Longitudinal Heating Gradient: Another Possible Factor Influencing the Intensity of the Asian Summer Monsoon Circulation

Song Yang
Atmospheric and Environmental Research, Inc. 840 Memorial Drive, Cambridge, Massachusetts 02139, USA
Peter J. Webster
The Pennsylvania State University, University Park, Pennsylvania 16802, USA
Min Dong
Chinese Academy of Meteorological Science, State Meteorological Administration, Beijing 100081
Received January 7, 1992

ABSTRACT

The largest longitudinal heating gradients in the tropics exist between the African desert and Asian convective regions during summer once the South Asian monsoon is established. The heating gradients are anchored by the latent heat release and net radiative flux convergence over the monsoon region, and by the dominant net radiative flux divergence over the desert.

An apparent relationship is found between the intensity of the Asian summer monsoon circulation and the longitudinal heating gradients mentioned, in addition to the latitudinal heating gradients cross the monsoon region. The monsoon circulation measured in terms of the zonal wind component is stronger when the longitudinal heating gradients are large, and vice versa. Thus, we claim that the longitudinal heating gradient may be another important factor which influences the intensity of the Asian summer monsoon circulation. There is little evidence that the interannual variability of the longitudinal heating gradients between Africa and Asia and, thus, the intensity of the Asian summer monsoon circulation, is a strong function of the El Nino / Southern Oscillation cycle.

I. INTRODUCTION

1. *Latitudinal Heating Gradients and Monsoon Circulation*

It has been a prevailing theory since the time of Halley (1686) that the major forcing function of the monsoon circulation is the latitudinal heating gradients between the heated continents of the summer hemisphere and the cooler oceans of the winter hemisphere. Halley recognized the importance of the differential heating between land and sea as the driving force in the atmospheric circulation including the monsoon circulation when he tried to explain the physics and dynamics of the northeast trade winds. In the summer hemisphere, the continents are heated more quickly than the oceans and the atmosphere over the continents is heated more strongly than that over the oceans. The heating contrast leads to a difference in atmospheric density and, thus, in pressure field between the continents and the oceans. Then, the resulted body force causes the wind to blow from the oceans to the continents at the lower troposphere and in an opposite direction in the higher troposphere. Hadley (1735) included the earth's rotation effect, thus completing a picture which possesses many of the features of the real monsoon system.

Indeed, modeling studies (e.g., Webster and Lau 1977; Webster and Chou 1980a, b; Webster 1983; Yang and Dong 1992) have shown that, in conjunction with the hydrological cycle, the latitudinal heating gradients between land and oceans are of primary importance in the monsoon circulation.

2. *Effect of Other Factors on Monsoon Circulation*

Besides the latitudinal heating contrast between land and sea, many other factors may affect the characteristics of the monsoon circulation. High mountains are thought to play an important role in determining the intensity of the monsoon circulation. Hahn and Manabe (1975) carried out a study of the importance of the Himalayas on the Asian monsoon circulation and found that the model monsoon circulation in their nonmountain case was very different from that in their mountain case. Murakami (1987) claimed that the Asian monsoon circulation would be much less distinct if there were no Himalayas. The existence of the high orographic features in the summer hemisphere is usually thought to enhance the diabatic heat sources in the atmosphere, and thus increase the atmospheric heating gradients between the mountains and the nearby regions.

Over one hundred years ago, Blanford (1884) suggested that unusually heavy snow in the Himalayas preceded a failure of the Indian summer monsoon. Blanford's result was confirmed later by the studies of Hahn and Shuckla (1976) and Barnett et al. (1989), among others. In particular, the winter / spring snow cover over Eurasia is negatively correlated with the succeeding Indian summer monsoon rainfall. When snow cover, whose high albedo reduces solar energy available to the surface and the lower atmosphere, is unusually thick in the high latitudes of the continent, additional heat is required to melt the snow and, thus, there is less available heat to warm the land's surface. Furthermore, the melted snow adds to the colder reservoir of ground water and reduces the large–scale temperature gradient between land and sea.

Since the time of Walker (1924), the relationships between El Nino / Southern Oscillation (ENSO) and the Asian monsoon rainfall have been sought (e.g., Angell 1980; Shukla and Paolino 1983), although the interrelated physics between the two phenomena is, as yet, unclear. Angell (1980) found less Indian summer monsoon rainfall preceding periods of warm sea surface temperature (SST) in the tropical central–eastern Pacific Ocean or more Indian rainfall following warm zonally averaged tropospheric temperature between 30°N and 30°S. Shukla and Paolino (1983) found that when the Darwin sea–level pressure trend (spring minus winter) changed from negative (positive) to positive (negative), the Indian monsoon rainfall in the following summer was deficient.

Hydrology also appears to be important in determining the mean nature of the monsoon circulation. Webster and Chou (1980a, b) found that the "dry monsoon" is much weak than the "moist monsoon". In fact, Yang and Dong (1992) have speculated that the differences in subseasonal time scales between the Asian monsoon and the African monsoon may be due to the difference in moisture content within the two monsoon systems.

