Mechanisms Effecting the State, Evolution and Transition of the Planetary Scale Monsoon[1])

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Summary – An attempt is made to identify the basic structure of the planetary scale monsoon and to define and describe the various mechanisms effecting or producing that structure and its variation on a variety of time scales. Both observational and theoretical descriptions are used to define the monsoon system or to point towards problems that require clarification.

The basic mean summer and winter monsoon regimes are described and a discussion presented on their degree of dependence on the distributions of orography, the form and distribution of land and ocean and the state of the interactive oceans adjacent to the major land masses. It is suggested that orography plays a critical role in determining the state of the mean summer and winter circulations behaving principally as an elevated heat source in summer and a mechanical perturber of the mean flow in winter. The spatial variation of sea-surface temperature and the distribution of land are shown to be the *primum mobile* of the mean seasonal monsoon.

Various modes and time scales of the variable monsoon are defined. It is shown that the principal time scale variation or phasing of the *interseasonal* transition is determined by the different response times of the oceans and land areas to variable solar heating. The reaction of the ocean is a strong function of latitude because of its ability to be heated by solar radiation and mixed mechanically by the atmosphere. The return heating distribution imposed on the atmosphere by the ocean strongly effects the circulation. The inclusion of a hydrology cycle adds a further time scale to the monsoon system allowing much more rapid variations to be superimposed on the modulations caused by radiative and sensible heating effects. The inclusion of clouds introduces a further *intra-seasonal* time-scale into the monsoon circulation causing 'break-like' variations of the precipitation pattern and circulation fields during the late summer.

Finally, a brief discussion is given on the various forms of interhemispheric interaction. It is suggested that contrary to earlier theoretical speculation, the equatorial latitudes may be considerably more 'porous' to mid-latitude disturbances than hitherto anticipated.

Key words: Monsoon evolution, monsoon structure.

1. Introduction

Occupying an extremely large part of the variance of the total energy in the global annual cycle, the planetary scale monsoon circulation is a phenomenon with influences and effects which dominate the meteorology of large sections of the globe. For example, the southwest summer monsoon of India and Southeast Asia cannot be thought of as a regional phenomena, but as one part of a system extending from the

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southern oceans of the southern hemisphere to the northern reaches of Eurasia. Furthermore, influences of the system extend in longitude across Africa to the west and probably assert some measure of control to circulations across the Pacific Ocean. The extent of the winter monsoon circulation possesses a scale of influence at least as large as its summer counterpart. Both monsoon regimes exercise enormous economic and social control over large segments of global society. Indeed, the very existence and life of more than 50 percent of the total population of earth is almost completely dependent upon the monsoon circulation.

The objectives of this paper are extremely modest and by no means exhaustive either in the delineation of a problem nor in its explanation. Basically, we will attempt to review observational studies which define the structure of the planetary monsoon circulation or point towards problems that require clarification. We then follow with a discussion of models, both observational or theorectical, which suggest a dominating process or phenomena.

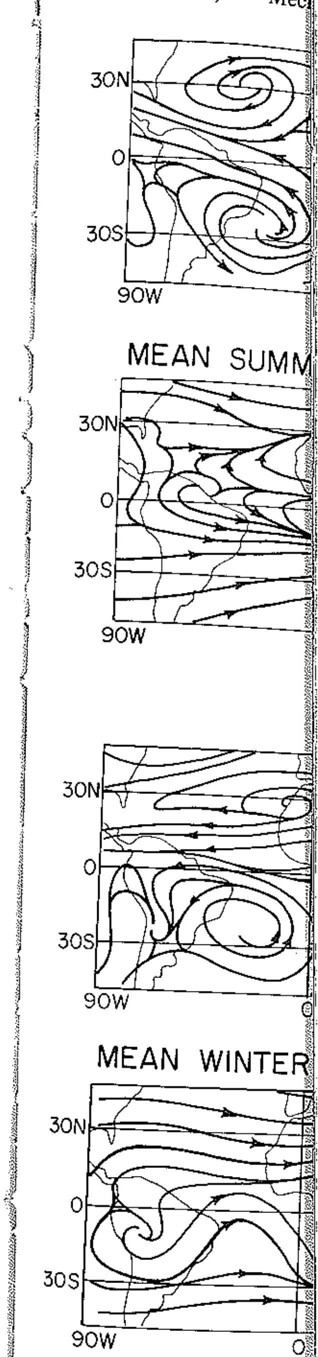
In the next section we will consider the seasonal mean structure of the monsoon system and attempt to break down into a phenomenological substructure the components which constitute its identity. In the third section we consider the transient behavior of the monsoon system on time scales ranging from the variation of the system between its mean solstitial states to shorter period interactions between phenomena of different hemispheres.

To define spatial and temporal bounds for the study, we will consider only planetary scale motions possessing time scales from longer than synoptic to seasonal and annual. This means that disturbances with somewhat shorter time scales, for example, Bay of Bengal disturbances, will not be discussed, even though it is conceded that they form an integral part of the overall monsoon circulation. Rather, we will concentrate on describing the seasonal modulation of the monsoon circulation and planetary scale interaction with other circulation regimes.

2. *Seasonal mean circulation*

(*a*) *Description*

In the simplest sense the monsoon circulation may be thought of as possessing two major regimes. These may be termed the northern hemisphere summer monsoon and the northern hemisphere winter monsoon (hereafter referred to simply as the summer and winter monsoons). The gross features of these regimes may be discerned from Fig. 1 which shows seasonal mean streamline patterns for 850 mb and 200 mb for the periods June, July and August (JJA) and December, January and February (DJF). Such analyses are typical of a number of observational studies of the large-scale tropospheric circulation. Of particular interest are the studies of NEWELL *et al.* (1972), RAMAGE and RAMAN (1972), KRISHNAMURTI (1971 and SADLER (1975). The latter two works utilize commercial aircraft flight path observations which fill in



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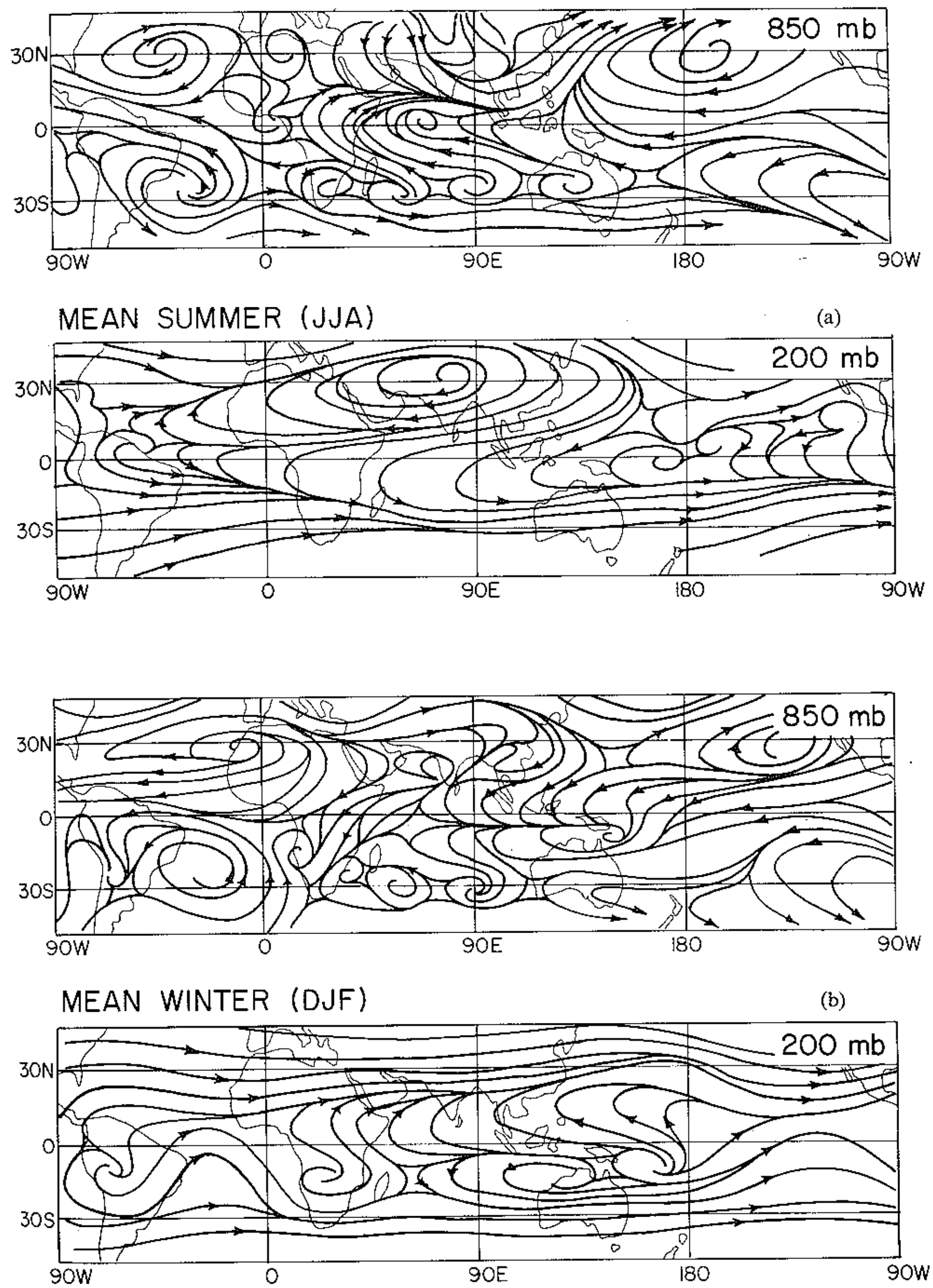

Figure 1
Mean seasonal streamline pattern at 850 mb (upper panel) and 200 mb (lower panel) for (a) the northern hemisphere summer and (b) the northern hemisphere winter. Adapted from NEWELL *et al* (1972) and SADLER (1975).

many of the dataless regions of the ocean areas. Ramâge and Raman's analyses refer specifically to data collected in the International Indian Ocean Expedition. In all analyses, the data in the southern hemisphere is relatively sparse compared to its northern counterpart.

