Reply

Douglas G. Hahn and Syukuro Manabe

Geophysical Fluid Dynamics Laboratory, NOAA, Princeton University, Princeton, N. J. 08540

9 April 1976 and 7 July 1976

1. Introduction

In their critical remarks, Sadler and Ramage (hereafter are referred to as SR) identified some unrealistic features in the structure and the evolution of the monsoon circulation which were produced by our global model of the atmosphere. Most of these features were acknowledged in Hahn and Manabe (1975, hereafter referred to as HM), and the reader is directed to the discussion in the original paper as indicated below:

1) Surface pressure and wind; misplacement of the surface low to Tibet (p. 1519, ¶1). It is pointed out in HM (p. 1519, ¶1; p. 1526, ¶1) that the small-scale features of the sea level pressure distribution over Tibet are suspect because of the large extrapolation involved in sea level reduction. (Also see Corrigendum, *J. Atmos. Sci.*, **33**, in press.)

2) Surface pressure and wind; trough over tropical depression track in western Pacific (p. 1518, ¶3; p. 1519, ¶4).

3) Rainfall; too little at Bombay (p. 1519, ¶6).

4) Rainfall; too little in Ganges Valley (p. 1519, ¶7, 8).

5) Upper tropospheric flow; anticyclone over western Pacific (p. 1520, ¶3).

6) Upper tropospheric flow; location of center of anticyclone (p. 1520, ¶2).

7) Vertical distributions; sinking motion and little precipitation along west coast of India and Burma (p. 1519, ¶6, 7, 8).

8) Monsoon onset; later than average (p. 1524,¶2; p. 1525, ¶2; p. 1537, ¶6).

9) Surface pressure and wind; lack of fresh southwesterlies during May; manifestation of item 8.

10) Monsoon onset; rainfall increases dramatically from April to May not May to June; manifestation of item 8.

Unfortunately, we have not been able to pinpoint causes for many of the difficulties in the simulation.

However, we believe that many of these difficulties are related and deserve further discussion. In this response, we shall attempt to identify some of the interrelations, discuss their possible causes and indicate our future strategy for overcoming them. In addition, we shall attempt to remove some of the misunderstanding from which some of their criticism originates.

2. Monsoon trough

Some of the critical comments of SR are related to the failure of the model to reproduce the monsoon trough with sufficient intensity. As is well known, monsoon depressions are partly responsible for the statistical intensification of the monsoon trough and heavy rainfall over the Ganges Valley. They move northwestward from the Bay of Bengal into the plain region. Though monsoon depressions in the M-model[1] atmosphere often form over the northern portion of the Bay of Bengal, they fail to move westward (HM, p. 1519). Instead, most of the rainfall along the monsoon trough results from rather weak disturbances, which are generated *in situ* and slowly move eastward. Accordingly, the rate of precipitation in this region is much less than the observed rate as noted by both SR and HM (p. 1519). Analysis of the M-model, which is in qualitative agreement with recent analysis by Krishnamurti *et al.* (1975b), reveals that monsoon depression systems are not limited to the lower troposphere, but extend to the upper troposphere. Therefore, the failure of the model to simulate the penetration of easterly flow to the mid-troposphere over northern India (Fig. 4.3) may imply the inaccurate simulation of the steering currents (HM, p. 1520). This may be responsible for the movement of these depressions in the wrong direction in the M-model. In addition, HM suggested that the failure of the model to accurately resolve the effects of the steep southern slope of the Himalayas (p. 1539) and Burmese Mountains (p. 1519) may be responsible for the weakness of the model monsoon trough. Currently, we are planning to repeat similar experiments after improving the resolution for finite differencing in the general region of southern Asia.

