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Low-Frequency Transitions of a Simple Monsoon System

PETER J. WEBSTER

CSIRO Division of Atmospheric Physics, Aspendale, Victoria, Australia

LANG C. CHOU

Department of Meteorology, Naval Postgraduate School, Monterey, CA 93940

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ABSTRACT

Seasonal, diurnal and synoptic time scales, plus a subseasonal modulation of synoptic events, appear to dominate the temporal structure of monsoon systems. Observational studies indicate that the latter low-frequency variation modulates or groups synoptic disturbances, producing periods of intense activity (the "active" monsoon) separated by distinct lulls (the "break" monsoon). Together with the "onset" and "retreat" of the monsoon, the modulations introduce into the system time scales which are far more rapid than that which would be expected from the evolving latitudinal variation of insolation.

As observations indicate that the seasonal cycle and low-frequency transients occur in large spatial scales a model is used which appears to simulate the large-scale mean seasonal structure of the monsoon. Such a model is a zonally symmetric moist primitive equation model coupled to an interactive and mobile ocean. With such a model the hypothesis is tested that the basic character of the low-frequency subseasonal transients of the simple monsoon system are a result of feedbacks between the hydrologic cycle and the differential heating between the interactive ocean and continental regimes.

With multi-annual integrations of the joint ocean-atmosphere model, monthly and daily variations are studied with the aim of testing the hypothesis. With full hydrology and an ocean-continent contrast, the model monsoon deviates substantially from the smooth transitions noted in dry experiments. Not only do the onsets and retreats of the monsoon system accelerate but rapid, orderly transitions occur during the Northern Hemisphere summer. During such transitions, the meridional monsoon cell periodically migrates inland causing rising motion north of 30°N and subsidence near the coastal margin which is the location of mean seasonal ascent. The transition is seen to possess biweekly time scales and show some characteristics of the monsoon "break". Similar transitions occur in subsequent years of the integration and differ in timing and intensity but maintain the same basic period. Comparison of the results with those of other studies are made and further studies detailed.

1. Introduction

Four major temporal scales appear to dominate monsoon circulations. These are the seasonal, synoptic and diurnal variations plus an aperiodic subseasonal or supra-synoptic modulation. Whereas the gross features of the seasonal monsoon have been recognized at least since the time of Halley (1686), significant progress in cataloging and identifying important seasonal and subseasonal structure have been achieved by the study of a long meteorological measurement record [see Ramage (1971) for a summary] and by utilizing data sets obtained from intensive observational efforts and subsequent detailed analyses (e.g., Ramage 1971; Ramage and Raman 1972, Sadler 1975; among others). Emerging from the analyses is a picture of a generally recurring slowly varying structure on a seasonal time scale with a variety of transients existing over various space and time subscales. Fig. 1, which shows the latitude-time sections of the evolving mean monthly 200 mb wind field in the south Asia region, illustrates the slowly varying envelope of the seasonal monsoon. The transients (i.e., low-frequency transients and synoptic events) which exist within such a seasonal envelope may show great variability in both character and occurrence from case to case or year to year (Ramage, 1971). Indeed, the variation in the form and the occurrence of the transient features probably accounts for the majority of the interannual variation of the monsoon.

For convenience and simplicity, the monsoon systems are usually divided into summer and winter components.[1] Here we will concentrate on an extremely simple interpretation of the Asian summer monsoon principally because it is the system on which most work has been undertaken and for which most data exist. The general observed character of

[1] A Northern Hemisphere chronology will be used throughout the paper.

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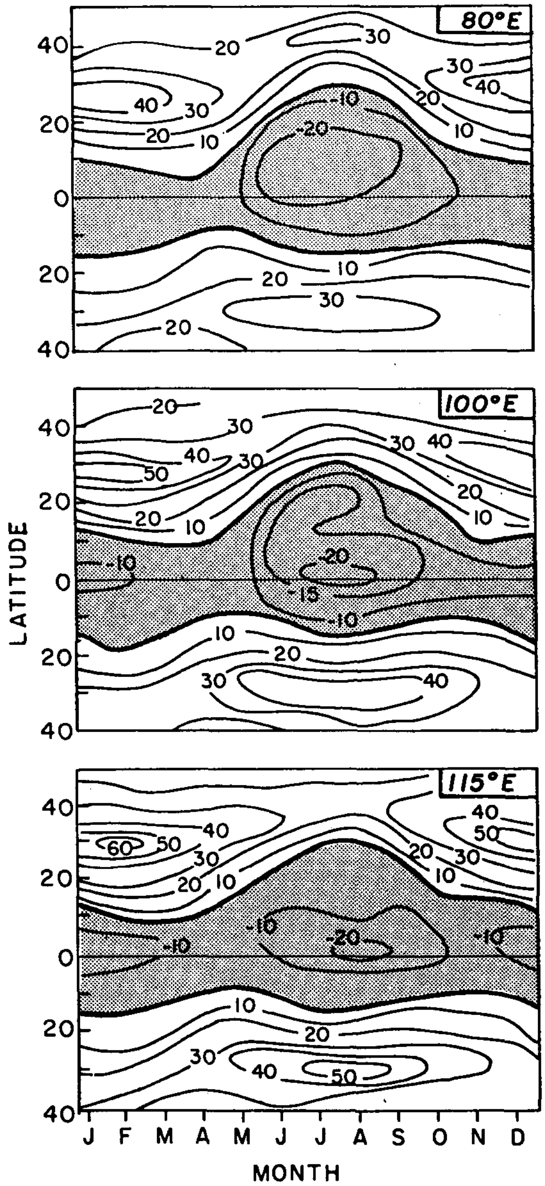

FIG. 1. Latitude-time sections of the 200 mb zonal velocity component (m s^{-1}) along the 80, 100 and 115°E meridians. Shaded area denotes easterly wind component.

the evolving summer component of the system may be summarized as follows: In late spring a vaguely defined cross-equatorial flow emerges from the Southern Hemisphere subtropics to the south of Asia and by early summer becomes both intense and steady in its southwesterly direction. The development and organization of the cross-equatorial flow coincides with the deepening of the monsoon trough over the north of India. Meanwhile, disturbed conditions are established over the Indian subcontinent, the arrival of which is often termed the monsoon "onset". Rapid modification of the upper level flow occurs and intense easterly winds in the upper troposphere ensue. Often in late summer, the disturbed pattern over India is interrupted with the cessation of precipitation over central India and an emergence of maximum precipitation to the north and south of the Indian plains. This is sometimes referred to as a monsoon "break." At such times the southern disturbed region is located nearer to its climatological spring position (i.e., just north of the equator) than to its summer position (Ramage, 1971). During autumn, with the accelerating retrogression of the latitude of maximum insolation, disturbed conditions diminish over India, the upper anticyclone weakens as do the upper easterlies and the monsoon is said to "retreat."

A number of observational studies have considered the transient nature of the monsoon. For example, Koteswaram and Rao (1963) suggest a sequence of states, explained in terms of meridional cell locations, to represent differences between active and break periods. For the active period Koteswaram and Rao suggest that two cells, a monsoon cell located from the equator to 20°N and a Hadley[2] cell lying between 20 and 40°N, possess a common ascending branch over the heated summer continent. During break situations, when precipitation is significantly diminished over central India, they propose that the ascending branch near 30°N in the active monsoon is replaced by a descending branch. Asnani (1967) embellished Koteswaram and Rao's monsoon cell by a combination double cell in order to account for the summer mean cloudiness minimum over the equator.

Recent observational studies by Krishnamurti and Bhalme (1976) and Murakami (1977) hint at strong interhemispherical coupling of the monsoon system on *subseasonal* (approximately bi-weekly) time scales. Variations appear to be communicated between the southern Indian Ocean and the precipitating regions over summer southern Asia. Although there may be some debate regarding the relative phases and lags of the variations, it is important to note that the communications appear to be accomplished by features reconcilable with the Koteswaram and Rao-Asnani descriptions of the monsoon. Most importantly, the observational studies suggest that the synoptic-scale disturbances of the monsoon flow (e.g., Bay of Bengal disturbances) occur in groups with time scales and sequences similar and consistent with the macroscale low-frequency variations, that is, they infer that the disturbances are modulated by the macroscale submonthly variations. As the disturbances are responsible for the majority of the precipita-

[2] The term "Hadley" cell is used loosely by Koteswaram and Rao as it infers a hemispheric scale and was probably chosen to distinguish the meridional circulation from their "monsoon cell" located further to the south.

tion in the summer monsoon, an understanding of the mechanisms which control or modulate their occurrence is of some importance.

