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Some Problems on the Monsoons of East Asia

KAO YU-HSIE et al

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SOME PROBLEMS ON THE MONSOONS OF EAST ASIA

东亚季风的若干問題

by

Kao Yu-hsie et al

高由禧等

Collected Papers of the Institute of Geophysics and Meteorology, Academia Sinica, Peking, China
No. 5, 1962. 106 p.

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ABSTRACT

This monograph presents some results of recent investigations concerning the monsoon problem in East Asia by members of the Institute of Geophysics and Meteorology, Academia Sinica. It consists of seven papers covering three major aspects of the problem. A detailed discussion on the contributions of previous investigators is first presented with particular emphasis on their viewpoints and methods of treatment in comparison with the approach adopted by the present authors. Secondly, factors governing the formation of monsoons over East Asia and their characteristics were examined and a classification of the monsoon regions of China was made. Thirdly, the advance and retreat of the winter and summer monsoons over East Asia were studied in relation to the seasonal variation of the general circulation. The onset and termination of the rainy season over China were also examined in terms of the march of the synoptic seasons. Based on the studies outlined above, tentative conclusions on the problem are given.

This reference monograph is intended for researchers in the fields of meteorology, climatology and geography.

FOREWORD

"Monsoon" is an important problem in climatology. This is particularly so in China, because it lies in a region which is noted for its pronounced monsoons. If we wish to analyze the characteristics of the weather and climate over China, a study on the monsoon problem is indispensable. In order to improve the long-range prediction of droughts and floods which is closely related to national economy, a knowledge of the anomalies of the monsoon is also vital. Thus "monsoon" has been a subject of extensive study in China for many years. However, due to the limitations of available information, the results so obtained on the characteristics of the monsoons over China cannot be regarded as completely conclusive.

After the liberation of the mainland of China, the amount of observational data increases significantly, particularly in the upper-air. Thus we are in a better position to study the monsoon problem of East Asia in a more comprehensive manner. Such studies have been carried out at the Institute of Geophysics and Meteorology for two consecutive years since 1955. These investigations include analyses of the three-dimensional structure and properties of the monsoons over East Asia and discussions on the inter-annual variations of the monsoonal characteristics of the weather and climate as well as the wind and temperature fields for this region. The advance and retreat of the monsoons in relation to the seasonal variation of the general circulation are also examined. In addition, the monsoons over the Tibetan Plateau are discussed and compared with those in other regions of East Asia (e.g., India, the Indo-China Peninsula and Japan).

Seven reports of the investigations carried out at the Institute of Geophysics and Meteorology covering three major aspects of the monsoon problem are included in this monograph. A detailed discussion on the contributions of previous investigators is first presented with particular emphasis on their viewpoints and methods of treatment in comparison with the approach adopted in the present study. Secondly, factors governing the formation of monsoons over East Asia and their characteristics were examined and a classification of the monsoon regions of China was made. Thirdly, the advance and retreat of the winter and summer monsoons over East Asia were studied in relation to the seasonal variation of the general circulation. The onset and termination of the rainy season over China were also examined in terms of the march of the synoptic seasons. Owing to limitations dictated by the relatively short period of available data and the time factor, it has only been possible to draw tentative conclusions from the analyses. Thus lack of penetration into the problem is almost inevitable. In this respect the authors shall be glad to receive comments and suggestions to pave the way for a more comprehensive study on the subject.

We are grateful to the supervisory organization of the Institute of Geophysics and Meteorology for sponsoring this work and for their guidance and interest. Grateful thanks are also due to Yeh Tu-cheng, Koo Chen-chao, Yang Chien-chu, Zhu Gang-kun, Chang Pao-kun and Chu Pao-chen for their constant encouragement and kind help, particularly to Dao Shih-yen for his critical review of this monograph and other valuable suggestions, to Tseng Yu-ssu for the preparation of charts and diagrams and to Chang Chih-ying, Li Chih-tung, Huang Hsüeh-fu and Chang Chieh for assistance in statistical computations. Finally, we wish to express our indebtedness to the Institute of Meteorology of the Central Weather Bureau (China) for permission to draw freely from the available data.

CONTENTS

THE MONSOON PROBLEM*

by

Kao Yu-hsie

The monsoon phenomenon had been known for a very long time. In the medieval ages, Arabic merchants had made good use of the seasonal change of the prevailing wind direction to plan their sea voyages. However, a scientific treatment of the monsoon was not carried out until the last two or three centuries. In 1686, Hadley [1] first attempted to explain the phenomenon as a direct circulation induced by the differential heating over land and sea. This classical approach was adopted by many standard textbooks on meteorology and climatology to account for the formation of the monsoon, and to describe the climatological features of certain regions and their seasonal variations. Voyeikov [2], a Russian meteorologist, was the first to give a concise definition of the monsoon. He was of the opinion that the monsoon phenomenon was a direct consequence of the differential heating between land and sea. He also attempted to interpret the seasonal variation of cloud and rain in terms of the monsoon activities. However, extensive investigations on the monsoon problem did not begin until the early twenties of the twentieth century. Furthermore, the basic approach adopted during the period from the twenties to the fifties did not extend beyond the conceptual boundaries set by Hadley and Voyeikov.

Since precipitation over East Asia has been found to be closely related to the monsoons of the region, many investigators have tried to predict the dates of onset and ending of the monsoons and their intensities, and then the

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* Manuscript completed in September 1959.

amount of precipitation during different phases of the monsoons from correlational studies. Blanford [3] was the first to draw our attention to the fact that summer droughts over India are related to the amount of snow accumulation over the Himalayas in the previous winter and spring. Eliot [4] has made a further effort to examine the relationship between monsoon rain over India and the temporal distribution of the various meteorological elements over Australia and Africa. A comprehensive correlational study of monsoon rain over India and the distribution of meteorological elements over various places on a global basis has been carried out by Walker [5]. On the other hand, many investigators in different countries over Southeast Asia have conducted similar studies in search of relationships between precipitation and the various meteorological factors in their own countries. The results presented by Tu [6] are particularly concise and systematic with enlightening explanations on the problem. However, in spite of the intensive research on this problem, only very few correlation coefficients have been found as useful predictors. It is too plain to need emphasis that much difficulties will be experienced in the statistical prediction of monsoon rainfall before the physical processes of precipitation are clearly understood. Some investigators [7 - 10] have introduced various types of indices to describe or compare the monsoon activities over different regions.

The concept of Hadley has been used by many workers to study and formulate theories on the formation of monsoons. Malkus and Melvin [11] have computed the annual range of the pressure field of the summer and winter monsoons induced by the periodic variation of heating under the assumption of a zero horizontal pressure gradient while others [12 - 15] have calculated the seasonal variation and the distribution of the pressure, wind and humidity fields resulted from the annual variation of the differences in heating (or temperature) over land and sea for a circular lake surrounded by an infinitely large land mass or for a circular land mass surrounded by an infinitely large sea. Reference [16] deals with monsoon phenomena generated by the differential heating over land and sea with an infinitely long and straight coast line. Description and prediction of monsoon phenomena have also been made for certain actual distribution of land and sea [17, 18]. Some investigators [19 - 22] have made computational analyses on the formation of atmospheric action centers or planetary waves induced by the differential heating over land and sea and its seasonal variation. No matter what type of theoretical computation is involved, all mathematical treatments tend to introduce some degree of "distortion" on the problem by their corresponding simplifying assumptions. The principal disadvantages are either an over-idealization of the land-sea distribution or the negligence of the fact that the monsoon is a composite phenomenon resulted from the land-sea distribution under the influence of various planetary flow configurations and topography.

On the other hand, References [23 - 27] present discussions on the monsoons over India and the structure of the Indian Low in summer, using upper-air data. Reference [28] gives a passing mention of the monsoons over North America in the discussion on the upper-air climatology over that region.

Summing up, we may say that progress of the studies on the monsoon problem has been very slow before the end of World War II. A relatively large change has been taking place since the beginning of the fifties and the principal contributions during this period cover the following three aspects of the problem:

Firstly, Khromov [9] and Flöhn [30] have questioned the validity of the classical definition of the monsoon and have independently postulated almost at the same time that the seasonal displacement of the planetary wind belt is the principal factor governing the formation of the monsoons and that the effect of land-sea distribution only plays a part in bringing about a certain degree of modification on the properties of the monsoons caused by the former factor. On the other hand, some writers still make use of the concepts of Hadley and Voyeikov. In particular, Sorochan [31 - 34] has attempted to confirm the correctness of the definition proposed by Voyeikov by an analysis of the formation of the monsoons over East Siberia, the characteristics of the monsoon circulation and the evolution of precipitation and cloud systems.

Secondly, the advance and retreat of the monsoons are closely related to the seasonal variation of the planetary circulation at middle latitudes. In the past, investigations on the monsoon phenomenon are mainly restricted to the determination of the onset and ending of the monsoons in terms of the variations of cloud and rain [35, 36], air-mass distribution and pressure patterns [37]. Although the results are useful to some extent, they do not reveal the basic properties of the march and the variation of the monsoons to provide sufficient clues for medium-range and long-range prediction purposes.

However, during the late forties and the early fifties of the twentieth century, much valuable work has been done in this field. Yin [38] has first related the outburst of the southwest monsoon over India to the abrupt dissolution of the jet stream over the southern part of East Asia and the westward displacement of troughs in the Bay of Bengal. He has also pointed out that this variation is a radical rearrangement of the planetary waves over the northern hemisphere. The findings of Sutcliffe [39] also indicate that the seasonal variation of the planetary circulation in early June is accompanied by the outburst of the southwest monsoon over India. Subsequently, Yeh, Dao and others have presented several articles [40 - 45] on the abrupt changes of the general circulation in early June and mid-October in relation to the establishment and the retreat of the summer monsoon over Southeast Asia. Hsu [46] has recently examined the close relationship between each phase of the advance and retreat of the monsoons and the seasonal variation of the westerly circulation. Their findings clearly point to the correct line of approach for the preparation of medium- and long-range predictions for the advance and retreat of the monsoons.

The third aspect is revealed by the efforts of achieving a better understanding of the monsoon problem. Before the fifties, an overwhelming majority of writers thought that the monsoon phenomenon is purely and simply the result of the seasonal variation of differential heating over land and sea. Obviously, this postulation cannot satisfactorily account for the observed features over all monsoon regions. After the fifties, it has been repeatedly pointed out by various investigators that the monsoon phenomenon is a result of the combined effect of the mutual interaction of the planetary circulation and the land-sea distribution. Some writers [9, 30] have placed more emphasis on the seasonal variation of the planetary wind system, while others [47 - 50] consider that the monsoons are caused by the mutual interaction of the planetary circulation and the land-sea distribution so that they exhibit significant regional characteristics. Thus the observed phenomena cannot be satisfactorily ascribed to a single governing factor which is independent of time and space.

The diversity of approach in the research development on the monsoon problem after the fifties necessarily led to differences in viewpoint and method of treatment by scholars in different countries, which are presented in the following sections together with comments of the present authors.

I. VIEWPOINTS ON THE FORMATION OF MONSOONS

In the last decade or so, there exist two schools of thought on the formation of monsoons. In U. S. S. R. and Germany, they have taken part in relatively hot arguments. The classical school [17, 51 - 53] maintains the concept developed by Hadley more than two hundred years ago and looks upon the monsoons as a large-scale thermal convective phenomenon with a period of one year resulting from differential heating between land and sea. Their main points of discussion are the complete reversal of the land-sea pressure distribution in January and July shown on surface pressure charts. It is thought that this seasonal reversal of the pressure field must give rise to a corresponding change of the prevailing winds which are thus called monsoons (or seasonal winds). The other school emphasizes the importance of the general circulation [54 - 56]. They begin with the seasonal variation of the planetary circulation and then infer that monsoon phenomena may occur in many regions even without a land-sea contrast, which is only capable of modifying the planetary wind belt to a certain extent.

Apparently, both schools are correct in many respects but there are also shortcomings in their explanations because of the over-emphasis on some particular aspects of the problem. The first school adheres to the concept of land-sea distribution but because of the over-emphasis of its influence, the model developed cannot satisfactorily account for certain special features of the monsoon phenomenon over many regions. For example, deductions from this model indicate that the vertical extent of the southwest monsoon over India does not exceed 2 km [18, 52], whereas observations reveal that it reaches 5 - 6 km on average and much higher in individual cases.

Exaggeration of the influence of the seasonal displacement of the planetary wind belt likewise produce some unrealistic results. For example, in the surface layer over the East Asiatic Continent, the winter planetary wind belt is practically absent [57] from the latitude range within which the planetary circulation generally prevails. This includes the easterly wind regime (0° - 30°N), the westerly circulation (30° - 60°N) and the polar easterlies (60° - 90°N). Similarly, the normal summer circulation is not discernible over the continent, such as the easterly wind regime (south of 35°N), the west wind belt (35° -55°N) and the polar easterlies (north of 55°N). Consequently, we may say that the land-sea distribution over East Asia gives rise to intense monsoons which destroy the normal pattern of the planetary wind belt in the surface layer and thereby inhibit occurences of various phenomena associated with the seasonal displacement of the planetary wind system. Thus, even if it is maintained that the monsoons over East Asia are induced by the seasonal displacement of the planetary wind belt, this statement can only hold good at the upper levels above 3 - 5 km. We cannot deny that the planetary flow exerts some influence on the monsoons at low levels in much the same way as the planetary wind belt is affected by monsoon circulations generated by the land-sea distribution. However, the influence of the latter is so pronounced as to outweigh any meaningful overemphasis of the former.