3. *Hypothesis*

The brief review of the monsoon circulation, presented above, indicates that the intensity of the monsoon circulation is largely a function of the magnitude of the diabetic heating gradient in the atmosphere. It has been known that the atmospheric diabatic heating in the tropics varies significantly not only with time, annually and interannually, but with both latitude and longitude as well (e.g., Yang and Webster 1990). Newell et al. (1974) showed that, for the latent heating component, the longitudinal gradient may be as large as the latitudinal heating gradient. This has been collaborated by Stephens and Webster (1979), who calculated the net radiative flux at the top of the atmosphere from Nimbus 3 data and found that the longitudinal gradient of the net radiation is equal in magnitude to the latitudinal gradient in the Asian monsoon regions in summer. Thus, we may speculate with some confidence that *there should be considerable longitudinal dynamical structure associated with the* longitudinal heating gradients.

The longitudinal variability of the tropical atmospheric circulation has received considerable attention (e. g., Bjerknes 1969; Webster 1972; Paegle and Baker 1982). However, while many efforts have been made to understand the ENSO-related features of the zonal circulation, especially the Walker Circulation over the equatorial Pacific Ocean, little notice has been given to the east-west circulation between the arid Africa and the moist tropical Asian continent during the boreal summer. The strongest east-west cell of the Asian summer monsoon circulation was found over the latitude about 15°N, although the ENSO-related change in atmospheric heating shows a maximum near the equator (Yang and Webster 1990; Yang 1990).

From the results of numerous studies (e. g., Krishnamurti 1971; Webster and Lau 1977; Yang 1990), the Asian summer monsoon circulation can be shown schematically in Fig.1. In a simple sense, the three-dimensional structure of the monsoon circulation can be decoupled into a north-south vertical cell between the Asian continent and the Indian Ocean to the south, and an east-west vertical cell over the tropics between Asia and Africa. The tropical Asian continent is usually characterized by strong diabatic heating, strong vertical motion, and convection, while the arid African regions and the winter Indian Ocean, which are usually cool, are dominated by subsidence. The reason for these distributions of diabatic heating, vertical motions, and convection is fairly straightforward. In the arid African regions and in the cool winter oceans, where the release of latent heat is of secondary importance, radiative cooling of the column dominates. To balance the cooling, subsidence must prevail. Model calculations by Krishnamurti and Ramanathan (1982) show that the deserts over Saudi Arabia and northwestern India and Pakistan experienced a net radiative cooling in spite of strong lowlevel heating over the lowest kilometer. On the other hand, over the Asian monsoon region, the atmosphere is heated strongly by the release of latent heat and the convergence of radiative flux which is closely associated with the strong convective process. Thus, strong upward motions and heavy precipitation are found over the monsoon regions.

In this study, we assume that the east-west circulation over the African-Asian region is maintained, in part, by the longitudinal heating gradients between the African arid regions and the Asian convective regions, in addition to the latitudinal heating gradients between Asia and the southern oceans. Thus, we may ask: will the changes in the heating gradient between the convective and arid regions affect the east-west monsoon circulation in a similar manner to the way ENSO-related changes in the atmospheric diabatic heating over the western and eastern Pacific Ocean affect the Walker Circulation? In particular, is there a relationship between the intensity of the Asian monsoon circulation and the longitudinal heating gradients? We intend to answer these questions by examining the variability of the monsoon wind field and the diabatic heating field, inferred from distributions of outgoing longwave radiation (OLR), with emphasis on interannual time scales.

II. DATA DESCRIPTION AND ANALYSIS METHODS

The primary data set used in this study is the tropical strip winds and OLR data set from the operational tropical objective analysis of the United States National Meteorological Center (NMC). The data set consists of monthly mean values of the zonal and meridional wind components (U and V) from March 1968 to February 1988 and monthly mean values of OLR, but from June 1974 to February 1988. Some data are missing within the data set. For example, 200 hPa winds are not available for October and November 1972 and the OLR data are missing from March to December 1978. In addition, 850 hPa winds are not available before December 1974. The data are depicted on a 72x23 Mercator longitude–latitude grid. The longitudinal spacing is 5°, and the latitudinal spacing is variable, ranging from 5° at the equator to 3.5° at the two boundaries located at 48.1°S and 48.1°N, respectively.

The data sources used in the NMC operational objective analysis include rawinsondes, aircraft reports and cloudtracked winds from satellite imagery. As the NMC data set is the product of an operational system, a number of different objective analysis and initialization schemes have been used in their production through the years. The effects of the changes in analysis procedure may be noticeable on particular fields such as divergent winds. A detailed description of the data set has been given by Webster and Dong (1992).

This study is based primarily on the analyses of seasonal mean fields, which are computed from monthly mean fields. Composite means are simply the arithmetic average of the data over a defined period characterized by some criteria. The anomaly of a field, over a time period, stands for the value with the time mean removed.

The character and intensity of monsoons can be represented by many fields such as the wind field, precipitation, diabatic heating, convection, and clouds. Quite often precipitation over a particular region, for example, India (Shukla and Paolino 1983; Yang and Gutowski 1992), is used as an indication of the intensity of the monsoon. However, precipitation, while it is perhaps the most important climate variable from a social–economical point of view, may be a poor indicator of the intensity of the planetary scale monsoon because of its distinct regional character. Thus, in this study, we use intensity of the 200 hPa and 850 hPa zonal winds to infer the strength of the monsoon. By this definition, the planetary monsoons are considered to be strong (weak) when the winds are strong (weak). The phase of ENSO is found from the values of SOI (Southern Oscillation Index; the normalized sea level pressure difference between Tahiti and Darwin). That is, the terms of El Nino and La Nina (or non–El Nino) years are applied for the years when SOI $\ll 0$ and SOI $\gg 0$, respectively.