The most significant features of the monsoon circulation may be summarized as follows. In the lower troposphere in summer a strong cross-equatorial flow emerges from the southern hemisphere subtropics and forms a generally west to southwest flow across the Indian Ocean and Southeast Asia. This is joined by a relatively dry low-level flow from the northwest which merges, in a climatological sense, over the northern Indian Ocean. At 200 mb an intense anticyclone centered over the Tibetan plateau dominates the upper troposphere and provides intense easterly winds along and to the north of the equator from the South China Sea to eastern Atlantic Ocean. This region also corresponds to the return cross-equatorial flow to the subtropical westerlies of the southern hemisphere. In winter the basic circu-

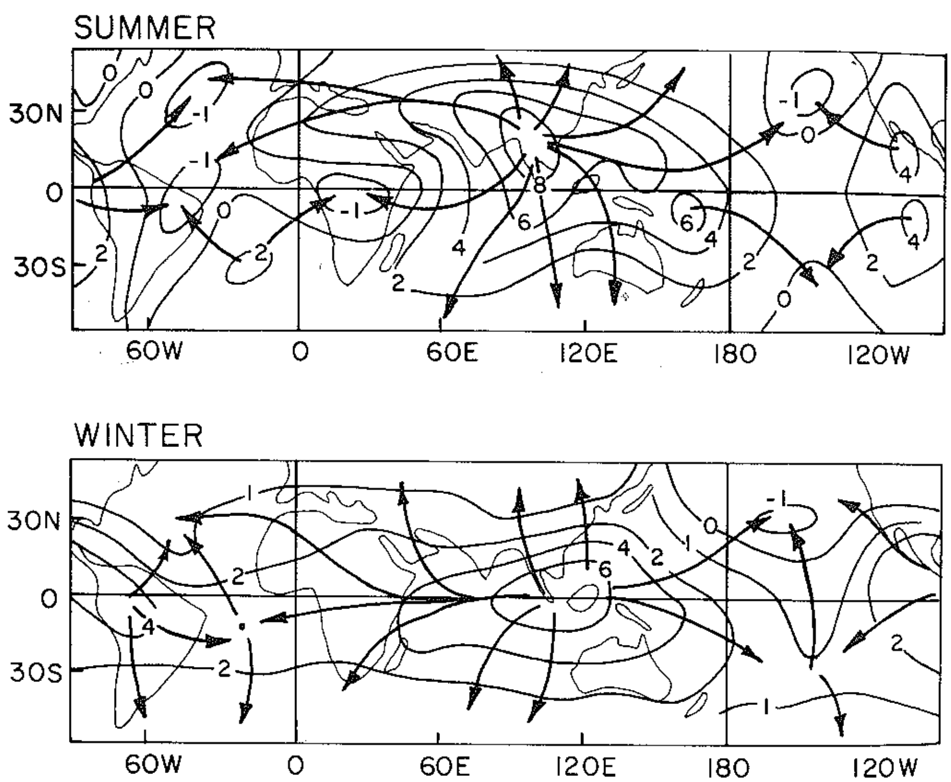

Figure 2
Mean seasonal distributions of velocity potential (solid lines) and vectors of the divergent part of the mean wind field (arrows) for the northern hemisphere summer (upper panel) and winter (lower panel). Adapted from KRISHNAMURTI (1971a) and KRISHNAMURTI *et al.* (1973).

lation reverses. At low levels the flow is off the Asian continent and forms the northeast monsoons of the China Sea and Indian Ocean regions. On crossing the equator the flow converges in the Indonesian–Australian region to provide the somewhat weaker southern hemisphere summer monsoon. A return flow at 200 mb emerges from the southern hemisphere tropics, but the intensity of the anticyclone, or the resultant equatorial easterlies, do not match that of the summer monsoon of the northern hemisphere. After crossing the equator the upper tropospheric mean flow merges with the strong northern hemisphere westerlies which are, in an average sense, to the south of the Himalayas.

Although much of the structure of the mean monsoon flow may be obtained from mean streamline analyses such as those shown in Fig. 1, a breakdown of the fields into divergent and non-divergent parts is instructive. Such analyses were performed by KRISHNAMURTI (1971a) and KRISHNAMURTI, *et al.* (1973). Figure 2 shows the mean velocity potential distribution (i.e., divergent part of the wind) at 200 mb for both July and January. The solid lines, which are orthogonal to the contours, denote the direction of the divergent part of the wind. The details of the distributions are not important and perhaps even the data does not permit us to note but the grossest features. However, from the analyses emerges a rather significant point. During each season there exists one major maximum in the monsoon region and a weaker maximum in Central America. In the summer the major maximum is located over the northern Bay of Bengal–Burma region with elongated maxima extending eastward across the Phillipines and the South China Sea. In winter the maximum velocity potential lies over the South China Sea–Indonesian region. So extensive are the summer and winter maxima that the mean divergent motions emerging from them are of planetary scale. This allows one to construe that the convective regions associated with both the summer and winter monsoons must be dominant over most phenomena in the tropics and perhaps even at higher latitudes.

(*b*) *Mechanisms*

The traditional explanation for the circulation distributions of both the winter and summer monsoons is based upon the interaction of the seasonally varying insolation and the distribution of the continents and oceans. Originating with HALLEY (1686), the concept of the monsoons as a planetary scale land–sea breeze circulation has dominated meteorology for nearly three centuries. Indeed, despite the accumulation of considerable new data and the rapid advancement of theoretical meteorology, the planetary scale rotational sea-breeze model has persisted. The improvements and modifications deal mostly with the nuances of the circulation and its seasonal phasing (WALKER, 1972).

Possibly four major factors render the winter and summer monsoons completely different phenomena each possessing its own distribution of parameters and character; neither being the mere mirror image of the other, as may be seen in Fig. 1. Three of the factors depend directly upon the physical form of the earth. They are the distribution of mountains in the monsoon region, the relative sizes of the continental land masses of Australia and Asia and the distribution of the large mountainous islands which contribute the 'maritime' continent (RAMAGE, 1968) made up of Sumatra, Borneo, New Guinea plus numerous other smaller islands characterized by equally high terrain, all of which are surrounded by extremely warm ocean water. The fourth factor is the distribution of sea surface temperature and, in particular, its temperature during the established summer and winter monsoon (July and January, respectively). Ultimately, of course, the sea-surface temperature depends considerably upon the distribution of land and sea.

The large orographic structures of the monsoon regions appear to possess dif-

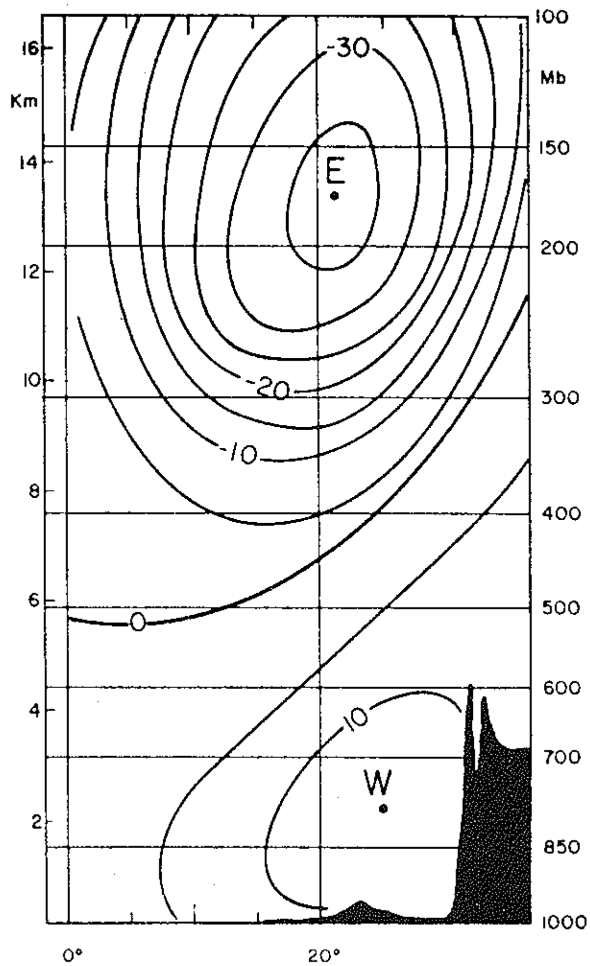

Figure 3
Latitude-height section of the computed zonal wind component distribution of MURAKAMI *et al.* (1970). Shaded region denotes the Himlayas.

ferent roles in the summer and winter monsoon regimes. For the summer monsoon, FLOHN (1950, 1960) was first to suggest that the role of the Himalayan region may be different from that of a mechanical deviator of the mean summer flow. It was suggested, in fact, that the Tibetan Plateau acted as an elevated source of sensible heat which was responsible for the generation of the warm upper tropospheric anticyclone which dominates this region in Fig. 1. Later RIEHL (1959) suggested that the latent heating was the dominant process. Using satellite data which showed the central and southeastern parts of Tibet to be snow free through most parts of the year, FLOHN (1968) was able to confirm earlier work (STAFF MEMBERS, ACADEMIA SINICA, 1958) that the Plateau was indeed a candidate for rapid heating with the advancing sun in spring but conceded that in the southeastern regions, where precipitation occurs many weeks before the onset of widespread precipitation over most of India, latent heating was probably the dominant process.

The role of the Himalayas in the maintenance of the mean summer circulation has stimulated a number of numerical studies. MURAKAMI *et al.* (1970) studied the role of orography by considering a number of numerical experimenta in which cases with mountains were contrasted with cases without mountains. The model was an 8-level two-dimensional hemispheric model extending from the equator to the pole. Cross-equatorial flow and deviations from the mean zonal state (disturbances) were precluded from the analysis. Without mountain effects only relatively weak easterlies in the upper hemisphere were produced. Only with a condensation cycle and the inclusion of the Tibetan Plateau was the mean field similar to the observed structure. The computed field of the zonal velocity component is shown in Fig. 3. It would seem that the model results of MURAKAMI *et al.* confirm the thermal role of the Tibetan Plateau.

More recently, WASHINGTON and DAGGUPATY (1975) and HAHN and MANABE (1975) have used global general circulation models to concentrate on the dynamics of the summer monsoon. In both studies many of the gross features of the monsoon are reproduced although computational problems arising from the inclusion of the abrupt Himalayan Plateau itself, affects some parameters such as the mean precipitation distribution, which, to varying degrees, often is too far southward of the observed maximum. A discussion of some of these points is given by SADLER and RAMAGE (1976) and HAHN and MANABE (1976).