In the descriptive models described above, the monsoon transitions (e.g., onset, modulation and retreat) are considered as rapid adjustments of the macroscale meridional circulation. However, the observational models do not provide mechanisms which explain the time scale of the transitions or the particular geographic location the structure occupies. Consequently, extensions of the Koteswaram and Rao-Asnani model or the testing of the Krishnamurti and Bhalme sequence must either rest on the compilation of sufficient data or on the development of theoretical and numerical models. At this stage the data base appears insufficient to sustain continued advances and perhaps we are forced to rely on models (albeit crude) to at least follow the initial leads prompted by observational studies.

In recent years, a variety of theoretical and numerical models have been developed and used for monsoon studies. Some success has been achieved in the simulation of the mean seasonal fields using general circulation models (e.g., Washington and Daggapaty, 1975; Hahn and Manabe, 1975) or simpler models (e.g., Webster, 1972, 1973; Webster and Chou, 1980, hereafter referred to as WC). Other studies have aimed at the modeling of specific phenomenology such as the role of topography on the evolving monsoon circulation (e.g., Hahn and Manabe, 1975), and the role of adjacent and interactive oceans (Webster and Lau, 1977, hereafter referred to as WL; Lau, 1978; Webster, 1979). However, studies aimed at the elucidation of subseasonal monsoon structures have received less effort. Their consideration constitutes a basic aim of this paper.

2. Hypothesis

WL tested the proposition that the interactive nature of the oceans adjacent to the Asian continent were crucial in determining the magnitude and spatial variation of the mean monsoon. Extensions of the study (Lau, 1978; Webster, 1979) showed that the phasing of the overall monsoon was a strong function of the different time lags exhibited by the oceans and land areas and the latitudinal variation of ocean response. We shall accept as a basic premise that the *addition of an interactive or specified time-varying ocean describes the lower limit of model sophistication necessary to study the longer term dynamics of the monsoon flow.*

The recent compilation of mean seasonal fields by WC indicates that at least the long-term dynamics rely critically on moist processes. By extrapolation we may suppose that moist processes are of equal importance to the evolving monsoon state.

The experiments of Lau (1978) and Webster (1979), neither of which contained a hydrology cycle, indicated only smoothly varying fields which simulated well the seasonal variability and allowed the atmospheric response to be coupled to the variable lag of the interactive ocean. However, subseasonal modulations and rapid transitions like these noted by Krishnamurti and Bhalme (1976) were not apparent. Furthermore the magnitude of various fields was less intense than those observed as were the mean fields for the dry experiments noted in WC. As precipitation and the release of latent heat appear to assume important roles in the cited observational studies, their neglect in Lau's and Webster's studies may have been critical. Consequently, we propose the hypotheses that it is *the feedbacks between hydrological processes and the basic dynamic elements of the monsoon system (i.e., the differential heating of the interactive ocean and the continent) which are responsible for rapid transitions and modulations of the system.*

To test the hypothesis we adopt the basic WL model but with the zonal symmetry assumption incorporated in WC. The assumption is consistent with a philosophy of attempting to construct the *simplest* possible model which may contain the physical ingredients necessary to produce structures described by the three observational studies. By an aggregation of physical processes, the complexity of the model is increased in steps and the hypothesis tested at each stage.

The study has the further purpose of attempting to understand the results of Hahn and Manabe (1975) from the confines of a considerably simpler system. The principal result of their study was an indication of the importance of the Himalayan Massif, a feature purposely omitted from the current study. On the other hand, the Hahn and Manabe study did not contain an interactive ocean which we stipulate as a basic physical ingredient in the monsoon system. Of course there are further differences such as the physical domain of the two models which add to the problems of comparison.

3. Experiments

The experimental format is identical to that described by WC except that it refers to evolving fields rather than seasonal means. The model was initialized as a horizontally isothermal atmosphere with an initially prescribed surface temperature. With the sun held in an equinoctial position the model was integrated for a year during which time it approached a steady state. Three experiments used the equinoctial equilibrium as initial data. These were a dry ocean-continent experiment (DOC) in which a continental cap was situated north of 18°N and an interactive and mobile ocean, a moist ocean continental experiment (MOC) which was identical to DOC except for the inclusion of hydro-

TABLE 1. Properties of the various experiments.

Experiment	Label	Hydrology cycle	Geography	Total integration period (years)	Period displayed Figs. 4–10 (years)
Moist Oceanic	MO	Yes	Global ocean	4	3, 4
Dry Ocean–Continent	DOC	No	Continental cap > 18°N	4	3, 4
Moist Ocean–Continent	MOC	Yes	Continental cap > 18°N	4	3, 4

logic processes, and a moist ocean experiment (MO) where the earth was assumed to be covered by ocean. A summary of the case abbreviations is presented in Table 1.

The rationale behind the three experiments is an effort to choose suitable cases with which to test the hypotheses posed in the last section. For example, a comparison between the DOC and MOC results provide indications of the role of the hydrologic processes. Similarly, comparing the fields of the MO and MOC experiments suggest the importance of continentality.

All experiments were run for a further three years after the generation of the initial state (i.e., in year 1). In subsequent paragraphs, the last two years of integration are discussed. The third year corresponds to the period over which the seasonal means of WC were calculated.

4. Results

a. *Monthly structure*

Indications of the transient state of the monsoon structure are first obtained from the evolving state of the mean monthly fields. Fig. 2 shows the variation of the upper tropospheric zonal wind field as indicated by the magnitudes and positions of the two westerly maxima (located in subtropical and middle latitudes) and the equatorial westerly minimum (or easterly maximum). Plots of the various extrema as functions of latitude are shown for both the MO (dashed curves) and MOC (solid curves) cases for the third and fourth years of integration. The numbers on each curve refer to the month of integration starting with the spring equinox of the third year.[3] Consequently, 7 and 19 refer to the two autumnal equinoxes and 1, 13 and 25 to the three spring equinoxes. The summer solstices (demarked SS) are shown on Fig. 2 for years 3 and 4. The vertical arrow on the abscissa (also on Fig. 3b) denotes the land-sea boundary at 18°N.

Considering first the MO case it can be seen that the two westerly maxima oscillate over a small range of latitude (between 25° and 30° in each hemisphere)

[3] In model nomenclature, month 1 refers to the 30-day period *following* the spring equinox. Subsequent months are defined similarly.

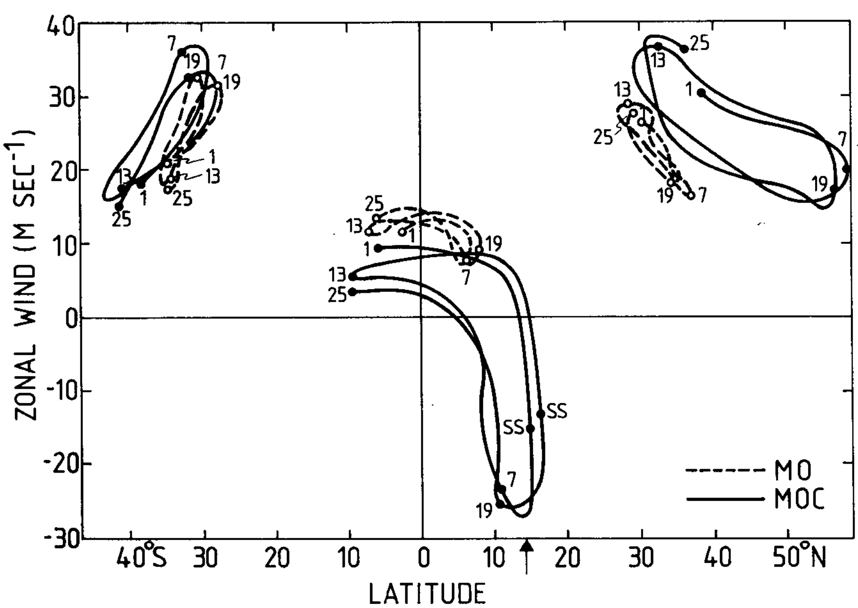

FIG. 2. Latitudinal variation of mean monthly 250 mb zonal wind component extrema during years 3 and 4 of integration. Curves connect mean monthly values of the two midlatitude westerly maxima (upper left and right) and the low-latitude westerly minimum (easterly maximum) for cases MO and MOC.

and a fairly small magnitude range. As expected, the two westerly maxima are out of phase. Maximum speeds occur near the spring equinox. The westerly minimum oscillates over a smaller amplitude range between 10°N and 10°S. Like the westerly maximum, the westerly minimum lags some months behind the solar declination.