III. CLIMATOLOGY OF THE DIABATIC HEATING AND THE MONSOON CIRCULATION

The climatology of the diabatic heating and circulation pattern of the Asian summer monsoon system have been summarized schematically in Fig.1. Consider a very simple atmosphere represented by three columns (A, B,and C), representative of the North African desert region, the Asian monsoon region, and the South Indian Ocean. In the convective regions (i.e., the Asian monsoon region and the Indian Ocean), the clouds are assumed to be sufficiently thick so that they radiate to space at the cloud top temperature T_t and to the surface at the cloud base temperature T_b. The boundary layer below the clouds is assumed to be moist with an emissivity ε_{bl} so that partial absorption of the downward long wave flux from the clouds occurs in the boundary layer. The temperature of the boundary layer is set at T_{bl}. An expression for the radiative flux convergence (Q_R) in these columns is:

$$Q_R(B,C) = S(1-\alpha_c)a + \sigma T_e^4 - (1-\varepsilon_{bl})\sigma T_b^4 - \varepsilon_{bl}\sigma T_{bl}^4 - \sigma T_t^4 \ . \tag{1}$$

Where S, α_c, a, σ and T_e, are successively incoming solar flux, cloud albedo, absorption coefficient of solar radiation in the cloud, Stefan–Boltzmann constant, and surface temperature. In the dry African desert region, where atmospheric emissivity (ε) is less than unity, the radiative flux convergence is expressed as:

$$Q_R(A) = \sigma T_e^4 - (1-\varepsilon)\sigma T_e^4 - 2\sigma\varepsilon T_a^4 \ , \tag{2}$$

Where T_a is the mean temperature of the air column. Using the data given in Table 1, we obtain the values of radiative flux convergence (calculated from equations (1) and (2)) and release of latent heat (Q_L ; see Holton *1979, page 328)* for the three atmospheric columns, as shown in Fig.1. A number of important features can be found from Fig.1 and Table 1. First,the gradient of radiative heating between cloudy and cloudless regions, which is a non−negligible portion of the total heating ($Q_L + Q_R$) gradient, is of the same sign of the gradient of latent heating. This is consistent with the result of Ramanathan (1987). Second, the total heating gradient between the Asian monsoon region and the North African deserts is of the same order of magnitude as the latitudinal total heating gradient across the monsoon region. These features provide strong arguments for the importance of the longitudinal heating gradients between the African deserts and Asia on the Asian summer monsoon circulation, as discussed later.

Table 1. Value of Parameters for the Three Columns of Fig.1 Which Represent North Africa, Asian Monsoon Region, and South Indian Ocean. Conditions are for Boreal Summer

Quantity		Columns		
		A	B	C
Surface temperature (K)	T_e	308	303	298
Boundary layer temp. (K)	T_{b1}		298	289
Cloud base temperature (K)	T_b		285	285
Cloud top temperature (K)	T_t		220	260
Mean atmos. temp. (K)	T_a	270		
Cloud albedo	α_c		0.3	0.2
Absorption coefficient	a		0.15	0.1
Emissivity	ε	0.6	1.0	0.6
Boundary layer emissivity	ε_{b1}		1.0	0.8
Precipitation (mm / day)	r	0	10	3
Latent heating (W m^{-2})	Q_L	0	289	82
Radiative heating (W m^{-2})	Q_R	−98	24	−130
Total heating ($Q_L + Q_R$; W m^{-2})	Q	−98	313	−48
Latent heating gradient	∇Q_L	289(B−A)		207(B−C)
Radiative heating gradient	∇Q_R	122(B−A)		154(B−C)
Total heating gradient	∇Q	411(B−A)		361(B−C)

Let's turn to the observations shown in Fig. 2 which depicts the 14 years' average of the OLR for June, July and August (JJA). Clearly, the OLR is a minimum over the Asian monsoon regions and a maximum over the North African arid regions, allowing it to be inferred that a substantial heating gradient lies between the Asian monsoon regions and the African desert regions. In addition, another OLR maximum is found over the tropical southern Indian Ocean. Over the relatively cool ocean, radiative cooling is important in contributing to a total diabatic cooling of the atmosphere, similar to the situation over the African desert regions.

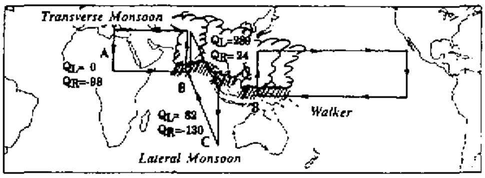

Fig.1. Schematic daigram of the zonal and meridional vertical cells of the Asian summer monsoon circulation and their association with the fields of diabatic heating and deep convection. Units for latent and radiative heating (Q_L and Q_R, respectively) are Wm^{-2}.