The study of HAHN and MANABE (1975) is especially geared towards studying the effects of mountains on the summer monsoon which is achieved by comparing mountain and no-mountain cases. It was found that the mountains were responsible for the formation and maintenance of the warm upper-tropospheric anticyclone via latent heating. Without orography, the summer monsoon was found to slowly extend into Asia and arrive at a poleward limit some 10° short of the case with mountains. Whereas the conclusions cannot be considered definitive until the climatology of the control case matches more closely the observed climatology, the results are suggestive of an extremely important control exerted by the large-scale orography.

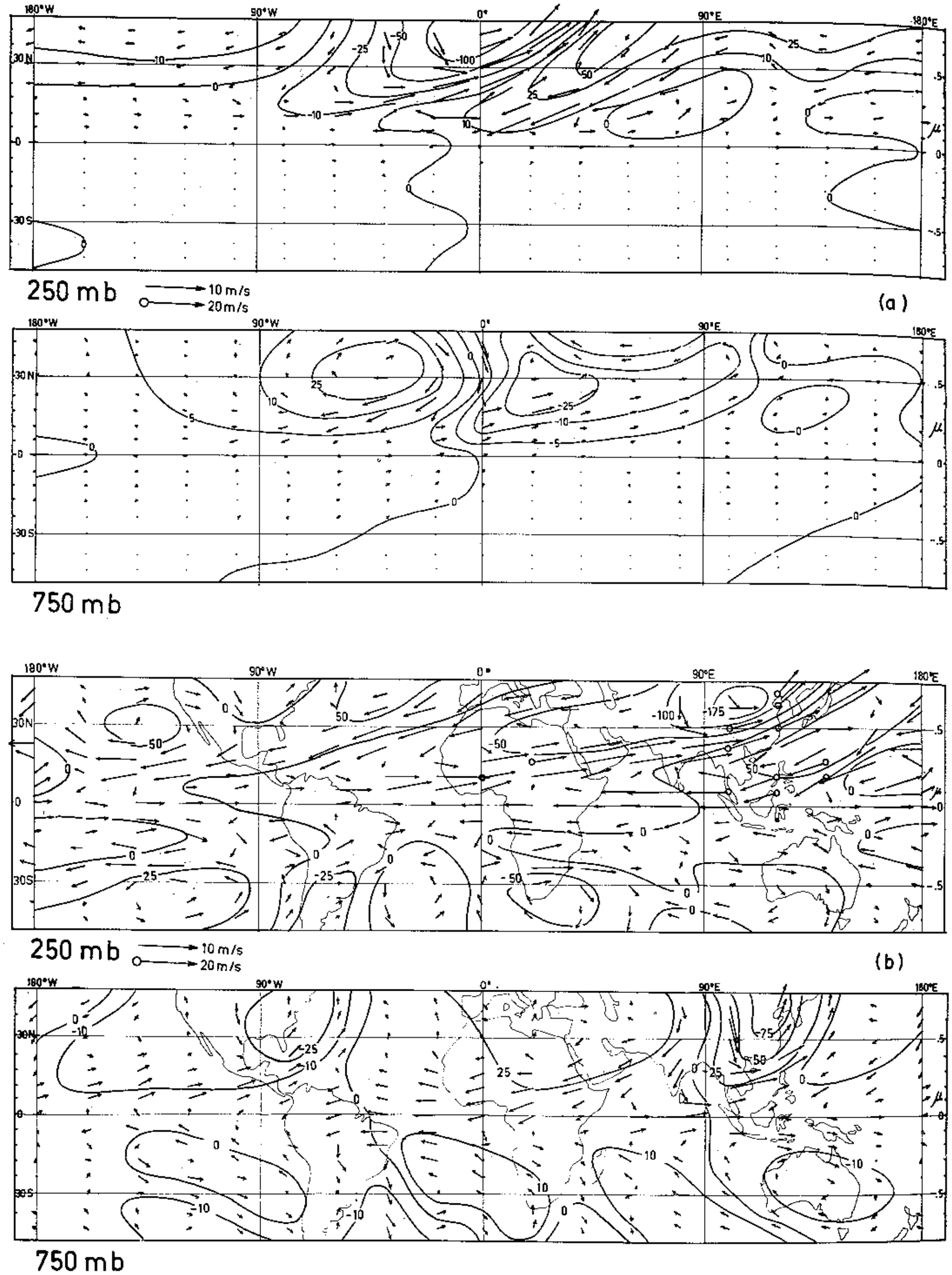

Figure 4
Perturbation velocity distributions at 250 mb and 750 mb for (a) imposed orographic barrier located at 25°N and (b) imposed realistic distributions of heating and orography. Both (a) and (b) refer to the northern hemisphere winter. From WEBSTER (1972).

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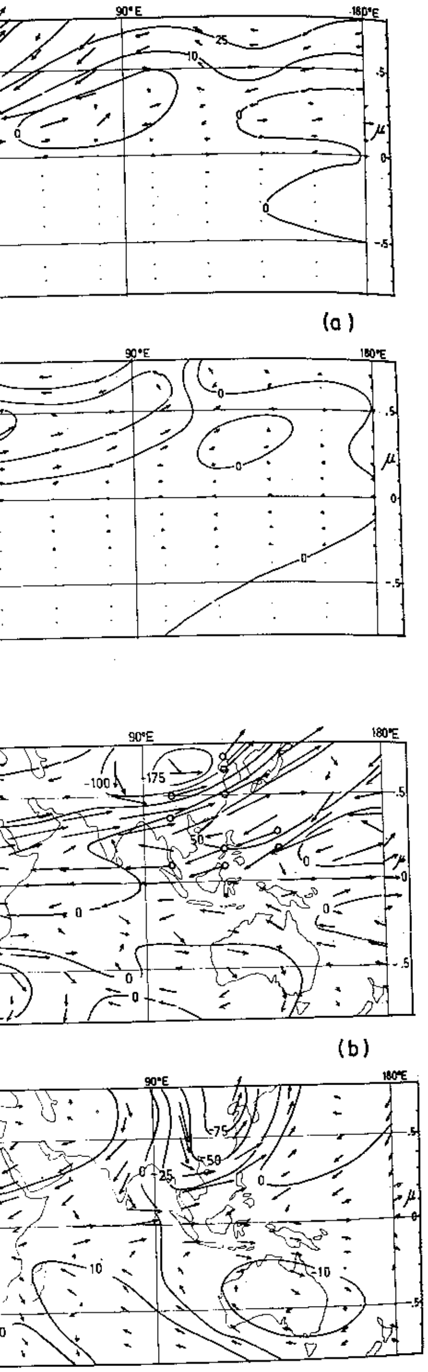

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During the northern hemisphere winter the role of the Himalayas changes substantially. With the maximum insolation occurring in the summer hemisphere, the elevated regions of Asia lose their ability to anomalously heat the middle and upper troposphere through winter. With the mean westerlies considerably poleward of their summer position, the Himalayas act as a strong mechanical perturber of the mean zonal flow. Such an effect is evident in Fig. 4 from WEBSTER (1972). Using a simple linear primitive equation model, Webster attempted to assess the relative roles of orography and heating at low latitudes. Figure 4a shows the 250 and 750 mb perturbation (i.e., with the zonally averaged flow subtracted out) circulation fields resulting from the interaction of a mean flow and an orographic structure situated at 25°N and the prime meridian. For our purposes it is sufficient to note the extremely high amplitude trough-ridge system which dominates flow downstream and in the vicinity of structure. In Fig. 4b, a realistic orographic distribution has been included together with the seasonal heating field. Comparing Figs. 4a and 4b, WEBSTER (1972) concluded that the orographic forcing from the Himalayas was a dominant process during the winter, at least north of the equator in the South-east Asian region.[4])

An understanding of the degree of control imparted by the distribution of ocean and continents on the earth's climate emanates from HALLEY (1686). The importance of the longitudinal differences produced by the ocean–land contrast and the effect on the lower boundary heating caused by the different character of the lower boundary was initially emphasized by ROSSBY *et al.* (1939) and SMAGORINSKY (1953), at least for the mid-latitudes. Considering principally the very low-latitude flow, WEBSTER (1972) suggested that the variations in the latent heating distribution arises, in part, from the land–sea contrasts and the variations in sea-surface temperature were responsible for much of the character of the large-scale flow. The basic character of the circulation was explained in terms of the equatorially-trapped Kelvin wave which matched both the selective response and the horizontal scales of the motion. However the vertical scale of the Kelvin wave did not match observations and it was not until the work of CHANG (1977), who illustrated the effect of dissipation on the vertical structure of equatorial modes, that the great east–west overturnings shown in Fig. 2 which dominate much of the mean low-latitude, could be seen to be atmospheric manifestations of viscous controlled Kelvin waves.

In an attempt to investigate further the role of land and ocean and the dynamic interaction of the oceans and the atmosphere, WEBSTER and LAU (1977a, b) and LAU and WEBSTER (1977) developed a simple coupled model. Much more complicated coupled ocean–atmosphere models have been developed (e.g. MANABE *et al.*, 1975) but most of these are sufficiently complicated to deny extended integration or too

[4]) Professor T. Murakami (private communication) contends that the different mean circulation features of early and late winter are a result of orographic forcing in the region to the east of Southeast Asia. In later winter, the mean zonal westerlies are over and to the south of the Himalaya such that the position and strength of the trough constrain disturbances in the South China Sea. In early winter the westerlies are further north and the South China Sea is disturbed.

simple (e.g., PIKE, 1971) to include ocean–land contrasts. The domain-averaged model fits between these two extremes. The physical structure of the model is shown in Fig. 5. Basically, the simplest prototype of the model, shown here, consists of two atmospheres of finite longitudinal extent surmounting, respectively, a land mass which is truncated in latitude at 18°N and a dynamic and interactive ocean. A similar ocean lies to the south of the truncated continent. The equations of motion which represent the atmosphere are averaged over the longitudinal extent of each domain. The domains 'communicate' via interaction terms at each domain. The ocean domains are also dynamic and are represented by an advective mixed layer model which allows transport of heat by the ocean. The primitive equation atmospheric model and ocean model are presented in detail by WEBSTER and LAU (1977). A later version of the ocean model allows for wind-driven circulations and longitudinal temperature variations (LAU, 1977).