The continental cap has a dramatic effect on the transient state of the monsoon flow. Sensible heat input during the summer is extremely large and substantial evaporation takes place over the warm continental region. The result is significant latent heat release in the atmosphere column with considerable warming in the upper troposphere. During winter, the lower boundary cools substantially, especially at higher latitudes, resulting in the enhancement of the latitudinal temperature gradient. Thus with a continental cap, the Northern Hemisphere westerly maximum now varies over a much larger latitude range and possesses a greater seasonal variability in magnitude. In summer the westerly winds are considerably further north with maxima near 50°N. The winter westerly maximas of *both* hemispheres are $\sim$5–10 m s^{-1} stronger than in the oceanic (MO) case. The most significant effect, however, occurs in the region between the equator and the continental cap with the development of a 27 m s^{-1} easterly jet stream in the upper troposphere with rapid acceleration occurring prior to the summer solstice (i.e., during months 1, 2 and 3 and 13, 14 and 15).

In a sense, Fig. 2 may be considered sets of hysteresis curves and the space between the extrema at successive (say) spring equinoxes (i.e., points 1, 13 and 25) define a hysteresis creep which may be viewed as a measure of the deviation from an annual cyclic equilibrium. Whether or not the deviations are a result of a "natural" interannual variability or because the model has yet to achieve a state of cyclic equilibrium is difficult to resolve at this stage. On one hand, points 13 and 25 lie closer together than points 1 and 13, indicating an approach to cyclic equilibrium. However, the model does contain mechanisms which potentially may have time periods longer than one year or allow "memories" of the previous annual state. Two examples are the deep ocean temperature of the ocean model (see WC for details) and the hydrologic cycle which possesses an annual memory over the continent via the amount of stored moisture. The solution to the problem of interannual variability may be resolved by extended integrations of the model ($>$10 years) which are logistically feasible because of the simplicity of the model.

Figs. 3a and 3b show the variations between the monthly means of the vertical velocity profiles at 500 mb as a function of latitude for the MO and MOC cases, respectively. Rather than a linear trend of latitudinal position with time, the MO profiles oscillate about the equator some months out of phase with the solar insolation. The vertical velocity, in fact, is coupled strongly to the evolving position of the sea surface temperature maximum. The positions of the maximum ascent are similar to the climatological location of the "intertropical convergence zone" or "near-equatorial convergence zone". (In the simple model discussed here, either term may be used unambiguously as the region of maximum vertical velocity corresponds to the region of maximum convergence.) Furthermore, the intensity of the mean monthly vertical velocities which averages near -4×10^{-4} mb s^{-1} or 35 mb day^{-1} appears consistent with determinations from general circulation studies (e.g., Hahn and Manabe, 1975).

For the MOC case, the mean monthly vertical velocity profiles show rapid transitions from a Southern Hemisphere location in month 1 to over the continent in month 2 where substantial intensification occurs. The ascent over the continent is maintained until the autumnal equinox although subsequent months show some variability. The most northward extent of the ascending branch of the meridional cell occurs after the summer solstice. At that time the circulation is broad and intense and associated with substantial subsidence to the south of the continental cap and in the subtropics of the Southern Hemisphere. After the autumnal equinox, the ascent reestablishes just north of the equator similar to the oceanic convergence zone during month 7 before rapidly retreating southward.

The double maximum at the equinoctial months (1 and 7) is similar to the observed near-equatorial trough positions (see Ramage, 1971, Fig. 5.18) although the model appears to show a premature establishment of the northern trough in spring. A comparison of the winter profiles of the MO and MOC show some similarity except for a generally stronger circulation resulting from the enhanced temperature gradient—a reflection of the rapidly cooling continent.

The position of the ascending branch of the monthly mean meridional circulation is closely akin to the observed location of Koteswaram and Rao (1963) for the active monsoon (i.e., as distinct from break) over India. It is interesting to note that the agreement has occurred without mountains being included in the model which provides contrast to the findings of Hahn and Manabe (1975) although, as mentioned previously, model result comparison is difficult.

b. Transient structure

The transitions between the monthly mean vertical velocity distributions of the MOC case mask an even greater variation occurring in the submonthly structure. To illustrate this behavior latitude-time

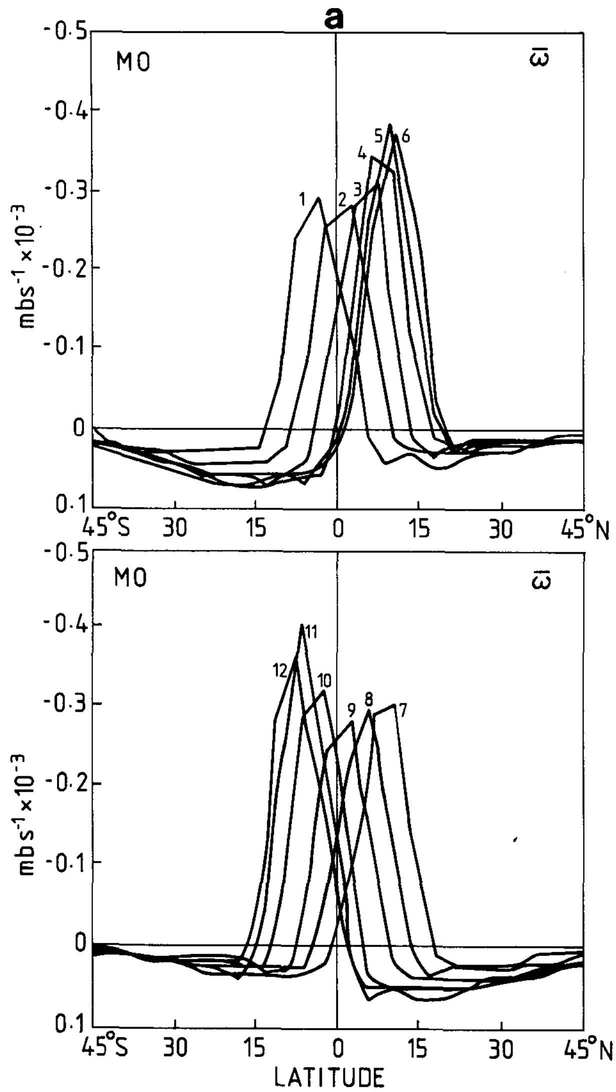

FIG. 3. Latitudinal variation of the mean monthly vertical velocity profiles for (a) MO and (b) MOC (units, mb $s^{-1} \times 10^{-3}$). Upper panel in each diagram refers to the first six months following the spring equinox of year 3 of integration. The lower panel refers to the second six months. Arrow on abscissa marks ocean-continent boundary.

sections of surface temperature, 250 mb temperature, the 250 mb zonal wind speed and the vertical velocity are shown for the MO, DOC and MOC cases in Figs. 4–7. The precipitation and specific humidity fields of cases MO and MOC are displayed in Figs. 9 and 10. Fields are shown for the 2-year period from Day 730 to Day 1460.

The surface temperature of the MO case shown in Fig. 12 has a smooth small-amplitude seasonal period with maximum temperatures near the equator[4] as distinct from the MOC and DOC cases where large continental changes tend to dominate. The Southern Hemisphere temperature variation is quite similar in all experiments except that the phase of the temperature maximum in MOC slightly leads that of the MO case in the subtropics. Furthermore, the subtropics in the MOC case are cooler (~2 K) than in the MO case. Both factors stem from the interhemispheric influence of the continentality of the Northern Hemisphere. The cooler temperatures result from the increased wind stress caused by the fresher low-level winds in the Southern Hemispheres of the MOC experiment and the subsequent modification of the mixed layer. The apparent phase advance of the Southern Hemisphere surface temperature maximum is caused by an earlier start to the cooling trend, again instigated by the increased wind stress. Both factors indicate the large-scale influence of the monsoon system and, at the