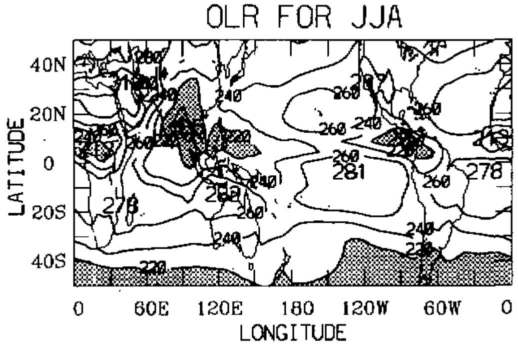

Fig.2. Climate mean pattern of OLR for JJA. The data are averaged for the period from 1974 to 1988. Regions where OLR < 220 Wm^{-2} are shaded. Note the OLR minimum over the Asian monsoon region and maximum over the African desert region.

Another feature of Fig. 2 is that significant heating gradients extend away from the Asian summer monsoon system in both the north–south and east–west directions. Fig.3 shows the OLR distributions along 20°N and 90°E for JJA, allowing a comparison of the magnitude of the longitudinal and latitudinal gradients of atmospheric diabatic heating. The OLR difference between 40°E and 90°E along 20°N is about 105 Wm^{-2} and about 70 Wm^{-2} between 15°S and 15°N along 90°E. Thus, the magnitude of the longitudinal heating gradient is of the same order as the latitudinal heating gradient across the monsoon regions. This is consistent with the result obtained by Stephens and Webster (1979).

To gain a further insight into the mean heating–monsoon relationship, we plot the 200 hPa and 850 hPa divergent wind fields for JJA in Fig. 4. The figure shows the mean pattern of the divergent winds averaged between 1978 and 1988. A comparison of Fig.4 with Fig.2 shows that, in the tropics, the divergent winds blow from the heat sinks to the heat sources at the lower troposphere and in the opposite direction at the upper troposphere. The most interesting feature is the divergent winds over the Asian summer monsoon regions where the divergent westerly wind is seen at the lower level but the divergent easterly wind prevails at the upper level between the western Pacific Ocean and the African arid regions. This east–west divergent circulation is presumably associated with the heat sink over the African desert and the heat source over the monsoon regions. Clearly, the arid and convective regions are dynamically connected.

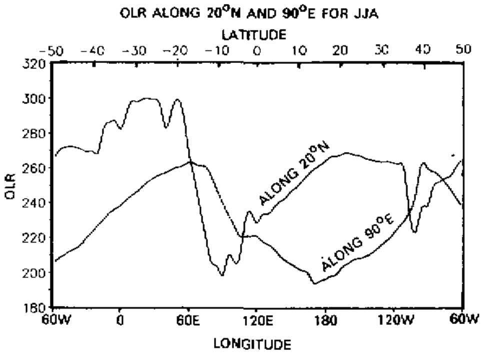

Fig.3. Distributions of OLR (Wm^{-2}) along 20°N and 90°E for JJA. The two curves show the relative magnitude of the zonal and meridional gradients of OLR across the Asian monsoon region. The upper abscissa shows latitude and the lower abscissa shows longitude.

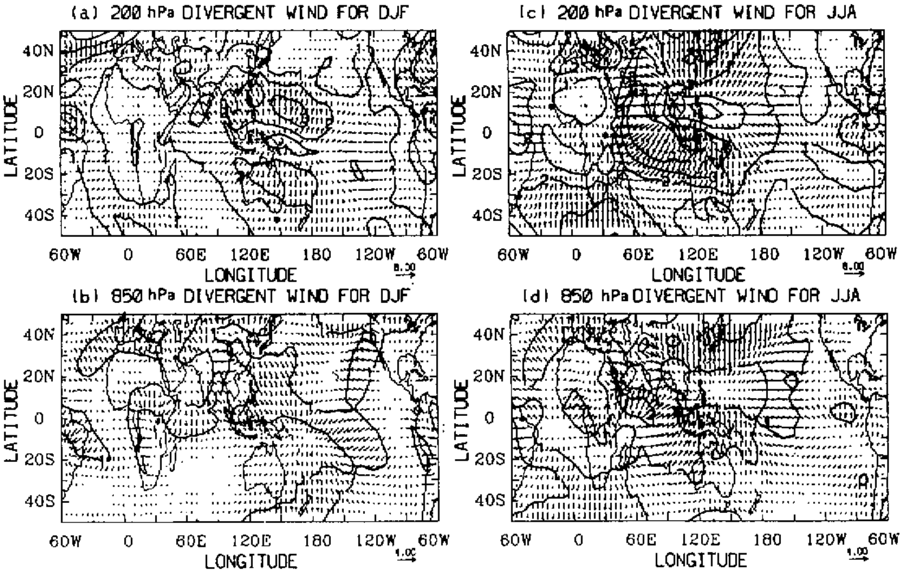

Fig.4. Mean JJA divergent wind (ms^{-1}) patterns for (a) 200 hPa and (b) 850 hPa. Data are averaged between 1978 and 1988. Contours show the magnitude of the total divergent wind.