Exaggeration on any aspect without a comprehensive appreciation of the whole problem may also lead to a biased approach. In fact, the land-sea distribution in many regions has given rise to monsoon phenomena which are concordant with the seasonal variation of the planetary wind belt aloft, resulting in an intensification of the monsoon flow. The monsoons over India and Southeast Asia act in this way. In winter, India lies south of the cold Continental High and within the trade-wind belt; hence the northeast monsoon is very stable.

In summer, India is located south of the Continental Heat Low and is affected by the passage of the equatorial westerlies during their northward excursion. This configuration not only gives rise to an intense and stable southwest monsoon over India but also enhances the pronounced seasonal variation of the monsoon flow (northerlies turning into southwesterlies). Conditions are somewhat different over China where the influence of the planetary circulation is discordant with the actual surface configurations. For example, during July and August in high summer, southwesterlies or southerlies prevail over the mainland south of the Yangtze with prevailing southeasterlies to its north over the southern part of North China. However, easterlies and westerlies or southwesterlies would be expected in the corresponding regions under the influence of the planetary circulation. Thus if the main factor in causing the monsoons is the planetary wind belt for India, then it will be the "influence of land-sea distribution" for China.

The monsoons over a particular region are, in fact, governed by the general circulation aloft, the land-sea distribution and the local topography. On average, a large low pressure system exists over the mainland of China in high summer. However, there are marked differences in properties within the same system between regions south and north of 35°N. Since upper westerlies dominate the regions north of 35°N in high summer, the frequency of occurrence of activities associated with aperiodic pressure systems is higher over this region than that of the south. These systems are, however, less frequent than in other seasons. As a result, the monsoons are not pronounced in the region north of 35°N and the prevailing winds become less stable with a lower frequency of occurrence. In the region south of 35°N the monsoonal influence is more significant because of the small meridional temperature gradient in the middle and lower troposphere, resulting in only weak atmospheric transportations in the east-west direction, and the infrequent occurrences of aperiodic pressure systems. Thus the pressure systems in the region are more "thermal" in nature and the pronounced monsoonal characteristics are reflected by stable prevailing winds with a high frequency of occurrence. The observed configurations reveal that with the same land-sea distribution marked differences in monsoon activities may be induced by meridional variations in the intensity of the general circulation over the affected region.

The effect of topography should not be neglected in examining the monsoons over a particular region. For example, techniques which have been found suitable for treating the monsoons over the plain areas in Central and South China may not be applicable in the Loess Plateau (1000 - 1500 m above MSL). The monsoon flow is most marked in the surface layer and extends to 2 km in winter and 3 - 5 km in summer over China. Thus the prevailing surface flow configuration over the plateau is similar to that in the free atmosphere above the friction layer over the plain regions of the mainland, and a strict comparison is therefore not justified. Furthermore, if a simple model of land-sea distribution is used in treating the monsoon phenomena over the Tibetan Plateau which is more than 3 km above MSL, then the results cannot be regarded as conclusive, because the plateau is already at the level of the planetary layer in the free atmosphere. The seasonal variation of the prevailing winds over the plateau is more complicated than the land-sea distribution or the difference in the flow cinfiguration between the plateau and the free atmosphere alone could account for, and may essentially be the result of the seasonal displacement of the planetary wind belt. In the present study, we have no intention to dispute the importance of the plateau as a heat source or sink to influence the monsoons. On the contrary, this feature will be emphasized in certain aspects of the monsoon problem.

In any investigation on monsoons, it is necessary to pay due attention to the size, shape and the distribution of land and sea. The findings of Khromov on the banding shape of monsoon regions successfully reveal the importance of the seasonal displacement of the planetary wind belt on the formation of monsoons. However, overestimation of its influence should be avoided. Two observed features may be used for illustration. The displacement of the planetary wind belt is the same in both hemispheres, but the banding shaped monsoon regions are all very small in size in the southern hemisphere and their monsoonal characteristics are not marked. In addition, the monsoons over the Eurasian Continent are more extensive and pronounced than over America. The l [illegible] re can only be accounted for in terms of the effect of the land-sea

We may now recapitulate that the monsoons are the result of the joint influence of the land-sea distribution, the general circulation and the underlying topography as a whole. In studies on the monsoons over any particular region, it is necessary to grasp these factors and their relative importance; otherwise the results are of superficial value only and do not enable us to gain a thorough understanding of the physical processes and the basic properties of the monsoons.

II. METHODS FOR THE STUDY OF MONSOON PHENOMENA

The first method relates the monsoons with the marked seasonal change of the prevailing wind systems. The change of wind direction may be ascribed to the influence of land-sea distribution, the seasonal displacement of the planetary wind belt or a combination of these and others factors. Investigators working along this line devise monsoon indices to account for the intensity of the monsoons over various regions. Hann [7] has made use of the sum of the differences between the frequency of occurrence of the prevailing winds and that from the opposite direction in January and July to evaluate the monsoonal influence over a particular region. Although such indices tend to suppress the influence of migratory pressure systems to some extent, they have no physical meaning in regions with complicated topography.

rad [8] analyzed the frequency of occurrence of winds from the ection between January and July, and has taken the sum of the m of their absolute values as a monsoon index. This index is capable of eliminating the effect of topography but is unable to filter the influence of migratory pressure systems. Subsequently, many investigators have combined various wind parameters to produce other forms of monsoon index. Although they look similar to the index of Conrad and are in a convenient form for practical use, they differ greatly from the latter in that the influence of topography have not been effectively suppressed as much as the investigators would wish to. Recently, Khromov [9] has introduced a new monsoon index in terms of the seasonal variation of the surface pressure gradient for January and July to represent the relative significance of the monsoon phenomena over a particular region together with a stability discriminant defined by half the sum of the frequency of occurrence of the observed prevailing wind direction in January and July for that region. The relative significance of the monsoon is depicted by the mean wind speed. Using this index, he has constructed a world map of monsoon regions. On the other hand, Kao and Hsu [58] have presented another monsoon index by taking advantage of the good points of the findings of Conrad and Hann and eliminating their shortcomings. The index is the sum of the maximum absolute values of the differences between the frequency of occurrence of winds from the same direction and that from the opposite direction in January and July and its value lies in the elimination of the effect of topography and the influence of migratory pressure systems. The resultant angular change is called the "monsoon angle". It may be noted that although the monsoon indices proposed by various investigators differ from one another and possess distinct advantages as well as shortcomings, they aim at the same goal in depicting the seasonal variation of the monsoons.

The second method deals with the intrinsic properties of the monsoons. In this approach, the march of the monsoons is necessarily related to the seasonal change of the prevailing wind direction and predominant air masses and the corresponding variations in weather and climate [59 - 60]. In other words, if there is no seasonal change in the prevailing wind direction, monsoon phenomena will not be observed. Furthermore, if the seasonal variation of prevailing winds is not accompanied by marked differences in weather and climate, the use of the term "monsoon" is not justified. Thus the existence of monsoons over any particular region must necessarily be characterized by marked seasonal changes of both elements.

Using the above concept, Conrad [8] has conducted comparative study of the monsoons over different regions. His analysis of the monthly variations of the frequency of the occurrence of prevailing winds, and the corresponding changes in cloud amount, rainfall, temperature and duration of sunshine in various European regions reveals that the seasonal changes of these important meteorological elements are not sufficiently marked to fulfill the criteria for the existence of monsoons.

Some writers ignore the seasonal changes of prevailing winds and restrict their scope of studies to the seasonal interchange of air-masses and weather or climate regimes. For example, Tu et al [61] have made use of the seasonal variations of the tropical maritime air-mass and the equatorial air-mass over the mainland of East Asia to discriminate the advance and retreat of the summer monsoon over China. Indian meteorologists have also taken the onset dates of the rainy season during the period May to June as the outburst of the southwest monsoon. Obviously, these methods can only hold good in regions where monsoons are known to exist.

In the third method, the monsoon is considered as the result of energy transportation through the circulation at both the lower and upper levels in opposite direction produced by differential heating over land and sea, and the relevant factors are then studied. This concept was first conceived more than two hundred years ago [1] and still attracts many followers [51 - 53] up to the present.

In this way, two distinct definitions of the term "monsoon" have evolved. The first requires marked seasonal changes in the prevailing wind direction over the affected regions, while the other permits relaxation on this point as long as the influence of the land-sea distribution is included in the configuration of prevailing wind flow.

III. DIFFERENT ASPECTS OF THE IMPORTANCE OF THE INFLUENCE OF MONSOONS

Since the constancy of the monsoons does not show up very well on the daily weather maps, some people may be of the opinion that the postulation of the existence of monsoons possesses no physical meaning. They may also think that the monsoons are low-level phenomena caused by differential heating of the underlying surface and therefore their influence on weather and climate are restricted to the low layers. Hence their contributions to the general circulation are not considered to be important. However, we feel that the above view is certainly open to criticism.

In the present study, the land-sea distribution is considered to be one of the deterministic factors governing the formation of action centers and their seasonal variations. If there is no uneven distribution of land and sea, then it would be difficult to imagine what form of action centers would prevail over specific regions on the global surface. Pogosian [47] has related the formation of atmospheric action centers and their seasonal variations to monsoon activities. He also attempts to explain the formation of the steady upper-air circulation pattern in terms of the influence of land-sea distribution. His findings lead us to consider that the appearance of these action centers and their evolutionary processes on the one hand give rise to the monsoons with an intensification of the meridional transportation in the atmosphere and on the other reveal the influence of the monsoons throughout the entire troposphere (though the monsoons are less pronounced at the upper-levels than at the surface). Moreover, monsoons are characterized by large-scale features in the form of action centers and do not belong to the domain of small-scale and local phenomena.

The characteristics of the monsoons must not be neglected in the discussion on the weather and climate over East Asia, because the development and maintenance of the semi-permanent action centers are basically associated with the processes of the monsoons. Since particularly intense monsoons over East Asia cause a breakdown of the planetary circulation in the middle and lower troposphere, exceptional features in the flow field and the vertical structure of the pressure systems are most prominent in these layers. Thus the weather and climate over China exhibit distinct characteristics which are not observed elsewhere at corresponding latitudes.

Although it is not yet possible to describe the current weather and the development of synoptic systems in terms of the variations of the action centers which are closely related to the activities of the monsoons, we believe that the behavior and development of these centers form the prime factor in determining the characteristics of the seasonal weather pattern for a particular region. This implies that the influence of the monsoons can be traced in the weather sequence observed even for a short period. In winter, differential heating between land and sea is particularly intense over East Asia. The Mongolian High and the Aleutian Low undergo marked development and cold anticyclones form the controlling synoptic systems over the mainland of China resulting in dry and cold weather. In summer, the pattern of heating becomes different. The Indian Low over the continent and the Pacific High show maximum growth, and become the dominating features on the daily weather map with sultry weather and plenty of thundery showers. Spring is the transitional season from the winter to the summer monsoons and is marked by the dissipation of winter action centers and the formation of the summer ones. The process is repeated in the reverse direction for autumn. The weather during the transitional seasons is characterized by an irregular interplay of mild and cool spells accompanied by the random occurrence of sunny and rainy periods. Thus if we neglect the influence of the monsoons over East Asia, it would be difficult to understand and account for the exceptional synoptic features observed during the various seasons.

At this stage it should be pointed out that the monsoon activities must be examined over periods ranging from one month to a season for prediction purposes because of the necessity to allow for the seasonal variations of the action centers. The formation development, maintenance and dissipation of these centers are closely and directly related to the seasonal variations of the monsoon elements.

Finally it should be borne in mind that the monsoons over East Asia exert an important influence on the national economic development of the affected countries. For example, the subtropical anticyclone dominates the regions south of the Yellow River including the entire Yangtze basin and most part of South China. Without the influence of the monsoons the climate over the productive regions of the Yangtze basin which is called the "Food Store" of China may become very dry and hot as in the principal deserts within the same latitude range elsewhere on earth, i.e., the Sahara, the Arabian, the Iranian and Afghanistan Deserts and those in Northwest India. Thus the monsoons tend to prevent the summer weather from becoming too hot and dry. Furthermore, after spring in each year, the summer monsoon brings in cool and moist air from the oceans, resulting in plenty of rainfall in South, Central and North China. Thus a vast area of land over China, which might have been a desert region is transformed into the "Food Store" of the country with two or three harvests each year. Of course, if the occurrence of droughts, floods and normal rainfall in China is related to the intensity and the times of onset and termination of the monsoons, then the "monsoon problem" will undoubtedly become even more important.

IV. THE APPROACH OF THE PRESENT STUDY ON MONSOONS OVER EAST ASIA

(a) A Resume of the Previous Investigations by Chinese Meteorologists on Monsoons

In China the monsoon phenomenon was first taken as a formal subject of study by Chu [62], who has successfully deduced from a limited amount of meteorological data the complicated relationship between precipitation in China and the monsoons of Southeast Asia. His findings indicate that under the monotonic influence of the pronounced southeast monsoon the weather is essentially dry but its interaction with the intruding cold air from the north results in plenty of rainfall. These conclusive statements have been proved to be correct by subsequent studies [63 - 64]. In 1938, Tu [65] has made use of a limited amount of upper-air data to describe the three types of monsoon air-mass over China and their interaction. Chang [66 - 67] then carried out a further study on the properties of the monsoon air-mass with more upper-air data. Although the latter investigation is more comprehensive in nature, the results are basically similar to those of Tu.