SR pointed out that the extension of the monsoon trough over the Bay of Bengal coincides with the area of warmest sea surface temperature. In HM, it was pointed out (p. 1532) that the tropical rainbelt (ITCZ)[2], which extends from equatorial Africa across the southern tip of Indian peninsula toward Indonesia, is located in the general area of the warmest sea surface temperatures. However, the closer inspection of the model results (Fig. 5.2 and Fig. A1) reveals that, over the Bay of Bengal, the monsoon trough rather than the above-mentioned rainbelt appears to coincide with the region of warm sea surface temperatures along the east coast of India as pointed out by SR. It should be emphasized, however, that this coincidence may not necessarily imply the causal relationship between warm sea surface temperatures and maintenance of the monsoon trough in view of the important role which the Himalayas play in maintaining the monsoon trough as discussed in HM (pp. 1526, 1531, 1538, 1539). For another viewpoint see Hobbs (1974).

3. Onset and evolution

SR noted the lack of "fresh" southwesterlies over the Bay of Bengal in May in the M-model atmosphere. They also point out that rainfall does not increase dramatically from April to May in the M-model. These features are consistent with the late arrival of monsoon in the M-model (discussed on p. 1524, ¶2; p. 1525, ¶2; and p. 1537, ¶6 of HM). According to the results of our analysis, the dates of monsoon onset in the last two years of the numerical time integration of the M-model are significantly different from one another (HM, p. 1524), i.e., the onset of the final year is approximately two weeks later than that of the previous year. It is not clear whether this difference results from an interannual variation or a systematic trend in the time-integration of the M-model. It is interesting to note, however, that mean surface temperatures over Tibet and parts of Eurasia during the last summer of the M-model integration are significantly lower than those of the preceding summer. As suggested by results from other numerical experiments (at GFDL, not yet published), the monsoon's onset date might be related to the surface temperatures of southern Asia. Currently, we are planning to investigate how the onset date is affected by the extent of snowcover over the Eurasian continent and the Himalayas.

For the M-model's July mean 190 mb flow field, Fig. 3.4, both SR and HM (p. 1520) noted that the anticyclone over Tibet is located too far south. Inspection of the 200 mb flow for June and July (Sadler, 1975) indicates that the center of the Asian anticyclone in the M-model (July, Fig. 3.4) being extended southward from the climatological mean is also consistent with the notion of a late onset and evolution of the monsoon circulation in the M-model. For the early days of July (Fig. 3.7), easterly flow near 20°–30°N was not fully developed and 200 mb flow patterns look somewhat like the June mean (Sadler, 1975), when the center of the Tibetan high is shown approximately 5° farther south. In the August mean of the M-model simulation (which was time-integrated since publication of HM) shown in Fig. 1, the easterly jet is more intense over the northern Indian Ocean and the center of the Tibetan high is located very near to 30°N.

There are some significant differences between July, August simulated 190 mb flow patterns and the

[1] In HM, the general circulation models with and without mountains are identified as the M-model and NM-model, respectively.

[2] For the M-model simulation, HM identified the ITCZ as a distinct feature apart from the monsoon trough. See HM (pp. 1532, 1535, 1536, 1539) for a more detailed description.

