The delay in the cooling associated with the onset will also help. The onset of the monsoon was also late in 1983. During this year the sea surface temperature anomaly in the Arabian Sea was +2 K just prior to the onset. This suggests that the onset in 1983 would be a good case on which to perform a prediction experiment similar to the one reported here. This preliminary research suggests that the following phenomena may be correlated and should be the subject of further study: late onset of the monsoon, warm anomaly in the Arabian Sea, sudden onset of the monsoon (rapid northward progress), and occurrence of an onset vortex. Some of these relationships may have predictive value.

6. **Problems**

The cooling of the Arabian Sea during the onset casts some doubt on the results reported here. For although the sea surface temperatures imposed in the anomaly experiment are realistic for 11 June, it has been suggested that they are too high for 19 June. To some extent this is true; the model is being forced by too strong an anomaly. This may explain why the growth of kinetic energy over the Arabian Sea is overestimated in the last half of the anomaly experiment (Fig. 4). However, the error is not very great; even on 19 June ships in the eastern Arabian Sea were reporting temperatures in excess of 303 K. The cooling of the Arabian Sea starts in the west and south and then spreads northwards and eastwards. The cooling in the northern part occurred only after the period of this study (see Krishnamurti 1985, Fig. 9(d)). The warm anomaly remained in the north and the anomaly imposed in the model is an underestimate in this region. So, although the imposed surface temperatures are too high in the south they are too low in
the north. These errors will tend to act in opposition. In fact, incorporation of this northern anomaly might even improve the prediction by inducing northward movement of the jet and the vortex.

7. CONCLUSION

An example of a forecast for the tropics has been presented in which a sea surface temperature anomaly had a substantial impact; such anomalies will also be important in other cases. Several major weather forecasting centres already use analysed sea temperatures in their operational prediction models; this study confirms that it is important to do so. The main conclusion of this study is that the warm anomaly in the Arabian Sea did affect the onset of the monsoon. It increased the rate of generation of kinetic energy, speeding up the onset and causing rapid northward progression of the monsoon. It also promoted the formation of the onset vortex, by enhancing the release of latent heat and perhaps by promoting barotropic instability. Latent heat release in the Arabian Sea played a very important part in the mechanism of the onset in the prediction model. It had a crucial role in the two feedback loops which were identified as dominant, one involving moisture-flux convergence, the other involving the surface fluxes. The same mechanisms probably operated in the real atmosphere, but without more observations (especially of surface fluxes) this cannot be verified. Various diagnostic techniques have shed considerable light on the workings of the prediction model in this case. Firstly, Gill's linear model helped to identify the crucial connection between the release of latent heat and the response of the circulation. Secondly, the diagnostics E and Q, developed by Hoskins, helped to clarify the roles of barotropic processes and baroclinic instability in the formation of the onset vortex.

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