Meanwhile, Chu [68] and Tu [69] have separately described the general monthly activities of the monsoons by means of the prevailing wind flow and the positions of frontal zones in the surface and lower layers of the troposphere. Dao [70] and Kao [71] have presented monthly streamline charts with more surface and upper-air observations, which reveal the general structure of the monthly flow in relation to the advance and retreat of the monsoons.

In 1944, Tu and Hwang [61] discussed the advance and retreat of the monsoons in terms of the variations of the pentad mean of wet-bulb potential temperature, and brought out an average picture about the change of monsoon activities with time over China. Lu [72] describes the monsoons over China from a fresh approach and touches upon their influence on the weather and climate of the affected regions.

Because of the rapid development in meteorology after 1949, meteorologists have carried out much more research work on the general circulation over East Asia. Yeh and Dao [40] have related the abrupt changes of the general circulation in early June and mid-October with the onset and ending of the southwest monsoon over India. Dao et al [41] have presented a discussion on the mei-yü in China and examined the relationship between this phenomenon and the outburst of the southwest monsoon over India as well as the onset of mei-yü over Japan. Hsieh et al [29] have analyzed the structure of the low pressure system associated with the summer monsoon over China and the equatorial "front" by means of dynamical methods. Koo [73] has discussed the effect of plateau on the monsoons in his study on the dynamical influence of the Tibetan Plateau. Recently, Chang [74] has also presented an analysis of the properties of the monsoons over China.

Summing up, we note that although investigations on the monsoons of East Asia have been relatively few in the past years, they offer valuable guidance for the present study.

(b) The Present Approach to the Monsoons of East Asia

An analysis of the summer and winter isobaric patterns at various levels over East Asia and the characteristics of their seasonal changes shows that the surface pressure distribution is basically caused by the existing land-sea distribution, because the planetary wind belt cannot be found within the latitude range of its usual occurrence over this region. Thus the seasonal displacement of the planetary wind belt cannot be determined. If we follow the postulation of Flöhn [30] and maintain that the monsoons over Southeast Asia are the result of the seasonal displacement of the planetary wind belt, then we may quote his words that the absence of the subtropical easterly wind belt (or the trade-wind belt) in the surface layer over southern Asia in summer is a manifestation that the easterlies are lifted to the upper levels by the southwest monsoon, associated with the formation and development of the surface heat low. However, the confronting problem is: "Why do the southwesterlies (or equatorial westerlies) occur in southern Asia?" If the appearance of southwesterlies is thought to be the result of the northward displacement of the equatorial westerly wind belt, then it will be necessary to search for a reason for the exceptional northward excursion of the equatorial westerlies over the region. We are of the opinion that this is fundamentally the result of the influence of the distribution of land and sea. It may be seen from Khromov's [54] maps depicting the classification of monsoon regions that the monsoons over East Asia are more pronounced longitudinally than latitudinally. Hence we may say that the monsoons in the lower layers of the troposphere are essentially produced by the distribution of land and sea.

Of course, it is unreasonable to dispute the possible influence of the displacement of the planetary wind belt on the formation of the monsoons over East Asia. At upper levels above 3 - 5 km, the development may become a principal factor to bring about the seasonal change of wind direction.

In fact, both controlling factors form a unifying whole to account for the existence of the observed phenomena. However, as far as the formation of the monsoons over East Asia is concerned, we are of the opinion that the influence of land-sea distribution plays a major role in the surface layer in as much the same way as the planetary wind belt affects the circulation at the upper-levels. The relative importance of the factors will depend on the type of problem examined. For example, if we neglect the influence of land-sea distribution, difficulties may arise in accounting for the formation of the monsoons over East Asia or in explaining the flow fields at low and high levels and their variations. On the other hand, the introduction of the evolution of the upper planetary wind system facilitates the explanation of the advance and retreat of the monsoons at the various stages. The effect of the seasonal variation of the planetary flow has, in fact, been emphasized during our discussion on the march of the monsoons. Observations indicate that the occurrence of large amplitude horizontal displacement of the monsoons in the lower troposphere in mid-June, mid-July and early September is accompanied by marked changes in the upper-westerly circulation. Obviously, the two features are mutually related and they operate as elements of a unifying system. Variations in one will induce corresponding changes in the other though the degree of change may not be the same in both cases.

Thus we may specify the monsoons over East Asia by the following characteristics: (i) There must be significant seasonal changes in the prevailing wind direction or in the pressure patterns; (ii) the seasonal variation of wind direction and pressure systems gives rise to marked changes in weather and climate. Hence small seasonal changes in the pressure and flow patterns cannot produce significant variations in the weather and climate. For regions in East Asia and particularly in China, the land-sea distribution and the planetary circulation form a harmonious system in which one factor reinforces or supplements the other. However, in regions of complicated topography where the representiveness of surface winds is low, the latter factor is more effective in providing solutions to various problems.

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A DISCUSSION ON THE FORMATION OF MONSOON CLIMATES IN EAST ASIA*

by

Kao Yu-hsie and Kuo Chi-yunn

Although the problem on the causes of monsoon climate has been studied for more than 200 years, no satisfactory theory has evolved so far and a great divergence in opinion is still prevalent. Hadley [1] was the first to point out in 1686 that monsoons are caused by the differential heating over land and sea. His explanation was accepted by many subsequent workers. In recent years, Flöhn [2], Khromov [3] and others have established a completely different view, which necessitates a modification of the early conception. They consider that the phenomenon of the monsoons is mainly the result of the seasonal migration of the planetary wind belts. However, the school headed by Shuleikin [4] is still in favor of the theory of differential heating. Vitvitskii [5] has recently stated that the difference in heating between land and sea is not the main cause of the monsoons over South Asia. The phenomenon is rather due to the combined effects of differential heating, topography of the Asiatic mainland and the planetary wind systems produced by the thermal gradient between the equator and the poles. The views of the present authors on this problem have already been expressed in Reference [6]. As far as monsoons over East Asia are concerned, the difference in heating between land and sea is thought to be the most important and direct cause of their formation.

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* Manuscript completed in September 1959.

The seasonal displacement of the planetary wind belts may only be one of the less important factors since the monsoons over East Asia are extremely extensive and cover almost 49 degrees in latitude. Likewise, topography can only be ranked as a factor of secondary importance because its effects are generally localized in nature. These three aspects will be discussed in the following sections.

I. THE SEASONAL MIGRATION OF THE PLANETARY WIND BELTS IN RELATION TO THE FORMATION OF MONSOONS OVER EAST ASIA

Because of the influence of land and sea over East Asia, the planetary wind belts are generally seriously interrupted or broken up at the surface. The postulation of the seasonal migration of these wind belts as the main cause of the formation of monsoons over the region is therefore of no physical significance, since there is no available evidence in the lower troposphere to support the theory. If this theory is not without grounds, then the only place where they may be found is in the upper air at 3- or 5-km levels. For this reason, the planetary wind systems are not considered as the main cause of the monsoons.

(a) The Absence of Planetary Wind Belts on Surface Isobaric Charts over East Asia

Assuming that the average zonal pressure is independent of the effects of land and sea, then the distribution of planetary pressure belts can be identified on isobaric maps. Figures 1.1 and 1.2 prepared under such an assumption depict the long-period mean meridional distribution of surface pressure over Eurasia, East Asia, North America, the Pacific and Atlantic Oceans and the northern hemisphere. If the distribution of the planetary wind belts can be represented by the meridional pressure profile and the variations of this profile can be taken to indicate the seasonal change of the positions of the wind belts, then it is clearly seen from these figures that the curves for East Asia (and also Eurasia) are distinctly different from those for the northern hemisphere. Pressure variations over East Asia in January and July are not related to changes in the position of pressure systems but rather due to the transformation of their properties.

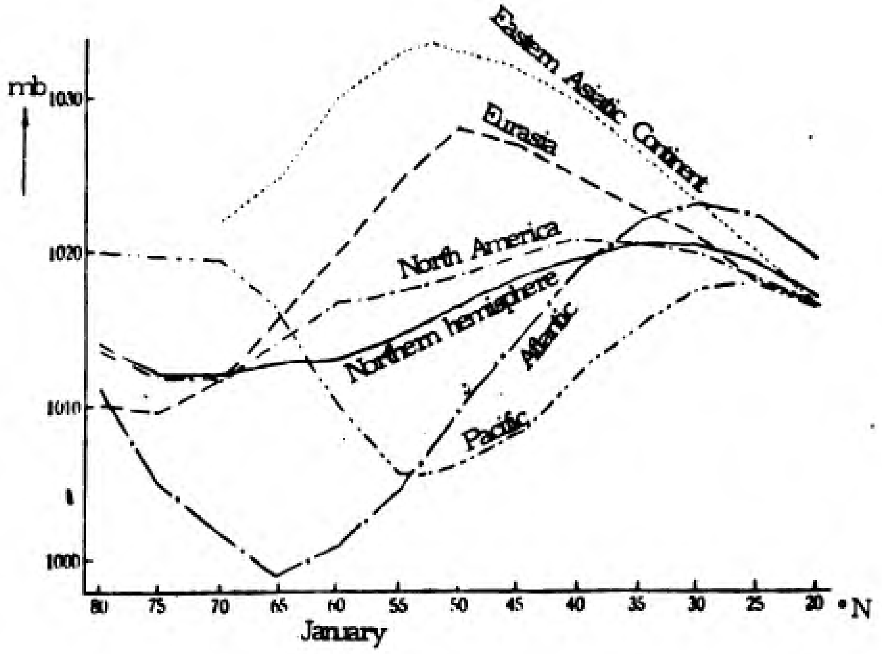

Figure 1.1

Meridional distribution of MSL pressure averaged along latitude circle in January.

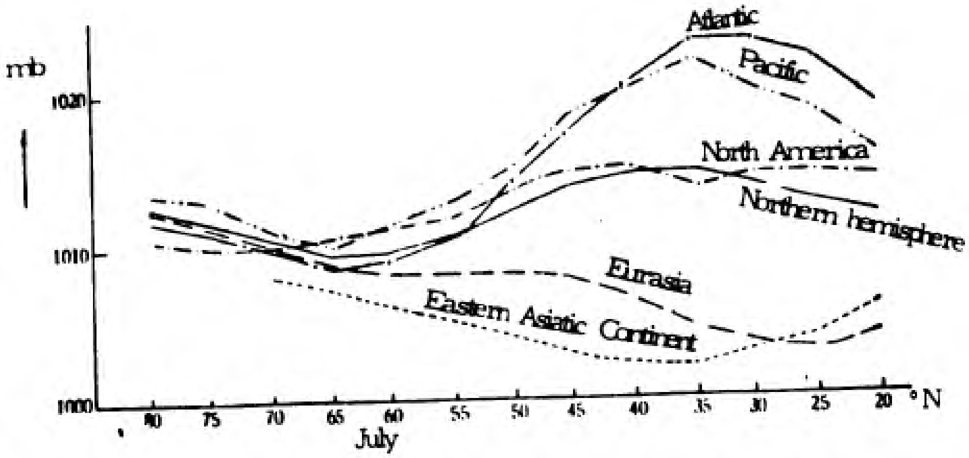

Figure 1.2

Meridional distribution of MSL pressure averaged along latitude circle in July.

In January, pressure is high near 35°N in the northern hemisphere with a low pressure belt at 70°N; over Eurasia, the high pressure belt lies between 50°N and 55°N. This indicates that the normal westerly wind regime is replaced by the easterlies and the normal polar easterly wind belt becomes a region of westerlies over the eastern Asiatic Continent. The situation is even more significant on the July chart which shows that 38°N is the latitude of highest pressure averaged over the whole northern hemisphere, but this latitude in fact passes through the belt of lowest pressure over East Asia. With the exception of the polar region, the same tendency is observed for all other places. Thus, the distribution characteristics of the planetary wind belts cannot be found over East Asia or even Eurasia.

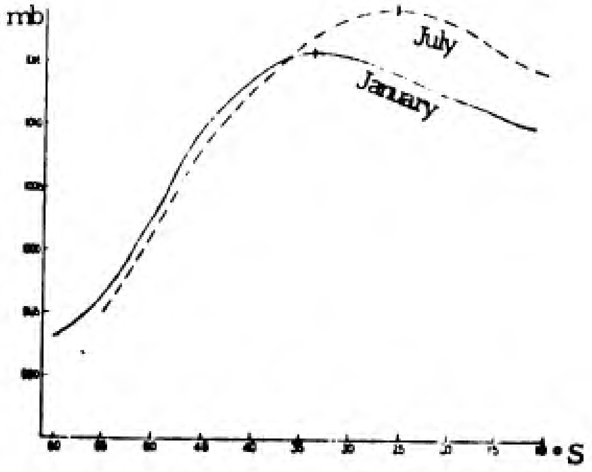

Figure 2

Meridional distribution of MSL pressure averaged along latitude circle for the southern hemisphere.

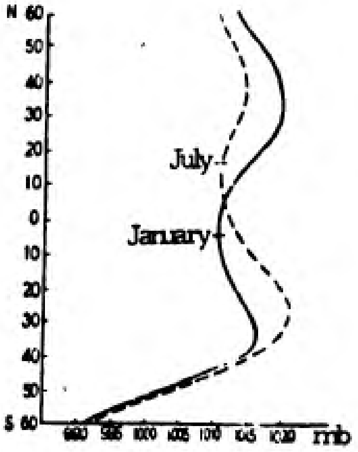

Figure 3

Meridional distribution of MSL pressure averaged along latitude circle [20].