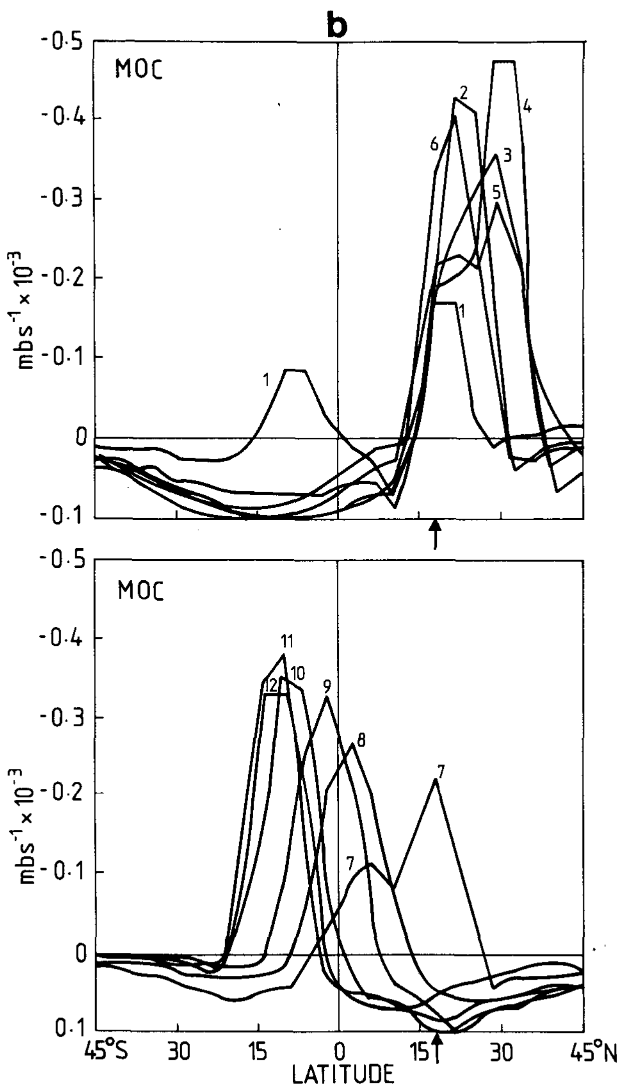

FIG. 3. (*Continued*)

[4] The version of the model used does not contain wind-driven Ekman transports capable of producing a weak relative temperature minimum along the equator. Such extensions are developed by Lau (1978).

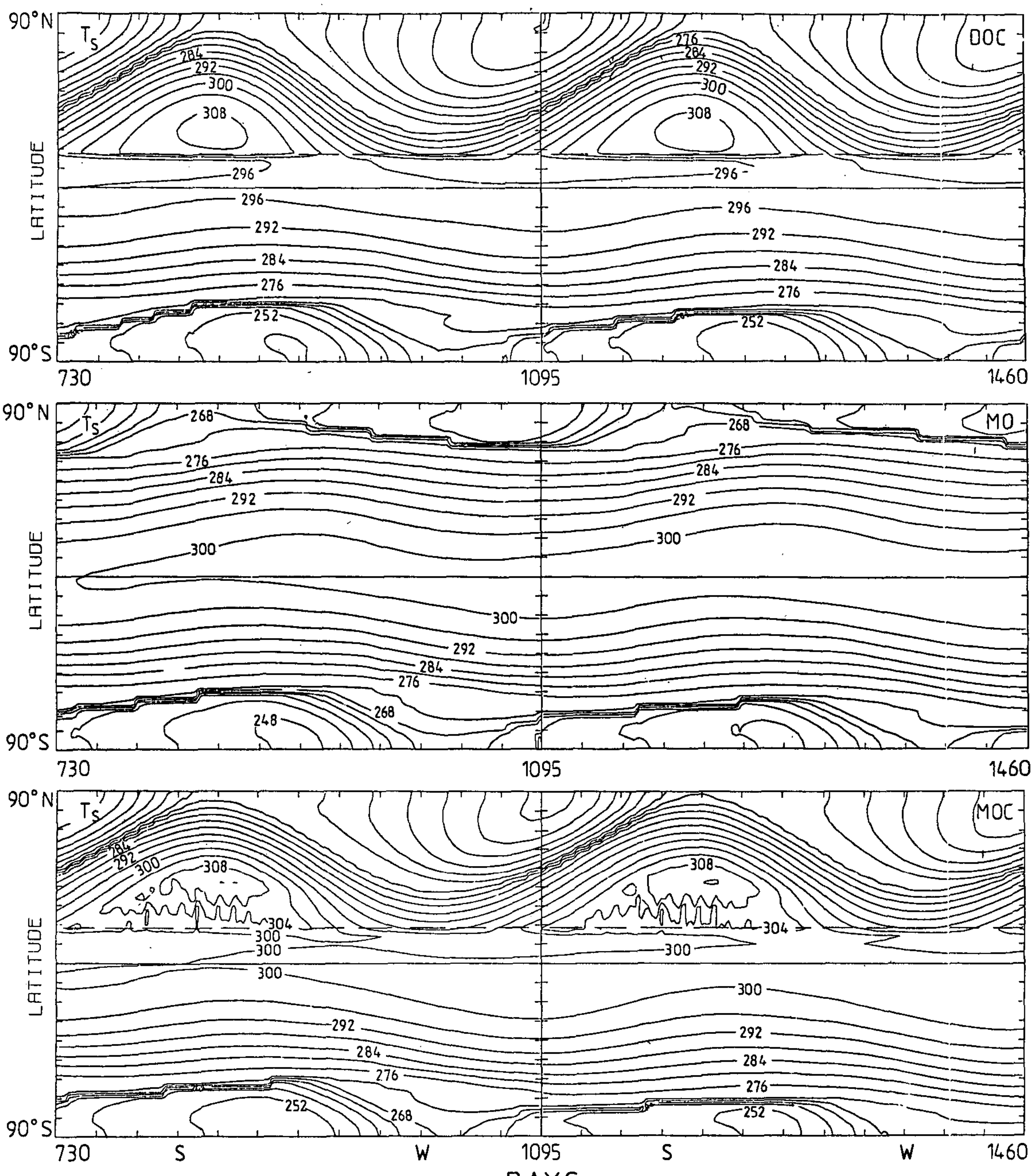

FIG. 4. Latitude-time sections of the surface temperature of cases MO, DOC and MOC for years 3 and 4 of integration. The summer and winter solstices are denoted by S and W, respectively, on the abscissa. Horizontal dashed line refers to coastal margin in the DOC and MOC cases. Solid line denotes the equator. Units are K with contour spacing 4 K.

same time, the interdependency of the ocean and the atmosphere.

Without hydrology (DOC) the maximum surface temperature appears much closer to the land-ocean margin than when hydrology is allowed. The displacement poleward of the region of maximum temperature when hydrology is included is due to the effect of groundwater accumulation and subsequent enhanced evaporation in the region of maximum precipitation. The magnitude and phase of the maxima appear little changed between the two cases.

An important feature of the MOC surface temperature structure is the rapid variation near and to the north of the continental margin of the Northern