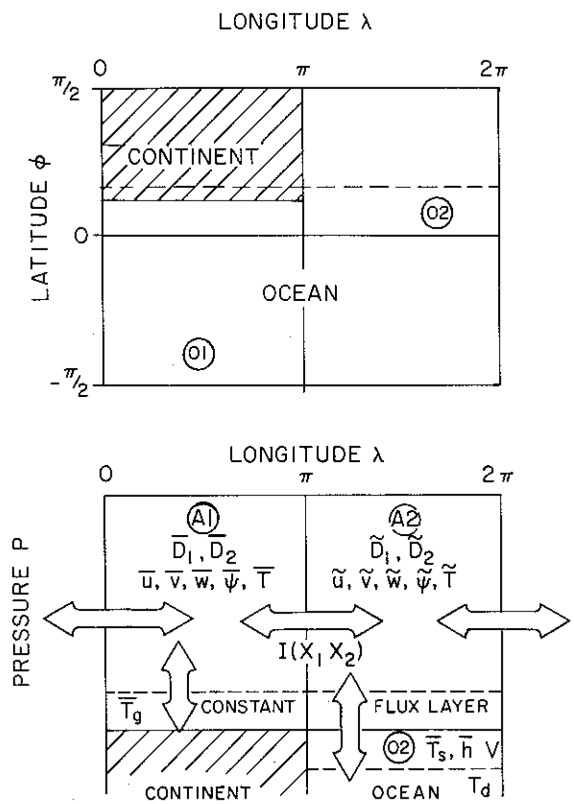


Figure 5

Latitude-longitude (upper diagram) and latitude-pressure (lower diagram) of a two-atmosphere and two-ocean domain-averaged model. One atmosphere A1 surmounts a truncated continental region with an ocean to the south (O1), whereas A2 lies over O2. Variables are defined and interdomain fluxes and interaction terms are denoted by arrows. From WEBSTER and LAU (1977).

Figures 6 and 7 show the temperature and zonal velocity component response of the domain averaged model described in Fig. 5 to summer solstice and winter solstice insolation. The vertical dashed line in all cases denotes the land–sea boundary. Solid lines signify the fields over the ocean–continent domains (A1 in the lower section of Fig. 5) and dashed lines for the atmosphere which lies over the uninterrupted ocean (A2). The surface temperature field emphasizes the importance of ocean–atmosphere contrast. At the land–sea boundary an abrupt change in surface temperature is apparent during the summer and the land temperature remains significantly warmer than that of the ocean further to the north. At the winter solstice the land region is considerably cooler than either the adjacent ocean or the ocean equatorward. To

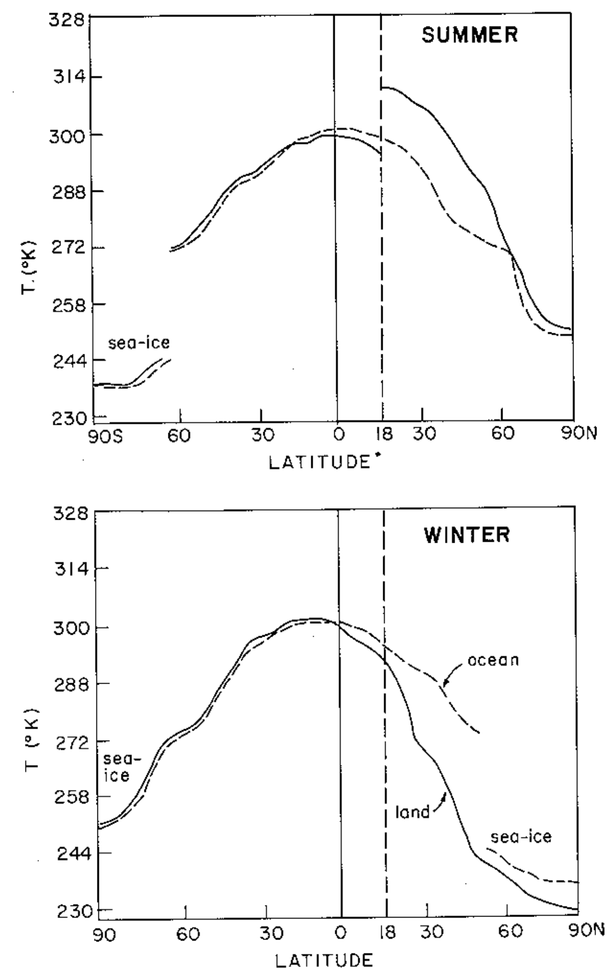


Figure 6

Surface temperature for the two atmospheric domains A1 and A2 defined in Fig. 5, for the winter and summer solstices. Dashed lines refer to A2 and solid lines to A1. WEBSTER and LAU (1977b).

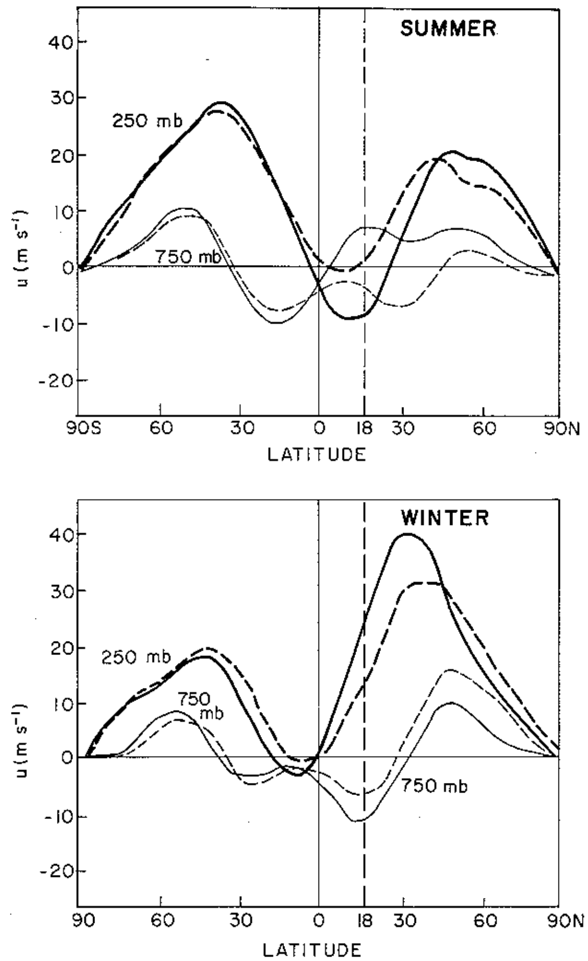
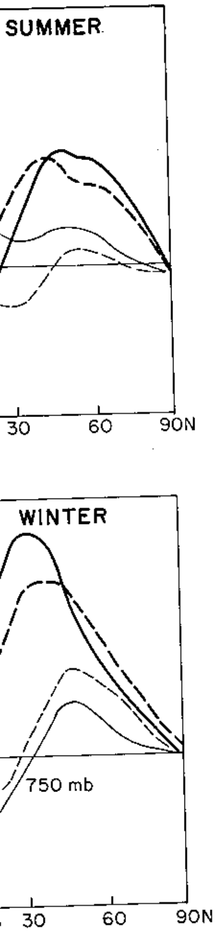


Figure 7
Same as Fig. 6 but for the zonal velocity component. WEBSTER and LAU (1977b).

the north the land has cooled at a much more rapid rate than the ocean although the sea–ice margin has advanced considerably equatorward. Between the solstices, the equator to land-margin temperature gradient has reversed and the dramatic effect this has had upon the upper tropospheric mean flow is apparent from Fig. 7. During the summer a relatively strong easterly maximum[5]) exists at the land margin whereas during winter it has been replaced by a substantial westerly wind. Similarly at 750 mb the strong summer westerlies have been replaced by winter easterlies. Over the ocean

[5]) The results shown in Figs. 6 and 7 refer to a case where moisture has been omitted. Inclusion of a hydrology cycle, including ground water immediately reduces the surface temperature over land between 18°N and 30°N by 5°C but simultaneously increases the magnitude of the upper troposphere easterly maximum by nearly a factor of three.



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(A2) less dramatic but similar variations have occurred but mainly through zonal interaction than by variations in the surface temperature, which in the equatorial regions (see dashed line in Fig. 6) remains effectively constant throughout the year.

3. *Transient circulations*

a. *Description*

Consideration of only the mean seasonal structure of the monsoon circulation, as displayed in Figs. 1 and 2, belies the importance of the transient stage through which the atmosphere must transgress during the equinoctial (or transitional) seasons and also the variations which occur at time scales less than seasonal. SADLER's (1975) careful analyses of the mean *monthly* upper tropospheric circulation are important and helpful in understanding the transition. The picture which emerges from Sadler's analyses is not simple, however, and does not display mere linear variation between the summer and winter seasons. Instead the transitions are characterized by rapid variations from one state to another with timing that varies spatially. Such variability in other atmospheric parameters through the annual cycle has been well documented by RAMAGE (1971) and RAMAGE and RAMAN (1972).

Figure 8 shows time-latitude sections of the zonal wind component adapted from Sadler's analyses. Six latitude-time sections of the zonal velocity component (ms^{-1}) are shown for the meridians 80°E, 100°E, 150°E, 180° and 150°W. Heavy lines denote ridge positions and the dashed lines represent troughs. All six sections show extreme seasonal variation with distinctive winter and summer distributions and rapid transitions between them. The sections may be divided up into two groups which roughly categorize their annual variation.

The first group (80°E, 100°E, and 115°E) are characterized by a rapid development of easterly winds between the equator and 25°N to north of 40°N. In the southern hemisphere a rapid increase in the westerlies appears to match, or perhaps lead, the development of the easterly maximum in the northern hemisphere. The suggestion that the development of the westerlies in the 80°E–180° sector leads the easterly maximum in the northern hemisphere was originally made by RADOK and GRANT (1957). They further suggested that the diminuation of the northern hemisphere summer easterlies, and therefore the return of the northern hemisphere westerlies, leads the decay of the southern hemisphere westerlies.