IV. INTERANNUAL RELATIONSHIP BETWEEN THE ASIAN SUMMER MONSOON AND THE LONGITUDINAL HEATING GRADIENTS

To test the hypothesis posed in the introduction, we will study the anomaly fields of OLR and the winds at 200 hPa and 850 hPa. In particular, the relationship between the changes in the Asian summer monsoon circulation and those in OLR over the African desert and the Asian monsoon regions during each JJA are emphasized. Over the African desert, radiative cooling is dominant in the atmospheric total diabetic heating because latent heat release can be neglected. Thus, an increase (decrease) in OLR over the desert regions implies an increase (decrease) cooling in the atmosphere. In contrast, over the strongly convective Asian monsoon regions, heavy precipitation and thus strong latent heat release occurs during JJA. Then the changes in OLR reflect the changes in both latent heat release and radiative flux convergence, especially latent heat release. Thus, a decreasing (an increasing) OLR is associated with an increase (a decrease) in latent heat release and radiative flux convergence.

When considering the OLRheating association, one should remember that the changes in heating rate calculations are not only functions of the changes in OLR, but also the magnitude of the OLR. For example, over a particular region, a change in OLR from 200 Wm^{-2} to 260 Wm^{-2} may result in greater heating changes than a change in OLR from 260 Wm^{-2} to 320 Wm^{-2} due to the effect of latent heating, although the OLR increases by the same value (60 Wm^{-2}) in both cases. These nonlinear changes in heating with OLR may necessarily cause substantial problems in our analysis. Here, we only qualitatively compare the different state of the Asian summer monsoon circulation when the OLR anomaly is positive over Asia and negative over Africa and when the OLR anomaly is negative over Asia and positive over Africa. It is safe to say that a relatively large positive (negative) OLR anomaly over Asia implies a weaker (stronger) local diabatic heating and that a relatively large negative (positive) OLR anomaly over Africa means a weaker (stronger) local cooling.

The year–to–year changes of the OLR anomaly (along 20°N), the 200 hPa U anomaly (along 15°N) and the 850 hPa U anomaly (along 15°N) are displayed in Fig.5. The impressive feature shown by the figure is that the Asian summer monsoon becomes stronger when the longitudinal heating gradient between Africa and Asia becomes larger, and vice versa. It can be seen from Fig.5a that the OLR anomaly is positive over the Asian convective regions and negative over the African desert before JJA of 1980. In contrast, the OLR anomaly becomes negative over Asia and positive over the African desert after JJA of 1982. Remember that the negative (positive) value of the OLR anomaly over the African desert means a decrease (an increase) of the OLR value over the heat sink and thus less (more) cooling in that region. Similarly, the positive (negative) value of the OLR anomaly over the Asian convective regions means an increase (a decrease) of the OLR value over the heat source and thus less (more) heating there. Thus, Fig.5a shows that the longitudinal heating gradient between Africa and Asia is smaller than normal before JJA of 1980 and larger than normal after JJA of 1982.

The changes in the intensity of the Asian summer monsoon circulation show close relations to the variations of the longitudinal heating gradient. Before JJA of 1980 the Asian summer monsoon is weak, which is consistent with the weak longitudinal heating gradient. A decrease of 4 ms^{-1} of the monsoon westerlies at 850 hPa can be found in Fig.5c. The weakening of the easterlies at 200 hPa can also be seen (Fig.5b), although the change is less significant than at 850 hPa. On the other hand, the Asian summer monsoon is much stronger after JJA of 1982 when the longitudinal heating gradient is larger. It can be seen that the monsoon wind increases by 8 ms^{-1} at 850 hPa (Fig.5c) and by 4 ms^{-1} at 200 hPa (Fig.5b). Therefore, a very close relationship is found between the longitudinal heating gradient and the Asian summer monsoon on the interannual time scales.

An examination of Fig.5c shows that the changes of the winds over the Pacific Ocean and the Atlantic Ocean are out of phase with respect to the Asian monsoon current. That is, at 850 hPa, when the monsoon is more easterly (weaker monsoon; e.g., before 1980), the trade winds over the Pacific and Atlantic oceans are more westerly (weaker trade winds); and when the monsoon is more westerly (stronger monsoon; e.g., after 1982), the winds over the oceans are more easterly (stronger trade winds). A similar relationship can also be found from the 200 hPa wind field. This feature in the changes of the wind field is associated with the variations in the heating field. For the thermally driven Asian summer monsoon, the variations of the winds on the east side of the heat source and the west side of the heat sink should be out of phase with the winds between the heat source and sink at each level. For example, stronger low-level westerlies between the heat source and sink will be accompanied by weaker westerlies (or stronger easterlies) on both east of the heat source and west of the heat sink.

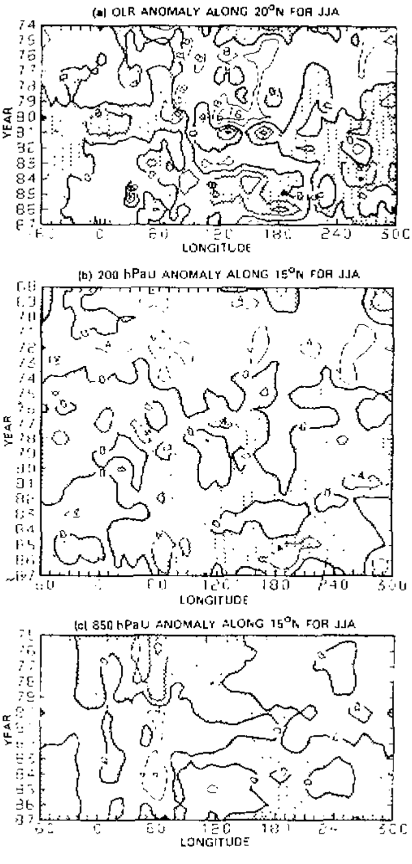

Fig.5. Time-longitude sections of (a) OLR anomaly (Wm^{-2}, along 20°N), (b) 200 hPa U anomaly (ms^{-1}, along 15°N), and (c) 850 hPa U anomaly (ms^{-1}, along 15°N) for JJA. Negative values are shaded.