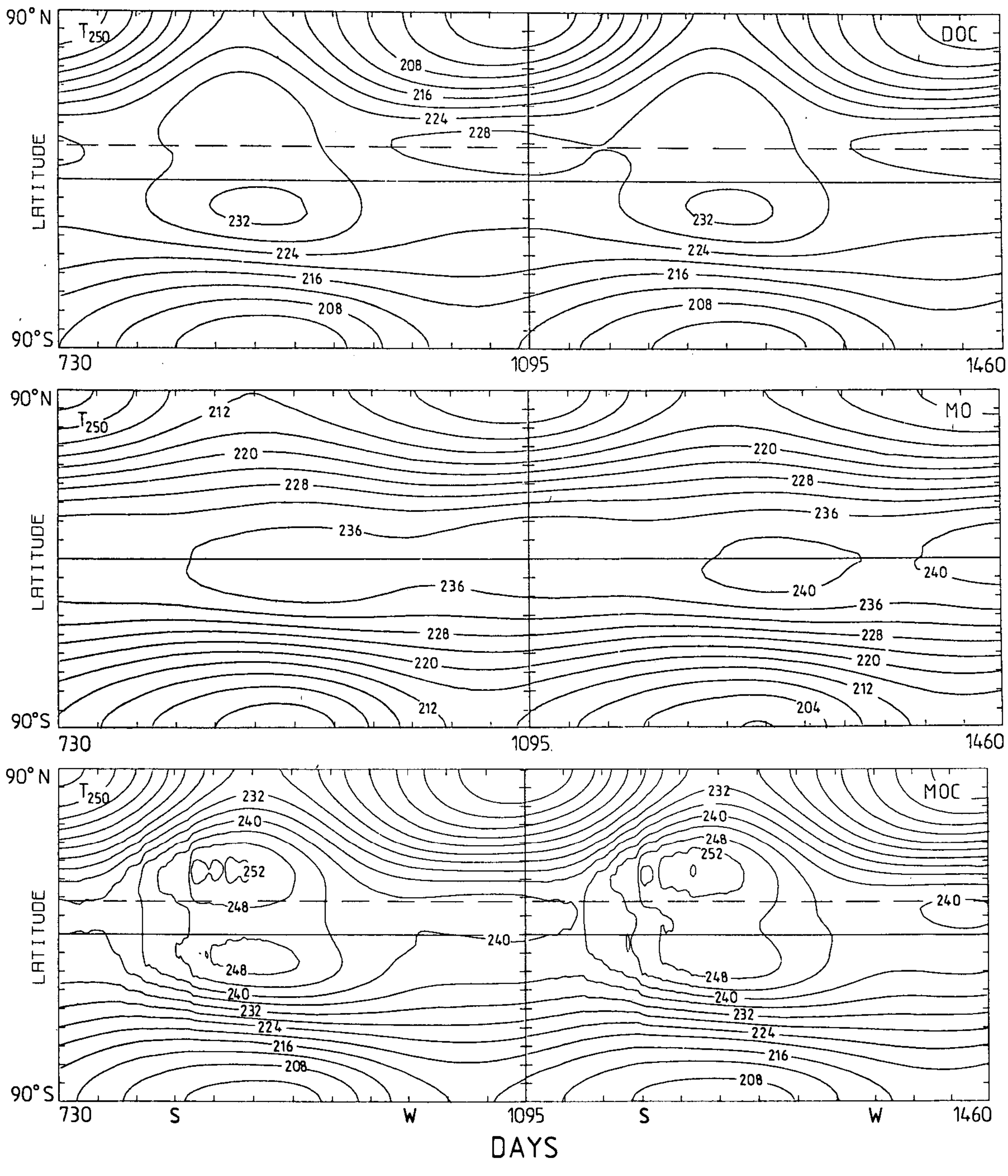

FIG. 5. As in Fig. 4 except for the 250 mb temperature field component.

Hemisphere during summer. Such variability is absent in both the MO and DOC experiments indicating that the modulations are associated with the hydrology cycle and the heated land mass. The variability has a fairly reproducible period of ~12–15 days and is reflected in an organized manner through the other variables of the model (see Figs. 5–10, especially Fig. 8).

The surface temperature for year 4 shows similar broad-scale features to year 3. Only the details of the summer modulations are different. Whereas the period of the modulation appears reproducible from year to year, its basic structure is different in detail which is perhaps indicative of the interaction of highly nonlinear processes existing within the model.

The effect of hydrology and continentality is best seen in the plots of the upper troposphere temperature (Fig. 5) and the 250 mb zonal wind (Fig. 6). With oceans and hydrology (MO), the upper

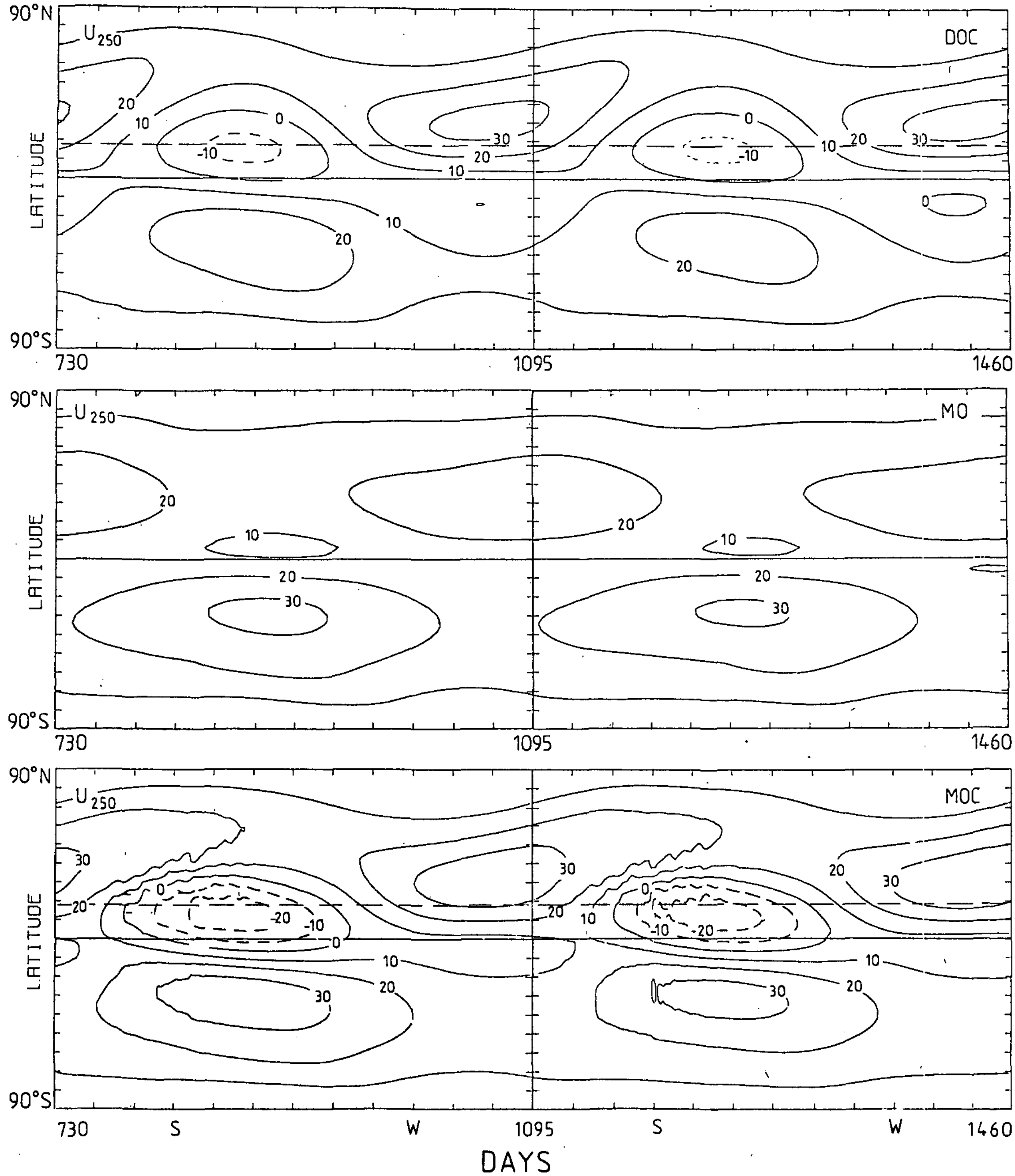

FIG. 6. As in Fig. 4 except for the 250 mb zonal wind component. Dashed lines denote easterlies. Units are m s^{-1} and contour spacing every 10 m s^{-1}.

tropospheric temperatures show an even seasonal variation, symmetric about the equator. With a continental cap but no hydrology (DOC) asymmetries are introduced into the structure with warmest regions coincident with maximum subsidence to the south of the equator in late summer. A secondary maximum occurs over the continent and corresponds to enhanced sensible heating. With both hydrology and continentality (MOC) the magnitude of the two maxima are increased substantially due to the increase in latent heating over the land and the increased subsidence in the winter hemisphere. It is interesting to note that the regions of maximum temperature in the MOC case are to the north of the strong ascent and correspond to significant subsidence in contrast to the DOC case. This agrees with the vertical velocity distribution over real deserts and suggests that desert regions