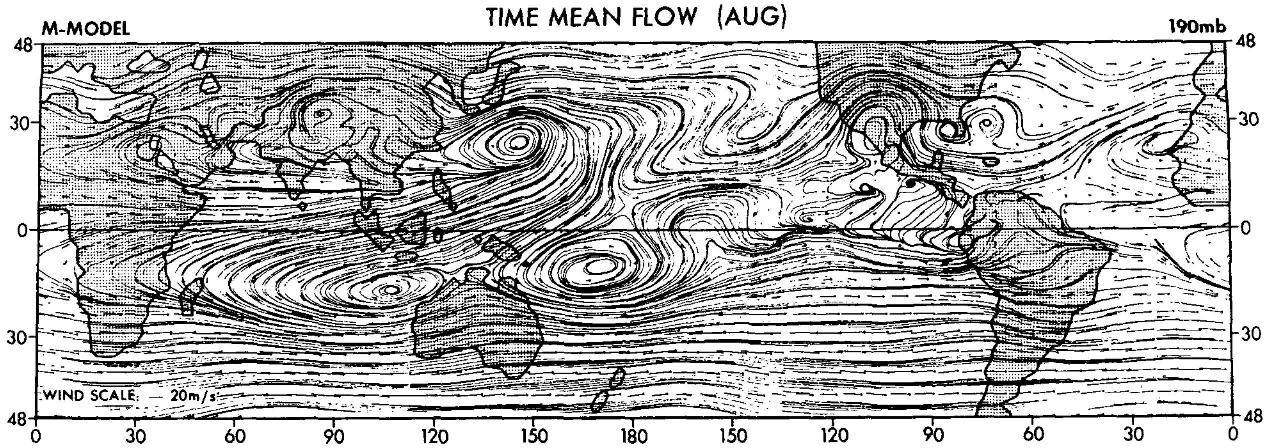

FIG. 1. August mean vectors and streamlines at the 190 mb level of the M-model.

climatological means published by Sadler (1975). However, many of these differences show up to some extent in one or both sets of mean 200 mb flow patterns derived for the individual years of 1967 and 1972 by Krishnamurti *et al.* (1973) and Krishnamurti *et al.* (1975a). The following comparisons are of interest:

1) In August the ridge tilts sharply equatorward from Tibet to Saudi Arabia in the M-model and during 1972, but not in Sadler's climatology or in July 1972.

2) In August Sadler's climatology shows the easterly jet extending across the Atlantic and South America over the eastern Pacific. It extends not as far westward in the M-model and in 1967, and only to the mid-Atlantic in 1972.

3) In Sadler's climatology the mid-Pacific trough (MPT) can be identified to extend from the northeastern Pacific westward to 140°E in June and almost to Asia (125°E) in July and August. During 1972 the structure of the MPT is more variable from June through August, and for the June-July-August mean of 1967, the MPT extends farther equatorward (more like the M-model) and cannot be identified west of 180°E.

4) Likewise, the June–August evolution of the subequatorial ridge over Central America and westward during 1972 is much more erratic than in the climatological evolution. By July, easterlies are found at nearly all longitudes at 5°N in Sadler's atlas, whereas, at 5°N in the M-model and during 1967 and 1972, westerlies are found at some longitudes.

In addition to reflecting a measure of interannual variability in the evolution of summertime flow patterns, some of these differences may be reflecting variations in data availability (or unavailability) and analysis procedures. Nevertheless, even though the large-scale 200 mb flow patterns are similar, differences in details between individual years of observed data and climatology are not quite, but almost as large as differences between the M-model's simulation and climatology.

HM have shown that the springtime northward shift of the position of maximum westerlies is more abrupt when mountain effects are included in their model (cf. Figs. 3.7 and 5.3). SR seem to indicate they prefer the more gradual northward shift of the subtropical jet found in the NM-model atmosphere for comparing with climatology (their Fig. 3). With the onset date varying from 1 May to 3 June (Orgill, 1967), 14-year averages of monthly mean winds (their Fig. 3) are hardly suitable for attempting to discuss the abruptness in the time evolution of the 200 mb wind field.

SR point out that climatology (mean conditions derived by averaging many years of observations) indicates, contrary to the M-model simulation, that there is a minimum of convergence and cloudiness at the equator during April, May and June with increasing convergence for July. Looking at observations of the position of the ITCZ over the Indian Ocean compiled for individual days of April–June, 1967–68, Hobbs (1974) reveals that the structure of the ITCZ is highly complex and constantly changing during these months, with axes of instability (determined from cloud pictures) sometimes very close to, if not on the equator. Likewise in Fig. 3.6 the M-model's simulated surface flow patterns are constantly changing from one 5-day mean to another, with regions of convergence usually found to lie slightly off the equator to the north and to the south. In short, the time-varying aspects of the convergence patterns near the equator in observations as well as in the M-model are very large. With gradients in climatological cloud data being very weak near the equator (SR, Fig. 2), deviations from climatology of the M-model's convergence patterns near the equator are probably no larger than those found within an individual year of observations. Furthermore, during April, May and June, the M-model simulates a rainfall maximum meandering about 5°S (Fig. 5.4) in qualitative agreement with Fig. 2 of SR. During June, both time sections indicate a second maximum near 15°N. In July the former (near 5°S) weakens or shifts southward, while the latter (near 15°N) intensifies and shifts northward.