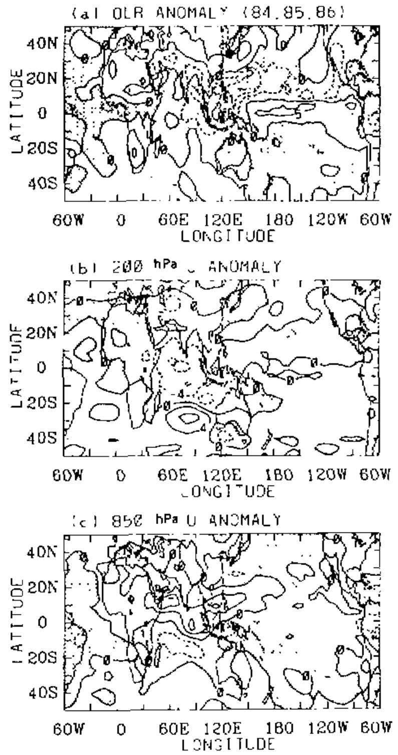

Fig.6. Composite patterns of anomaly fields for (a) OLR (Wm^{-2}), (b) 200 hPa U (ms^{-1}), and (c) 850 hPa U (ms^{-1}) for the JJA of 1984, 1985, and 1986 when the longitudinal heating gradients between Asia and Africa are large. Negative values are shaded.

V. COMPOSITE ANALYSIS

The relationship between the longitudinal heating gradients and the Asian summer monsoon can be shown more clearly by the composite patterns. Fig.6 shows the composite mean anomaly fields of the OLR, 200 hPa U and 850 hPa U for the JJA of 1984, 1985 and 1986, the years with large longitudinal heating gradients (LLHG) between the Asian convective regions and the African desert. It can be seen from Fig.6a that the OLR anomaly is positive in the African desert and negative in the Asian convective regions, which means a stronger than normal cooling in the former and a larger than normal heating in the latter regions. In response to the larger longitudinal heating gradient, the Asian monsoon circulation becomes stronger in these LLHG cases, with the 200 hPa U increasing by 8 ms^{-1} (Fig.6b) and the 850 hPa U increasing by 4 ms^{-1} (Fig.6c). Note that both the negative values of the anomaly of the 200 hPa easterlies and the positive values of the anomaly of the 850 hPa westerlies imply a stronger Asian summer monsoon circulation.

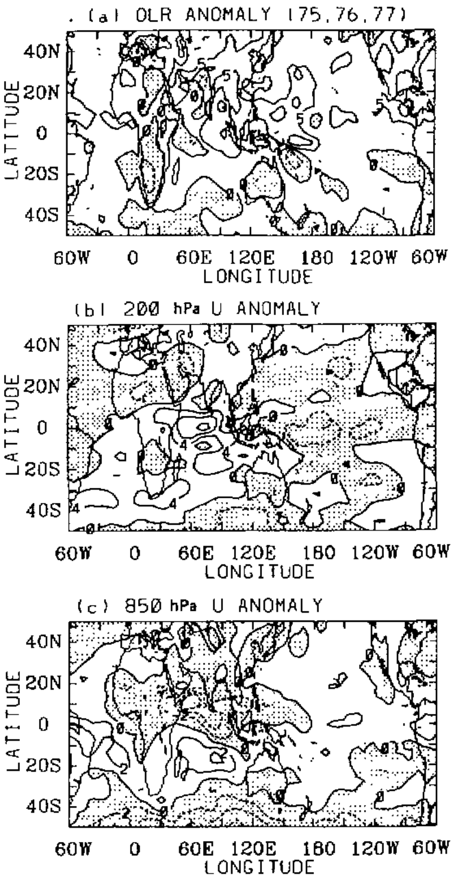

Fig.7. Composite patterns of anomaly fields for (a) OLR (Wm^{-2}), (b) 200 hPa U (ms^{-1}), and (c) 850 hPa U (ms^{-1}) for the JJA of 1975, 1976, and 1977 when the longitudinal heating gradients between Asia and Africa are small. Negative values are shaded.

Figure 7 shows the corresponding patterns of Fig. 6, but for the JJA (1975, 1976 and 1977) composite of small longitudinal heating gradients (SLHG) cases. It is found that the changes of the OLR and wind fields shown in Fig. 7 are out of phase with those shown in Fig. 6. It can be seen from Fig. 7a that the OLR anomaly is negative in the African desert and positive in the Asian convective regions, implying a decrease of the heating gradient between the two areas. Consistently, both the 200 hPa and 850 hPa winds become significantly weaker (Figs.7b and 7c). Apparently, the weaker winds indicate a weaker Asian summer monsoon circulation.