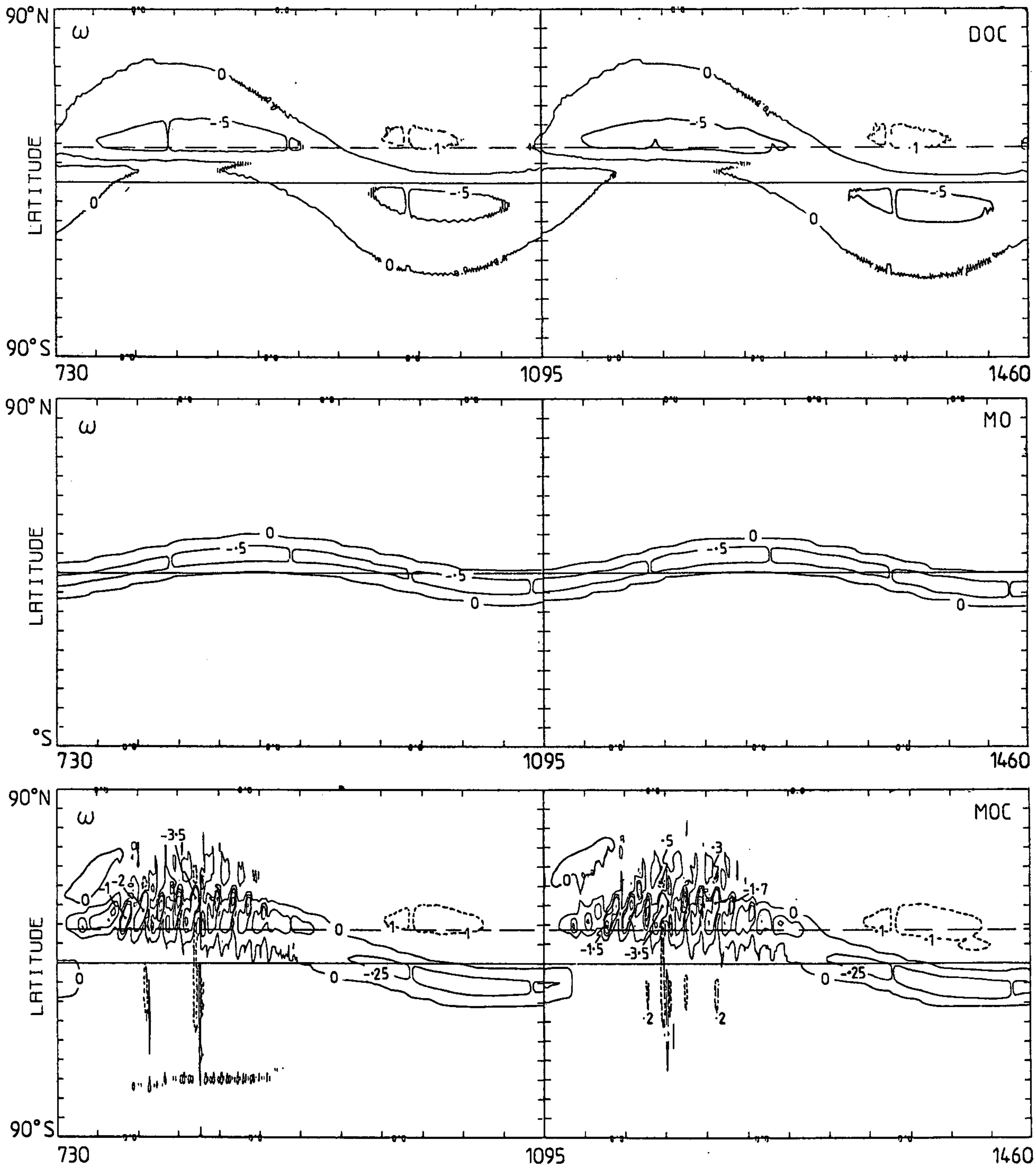

FIG. 7. As in Fig. 4 except for the 500 mb vertical velocity component (10^{-4} mb s^{-1}).

cannot be considered as isolated regions but are dynamically coupled to intensive moist-convective regions. A connection between the desert regions to the west of the Arabian Sea and the Bay of Bengal during summer was hinted at by Stephens and Webster (1979).

A most important effect occurs in the modification of the near-equatorial temperature gradient. In the MO case, the latent heating being relatively weak and fairly close to the equator, the low-latitude temperature gradient is rather flat. Consequently, the upper tropospheric winds (Fig. 5) show

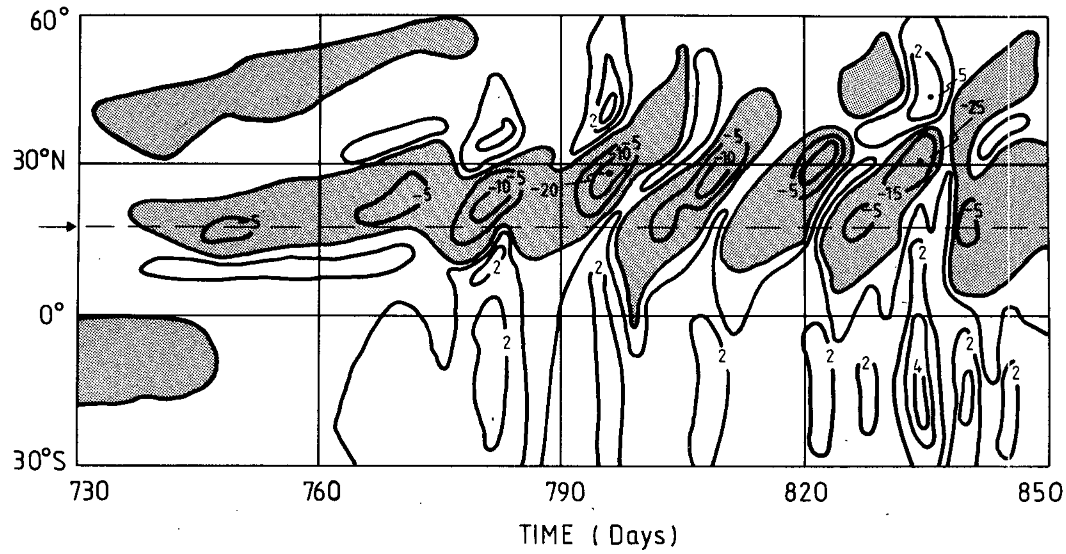

FIG. 8. Detail of the MOC ω field (500 mb) shown in Fig. 7 from day 730 to day 850 between 30°S and 60°N. Units are mb $s^{-1} \times 10^{-4}$. Shaded area denotes ascent.

only a relative westerly minimum. The double temperature maximum of the DOC case results in a reversed temperature gradient at low latitudes producing a weak easterly jet stream in summer to the south of the continental margin. The reverse is emphasized in the MOC experiment by the convective heating and a considerably stronger easterly jet stream is produced. Comparing Fig. 5 with the observed distribution of Fig. 1 shows the magnitude and phase of the seasonal variation of the zonal winds.

The low-frequency transients noted in Fig. 4 are also apparent in the upper troposphere temperature and zonal velocity fields shown in Figs. 5 and 6 for the MOC case. Such vertical consistency between the various variables is to be expected as in the convectively unstable regions the hydrologic cycle accomplishes a rapid and vigorous link between the lower and upper troposphere. Because of the absence of the modulations in either the MO or DOC experiments it is clear that the driving mechanisms result from interactions between the hydrologic cycle and continentality, the latter emphasized by the interactive ocean.

The vertical velocity fields are shown in Fig. 7 and keenly indicate the effects of hydrology and continentality. As expected from the monthly variation of the vertical velocity for the moist ocean, the maximum ascending motion is contained to near the equator in a narrow band which appears in phase with the maximum sea surface temperature. The dry ocean-continent experiment possesses a much more varied vertical velocity field. Fairly rapid latitude changes of the maximum occur in early summer coupled with the rapid heating of the continent (see Fig. 4) followed by a slower transition to the Southern Hemisphere during winter. The most striking differences between the MOC and DOC cases is the intensity of the vertical circulation produced by the release of latent heat and the variability of the circulation during the summer period. Furthermore, the "onset" and "retreat" of the monsoon appear more sudden in the MOC case than the gradual transition or slow meander which occurred in the DOC case.

To illustrate the subseasonal structure apparent in the MOC vertical velocity field during summer and to present a simpler picture, a detail of Fig. 7 is shown in Fig. 8. Plotted for only 120 days and commencing at the spring equinox of year 3 of integration, the diagram shows the evolving vertical velocity field between 30°S and 60°N. In the period immediately following the spring equinox, the maximum rising motion is to the south of the equator. Near Day 740, three weak vertical velocity maxima occur with the emergence of ascent near the coastal margin and near 40°N. With increasing insolation and the consequent rapid heating of the continent the rising motion increases at the expense of the Southern Hemisphere circulation. Further intensified by the increased flux of moisture from the ocean, the continental region of ascent finally quenches the southern maxima by increased subsidence.