The second group of sections (150°E, 180° and 150W) does not show the development of strong equatorial easterlies during the summer. On the contrary, the 150°W section exhibits westerlies during the entire year along the equator and only weak easterlies are evident in the other sections. North of the equator convoluted structures rapidly evolve in late spring and summer. At 180° and 150°W a double trough-ridge system develops with an attendant double easterly maxima and intervening strong westerlies ($\sim 20\ ms^{-1}$) near 20°N. Such complicated structures quickly replace the

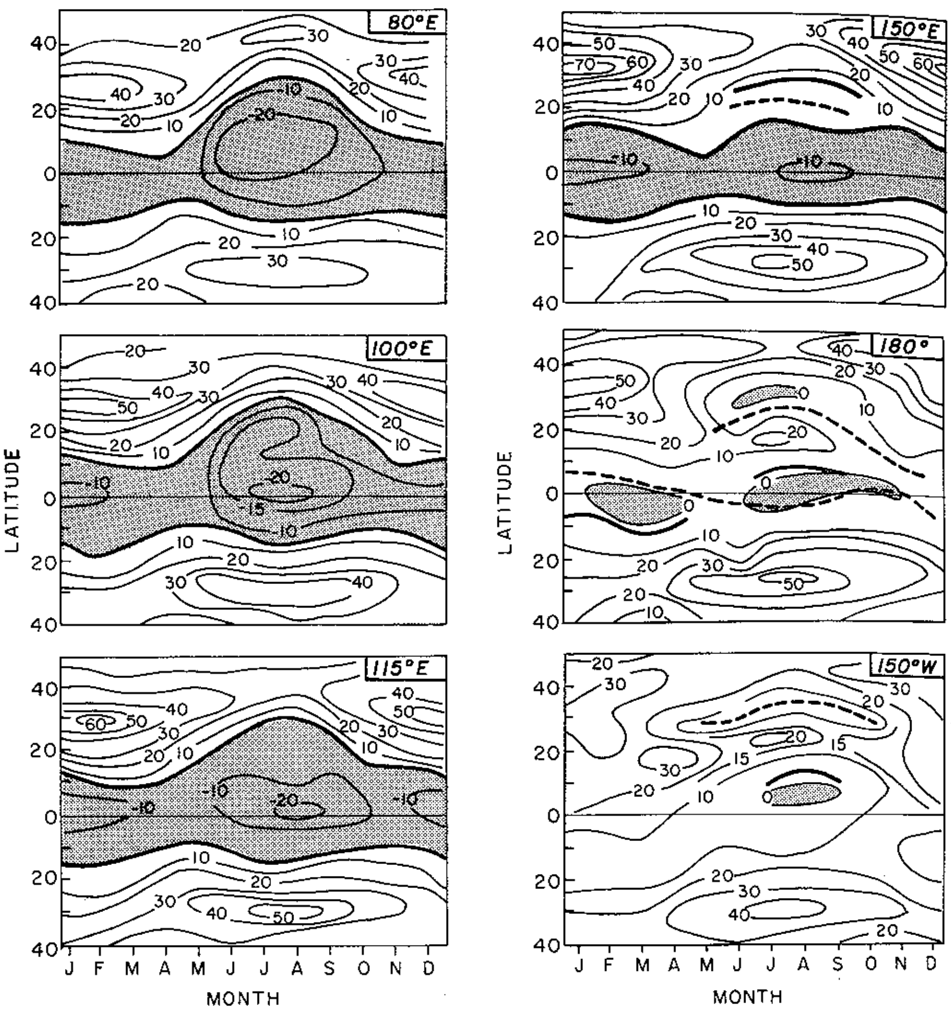
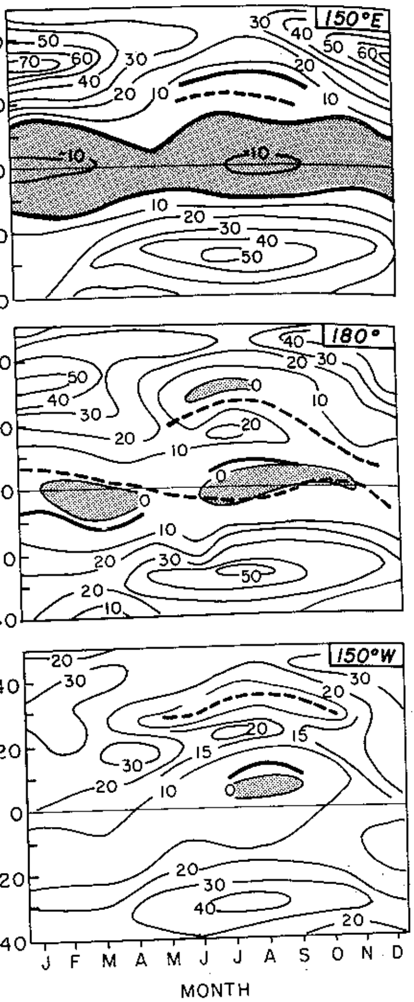


Figure 8

Latitude-time sections of the zonal velocity at 200 mb component for selected meridians. Dashed lines denote mean troughs and heavy solid lines show mean ridges. Units are m s^{-1}. Adapted from SADLER (1975).

intense westerly maxima in the most western sections, which reappear equally rapidly in the northern hemisphere fall. In the southern hemisphere the structure appears somewhat simpler although, in the same manner as the first group of sections, the westerly maximum assumes its southern winter position and magnitude with some rapidity.

The variations in the mean flow in the southern hemisphere cannot be well represented by the few latitude-time sections shown in Fig. 8. This is because during the southern hemisphere summer the westerly maximum is further to the south and west than can be seen in Fig. 8. Figure 9 shows a compilation of the hemispheric



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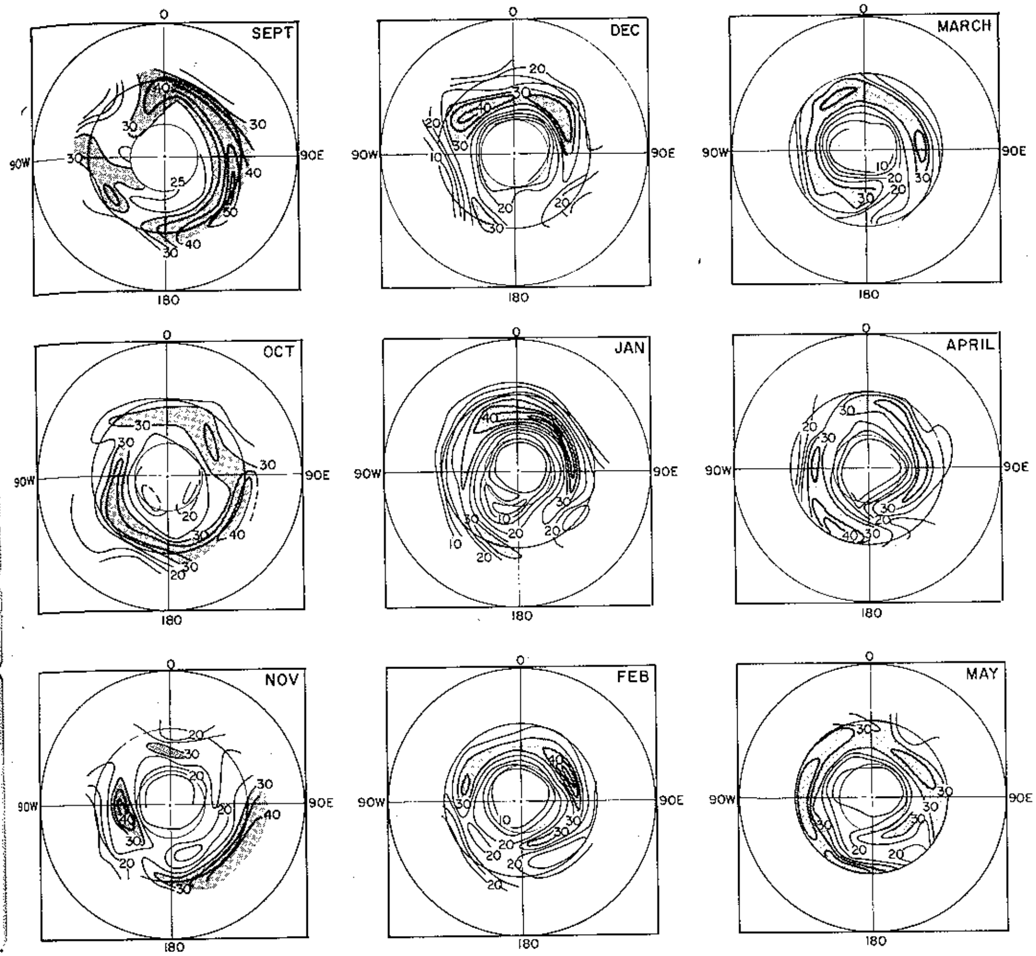

Figure 9
Mean monthly distribution of the zonal velocity at 200 mb as determined by WEBSTER and CURTIN (1975) from the EOLE epxeriment, September 1971 to July 1972.

distribution of the 200 mb zonal velocity component compiled from data obtained during the EOLE experiment (WEBSTER and CURTIN, 1975). Of particular interest is the rapid change of amplitude and phase of the monthly mean zonal flow. In the southern hemisphere summer, the westerly maximum is located in the sector between 90°W–0°–90°E and near 40°S. In the southern spring (and also in the southern winter) the maximum changes phase by nearly 180° and moves equatorward by 10° of latitude. In the northern hemisphere large-scale seasonal variations also occur in the strength and position of the westerly maximum. However, these changes are perhaps more associated with latitudinal shifts and variations in intensity that can be more easily attributed to variable seasonal forcing.

Besides the considerable change in the monsoon structure which occurs on a seasonal or even a monthly time scale, large variations occur interannually. This may be seen, for example, from SADLER *et al.* (1976) who has compiled and analyzed eight separate years of satellite data over the Pacific Ocean region. Whereas certain features are often discernable from year to year, large variations in cloud cover are also obvious. It is assumed that mean cloud cover and mean circulation are related, then the existence of the anomalies is indicative of variations in circulation. Similar variability appears to occur during the 'onset' and 'withdrawal' phases of the southwest monsoon. This has been emphasized by RAMAGE (1971) who finds, from observational studies of many years of data, including the International Indian Ocean Expedition, little evidence of a single or reproducible sequence of events but rather many phenomenological variations placed upon the background of a somewhat irregular annual variation. Whether this viewpoint is true or whether a unifying theory lurks amidst the data waiting elucidation is difficult to foretell.

A further factor which may play a strong role in determining the transient state of monsoon is the interaction between the two hemispheres. Unfortunately, interhemispheric relationships have received little observational and theoretical consideration, even though the interhemispheric nature of the monsoons, at least in its mean solsticial state, is readily apparent from Fig. 1.