A comparison of the profiles for the two hemispheres shown in Figure 2 on the other hand reveals that the subtropical anticyclone shifts 2 - 3 degrees latitude (from 35°N to 38°N) from January to July while the corresponding system over the southern hemisphere does 10 degrees latitude (from 25°S to 35°S). The equatorial trough is located at 4°S in January but it moves to 14°N in July giving a total displacement of 18 degrees latitude (Figure 3). If the conditions over the southern hemisphere can be taken to represent the typical distribution pattern of the planetary wind systems, then the small displacement shown by these planetary wind belts in the northern hemisphere would serve to indicate the important influence of the distribution of land and sea along the various latitude circles The asymmetrical displacement of the equatorial trough between January and July is also due to the non-uniform latitudinal distribution of land and sea between the northern and southern hemispheres. A study of the average conditions over both hemispheres shows that the planetary wind belts are greatly disturbed or broken up over the northern hemisphere, which is characterized by huge land masses and a complicated distribution of land and sea.

The problem can be more effectively illustrated by a comparison of the conditions over the southern hemisphere and Eurasia. The winter high pressure belt occurs between 50°N and 55°N over Eurasia and near 25°S in the southern hemisphere. In summer, the high pressure belt in the southern hemisphere (January) is situated at 35°S but the corresponding latitude of 35°N over Eurasia is the seat of a trough of low pressure. The summer trough in the southern hemisphere moves southward to only 4°S but the system travels as far north as 25° - 30°N over the Eurasian Continent with a displacement 5 - 6 times greater than that in the southern hemisphere.

If the displacement of the planetary wind belts is the main cause of the monsoons, then the monsoonal phenomena should be most significant and typical in the southern hemisphere. However, this is not found to be the case. Figure 4 is a map of monsoon regions prepared by Khromov [7]. A study of the size of the regions affected by the monsoons over both hemispheres would serve to demonstrate the importance of the land-sea distribution.

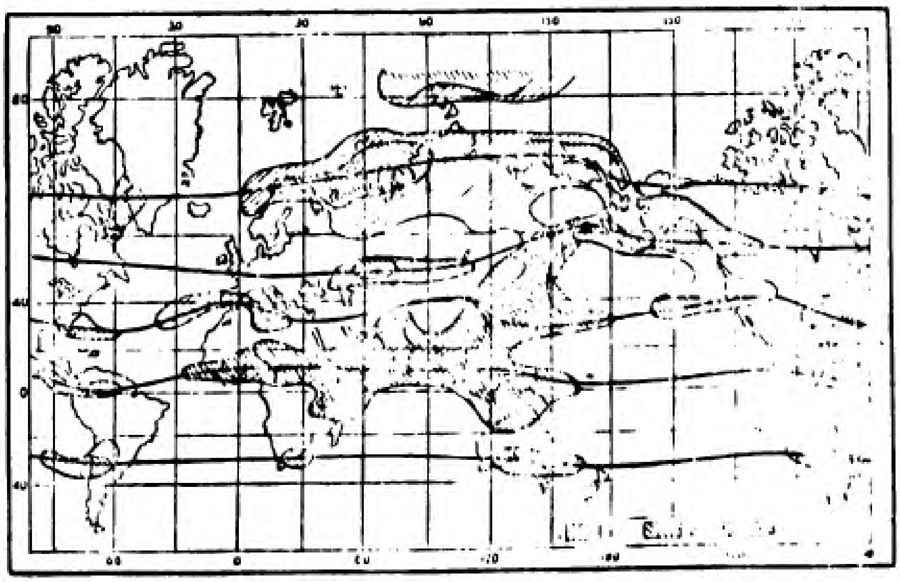

Figure 4

Geographical distribution of monsoons.

1. Frequency of prevailing wind in January and July < 40%.
2. Frequency of prevailing wind in January and July 40-60 %.
3. Frequency of prevailing wind in January and July > 60%.

Angular change of monsoon winds between 120° and 180° in all cases.

In most text books, the prevailing northeasterly winds in winter over the southeastern part of Eurasia, including the coastal regions of South China, Southeast Asia and the Middle and Near East, have often been referred to as the northeast trades. We consider that these wind regimes are distinctly different from the northeast trades over the ocean, i.e., regions in the southern part of the subtropical high pressure system, in terms of their causes, vertical structure and air mass properties. The only similarity lies in the fact that they all occur at about the same latitudes. Some investigators have advocated that the southwest monsoon over Southeast Asia is a result of the northward movement of the equatorial westerly wind belt (a planetary wind system). It is, however, difficult to explain why this belt should move so far north over Southeast Asia and India but not over other regions at similar latitudes. We feel that no satisfactory answer can be given without first considering the effects of the distribution of land and sea. We can also state with confidence that these effects are particularly important over Eurasia in view of its exceptionally huge land masses so that the planetary circulation is generally broken up in the lower troposphere. The feature is most noticeable over East Asia.

(b) Possible Effects of Land-sea Temperature Contrast on the Formation of Monsoons

Figures 5.1 and 5.2 show the mean temperature difference between the oceans and the Eurasian Continent at the various standard levels for the same latitude in January and July 1956 [8]. It is seen that:

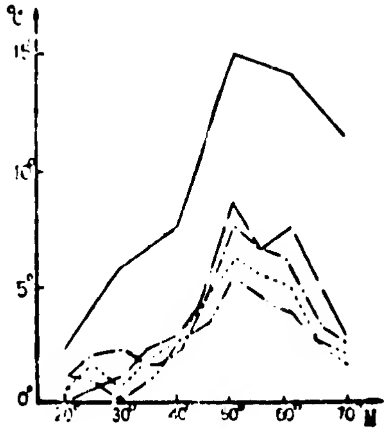

Figure 5.1

Meridional distribution of temperature difference between the Eurasia-Africa Continents and the oceans in January 1956 [8].

——— Sea level　　— — 850 mb
—·—· 700 mb　　..... 500 mb
—··—300 mb

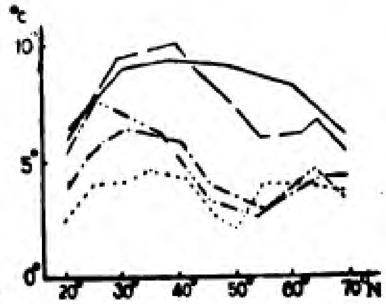

Figure 5.2

Meridional distribution of temperature difference between the Eurasia-Africa Continents and the oceans in July 1956 [8].

(Legend same as Figure 5.1.)

(i) The temperature difference is largest at the surface and decreases rapidly with height. The value is extremely small at 850 mb and becomes insignificant at 700 mb in summer.

(ii) A transitional "point" is located at 45°N. The values of temperature difference are higher to its north than to its south in winter with a maximum at 50°N. In summer, the opposite tendency is noted, which is particularly marked at 700 mb.

(iii) The values of maximum difference are found in regions of atmospheric action centers.

The above observed features lead to the following two tentative conclusions:

(i) The effects of land and sea are most pronounced in the lowest layer of the troposphere and they diminish with height. However, their influence is still noted at 10 km.

(ii) The regions of large temperature difference are in general the seats of stable atmospheric centers. It may therefore be said that the distribution of land and sea plays an important though not unique role in the formation of atmospheric action centers.

To show the effects of land-sea temperature contrast on the planetary circulation, Table 1 has been constructed to give the ratio of the temperature gradient between land and sea and the meridional gradient for the various latitude circles. Below 850 mb, the temperature gradient between the oceans and the Eurasian Continent is tighter in the south than in the north during summer. In winter, tighter gradients are found in the north. In the summer hemisphere south of 30°N, the temperature gradient between land and sea is greater than the meridional gradient by 1 - 2 times. However, the former value is smaller than the latter north of 30°N and the ratio of these two terms lies between 0.4 - 0.6. The ratio is generally much smaller at 700 - 300 mb although large values are still found in low latitudes in summer. It is evident from the above statistics that in the lower troposphere, the magnitude of the temperature gradient produced by the distribution of land and sea is sufficient to destroy the planetary wind systems. In winter, the planetary wind belts are also liable to be disturbed by this factor at high latitudes.

The evaluation of the effects of the distribution of land and sea in terms of land-sea temperature difference is subject to many limitations and shortcomings. This is particularly true when the inferences presented are based on the discrete conditions of individual years only. Thus, it is not possible to explain why a bigger difference should occur in the upper instead of the lower levels at 25°S as shown in Table 1. However, an analysis of the situations from individual years does provide certain factual evidence which is not generally expected from theory. For the problems under discussion, it may be stated that the effects of land and sea should be considered as the most important parameter in the lower troposphere.

(c) Evaluation of the Effects of the Distribution of Land and Sea on the Formation of Monsoons by Means of Annual Pressure Variation

If the mean annual pressure range for the various latitude circles is not affected by the distribution of land and sea but is only due to factors such as seasonal variation of the planetary circulation and the difference in size of the land masses between the northern and southern hemispheres, then the annual pressure variation caused by the influence of the land and sea at any level of a location may be evaluated by subtracting the mean annual range for the latitude of that location from the observed total range. A computation of this parameter for the surface and 700- and 500-mb levels has been carried out for stations in the eastern part of the Asiatic Continent but only two stations with long-period records, namely Tientsin and Shanghai are listed here for illustration. Since the length of upper-air records is short for most stations, the longer series of observations from Nanking and Peking (1953 - 1957) have been used instead. The results are given in Table 2, which indicates that the annual pressure variation caused by the effects of land and sea is greater by 1 - 3 times than that contributed by other factors and constitutes a major portion of the observed variation at the surface. At 700 mb, the land and sea effects are considerably diminished and only produce an amount which is 0.38 - 0.63 of that caused by other factors. At 500 mb, the effects of land and sea become insignificant. If the planetary circulation is considered to be fully established at the upper levels in January, then the ratio for 700 mb for this month (given in the table) may be used as a reference standard to assess the relative importance of the land-sea effects. A study of the values of the ratio for various months and levels leads us to conclude that the phenomenon of the monsoon climate produced by these effects is most predominant near the surface.

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The evaluation of the effects of the distribution of land and sea in terms of land-sea temperature difference is subject to many limitations and shortcomings. This is particularly true when the inferences presented are based on the discrete conditions of individual years only. Thus, it is not possible to explain why a bigger difference should occur in the upper instead of the lower levels at 25°S as shown in Table 1. However, an analysis of the situations from individual years does provide certain factual evidence which is not generally expected from theory. For the problems under discussion, it may be stated that the effects of land and sea should be considered as the most important parameter in the lower troposphere.

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TABLE 1

Ratio between temperature gradient over the Eurasian Continent and the oceans and meridional temperature gradient

Mean of surface and 850 mb (1956)									Mean of 700, 500 and 300 mb (1956)								
Latitude / Ratio / Time	25	30	35	40	45	50	55	60	Latitude / Ratio / Time	25	30	35	40	45	50	55	60
July	2.2	1.2	0.6	0.4	0.4	0.4	0.6	0.6	July	3.4	0.6	0.4	0.2	0.2	0.2	0.3	0.3
January	0.1	0.1	0.1	0.1	0.2	0.3	0.4	0.4	January	0.0	0.0	0.0	0.1	0.2	0.4	0.4	0.2

TABLE 2

Station	Length of record	Altitude	Annual pressure range (I)	Pressure difference due to land-sea effects (II)	Pressure difference due to other factors (III)	Ratio of (II)/(III)
Tientsin	28 years	Surface	24.8 mb	19.6 mb	5.2 mb	3.77
Peking	5 years	700 mb	152 m	42 m	110 m	0.38
Peking	5 years	500 mb	392 m	72 m	324 m	0.22
Shanghai	68 years	Surface	21.9 mb	15.3 mb	6.6 mb	2.32
Nanking	5 years	700 mb	67 m	21 m	46 m	0.45
Nanking	5 years	500 mb	257 m	75 m	182 m	0.41

The above method has also been used to determine the influence of land and sea on the formation of the four action centers in East Asia. The results are depicted in Table 3 and are basically similar to those given in Table 2. In the surface layer (from surface to 700-mb level), the ratio of land-sea effects to other factors is less than unity only in the center of the Pacific Anticyclone. However, at 700 mb, the distribution of land and sea on the formation of the intense low pressure area centered over India is considerably more important than all other factors.

In the light of the above findings, we consider that the influence of land and sea is most marked over East Asia and it has also a two-fold effect. On the one hand, it tends to break up the planetary wind circulation over the region and on the other it assumes the role of a controlling factor for the formation of pressure systems in the lower troposphere.