No claim was made by HM that all of the features of the break monsoon are duplicated by the NM-model and during 5–9 July of the M-model. The comparison of the meridional circulation and precipitation patterns of the NM-model with those of the break monsoon was made for the purpose of contrasting these fields with those of the M-model. It is felt that more convergence, rainfall and upward motion along and to the south of the monsoon trough in the M-model's monsoon circulation resemble active monsoon conditions. To the contrary, absence of the monsoon trough in the NM-model, coupled with downward motion, northerly flow near the surface and less rainfall over central India, resembles break monsoon conditions to some extent. Even a stronger easterly jet in the NM-model (Fig. 4.3) is somewhat consistent with the notion of break monsoon conditions persisting in the NM-model (Ramamurthy, 1969).

4. M-model vs NM-model

SR have pointed out many features where they feel the M-model is inferior to the NM-model simulations. For example, they seem to have been disturbed by the fact that the easterly jet is stronger in the NM-model than in the M-model (Fig. 4.3). Here one should note that the intensity of the easterly jet in the M-model (maximum averaged from 80°–95°E: 28.9 m s^{-1} in July, 30.8 m s^{-1} in August) compares quite well with July ($\sim$32 m s^{-1})[3] and August ($\sim$30 m s^{-1})[3] mean observations of Ramage and Raman (1972). SR have pointed out that they feel the results of Murakami *et al.* (1970) are more realistic because they simulated a much weaker easterly jet when mountains were taken out of their 2-D model. For support of this statement, SR infer the intensity of the easterly jet for the hypothetical no-mountain monsoon by observing the regimes equatorward of South China and Australia. In view of the many factors which make the South Asian monsoon circulation unique, it is felt that mountain effects cannot be satisfactorily isolated from other effects (effects of land-sea distribution, etc.) in such comparison studies of mountain and no-mountain regimes of the monsoon circulation. Moreover, there is no reason to believe that the intensity of the easterly jet observed equatorward of South China and Australia is not affected by the large-scale orography of Asia.

In regions where local climatology is dominated by the effects of tall narrow mountain ranges, SR have made many detailed evaluations of the M-model's rainfall and vertical motion fields. It was pointed out by HM (p. 1517, Fig. 2.1) that the M-model's horizontal resolution was too coarse to resolve tall narrow mountain ranges. As a result, detailed comparison with climatology reveals the M-model's simulation is poor in these regions, as one might expect. Consequently, we agree with SR that the NM-model (which has no mountains anywhere) performs no worse in these regions. The western Ghats and Burma mountains (which were effectively left out of the M-model: see p. 1517, ¶2; Fig. 2.1; p. 1519, ¶6, 7, 8; p. 1532, ¶5) are probably essential (though not necessarily sufficient) to a more accurate simulation of the detailed monsoon circulation and precipitation patterns. Higher resolution general circulation models or perhaps nested grid models are necessary in order to include these smaller scale mountain effects.

While concentrating on the M-model's shortcomings, it is easy to get the false impression that the M-model's simulation is less realistic than that of the NM-model. However, one can identify many large-scale features which indicate the contrary. A few examples: the South Asian low-pressure belt forms along 30°N in the M-model; it does not exist in the NM-model (Figs. 4.1 and 5.2). Near the surface, moist southwesterly flow extends northward to the mountain region (along 30°N) in the M-model, but in the NM-model only to 10°–15°N, with dry northwesterly flow and little precipitation persisting from Tibet to central India (Figs. 5.1 and 5.4). Upward motion persists over Tibet in the M-model, downward in the NM-model (Fig. 4.6). In the middle troposphere, high temperatures are simulated by the M-model over Tibet; in the same regions, much colder temperatures are simulated by the NM-model (Fig. 4.4). Similar comments can be made about the simulations of the Somali jet (Fig. 4.8), the meridional circulation (Figs. 4.7 and 5.6), etc. In short, the large-scale features of the distributions of pressure, precipitation and flow patterns of the M-model are much more similar to the observed monsoon phenomena than those of the NM-model.