The mean interhemispheric flow in summer in the Indian Ocean region is confined to a narrow low-level jet off East Africa whereas a much broader return flow exists in the upper troposphere (FINDLATER, 1969, 1971). In winter a southward flow in the western Indian Ocean is still characterized by a maximum stream off the African coast. However, in the winter FINDLATER (1969) estimates that the flow in this region accounts only for 10 percent of the total cross-equatorial flow in the Indian Ocean whereas its stronger counterpart in summer accounts for some 50 percent.

An intra-seasonal variability in the low-level cross-equatorial flow in the Indian Ocean region has been suggested by KRISHNAMURTI and BHALME (1976) who infer a phenomenological linkage from the southern Indian Ocean to the Indian subcontinent during the summer monsoon. The meridional teleconnection was apparent from variations in the Mascarene High which were transmitted northward via the cross-equatorial Somali jet stream and then across the Indian Ocean to the monsoon trough where it was speculated that imposed variations in stability effected the development of monsoon depressions. At this stage the cross-equatorial teleconnection should be treated as tentative and speculative until considerably more data becomes available in order to enhance the statistical significance of the chain of events. However, even with the small data set that is available, the relative phase relationships of the various quantities do establish a sequence which probably allow one to conclude that teleconnections did exist, at least during the data period. Furthermore, long-period variations have been noted in the southern hemisphere troposphere by WEBSTER and KELLER (1975) which could effect the structure of the Mascarene High.

Less observational definition exists for the cross-equatorial aspects of the winter monsoon mainly because of the data scarcity in the central Indian Ocean and over Indonesia. Outflow from the cold Asian landmass towards the warm equatorial regions and the southern hemisphere occurs at low levels along its southern and eastern flanks. Rather than being a continuous stream flowing equatorward to form the northeast monsoon, the outflow occurs as irregular spurts or surges with intervening lulls which appear to be tied to the dynamics of the entire monsoon system. Such cold surges are reflected in the variations of atmospheric parameters at least as far south as Indonesia and occasionally into the summer hemisphere. Their longitudinal extent appears limited in the west due to the barrier of the Himalaya which shields India from the influence of the low-level surges. The development or modification of near-equatorial disturbances in the South China Sea appears to be connected to the occurrences of cold surges although the details of the connection are obscure. In turn, the development of disturbances in the northern hemisphere often corresponds to a temporary cessation of activity in the North-Australian monsoon region.[6])

The variations of the monsoon cycle and the cross-equatorial flow described above could be thought of as local manifestations of meridional circulations. With a meridional circulation model in mind, a number of observational studies have emerged to account for the position of the mean trough(s), precipitation distributions and cloudiness structures, especially for the summer monsoon. Basically each model pattern shows strong ascending motion over the land areas with descent occurring equatorward and in the winter hemisphere. This simple picture, established for the summer circulation by KOTESWARAM (1958) and many others, has been subject to a number of complicating adaptations, for example by ASNANI (1967), in order to account for nuances of the gross circulation which have emerged via an ever improving data set. Asnani replaces Koteswaram's single equatorial cell for the summer monsoon with a combination double cell which enhances descent near the equator and ascent near 5°N. Further, to account for disturbed conditions to the south of the equator, Asnani adds a second and distinct southern hemisphere circulation. Whereas some insight may be gained by continuing to add mean circulations to account for all new variations of differences introduced by new data, it must be considered a dubious practice which does not even allow for the consideration of rectification of shorter period phenomena which may play an important role in producing the discerned mean state.

An appreciation of the complication involved in even the structure of mean meridional circulations in the winter monsoon may be gained from BLACKMAN *et al.* (1977) who have compiled schematic representations of the *winter* mean meridional circulations along meridians through central Asia and through the central Pacific Ocean, or, upstream and downstream of the upper tropospheric jet stream located over Japan. Such a representation is reproduced in Fig. 10. For our purposes it is

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[6]) Personal communication: Professor T. Murakami, University of Hawaii.

sufficient to concentrate upon the zonal variation of the structure. Basically, the upstream (or continental) section is dominated by a strong thermally *direct* meridional circulation whereas the downstream (or oceanic) section is dominated by an equally strong but thermally *indirect* meridional circulation. Both circulations are consistent with the eddy momentum and heat fluxes and the distribution of the disturbed weather in the winter monsoon circulation. The complexity in the zonal direction has the immediate implication that the monsoon circulation is a strong function of longitude as well as latitude. Unfortunately, Blackmon *et al.*, have not as yet extended their analyses to include the northern hemisphere summer circulation.

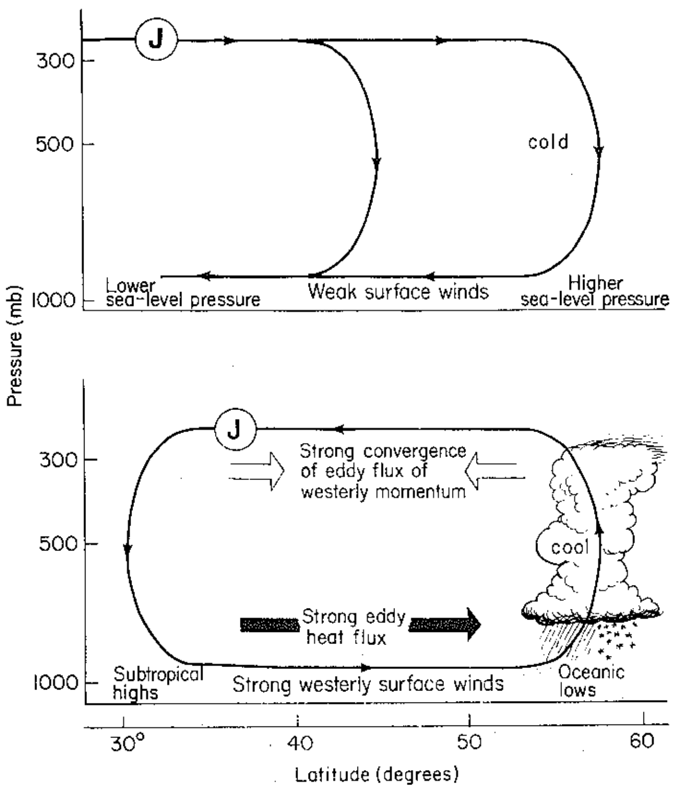

Figure 10
Schematic description of the wintertime mean meridional circulation through Central Asia (upper panel) and the Western Pacific Ocean (lower panel). From Blackmon *et al.* (1977).

Cross-equatorial interactions of a much shorter time scale which may influence the monsoon circulation is apparent from Fig. 11 where the 200 mb perturbation kinetic energy distribution calculated by Murakami and Unninayar (1977) is displayed. In the equatorial regions, the most notable feature is the relative maximum to the east of the date line which bridges the gap between the mid-latitude maxima of

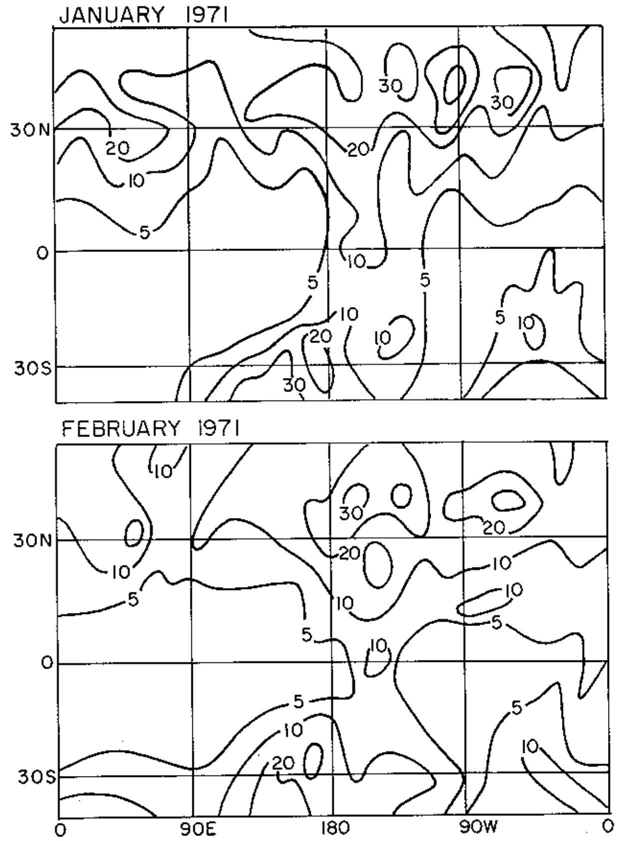

Figure 11

Perturbation kinetic energy distribution at 200 mb for January and February 1971. From MURAKAMI and UNNINAYAR (1977).

each hemisphere. A second maximum, which is somewhat smaller, exists near 45°W. Both regions correspond to locations of slowly varying and strongly tilted upper-tropospheric disturbances (WEBSTER and KELLER, 1975) and centers of anomalously large poleward momentum maxima (WEBSTER and CURTIN, 1975) especially in the Pacific Ocean. Both correspond to regions of mean seasonal westerly winds in the upper troposphere (see Figs. 8 and 12). The Pacific Ocean zone of maximum kinetic energy lies to the south of the upper tropospheric mid-Pacific trough of the northern hemisphere which has been linked to transients of the northern hemisphere summer monsoon (COLTON, 1973). The possible nature of an interhemispheric interaction in these particular locations will be discussed in the next section.