TABLE 3

Atmospheric action centers		Annual pressure range (I)	Contribution by land-sea effects (II)	Contribution by other factors (III)	Ratio of (II)/(III)
Surface	Mongolian High	30.0 mb	25.5 mb	4.5 mb	5.67
	Aleutian Low	16.0 mb	11.5 mb	4.5 mb	2.56
	Pacific High	10.0 mb	4.4 mb	5.6 mb	0.78
	Indian Low	21.0 mb	14.4 mb	6.6 mb	2.18
700 mb	Mongolian High	101 m	-95 m	195 m	0.48
	Aleutian Low	273 m	78 m	195 m	0.40
	Pacific High	113 m	-27 m	140 m	0.19
	Indian Low	-3 m	-50 m	47 m	1.06
500 mb	Mongolian High	472 m	79 m	383 m	0.20
	Aleutian Low	435 m	51 m	383 m	0.13
	Pacific High	223 m	-127 m	350 m	0.36
	Indian Low	116 m	1 m	115 m	0.01

II. THE EFFECTS OF HEATING AND COOLING BETWEEN LAND AND SEA IN RELATION TO THE FORMATION OF MONSOONS

We will now discuss the difference in heating between land and sea and its seasonal variation in relation to the establishment and seasonal changes of the pressure field in the lower troposphere.

(a) Differential Surface Heating between Land and Sea as an Important Factor in the Establishment of the Monsoon Pressure Field

The influence of land and sea on the distribution of pressure systems is well recognized by meteorologists. Many theories have been proposed to account for the observed features, but like many other major meteorological problems, no satisfactory explanation has yet come to light. From our experience and understanding in this subject, we are of the opinion that the phenomenon of monsoons is mainly caused by the interaction of high and low pressure systems produced by the unequal heating of land and sea. To show that these systems are related to the distribution of surface heating, Figures 6.1 and 6.2 have been constructed for comparison with Figures 7.1 and 7.2. The former two figures were taken from the article entitled "Steady-state Perturbations in the Westerlies in Relation to Large-scale Heat Sources, Heat Sinks and Topography" (Figures 7 and 8) by Chu Pao-chen [9]. The data on which these figures were based are completely different in respect of both time of observation and length of record. However, in the absence of homogeneous long-period data for accurate comparison, the two sets of figures presented still bring out certain salient features worthy of note:

(i) Both the temperature field and the pressure systems exhibit characteristics of seasonal variation. The pressure and heating configurations in winter (January) are completely opposite to those in summer (July). In winter, the continents are high pressure cold sources while the oceans act as low pressure heat sources. In summer the opposite conditions are observed with the continents as low pressure heat sources and the oceans high pressure cold sources.

(ii) Both the isobaric and heating patterns are characterized by individual centers of opposite tendencies with their boundaries near the coastline. However, the pressure centers do not coincide with the centers of heating or cooling although their positions are generally related.

Thus high (low) pressure is nearly always found to the right of a heat (cold) source. This is due to the dynamic processes involved in the formation of the pressure pattern caused by the heat and cold sources. When the pressure field is established, it would in turn exert some influence on the distribution of the heat and cold sources.

Figure 6.1

Distribution of heat sources and sinks in the lower half of the troposphere over the northern hemisphere in January [9]. (Unit: 10^{-5} cal · gm^{-1} · sec^{-1})

Figure 6.2

Distribution of heat sources and sinks in the lower half of the troposphere over the northern hemisphere in July [9].

(Unit: 10^{-5} cal · gm^{-1} · sec^{-1})

(b) Seasonal Variation of the Difference in Heating between Land and Sea in Relation to the Annual Variation of the Pressure Field

The latitudinal variations of the differences in pressure and heating between land and sea are depicted in Figures 8.1 and 8.2, which were based on data from References [10] and [11]. In constructing these figures, the mean values of the parameters for each latitude between

110°E and 115°E are taken to be representative of the conditions over land for that latitude while those over the Pacific Ocean between 145°E and 150°E are regarded as representative values over the sea.

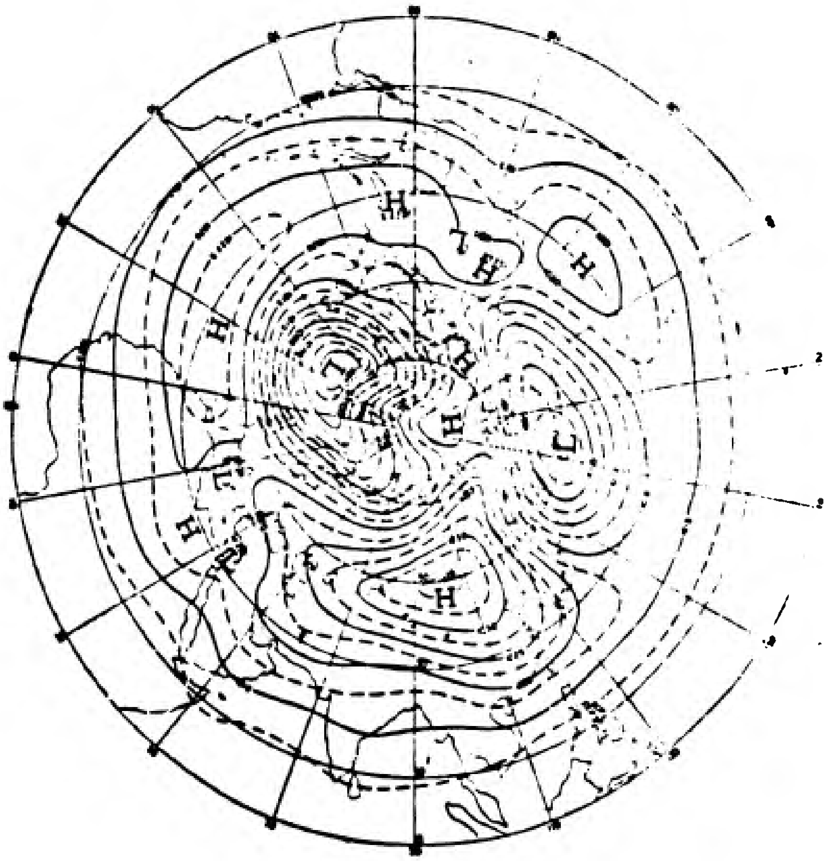

Figure 7.1

Mean sea-level pressure over the northern hemisphere in January [10].

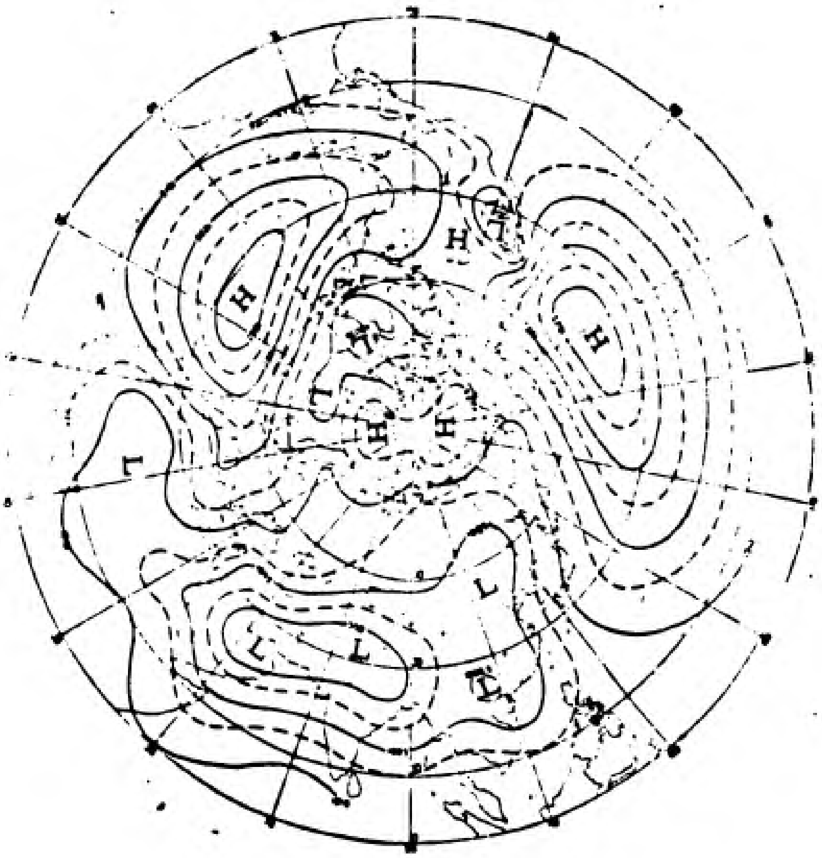

Figure 7.2

Mean sea-level pressure over the northern hemisphere in July [10].

Figures 8.1 and 8.2 show that:

(i) There is a well-marked seasonal variation in both entities. During the winter half year, the pressure difference is positive for nearly all latitudes, implying that pressure is higher over land than over sea. The sign for heating is negative to indicate that the continent is colder than the ocean. The opposite conditions are found in the summer half year when pressure difference becomes negative and heating positive.

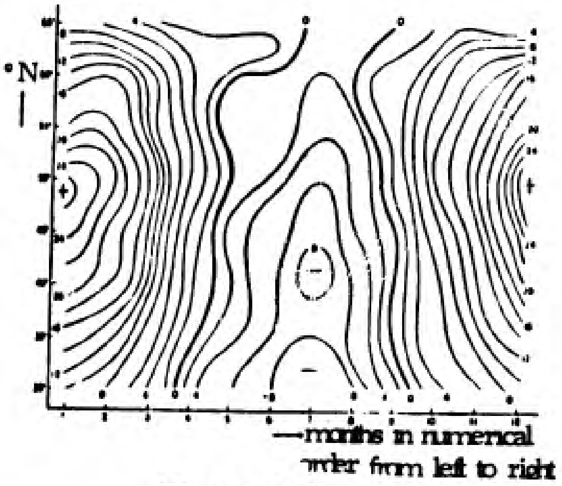

Figure 8.1

Latitudinal variation of pressure difference between land and sea for the various months.

(Unit: mb)

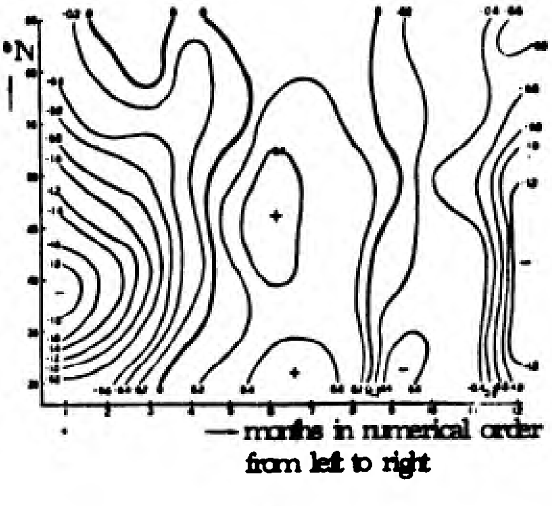

Figure 8.2

Latitudinal variation of the difference in heating between land and sea for the various months.

(Unit: 10^{-5} cal · gm^{-1} · sec^{-1})

(ii) The transition from positive to negative values or vice versa of pressure difference and heating occur simultaneously north of 45°N from summer to winter. South of this latitude, the heating difference changes sign to negative in September but the pressure difference does not become positive until October. From winter to summer, the change of signs for both entities occur generally at the same time over the whole region. However, it should be noted that the unit of time used in the analysis is the "month" and it appears that some variations may become noticeable when the "pentad" or "decade" is adopted instead.

(iii) The magnitudes of pressure and heating differences are larger in winter than summer and intense pressure systems are always associated with large values of heating difference. The monthly variations of these two entities are also much greater in winter than summer months.

(c) Seasonal Variations of the Effects of Land and Sea in Relation to the Evolution of the Monsoon Phenomenon

The above discussion of the effects of land and sea on the formation of monsoons has been confined to the conditions during winter (January) and summer (July). Their significance during the transitional seasons has not been described. In China, the continental influence is much more pronounced during the winter monsoon while the oceanic influence becomes more important in the summer monsoon. The governing factor which determines the relative importance of these influences is the variation of the incoming solar radiation during the year. The seasonal variation of the difference in heating between land and sea is most significantly reflected in the pressure field. In January, the high pressure systems over land and the low pressure systems over sea are most intense and this gives rise to a maximum pressure gradient from land to sea. In July, the pressure gradient is also near a maximum but directed in the opposite sense. The transitional period lies between late winter and mid-summer. Due to lack of data, the evolution of the monsoon phenomenon and the seasonal variations of the differential heating over land and sea can only be discussed in terms of the seasonal changes in the pressure difference. Figure 9 depicts the variations of the five-day mean of sea-level pressure over land (solid line) and sea (broken line) during the course of the year. The values over land are the averages of observations taken at the following six points: 45°N, 115°E; 40°N, 110°E; 40°N, 120°E; 35°N, 110°E; 35°N, 120°E and 30°N, 115°E. The six corresponding point over the sea are: 45°N, 150°E; 40°N, 155°E; 40°N, 145°E; 35°N, 145°E; 35°N, 155°E and 30°N, 150°E. The ordinate is pressure in mb and the abscissa time in pentad. This figure shows that from the third pentad in December to the second pentad in February, both curves are similar in shape and an almost steady pressure difference is maintained. From the fourth pentad in February to the fourth pentad in April, pressure is still higher over land than sea but the difference is gradually decreasing. From the fourth pentad of February onward, pressure begins to rise rapidly over sea and fall over land and a change of sign occurs in the fourth pentad of April. At this time, the summer monsoon begins to affect South China. Between the fourth pentad of June and the third pentad of August, both curves remain fairly steady and the pressure difference between land and sea reaches a maximum. This is the period when the summer monsoon attains its full development over China. Thereafter, pressure starts to rise again over land and becomes greater than that over the sea during the fourth pentad of September. The continental influence once again plays its role and triggers the onset of the winter monsoon over a great part of China. The regular interchange of the above processes during the course of the year and the uniformity in the dates of occurrence of the winter and summer monsoons provide further evidence of the effects of the distribution of land and sea on the formation of monsoons.