5. Summary and conclusions

In conclusion, we agree with SR that some of the difficulties in simulating monsoon circulation are significant and require a major effort to remove them. Nevertheless, we are very much encouraged by our present success in simulating some of the basic characteristics of the structure and evolution of the large-scale features of the monsoon circulation. In the future, despite the concern expressed by SR, we feel it is possible to improve our model so that more definitive numerical experiments can be made.

In their critique, SR pointed out the M-model simulation is less realistic than that of the NM-model. To the contrary, we feel the large-scale features of the M-model

[3] The cross section for the M-model in Fig. 4.3 of HM (averages from 80°–95°E) lies between two cross sections (at 73° and 100°E) of Ramage and Raman (1972). For comparison, maximum values in both observed cross sections are averaged.

atmosphere are much more realistic than those of the NM-model as discussed in the preceding section. Therefore, we feel it is meaningful to discuss the role mountains play in the South Asian monsoon circulation by comparing results from the two experiments. However, in view of the various difficulties discussed here, it may be useful and enlightening to repeat similar numerical experiments after significant progress is achieved in the simulation of the monsoon circulation. Results from such further experiments will be particularly interesting if the horizontal grid resolution is significantly improved so that the smaller scale orographic features can be resolved. Nested grid models may be useful in this effort.

REFERENCES

Hahn, D. G., and S. Manabe, 1975: The role of mountains in the south Asian monsoon circulation. *J. Atmos. Sci.*, **32**, 1515–1541.

Hobbs, J. E., 1974: A complex intertropical convergence zone—Some examples from the Indian Ocean. *Weather*, **29**, 122–143.

Krishnamurti, T. N., S. M. Daggupaty, J. Fein, M. Kanamitsu and J. D. Lee, 1973: Tibetan high and upper tropospheric tropical circulations during northern summer. *Bull. Amer. Meteor. Soc.*, **54**, 1234–1249.

——, Elford G. Astling and Masao Kanamitsu, 1975a: 200 mb wind field June, July, August 1972. Dept. Meteor., Florida State University, 116 pp. [Available from Florida State University.]

——, M. Kanamisu, R. Godbole, Chia-Bo Chang, Fred Carr and J. Chow, 1975b: A study of a monsoon depression. I. Synoptic structure. *J. Meteor. Soc. Japan*, **53**, 227–240.

Murakami, T., R. V. Godbole and R. R. Kelkar, 1970: Numerical simulation of the monsoon along 80°E. *Proc. Conf. Summer Monsoon Southeast Asia*, C. S. Ramage, Ed., Norfolk, Va., Navy Weather Res. Facility, 39–51. [NTIS AD-876-677.]

Orgill, M. M., 1967: Some aspects on the onset of the summer monsoon over Southeast Asia. Dept. Atmos. Sci., Colorado State University, Final Report, Contract DA28-043-AMC-0130 3(E), 75 pp. [NTIS AD819-959.]

Ramage, C., and C. V. R. Raman, 1972: *Meteorological Atlas of the International Indian Ocean Expedition.* Vol. 2. National Science Foundation, Washington, D. C. [Superintendent of Documents, Washington, D. C., Stock No. 3800-00124.]

Ramamurthy, K., 1969: Monsoons of India: Some aspects of the "break" in the Indian southwest monsoon during July and August. *India Meteorological Department Forecasting Manual*, Part IV, No. N-18, 3. [Available from Director General of Observatory, India Meteorological Department, Lodi Rd., New Delhi 3, India.]

Sadler, J. C., 1975: The upper tropospheric circulation of the global tropics. UHMET 75-05, Dept. Meteor., University of Hawaii, 35 pp. [Available from the University of Hawaii.]