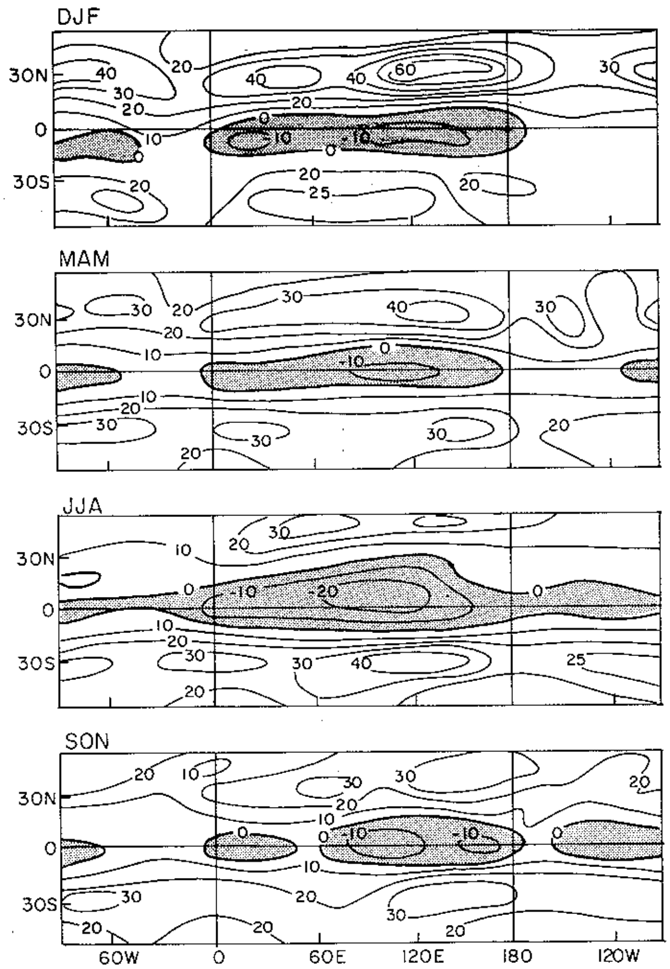

Figure 12

Mean seasonal distributions of the zonal velocity component at 200 mb as a function of longitude and latitude. Shaded areas denote easterly winds. Adapted from NEWELL *et al* (1972), and SADLER (1975).

b. Mechanisms

Few theoretical or numerical studies have considered the transient behavior of the monsoon system on annual time scales. This is because the models are either too simple to allow the inclusion of a land–ocean contrast, in which the ocean is capable of a dynamic feedback, or too complex to allow the required extended integration. As the domain averaged model described in WEBSTER and LAU (1977) was designed to specifically study large-scale phenomena on an annual time scale, we will use it to illustrate physical processes which may be important in determining the manner the large-scale monsoon flow may vary.

To accomplish the investigation we will consider three versions of the domain averaged model all of which contain similar domain configurations shown in Fig. 5. In the first case we will omit a hydrology cycle in order to isolate the annual cycle produced by purely land and ocean contrasts and transient variation in the state of the ocean. Similar experiments are discussed by WEBSTER and LAU (1977b) and LAU and WEBSTER (1977). Here we show results for a three-domain model which is the same as that shown in Fig. 5 except for a second ocean which is added to the east of O2 (LAU, 1977). This allows for a better ocean model in which wind-driven circulations and transportations are included. The improved ocean model is described in detail by LAU (1977). In the second experiment we include a hydrology cycle and in the third experiment allow clouds to interact with both the incoming and outgoing radiation fields. The inclusion of a hydrology cycle and its effects are discussed by WEBSTER and CHOU (1977). The domain configuration used in the latter two experiments is not precisely compatible to that of the first experiment as are the land–ocean domain of Fig. 5 which is considered. However, experiments with a dry version of the one-domain model roughly approximate the condition in the like domain in the three-domain version so that a meaningful comparison may be made.

Figures 13 and 14 show the upper tropospheric zonal velocity component and the surface temperature variation as a function of latitude and time for the three domains. A1 refers to the atmosphere over the eastern ocean, A2 to the atmosphere over the western ocean and A3 to the atmosphere over the truncated land domain. The dashed line in A3 denotes the 18°N coastline and the heavy isopleth in Fig. 14 represents zero zonal wind. Comparing Fig. 13 with Fig. 8, the observed latitude-time sections at specific longitudes, we find considerable similarity. Most importantly, variation in magnitude is a maximum over land where the equatorial easterlies and mid-latitude westerlies are maximum. A similar variation in latitude occurs although perhaps less abrupt in the simulated case. In A1 and A2 the variations are less severe and out of phase with the changes over land. Maxima tend to occur slightly later than over land, which is also apparent in the surface temperature structure shown in Fig. 14 and may be attributed to the phase lag of the ocean surface temperature. Interestingly, the upper atmospheric fields over the ocean tend to *lead* the ocean surface parameters slightly in phase but still lag their counterparts over land. This may be interpreted as an indication of the lateral effect of the continents on the oceans (and vice versa). An interesting similarity between observed and simulated may be seen between the zonal flow in A1 (the upper troposphere over the eastern ocean) and the observed section along 150°W in Fig. 8. Although the model cannot simulate the nuances of the mid-Pacific trough structure due to its extremely coarse longitudinal resolution, it is interesting to note that westerlies are maintained at the equator throughout the year.

Much of the surface temperature structure shown in Fig. 14 was alluded to in the discussion of the zonal velocity component. Most striking is the variation of the temperature over the land and the degree to which the temperature only slightly lags

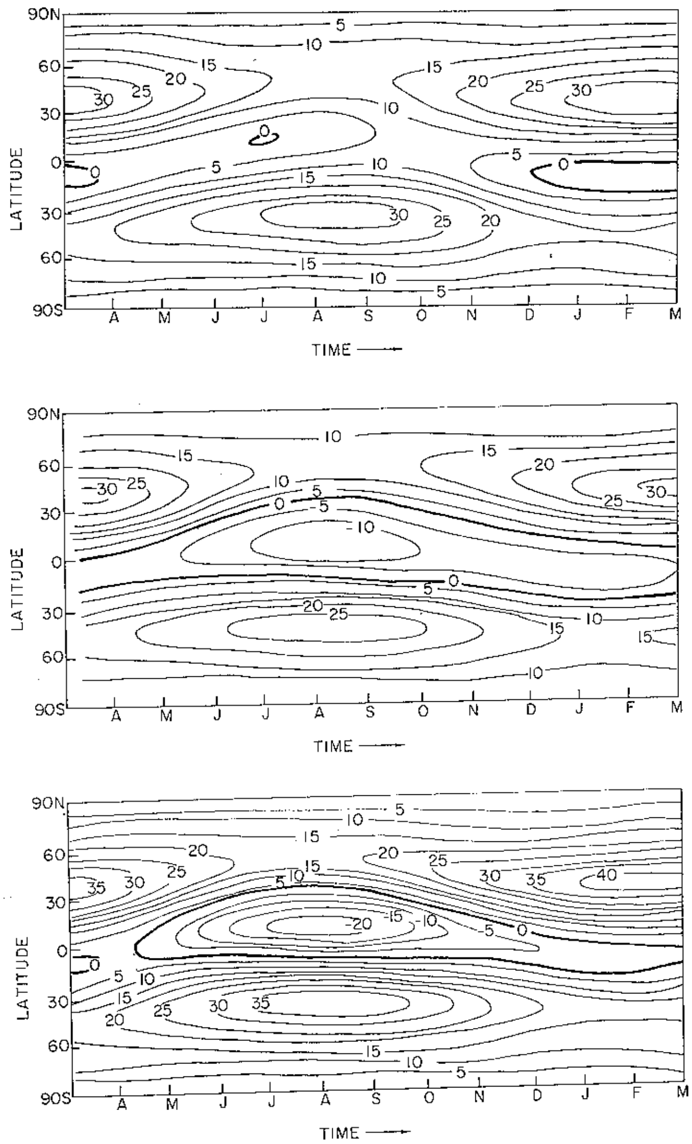

Figure 13

Time-latitude variation of the 250 mb zonal velocity component distribution computed by LAU (1977) for a three domain version of the domain averaged model. See text for details.

the solar variation. Of similar interest is the large lag of the ocean temperature. To the south of the land area a warm water pool develops due both to the thermal lag of the ocean and to the adjacent warm land. The anomalously warm ocean has the effect of extending somewhat the warm conditions over the land area into the fall (WEBSTER and LAU, 1977b) and therefore increasing slightly the lag of the surface temperature over land behind the solar variation.

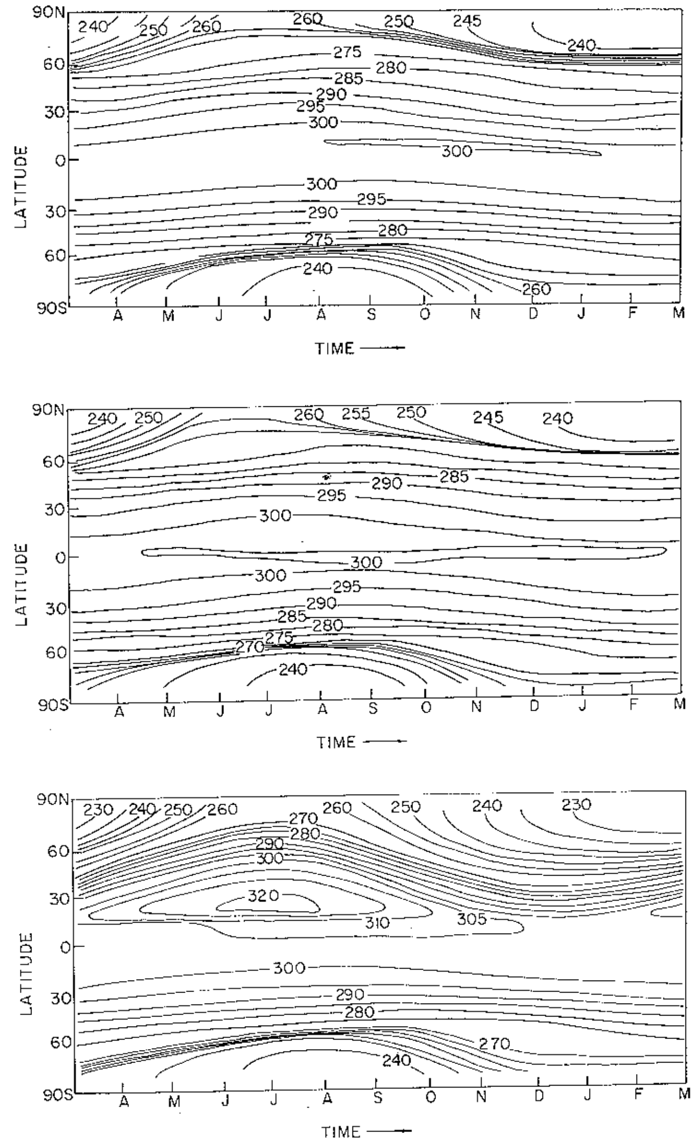

Figure 14
Same as for Fig. 13 except for the surface temperature.