The most interesting episodes occur even later in summer when period intensification of the circulation occurs. From Day 780 onward, the meridional circulation, as indicated by the vertical velocity distribution, shows periods of intensification some 12–15 days apart. While intensifying the meridional cell appears to propagate poleward from the coastal margin, reaching a maximum intensity usually between 15–20 mb s^{-1} (i.e., ~150 mb day^{-1} or 20

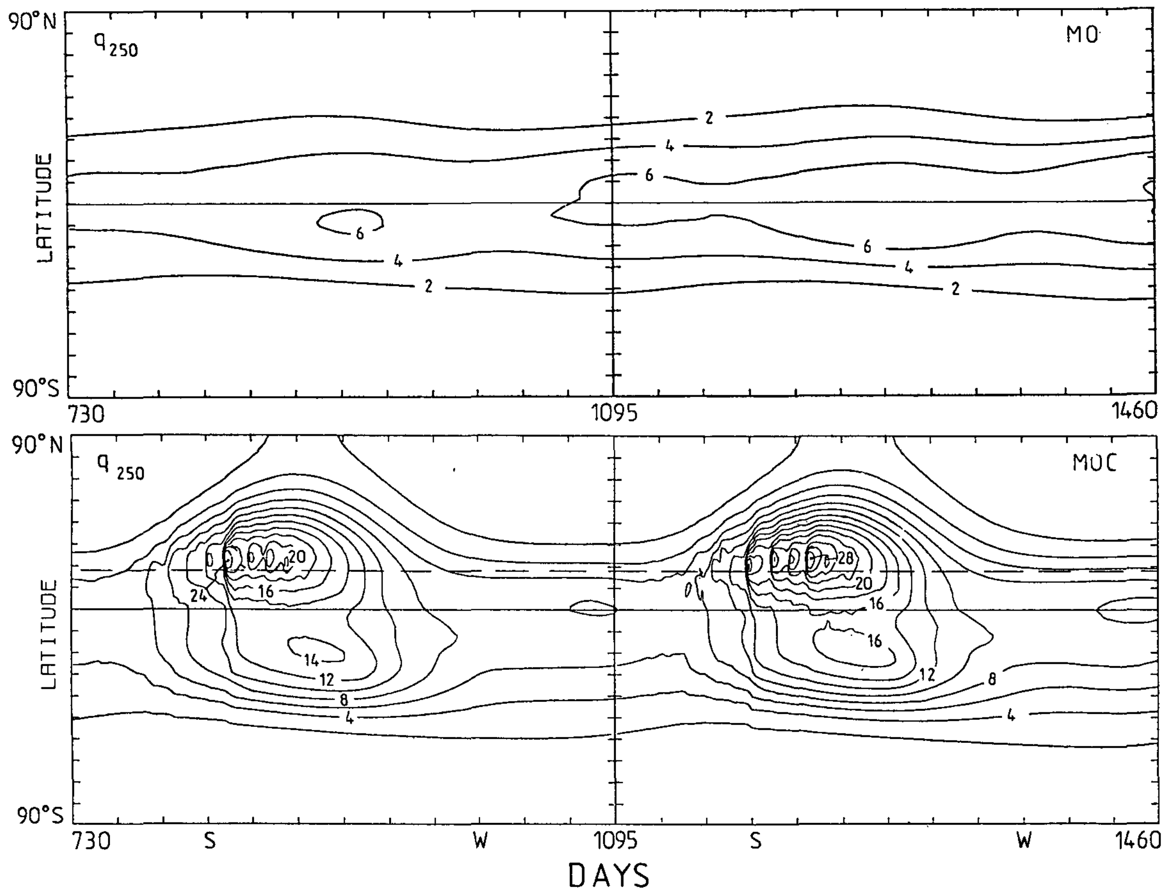

FIG. 9. as in Fig. 4 except for the 250 mb specific humidity fields. A scale factor of 10^4 has been applied.

cm s^{-1}) with considerable subsidence to the north and south. Rapid decay ensues, accompanied by the formation of a second cell. In turn the second cell intensifies, moves poleward and becomes the dominant circulation.

Such a sequence is repeated perhaps five or six times per simulated summer. In the next year (year 4) a similar periodicity of events occurred although with different timing, detailed position and intensity. The mechanisms which are responsible for the variations are related to a complex interaction between the hydrology cycle over the heated land mass and the effect of the ocean to the south. A detailed description of the mechanisms will be presented in a companion paper. It suffices here to note that the variations of Figs. 7 and 8 are coherent features of the model monsoons and are reproducible, at least in form, from one year to the next and possess many characteristics of the variation noted by Krishnamurti and Bhalme (1976) and Murakami (1977).[5]

The 250 mb moisture fields are shown in Fig. 9 for the MO and MOC cases. Following from the strong temperature dependency of specific humidity, some similarity exists between the form of the q fields and the temperature at 250 mb shown in Fig. 4 including the submonthly variability. The variations follow from the efficient upward transport of moisture by the convective processes described in WC. As the convective processes strongly correlate with the vertical velocity distribution and hence the release of latent heat, regions of maximum temperature and maximum specific humidity will correlate on the time scale of the submonthly variation (cf. Figs. 4 and 9).

Fig. 10 shows the precipitation fields for the MO and MOC cases. Except for the regions where the moisture has converged in the upper troposphere, the precipitation fields follow the vertical velocity

[5] We have recently become aware of a detailed observational circulation study by Sikka and Gadgil ("On the maximum cloud zone and the ITCZ over Indian longitudes during the Southwest Monsoon", Report 79 FM7, Indian Institute of Science, Bangalore, India May 1979) aimed specifically at low-frequency transients of the southwest monsoon. Sikka and Gadgil emphasize the establishment of the continental precipitation following breaks by ". . . northward moving epochs of the oceanic ITCZ. . . ." Similar time scales are indicated.

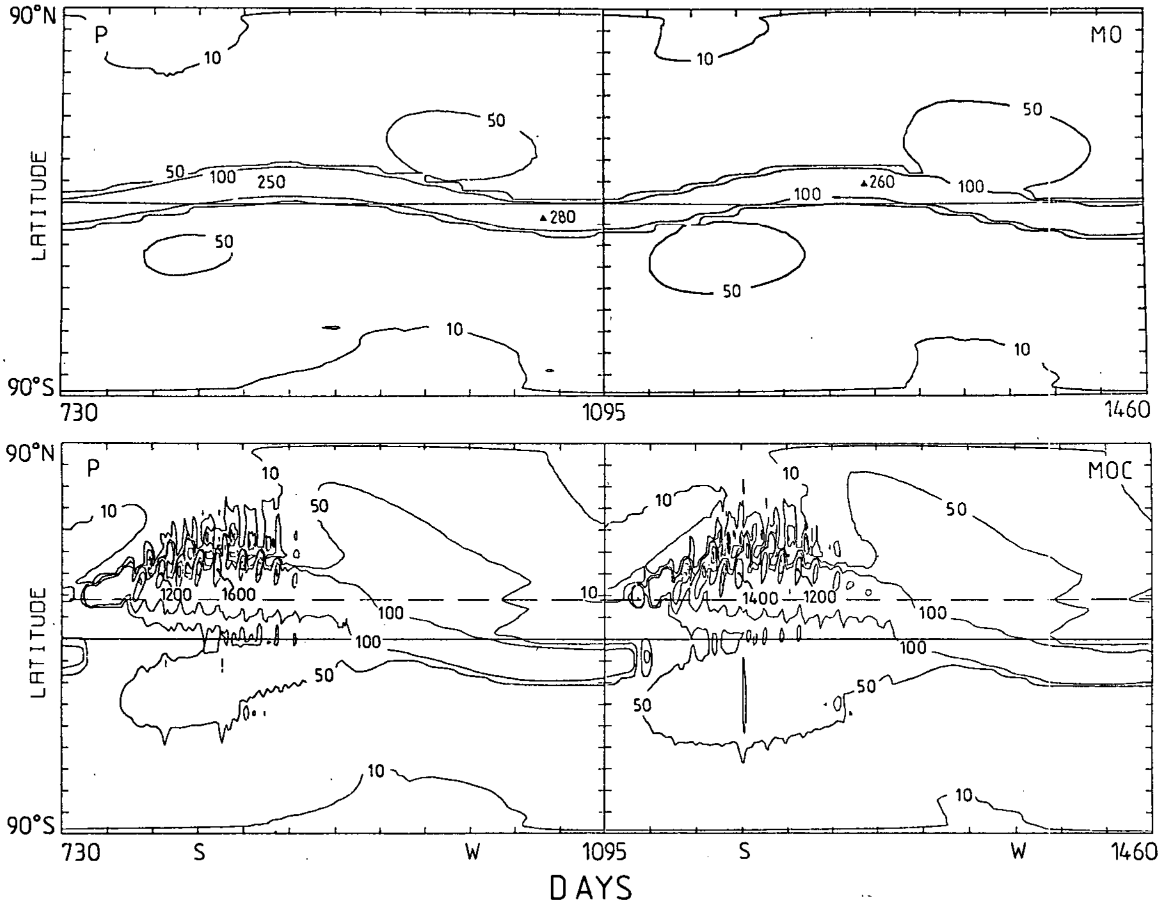

FIG. 10. Same as in Fig. 4 but for the precipitation rate fields. Units are cm day^{-1} and contours are drawn for precipitation rates of 10, 50, 100, 500 and 1000 cm $year^{-1}$.

as might be expected from the convective parameterization schemes. The light precipitation in the subsident subtropical regions of the winter hemisphere is probably an artifact of the neglect of eddies within the model. Without eddy transport, moisture tends to converge in regions of maximum meridional cell subsidence. Eddies, on the other hand, may move moisture most efficiently further poleward and possibly cause a second precipitation maximum. Even so, precipitation within the tropical regions follows observations quite well. For example, the MOC case estimates total annual precipitation between 22 and 31°N to be ~130 cm which is similar to the rainfall on the Gangetic Plains.