A study of the pressure variations over land and sea shows that the periods from mid-June to mid-August and from the beginning of December to mid-February are relatively stable and quiet while the periods from mid-February to mid-June and from mid-August to early December are periods of development. In terms of the seasonal variation of the effects of land and sea, it may be said that the influence of land is most prominent from early December to mid-February and the oceanic effects do not come into play until the end of this period when a rise of pressure is observed over sea and a fall over land. The process is, however, too weak to revert the sign of pressure difference and only results in a weakening of the pressure field associated with the winter monsoon. The change of sign takes place in mid-April as the summer monsoon slowly intensifies. The oceanic influence reaches its maximum between mid-June and mid-August, which is a stable period characterized by the full establishment of the summer monsoon pressure field. From mid-August onward, the continent begins to exert its influence once more and produces a rise of pressure over land with a corresponding fall over the ocean. The summer monsoon pressure field then weakens and is replaced by that of the winter monsoon after mid-September. The maximum intensity of the latter occurs in early December.

In terms of the intensity and evolutionary processes of the monsoon phenomenon, it may be said that early December to mid-February is the period of steady and intense winter monsoon over China; mid-February to mid-April represents the weakening phase of the winter monsoon; mid-April to mid-June marks the onset and intensification of the summer monsoon; mid-June to mid-August is the peak period of the summer monsoon; mid-August to mid-September is noted for the weakening of the summer monsoons and mid-September to early December marks the onset and intensification of the winter monsoon.

Several other points are worthy of note:

(i) The variation of the pressure gradient between land and sea is mainly produced by the variation of pressure over the continent since the pressure change over the sea is very small. However, when large variations are observed in the oceanic field such as in mid-March, the first decade of September and late November (Figure 9), some significant changes in the monsoons are also noted. Thus, in order to examine the evolution of the monsoon phenomenon, the behavior of pressure systems over land and the development of several oceanic systems should be considered.

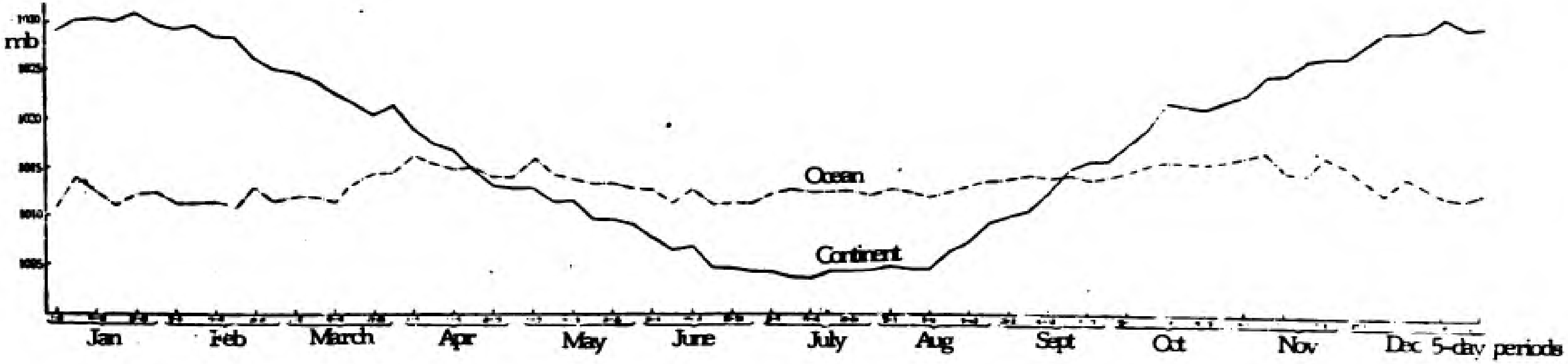

Figure 9

Annual distribution of 5-day mean surface pressure.

(ii) The mean variation of pressure over land is almost linear with time during the periods from late January to mid-June and from mid-August to early December. The contributions due to the influence of land and sea and the effects of the atmospheric circulation and topography may therefore be estimated from the annual profile.

(iii) The variation of pressure over the ocean is rather irregular and the surface pressure is lower in summer than spring or autumn. This may be due to the choice of the locations of observing stations included in the study. However, if the results are representative, then this phenomenon would possess some physical significance and should merit further investigations. It is also noted that abrupt rises and falls are commonly observed over the sea. Thus, nonlinear factors are more important over the ocean than the continent.

Although the above findings have been obtained by means of a relatively crude analysis, they provide sufficient evidence to account for the existence of different types of pressure fields in the same season. From the results of the above analysis, we are of the opinion that the distribution of pressure systems in the steady state is solely dependent on the differential heating over land and sea and that the seasonal variation of this heating difference gives rise to a similar variation in the pressure field.

III. THE INFLUENCE OF THE TIBETAN PLATEAU ON THE FORMATION OF MONSOONS

The influence of the Tibetan Plateau on the general circulation has been extensively studied by meteorologists [12, 13] in China. The problem of the existence of monsoon phenomena over the plateau has also been dealt with in detail in Reference [14]. In the ensuing sections, the emphasis of discussion is laid on the aspects of the heating and mechanical effects of the plateau on the development of monsoon climates over China.

(a) The Mechanical Effect of the Tibetan Plateau

Due to the huge size of its land mass, the Tibetan Plateau exerts a pronounced influence not only on the zonal currents of the atmosphere but also on the meridional circulations such as the northwest monsoon in winter and southeast and southwest monsoons in summer. The branching, weakening and intensification of these currents are all related to the presence of the plateau. Since the monsoon is a low-level phenomenon, the abovementioned effect is even more marked in monsoon situations.

(i) The eastern edge of the Tibetan Plateau is oriented from north to south, which facilitates warm or cold meridional advection over the plain to its east during the winter and summer monsoons. It is well-known that not only the monsoons over East Asia are more intense than those of other regions at the same latitudes but that the latitudinal limits affected by the Asian monsoons are also wider. This accounts for the high annual variability of the meteorological elements over the region. Figure 10.1 represents a meridional cross-section of temperature and pressure differences between January and July along 20°N and 40°N. Profiles of temperature difference along the vertical between these months are situated south of the plateau, are sheltered from the winter monsoon so that the temperatures over these regions are generally higher and pressure values lower than those to the east and west for the same latitudes. The annual ranges of temperature and pressure over these regions are also very small (Figure 10.1). In Kansu and Sinkiang, which lie to the north of the plateau, the weather is usually dry with little rain or cloud (even in summer) because of the blocking of the summer monsoon by the highland to the south. The above inferences may easily be verified by an inspection of the humidity observations at the four stations listed in Table 4.

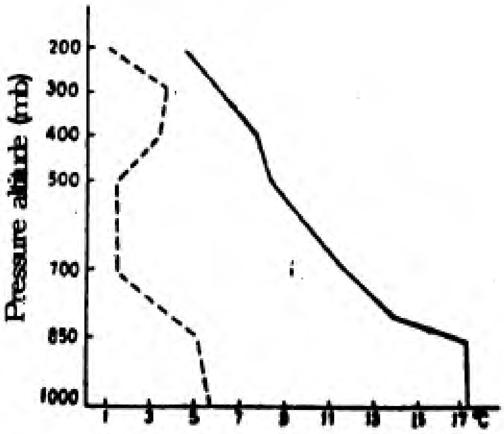

Figure 10. 2

Temperature difference between January and July at various pressure altitudes over Foochow and Miami.
(—— Foochow, ---- Miami)

Tengchung and Chiuchuan are located along the same meridian on the northern and southern sides of the plateau respectively. The mean values of absolute humidity at the various levels over Tengchung

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TABLE 4

Long-period mean relative humidity and absolute humidity over the various stations in July

Pressure altitude (mb) / Humidity / Station		1000	900	850	800	700	600	500	400
Tengchung	Relative humidity (%)				92	91	82	79	67
	Absolute humidity (gm/kg)				13.7	10.7	7.5	5.0	2.2
Chiuchuan	Relative humidity (%)				37	54	40	30	30
	Absolute humidity (gm/kg)				6.6	6.3	2.8	1.3	0.6
Swatow	Relative humidity (%)	87	84	78	67	57	50	42	39
	Absolute humidity (gm/kg)	19.7	17.1	13.3	10.6	7.2	4.7	2.7	1.3
Peking	Relative humidity (%)	80	71	65	64	58	61	45	36
	Absolute humidity (gm/kg)	15.3	12.5	10.3	9.0	6.0	4.2	2.1	0.8

are 1 - 3 times greater than those at Chiuchuan while the mean values of relative humidity of the two stations also differ by a factor of 2. It may be noted that Peking and Swatow are likewise situated along the same meridian, but due to the absence of any highland between them, their humidity observations are of the same order of magnitude except perhaps at the lowest level. In fact, no appreciable difference is observed for levels above 800 mb.

The Tibetan Plateau also exercises some control on the behavior of troughs and ridges in the upper westerlies, and indirectly increases the stability of the monsoons over China. This is manifested in the damping and modification of migratory systems from the west. Thus, some waves dissipate completely after passing over the plateau while others become considerably weaker. Figure 11 was prepared from data given in Reference [15] and shows the meridional distribution of the frequency of occurrence of cyclones along 40°N. It is seen that due to the blocking effect of the plateau, the relevant frequency of occurrence shows a marked decrease in its neighborhood. Consequently, the monsoon pressure field over the plains in East China is generally maintained with only infrequent interruptions.

(iii) The branching of monsoon currents over the Tibetan Plateau: The branching of the Indian summer monsoon by the southwestern part of the Tibetan Plateau has already been discussed by Yeh [16]. He has found that when the southwest monsoon reaches the Bay of Bengal, it splits into two branches. The first travels along the Himalayas and turns eventually into a cyclonic easterly to maintain a depression over the northern part of India. The other branch follows the mountain ranges and penetrates into the southwestern region of China as a slightly anticyclonic flow. This configuration is clearly seen from the surface up to 6 km.

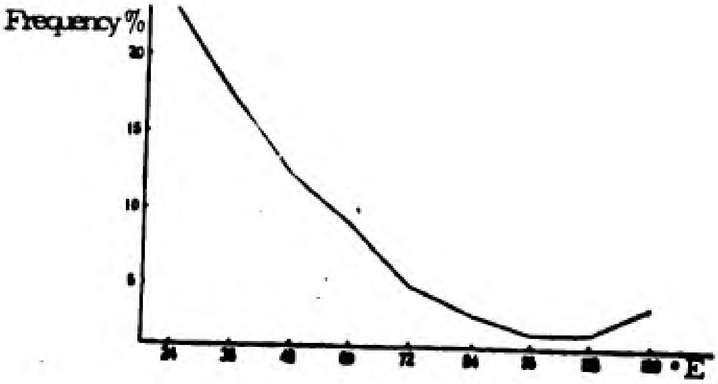

Figure 11

Frequency of occurrence of cyclones along 40°N versus longitude in January.

The branching of the winter monsoon is also observed over the northern part of the plateau although the phenomenon is generally less conspicuous than that in the south. When a northerly air-stream reaches Chilien Shan near 95°E, it frequently branches into two separate currents. The first penetrates into the Tarim Basin along the Altyn Tagh as an easterly wind while the other flows over the Hohsi Corridor* as a west or west by north air-stream. The former is greatly influenced by the topography of the Tarim Basin and usually appears as an anticyclonic circulation. This branching phenomenon is most apparent near the surface

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* Hohsi Corridor is also known as Kansu Corridor and represents the "corridor" region west of the Yellow River.

(Figure 12, but becomes undetectable at 3 km or above (Figure omitted). The anticyclonic flow over the Tarim Basin is a persistent feature on the weather map. Although there is no evidence to prove that this circulation is produced and maintained entirely by topography it is nevertheless considered that the branching effect of topography must play an important part during its formative stage.

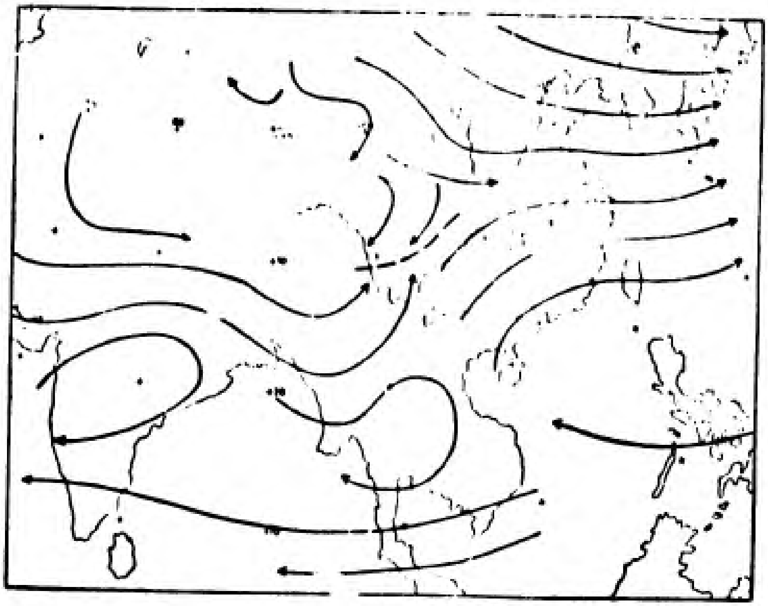

Figure 12

Mean flow field at 2 km in January.