The most important feature noted above is that the inclusion of land into a mobile interactive ocean produces a considerable spatial variation in the annual atmosphere flow, although the transitions between the solstices are somewhat smoother than those noted in Fig. 8. The addition of a hydrology cycle, which allows evaporation, precipitation, the transport of water vapor both in the vertical and the horizontal and the possibility of moisture build up in the land areas, alters the relatively smooth transition by imposing a significant intra-seasonal time scale on the monsoon flow.

The effect of the hydrology cycle may be best seen by viewing the precipitation rate distribution (10^2 cm yr^{-1}) in the land–ocean domain of Fig. 6. This is reproduced from WEBSTER and CHOU (1977) in Fig. 15. The precipitation rate is also indicative of the vertical velocity field.

The most significant feature of Fig. 15 is the rapid change of the position of maximum precipitation (and therefore maximum upward motion) from south of the equator in middle April to north over the coastal boundary (18°N). Little change in the location of the maximum occurs throughout summer with the precipitation pattern gradually returning to its winter position south of the equator. The rapid transition in late spring may be thought of as a competition between two processes although it should be remembered that this experiment is cloud-free. In the winter the land is cool and the sea surface temperature maximum lies just to the south of the equator. Strong subsidence dominates the northern hemisphere and the ascending leg of the Hadley circulation is highly convective. After the equinox, the northern land mass starts to become warmer (see Fig. 14) which starts to counteract the low-level southward flow of the Hadley circulation. The weak onshore flow is moist and the weak rising motion results in the release of latent heat which counteracts the existing circulation. During the fall as the maximum insolation moves southward the onshore flow gradually weakens but at this stage ground moisture accumulated during the summer provides a local moisture source which may tend to make the autumnal transition somewhat weaker. At this time of year the latitudinal temperature contrast is very weak and the effect of that is further diminished by the local latent heat source.

The results of an identical experiment to that described above, but with a cloud model included, is also shown in Fig. 15. The long-term variation is principally the same. Rapid jumps occur in both cases after the spring equinox but the effect of the clouds is to install a long-period modulation on the precipitation and dynamic fields. Such an effect can be best seen in late June and July which corresponds to the time of maximum heating. From this time on the precipitation pattern occasionally makes abrupt jumps for intervals of roughly five to ten days when the region to the north of the coast becomes dry and precipitation shifts southward to near the equator and also further inland.

The effect of cloud is two-fold. First, in regions of high precipitation, and therefore high cloudiness, the incoming solar radiation is reduced therefore depleting the insolation at the earth's surface. However, this is nearly compensated by the increased opacity of the atmosphere to terrestrial radiation. The overall effect is a slow decreasing surface temperature which instills a long-period reduction on the local heating rate. This continues until the heating of the atmosphere by the adjacent cloud-free regions dominates, shifts the regions of upward vertical velocity both equatorward and poleward and so changes the precipitation distribution. It is interesting to note that during these 'break' periods the precipitation in the equatorial regions is much

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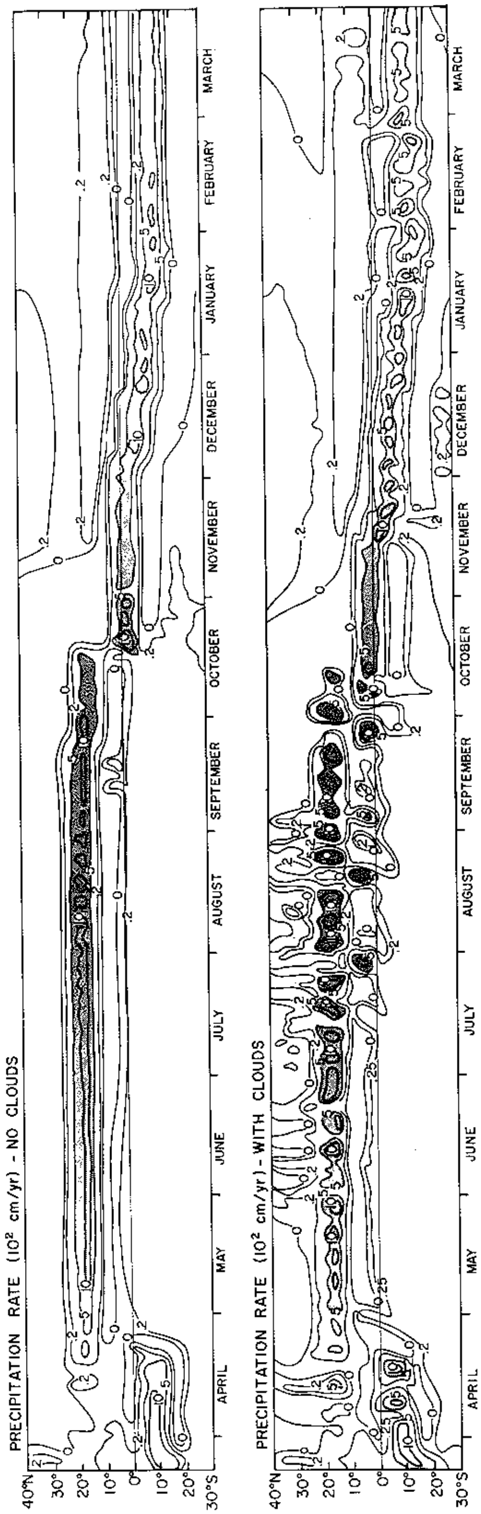

Figure 15
Time-latitude distribution of precipitation rate (10^2 cm yr^{-1}) for cloud-free case (upper panel) and for cloud-feedback case (lower panel). From WEBSTER and CHOU (1977).

stronger than that farther inland. Perhaps this is because the model contains no orographic variation. Because of that the general distribution of precipitation is similar to the 'no-mountain' case of HAHN and MANABE (1975).

The results shown in Figs. 13–16 were shown to indicate that a number of factors are of importance in determining the transient behavior of the monsoon system. They were not meant to illustrate attempts to simulate the precise structure of the monsoon but rather to indicate dominant processes.

In Section 3a we indicated that one of the most interesting features of the transient monsoon was the rapid change in the location and strength of the westerly maximum in the southern hemisphere. Some of that character can be noted in the model from Fig. 14 which shows the zonal velocity structure. The strongest westerly winds in the southern hemisphere occur in the lowest panel which corresponds to the land–ocean domain and over the ocean domain to the west. This distribution is consistent with the momentum advection across the equator from the summer monsoon and its convergence into the southern hemisphere. However, the rapid transition noted in Fig. 10 is absent from the model results. VAN LOON (1972) attributes the rapid change of the southern hemisphere circulation in May, not to the effect of the northern hemisphere summer monsoons, but to the rapid cooling of the Australian continent which is emphasized as a cold area by the extremely warm tropical oceans to the north. He concludes that the initial rapid changes in the circulation are instigated in this manner although the momentum flux from the northern hemisphere is of much greater importance later in the southern winter.

The interaction between the hemispheres on shorter time scales is not well understood. Earlier theoretical studies by CHARNEY (1969) and BENNETT and YOUNG (1971) indicate the difficulty of a wave disturbance passing through the equatorial regions which are characterized by basic zonal winds which move westward relative to the phase speed of the wave. The effect of this restriction was to suggest that equatorial regions should be impervious to incident waves. A corollary may be that the hemispheres could be considered independent on time scales less than seasonal. However, MURAKAMI and UNNAINAYEAR's (1977) data seems to suggest zones or corridors of perturbation kinetic energy bridging the gap between the mid-latitudes, credence to which is added by the observations of WEBSTER and CURTIN (1975). We are thus faced with a paradox. Theory, on one hand, suggests the impossibility of major interhemispheric interactions occurring on relatively short-time scales whereas data implies strong regional interactions. The link between observation and theory perhaps lies in Fig. 13 from which we noted earlier that the regions of high perturbation kinetic energy at the equator corresponded with mean *westerly* winds whereas regions of near-zero energy correspond to mean *easterlies*. If we assume that the basic zonal flow shown in Fig. 12 is maintained by forcing on a much longer time scale than that of the incident disturbances, it is conceivable to speculate that the waves are transmitted horizontally through corridors of weak easterlies or westerlies in much the same way as vertically propagating waves are ducted by distribution of zonal flow in the vertical (e.g., MATSUNO, 1970). Preliminary investigations are reported by WEBSTER (1977).

4. Some concluding remarks

We have attempted to document and place in perspective the current knowledge of some aspects of the planetary scale monsoon which has been developed from a number of observational and theoretical studies. We have placed particular emphasis on four features of the monsoon circulation: the mean seasonal structure, the seasonal variation, the intra-seasonal variation and interhemisphere interaction. Obvious from the discussion of the results of the various studies is the large degree of uncertainty remaining in a physical interpretation of the monsoon system, even though certain factors, for example the effect of variable oceans, emerge from various studies as critical processes. The following stand out as particular and pertinent problems:

(i) The precise role of the large-scale orography of southern Asia. It remains yet to decide if its influence emanates from its control on mid-tropospheric latent and sensible heat release or from some other, perhaps more obscure, dynamic effect. During winter we can but say that the orographic mechanical effect appears to be important but as yet we have difficulty in appreciating the nuances of atmosphere–orography interaction which may, for example, produce the vast differences between early and late winter in Southeast Asia.

(ii) The interseasonal variation of the southern hemisphere circulation and its relationship to the seasonal variations within the northern hemisphere. Whereas both observational and theoretical studies indicate that the planetary scale monsoons are of interhemispheric scale, the distinct seasonal structure of the southern hemisphere circulation is puzzling.

(iii) The interactive role of the ocean and atmosphere on the seasonal time scale. It appears that the monsoon circulation over land may be affected by oceans both adjacent in longitude and latitude but what are their precise influences on the phasing and variation of the monsoon?

(iv) The intra-seasonal variation of the monsoon circulation both in summer and winter. Variations in the evolving monsoon, such as breaks, are of extreme importance in the forecasting aspects of monsoon meteorology. However, their nature remains basically unknown. Lying in a similar state of uncertainty is the degree to which adjacent hemispheres influence each other on time scales less than seasonal.

Continuation of the progress made towards understanding some of the problems mentioned above may be achieved by further observational and theoretical studies, especially if the two approaches are carried out in concert. Some problems, however, defy extension merely because necessary data does not exist. For problems such as the influence of orography during various phases of the summer and winter monsoons, for example, there is a special need for detailed and precise measurements. It is hoped that during the winter and summer parts of the Monsoon Experiment to be held during the First-GARP Global Experiment in 1978–79, such data sets will evolve.

Acknowledgment

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