Again, the corresponding changes of the wind fields over the Pacific and Atlantic oceans, which have also been discerned in Fig.5, can be found in the composite latitude–longitude patterns. That is, over the Pacific and Atlantic oceans, the 850 hPa wind becomes more easterly (westerly) while the 200 hPa wind becomes more westerly (easterly) in the case of LLHG (SLHG). The only exception found is the 200 hPa wind over the Pacific Ocean in the case of LLHG (Fig.6b).

VI. DISCUSSION AND CONCLUSIONS

From this analysis, it has been found that in the tropics the largest longitudinal heating gradients exist between the African desert and Asian convective regions during summer once the South Asian monsoon is established. An apparent relationship is found between the intensity of the Asian summer monsoon and the longitudinal heating gradients between the Asian convective regions and the African desert. A stronger (weaker) monsoon is associated with a larger (smaller) longitudinal heating gradient. Thus, the eastwest heating gradient provides lateral linkages between the monsoon and other regions beyond the traditional continental–ocean connection. That is, the longitudinal heating gradient may be another important factor which influences the intensity of the Asian summer monsoon circulation.

The relationship can be easily understood if the monsoon is considered as a thermally driven system. In the longitudeheight cross section across the monsoon regions, easterly (westerly) winds are found at the upper (lower) levels, with upward and downward motions in the Asian convective regions and the African desert, respectively (see Fig.1). Thus, the heating and cooling over Asia and the African desert are very important in determining the variability of the zonal wind field and the associated vertical motions. One should also keep in mind that the eastwest component of the Asian summer monsoon circulation is much stronger than its northsouth component. Although the eastwest gradient does not possess a significant magnitude until convection commences in South Asia, variations in the radiational cooling over the North African deserts introduce considerable variability into the monsoon.

One may question that, besides the relationship between the monsoon and the longitudinal heating gradients, how is the Asian summer monsoon related to the latitudinal heating gradients across the monsoon regions? As stated in the introduction, the traditional view point emphasizes that the latitudinal heating gradients initiate the monsoon circulation. Thus, when we emphasized the relationship between the monsoon circulation and the longitudinal heating gradients, we did not exclude the importance of the latitudinal heating gradients on the monsoon, especially in initiating the monsoon circulation. However, the problem is that, once the monsoon circulation is established, the changes in both latitudinal and longitudinal heating gradients resulting largely from the changes in the strong convection will affect the intensity of the monsoon circulation. Furthermore, there is no reason to assume a simultaneous change in the latitudinal and longitudinal heating gradients because the changes in heating over the African desert and the southern Indian Ocean may alter the heating gradients to a certain extent.

It is difficult to conclude how the Asian summer monsoon circulation and the longitudinal heating gradients between Africa and Asia vary with the ENSO cycle. The monsoon is strong during some EL Nino years (e.g., 1982, 1983 and 1987), but weak during some other El Nino years (e.g., 1969, 1972, 1976 and 1977). On the other hand, the monsoon is weak during some La Nino years (e.g., 1968, 1971, 1973 and 1974), but strong during some other La Nino years (e.g., 1981). In addition, the monsoon is strong in JJA of 1986, which is a non–El Nino year. The current study by Webster and Yang (1992) shows that the ENSO cycle is closely related to the Australian (boreal) winter monsoon circulation, but not well with the Asian summer monsoon circulation. Although a relationship is claimed to exist between the Indian summer monsoon rainfall and ENSO (Angell 1981, Shukla and Paolino 1983), how significant the correlation is still remains a question. Also, the physics of the relationship, if it exists, is unclear. At least, it is safe to say that no acceptable conclusion on the monsoon ENSO relationship has been drawn yet.

It should be pointed out that the relationship between the longitudinal heating gradients and the Asian summer monsoon circulation is far from being well understood. In this study, we only show a simple relationship between the final state of the monsoon circulation and the longitudinal heating gradients in addition to the relationship of the monsoon circulation to the latitudinal heating gradients. We are unable to discuss the cause–effect relationship between the monsoon and the longitudinal heating gradients because of the limitation of the available data. At most, we just assume that the east–west component of the Asian summer monsoon circulation is, in part, the response to the longitudinal heating gradients, as many scientists (e.g., Bjerknes 1969; Webster 1972) thought for the Walker Circulation and the longitudinal heating gradients across the equatorial Pacific Ocean. However, the changes in the east–west circulation, regardless of the zonal Asian monsoon circulation or the Walker Circulation, will certainly alter the related longitudinal heating gradients. Thus, more studies, both diagnoses and modeling, are needed to understand this almost untouched subject.

At the present time, the processes of how the African desert affects the Asian monsoon circulation are unknown, although it seems that the desert plays a very different role in the African monsoon than in the Asian monsoon system. On one hand, the desert prohibits the development of a heat source in the regions (just like the situation in Asia), and, thus, reduces the intensity of the African monsoon circulation and confines the monsoon to extend northward. While, on the other hand, the desert seems to intensify the Asian summer monsoon because the downward branch of the (east–west) monsoon circulation is over the desert regions where a heating divergence exists in the whole atmospheric column. We hope this simple observational analysis will establish a research thrust into this subject in the near future.