5. Concluding remarks

A hypothesis which postulated that low-frequency sub-seasonal modulations or transients may result from the interaction of hydrological processes and the basic dynamic elements of a monsoon was tested. The minimal components of a monsoon system were derived by WL to include an interactive ocean to the south of an "Asian" continental cap. The tool chosen to test the hypothesis was a relatively simple ocean-atmosphere interaction model which was compiled to allow a tractable analysis. The aim of the analysis was the identification and relegation of the various driving mechanisms of the monsoon. The study was prompted by a number of observational studies, most notably Krishnamurti and Bhalme (1976) and M. Murakami (1977), who noted large-scale and low-frequency modulations of the monsoon system.

The evolution of the mean monthly structure (Figs. 2 and 3) for the MO and MOC cases indicate, by comparison, the interactive nature of the hydrologic cycle and the basic drive of the monsoon system. For example, in the MO experiment (Fig. 3a) the strong interaction between the ocean and the atmosphere is indicated by three-month phase lag behind the solar heating of the vertical velocity variation, thus following closely the sea surface temperature variation. The lag is reminiscent of the behavior of the near-equatorial troughs or intertropical convergence zones of the western

Pacific Ocean and underlines the importance of local moisture availability or evaporation from a warm ocean surface. With greater surface heating coupled with the availability of moisture from the wet continental region, considerable variability is added to the system (Fig. 3b). During the winter, the system maintains the same phase lag as the oceanic case. However, during spring the ascending motion rapidly advances its northward propagation to become in phase with solar insolation over the continental region. The result is an extremely rapid "onset" and a somewhat retarded "retreat."

The interplay of the heated land mass and the hydrology cycle appears most pronounced in the latitude-time sections shown in Figs. 3–10. Only the cases which possess both a hydrology cycle and an ocean-continent contrast appear capable of temporal variations which are of a time scale less than seasonal. Fig. 8 is particularly pertinent and shows two major transition types which apparently depend on the interplay. These are the rapid "onset" of the monsoon in late spring and the variability of the established monsoon itself. The latter variation shows a coherent transition between periods of strong ascent and subsidence as the ascending branch of the Hadley cell propagates northwards. Although the details of the mechanisms which cause the monsoon modulation warrant a separate study and consequently will be reported in a companion paper, it should be noted that the periods of intermittent activity and the lulls in between occur as low-frequency transients similar to those reported by Krishnamurti and Bhalme (1976) and Murakami (1977). In a simple sense, the modulations appear to be similar to the active-break cycle of Koteswaram and Rao (1963); the active period corresponding to vertical motion just to the north of the coastal margin and the break period following the propagation of the cell northward when strong subsidence exists near the coastal margin. Despite the similarity it would probably be excessive to speculate that the model utilized in this study contains all the physical ingredients of what must be a very complicated system. On the other hand, what is apparent is that some of the basic mechanisms which contribute to the large-scale modulation of the monsoon system must be included in the model. A vital test would be the extension of the model to allow longitudinal variability in the manner of WL.

It is interesting (although difficult) to compare the results of this study with the careful experiments of Hahn and Manabe (1975). Granted that the model, even with its interactive ocean, is significantly simpler than that of Hahn and Manabe, the time scales of transitions of the monsoon system appear to be consistent with their experiments which *included* mountains. This is a surprising result as Hahn and Manabe found that without orography, the transitions of the monsoon system were significantly slower, especially the onset of the monsoon. It should be noted that the fields presented in the last section were somewhat weaker than those found by Hahn and Manabe and whether this is due to the neglect of eddies or omission of the Himalayas is difficult to determine at this stage. Perhaps the most important result of the study is the indication that the monsoon system possesses low-frequency transients without the modification of the latent and sensible heating which would be introduced by orography. Nor is this supposition without some observational support as both the West African and the North Australian monsoon systems, both substantially removed from substantial orographic features, are characterized by low-frequency modulation.

It should be emphasized that clouds were held constant in the experiment so that a potentially strong feedback mechanism was omitted. It is difficult to know what modification the effects of an evolving cloud field would have had on the fields displayed in Figs. 4–10. However, as cloudiness appears as an important modifier of the local radiation budget its effect on the monsoon system would probably be significant. Thus the incorporation of cloud into the model is the next logical step. To include cloud variability a more sophisticated radiation model [such as that developed by Stephens and Webster (1979)] is necessary, plus consistent parameterization of cloud formation, maintenance and decay. At the same time, the findings of Hahn and Manabe insist that orography be included in the system. Consequently, the present paper should be considered as just one part of an ongoing research programme aimed at investigating monsoon dynamics using the "aggregate" philosophy. An ultimate aim is the establishment of a multi-domain version of the WL domain-averaged model so constructed as to produce a useful tool for the study of the total monsoon system and the interaction of the Asian, Australasian and African components.

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REFERENCES

Asnani, G. C., 1967: Meridional circulation in the summer monsoon in South-east Asia. *Nature*, **214**, 73–74.

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

Halley, E., 1686: An historical account of the trade winds and monsoons observable in the seas between and near the tropics with an attempt to assign a physical cause of the said winds. *Phil. Trans. Roy. Soc. London*, **16**, 153–168.

Koteswaram, P., and N. S. B. Rao, 1963: The structure of the Asian summer monsoon. *Aust. Meteor. Mag.*, **42**, 35–36.

Krishnamurti, T. S., and H. H. Bhalme, 1976: Oscillation of a monsoon system, Part I: Observational aspects. *J. Atmos. Sci.*, **33,** 1937–1954.

Lau, Ka Ming W., 1978: Experiment with a simple ocean-atmosphere climate model: The role of the ocean in the global climate. *J. Atmos. Sci.*, **35,** 1144–1163.

Murakami, M., 1977: Spectrum analysis relevant to Indian monsoon. *Pure Appl. Geophys.*, **115,** 1149–1166.

Ramage, C., 1971: *Monsoon Meteorology*. Academic Press, 296 pp.

——, 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, DC, Stock No. 3800-00124].

Sadler, J. C., 1975: The upper tropospheric circulation over the global tropics. Rep. UHMET 75-05, Department of Meteorology, University of Honolulu, 35 pp.

Stephens, G. L., and P. J. Webster, 1979: Sensitivity of radiative forcing to variable cloud and moisture. *J. Atmos. Sci.*, **36,** 1542–1556.

Washington, W. M., and S. M. Daggapaty, 1975: Numerical simulation with the NCAR general circulation model of the mean conditions during the Asian-African summer monsoon. *Mon. Wea. Rev.*, **103,** 105–114.

Webster, P. J., 1972: Response of the tropical atmosphere to local steady forcing. *Mon. Wea. Rev.*, **100,** 518–541.

——, 1973: Temporal variation of the low-latitude zonal circulation. *Mon. Wea. Rev.*, **101,** 803–816.

——, 1979: A model of the seasonally varying planetary scale monsoon. *Dynamics of Monsoons*, R. Pearce and J. Lighthill, Eds., Cambridge University Press, (in press).

——, and K. M. W. Lau, 1977: A simple ocean-atmosphere climate model: Basic model and a simple experiment. *J. Atmos. Sci.*, **34,** 1063–1084.

——, and L.. C. Chou, 1980: Seasonal structure of a simple monsoon structure. *J. Atmos. Sci.*, **37,** 354–367.