The branching effect of topography thus brings about a mechanical enlargement of the monsoon area over the southern and northern parts of the Tibetan Plateau.

(b) The Effect of Heating of the Tibetan Plateau on the Monsoons

In order to discuss the topic, it is necessary to examine the behavior of the Tibetan Plateau in the transfer of heat and its seasonal variation.

The problem has been investigated by many workers [17, 19] in China and their findings all indicate that the plateau is basically a heat source in summer and a cold source in winter (with perhaps the exception of a few localized regions). The following discussion deals with the influence of the Tibetan Plateau as a heat or cold source on the monsoon circulations and is based on the result of these research workers.

During the period of the prevailing winter monsoon, the mainland of China is covered by a cold anticyclone or ridge, which is associated with general subsidence in the atmosphere. The Pacific Ocean, on the other hand, is covered by a low pressure area where ascending motions are observed. The coast line forms a transitional zone between the high and low pressure systems. This may be clearly seen from the distribution of the mean vertical wind speed in the lower half of the troposphere over the northern hemisphere for January computed by Chu [9]. Thus, a mass transport from land to sea takes place in the lowest layer of the troposphere with a compensating flow from sea to land above the layer of monsoon winds. However, the transport from sea to land is in fact masked by a stronger flow from west to east in the upper air. If the vertical speed varies uniformly with height over the plateau, then regions of subsidence or ascending motion over the plateau must extend to greater heights than over the low-lying lands in its vicinity. Since the plateau is a cold source with respect to the surrounding free atmosphere in winter, this should give rise to an outward mass transport from the plateau and result in an intensification of the subsiding motions over the neighboring regions. Yeh [18] has computed the distribution of the vertical speed over the plateau and in its neighborhood and found that at both 3-km and 6-km levels in winter, ascending motions only occur over the southeastern part of the plateau with subsidence elswhere. Thus, the plateau acts to strengthen the surface cold ridge in this season and leads to an intensification of the winter monsoon. The mean surface map for January shows that the axis of the ridge is not closely associated with the coldest region along the same latitude over the mainland of China. The axis occurs at 110°E while the coldest region is situated near 120°E. The widespread subsidence in the neighborhood of the Tibetan Plateau may be one of the causes that give rise to the observed pattern. Subsiding motions tend to increase the stability of the atmosphere and thus usually result in drier weather with less precipitation in winter. However, conditions are somewhat unsettled in the southeastern part of the plateau where ascending currents prevail. Table 5 indicates that more cloudy days are recorded at Chengtu and Nanning in this region than other regions in China.

TABLE 5

Number of sunny and cloudy days and total cloud amount recorded at various stations in January 1953 - 1955

	Peking	Lanchow	Nanking	Hankow	Chengtu	Nanning
No. of sunny days (Cloud amount 0 - 2)	16.0	11.7	7.3	6.0	1.7	2.7
No. of cloud days (Cloud amount 8 - 10)	3.7	6.0	13.7	13.3	23.0	20.0
Total cloud amount	3.2	3.9	6.1	6.4	8.4	7.8

In summer, the mainland of East Asia is controlled by a heat low associated with the summer monsoon and ascending currents are found in most parts of the continent. Over the Pacific, pressure is generally high with widespread subsidence in the atmosphere. In the layer affected by the summer monsoon, the main transport of mass is directed from sea to land. At this time, the Tibetan Plateau is a heat source with an influx of air at the lower levels. The findings of Chu [9] indicate that the vertical currents are strong over the plateau with a mean value of 1.0 cm sec^{-1} in the central region, which is more than four times of that observed near Hankow (0.2 cm sec^{-1}). If the rate of decrease of vertical speed is uniform from 500 mb upward, then ascending motions will exist at greater heights over the plateau than in its neighboring regions. The presence of the plateau thus gives rise to a deeper layer of inflow, which necessarily causes the monsoon depression in its vicinity to deepen, and thereby enhances the effect of the summer monsoon on the weather and climate of the affected regions.

The above discussion only relates the effect of large-scale topography on the maintenance and intensification of the monsoons over East Asia. It may also be pointed out that during spring, the rapid warming of the Tibetan Plateau accelerates the process of disintegration of the westerly circulation over South Asia and indirectly brings about an early onset of the summer monsoon. In autumn, high temperatures are generally maintained over the plateau, thus delaying the establishment of the westerlies to its south and the retreat of the summer monsoon over Southeast Asia. The patterns of heating and cooling during mid-winter and mid-summer are favorable for the intensification of the upper-air planetary circulation, which acts to stabilize the winter and summer monsoon wind fields. These aspects are not discussed in detail in the present study.

Summing up, we may point out that many features of the monsoons over East Asia are produced by the mechanical and heating effects of the Tibetan Plateau.

The above analysis shows that the exceptionally large temperature difference produced by the distribution of land and sea over East Asia not only causes a breakdown of the planetary wind belts in the surface layer but also creates a monsoon pressure field in the lower troposphere. Hence, the differential heating over land and sea should be the main cause for the occurrence of the monsoons. Topography may exert a significant influence on the characteristics of the monsoons over the various regions, but it is hardly a basic governing factor for the formation of the monsoon phenomenon.

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SOME CHARACTERISTICS OF THE MONSOON CLIMATES IN CHINA

by

Kuo Chi-yunn and Kao Yu-hsie

In any discussion on the characteristics of the monsoon climates in East Asia or China, the notable seasonal variations of the prevailing winds are usually dealt with first. These variations are found to bring about different types of climate and weather. Of course, these are the basic features which may easily be detected and recognized by investigators. However, a deeper examination on the problem of the monsoon climates over East Asia (including China) reveals that there are additional features worthy of investigation. The monsoon circulation is found to contribute much to the evolution of the general circulation. Not only are its characteristics pronounced during its prevalent periods in January and July, but they also exert a definite influence on the annual distribution of the isobaric field, the seasonal change of the general circulation and the structure of the temperature and wind fields in the lower troposphere. These features are studied and discussed in the present paper as other aspects of the monsoon problem.

I. THE MONSOONAL INFLUENCE ON THE BREAKDOWN OF THE PLANETARY CIRCULATION IN THE LOWER TROPOSPHERE

To begin with, let us assume that the influence of land and sea may be basically eliminated by averaging along the latitude circle to obtain the meridional distribution of the pressure field. Thus this entity may be taken as the characteristic distribution of the planetary pressure field caused by the temperature gradient between high and low latitudes. Of course, this simplifying assumption is inadequate at any latitude with a large difference in land and sea areas such as the high and low latitudes in the northern hemisphere. However, it does not affect the basic problem of our discussion to satisfy the requisite degree of accuracy in the present study.

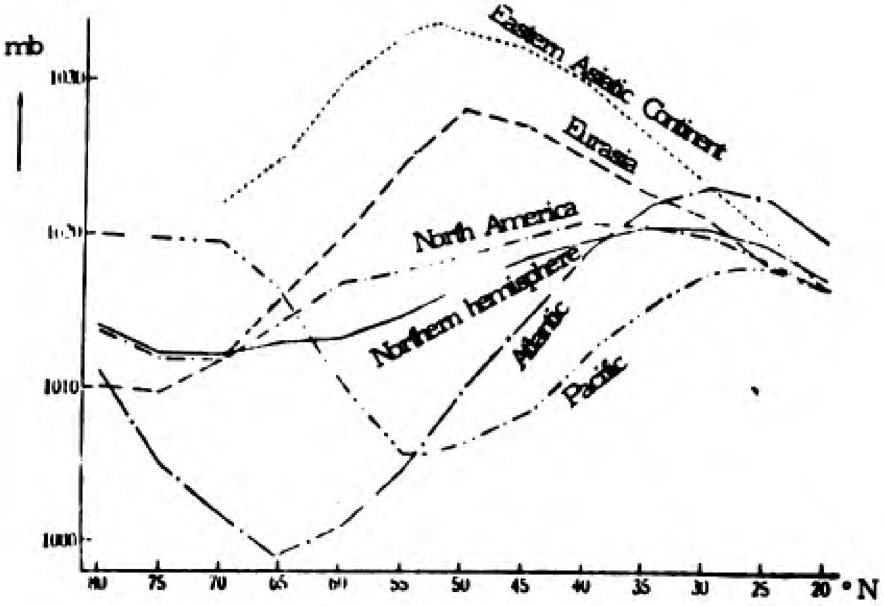


Figure 1.1

Meridional distribution of MSL pressure averaged along latitude circle in January.

Figures 1.1 and 1.2 were prepared under the above assumption. They depict the meridional distribution of mean atmospheric pressure over the Eurasian Continent, the eastern Asiatic Continent, North America, the Atlantic and the Pacific and the entire northern hemisphere.

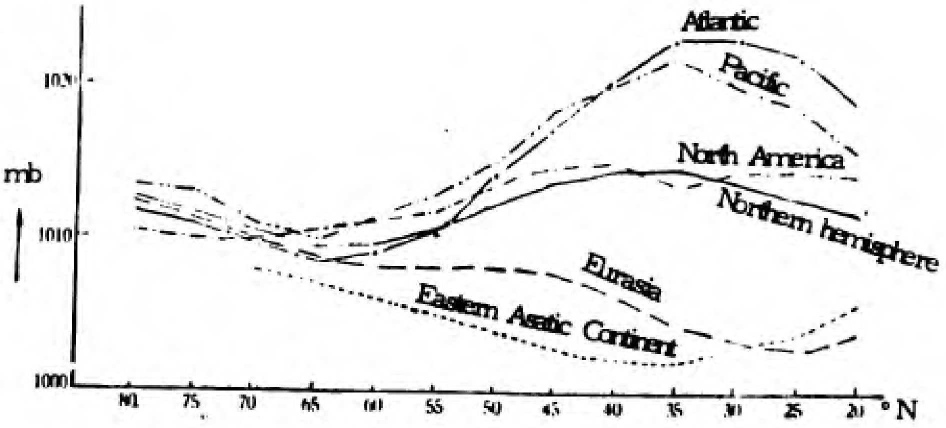


Figure 1. 2

Meridional distribution of MSL pressure averaged along latitude circle in July.

It may be noted from these figures that the profiles over Eurasia and the eastern Asiatic Continent are significantly different from those over America, the two oceans and the entire northern hemisphere either in January or in July. The curve depicting the meridional distribution of mean atmospheric pressure averaged along the latitude circle in the northern hemisphere in January is characterized by the existence of two high cells and two lows, namely: the subtropical high, the polar high, a high-latitude low and the equatorial trough. The major high pressure belt lies between 30°N and 35°N. The mean pressure configurations over America and the two oceans are basically similar to the profile of the entire northern hemisphere. However, the region of maximum pressure lies at higher latitude near 40°N, while the major high pressure belts over the two oceans are located in the neighborhood of 25°N. On the other hand, the special features over the Eurasian Continent and mainland China are characterized by the existence of one high cell and two lows. The center of high pressure lies between 50°N and 55°N with maximum values much higher than those over America and the entire northern hemisphere. A pressure difference of up to 15 mb is noted.

Significant differences are also observed in July. In broad general terms, the meridional distribution of mean atmospheric pressure in this month is rather similar to that of January. However, the position of the subtropical high is now located at higher latitudes between 35°N and 40°N. The maximum pressure is lower than that of January by about 6 mb. It is worthy of note that the pressure distribution over America is very close to that for the northern hemisphere in both magnitude and profile. In middle and low latitudes, the meridional distribution of mean pressure over China in July is completely out of phase with that over America and the northern hemisphere. It is also significantly different from the January situation. Over mainland China, a pronounced low pressure system prevails with the minimum pressure lying between 35°N and 40°N approximately, while pressure is high in the corresponding latitude belt in other regions over the northern hemisphere. If the meridional distribution of mean pressure over the northern hemisphere is taken as that caused by temperature difference between the Equator and the Pole, then the observed deviations from the profile for the northern hemisphere may be treated as the contributions essentially due to the distribution of land and sea. This pressure distribution over the Eurasian Continent and the eastern Asiatic Continent, which is characterized by the existence of a winter high and a summer low, may be ascribed principally to the influence of the distribution of land and sea. On the other hand, the latitudinal variation of pressure over North America is basically similar to that for the northern hemisphere in winter (January) and summer (July). In both January and July, only variations in position and intensity of the high and low pressure systems are noted. This may be mainly due to the fact that North America is a small continent and hence the influence of the land and sea distribution is weak. Thus we may say that the monsoon of East Asia is so intense as to give rise to a breakdown of the normal pressure distribution of the planetary wind belt in the surface layer over the eastern Asiatic Continent (or the Eurasian Continent). If we restrict our attention to the monthly mean pressure field over the mainland of China or the Eurasian Continent, then it may be noted that in winter there is no "west-wind" pressure gradient south of 50°N and no "east-wind" pressure gradient north of 50°N. The situation is exactly reversed in summer.