This study was partially supported by US NSF Grant ATM 8703267. We would like to express our gratitude to Dr. HaiRu Hang and Mr. Sanjay Dixit at The Pennsylvania State University and Dr. Ilana Wainer at National Center for Atmospheric Research for many interesting discussions. We thank Dr. P. A. Arkin of the U.S. NMC / CAC for the provision of the data used in this study, and Ms. Betsy A. Schaeffer and Ms. Mary Beth Phillips at Atmospheric and Environmental Research, Inc. for their technical assistance.

REFERENCES

Angell, J. K. (1981), Comparison of variations in atmospheric quantities with sea surface temperature variations in the equatorial eastern Pacific. *Mon. Wea. Rev.*, **109**: 230–243.

Barnett, T. P., L. Dumenil, U. Schlese and E. Roeckner (1989), The effect of Eurasian snow cover on regional and global climate variations. *J. Atmos. Sci.*, **46:** 661–685.

Bjerknes, J. (1969), Atmospheric teleconnections from the equatorial Pacific. *Mon. Wea. Rev.*, **97:** 163–172.

Blanford, H. F. (1884), On the connexion of the Himalayan snowfall with dry winds and seasons of drought in India. *Proc. R. Soc. London* **37** 3–32.

Hadley, G. (1735), Concerning the cause of the general tradewinds. *Philos. Trans. R. Soc. London* **39:** 58–62.

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

Hahn, D. G., and J. Shuck (1976), An apparent relationship between Eurasian snow cover and Indian monsoon rainfall. *J. Atmos. Sci.* **33:** 2461–2462.

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

Holton, J. R. (1979), *An Introduction to Dynamic Meteorology*, second edition, Academic Press, 391 pp.

Krishnamurti, T. N. (1971), Tropical eastwest circulations during the northern summer. *J. Atmos. Sci.* **28:** 1342–1347.

Krishnamurti, T. N., and V. Ramanathan (1982), Sensitivity of the monsoon onset to differential heating., *J. Atmos. Sci.* **39** 1290–1306.

Murakami, T. (1987), Orthography and monsoons. *Monsoons*, Eds. Jay S. Fein and Pamela L. Stephens, 331–364.

Newell, R. E., J. W. Kidson, D. G. Vincent and G. J. Boer (1974), *The General Circulation of the tropical Atmosphere and Interactions with Extra tropical latitudes*. Vol.2, The MIT Press, 371 pp.

Paegle, J., and W. E. Baker (1982), Planetaryscale characteristics of the atmospheric circulation during January and February 1979. *J. Atmos. Sci.* **39:** 2521–2538.

Ramanathan, V. (1987), The role of earth radiation budget studies in climate and general circulation research. *J. Geophys. Res.* **92:** 4075–4095.

Shuck, J., and D. A. Palomino (1983), The Southern Oscillation and long range forecasting of the summer monsoon rainfall over India. *Mon. Wea. Rev.*, **111:** 1830–1837.

Step hens, G. L., and P. J. Webster (1979), Sensitivity of radiative forcing to variable cloud and moisture. *J. Atmos. Si.*, **36:** 1542–1556.

Walker, G. T. (1924), Correlation in seasonal variations of weather, IV: A further study of world weather. *Mem Indian Meteor. Dep.*, **24:** 275–332.

Webster, P. J. (1972), Response of the tropical atmosphere to local,steady, forcing. *Mon. Wea. Rev.*, **100:** 518–540.

Webster, P. J. (1983), Mechanisms of monsoon lowfrequency variability: Surface hydrological effects. *J. Atmos. Sci.*, **40:** 2110–2124.

Webster, P. J., and L. Chou (1980a), Seasonal structure of a simple monsoon system. *J. Atmos. Sci.*, **37:** 354–367.

Webster, P. J., and L. Chou (1980b), Low frequency transition of a simple monsoon system. *J. Atmos. Si.*, **37:** 368–382.

Webster, P. J., and M. Dong (1992), The structure of low frequency phenomena in the tropics and its interaction with the extra tropics. *Adv. Atmos. Sci.* **9:** 1–16.

Webster, P. J., and K. M. Lau (1977), A simple oceanatmosphere climate model: Basic model and a simple experiment. *J. Atmos. Sci.*, **34:** 1064–1084.

Webster, P. J., and S. Yang (1992), ENSO and monsoons: selectively interactive systems. Submitted to *Quart. J. Roy. Meteor. Soc.*

Yang, S. (1990), Atmospheric teleconnections: Emphasis on El Nino / Southern Oscillation and monsoons. *Ph. D.*

Thesis, The Pennsylvania State University, 243 pp.

Yang, S., and M. Dong (1992), Differences in the Asian and African monsoons: Emphasis on surface hydrological effects. Submitted to *Chinese J. Meteor.*

Yang, S., and W. J. Gutowski (1992), On the relationship between tropical Chinese rainfall and the Indian summer monsoon. Submitted to *J. Meteor. Sci. Japan.*

Yang, S., and P. J. Webster (1990), The effect of tropical heating in the adjacent hemisphere on the extratropical westerly jet streams. *J. Geophys. Res.*, **95:** 18705–18721.