II. PHASE-LAG IN THE LOWER AND UPPER FLOW PATTERNS OF THE TROPOSPHERE DURING PERIODS OF SEASONAL TRANSITION AS A RESULT OF THE INTENSE MONSOONS

The seasonal change of the general circulation is usually manifested as the variations of the activities of action centers between summer and winter in the surface layer and the transformation of the planetary circulation pattern in the upper troposphere. However, previous investigators [1-3] mainly restricted their discussion to the evolution of the upper level pattern with respect to the advance of the season. Furthermore there were no systematic studies on the seasonal change of low-level patterns and the interrelationship between the flow configurations at high and low levels. These problems were only given a passing mention in most studies. For example, it was pointed out by Yeh [3] and others in their discussion on the sudden change of the general circulation in October that in the low-level flow over East Asia a sudden but significant change of cyclonic to anticyclonic flow was found to occur in September. However, no significant changes in the low-level flow pattern were found to accompany the abrupt establishment of the upper westerlies in October. This shows that the transition from summer to winter occurs earlier at low than high levels. Furthermore, Kao [4] has clearly shown that the "high autumn fine spell" (over China) mainly results from the time lag between the onset of the seasonal transition of circulation patterns at the low and upper levels from summer to winter. In the present study, such phenomena are examined in more detail with particular emphasis on the significance of monsoonal characteristics.

During the transition from summer to winter, the onset of strong winter monsoon gives rise to a breakdown of the surface pressure field resulting in a time lag between the changes of circulations at low and high levels. (High-level changes lag behind low-level ones.) Figure 2 shows the latitudinal position of the boundary between easterlies and westerlies on monthly mean surface pressure charts (solid line) and the axis of the subtropical ridge at 700 mb (broken line) for the months of July to October. Easterlies prevail north of the solid line and westerlies south of it. The distribution of winds with respect to the broken line is, however, just opposite; thus westerlies are prevalent north of the broken line and easterlies south of it. It may be noted from Figure 2 that there is relatively little difference in the latitudinal positions of these two types of boundary lines in July and August and that the axis of the upper ridge usually lies to the south of the boundary line separating the easterlies and westerlies on the surface chart. In September and October, the axis of the upper ridge is located to the north of the boundary between the surface easterlies and westerlies over East Asia; while their relative positions over other regions are essentially the same as in July and August. This phenomenon suggests that the circulation over East Asia possesses special characteristics which are not found elsewhere.

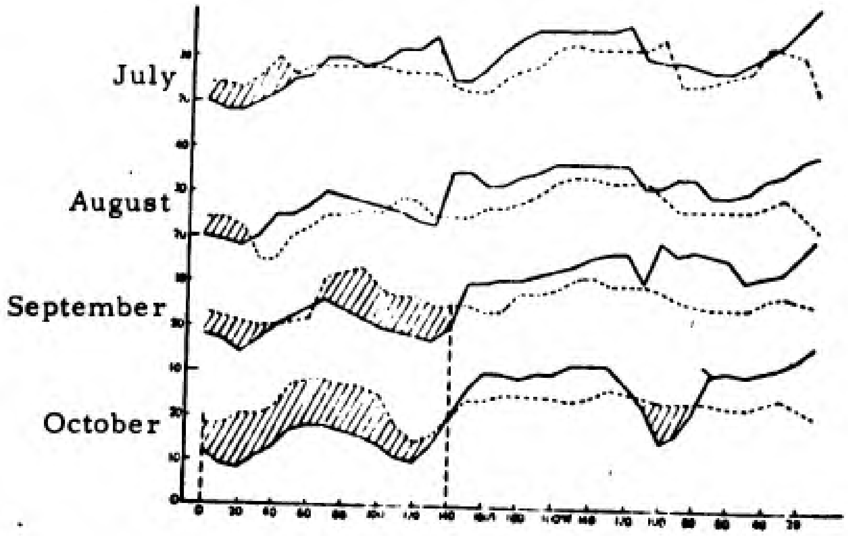


Figure 2

Latitudinal position of the boundary separating easterlies and westerlies (full lines) and axis of subtropical ridge at 700 mb (broken lines) versus longitude for the months of July to October.

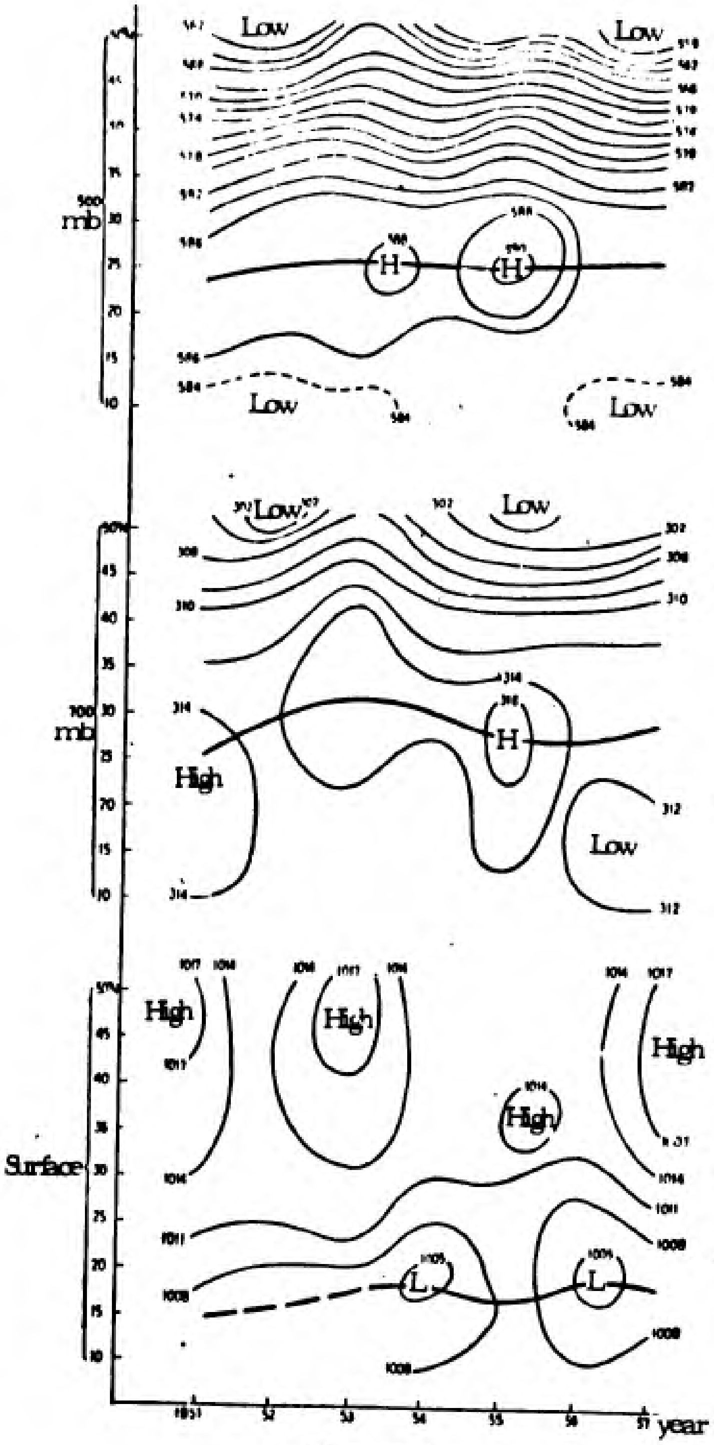


Figure 3

Meridional distribution of surface pressure, 700-mb and 500-mb contour heights averaged along latitude circle over the region from 105°E to 120°E for September 1951 - 1957.

(Legend: Thick full and broken lines indicate ridge and trough axes respectively.)

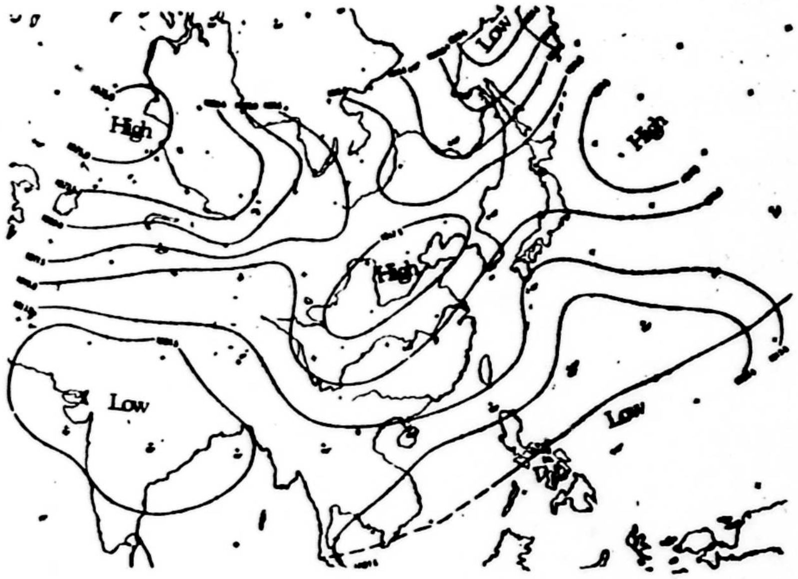


Figure 4.1

Mean surface pressure chart for the period 18 September - 2 October 1955.

In addition to Figure 2, charts depicting the meridional distribution of surface pressure, 700-mb and 500-mb contour heights averaged along the latitude circle over the region from 105°E to 120°E were prepared from September mean charts for the years 1951 - 1957 and are given as Figure 3. Ridge and trough axes are denoted by thick full and broken lines respectively in the figure. Obviously, the trough line of the surface pressure system always lies to the south of the ridge axes at upper levels in each September during the seven-year period from 1951 to 1957. This indicates that the low-level circulation pattern over a major part of the eastern Asiatic

Continent is under the control of the winter monsoon by September. Such a phenomenon has been pointed out in discussions on the seasonal change of the mean flow field in Reference [5]. It may also be noted from Figure 3 that during the seven-year period 1951 - 1957 the mean position of the ridge axis at 700 mb is located some 5° latitude north of that at 500 mb with an average slope of 1/250, which is much smaller than the normal value (on a global basis).

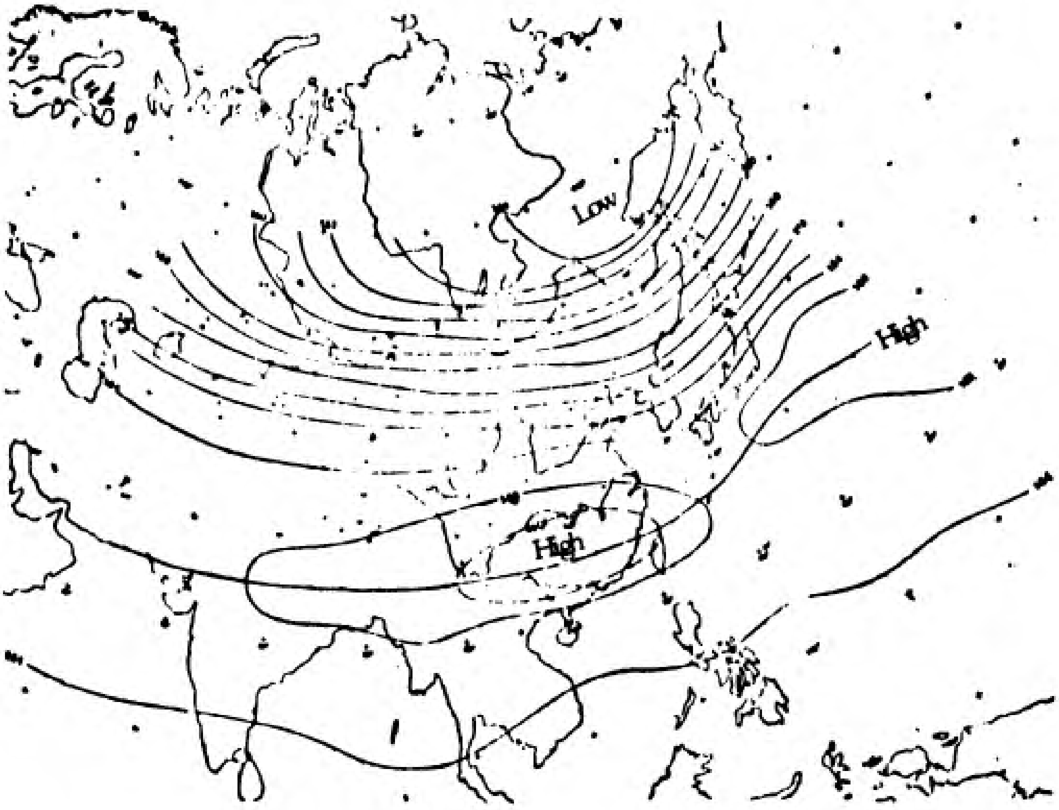


Figure 4.2

Mean chart of 500-mb topography for the period 18 September - 2 October 1955.

Figures 4.1 and 4.2 show the distribution of mean surface pressure and 500-mb contour height for the 15-day period from 18 September to 2 October 1955 respectively. It may be seen from these figures that the axis of the 500-mb ridge remains near 27°N with the steady southward advancement of the surface continental anticyclone. Thus, the trough line which lies equatorward of the continental anticyclone is displaced to low latitudes south of 15°N. In this way, anticyclonic flow prevails at both low and high levels in the lower troposphere south of 35°N, resulting in a deep layer of easterlies south of 27°N. This marks the maximum vertical excursion of the winter monsoon from the surface. An analysis of an additional example with 1957 data gives results similar to the above.

The weather associated with the aforementioned circulation configurations is characterized by a smaller number of rain-days or less precipitation over Central and South China than in eastern and western regions. In other words, the precipitation belt over China has changed its orientation from east-west to north-south. Figure 5 shows the distribution of the total amount of precipitation over various regions in China during the decade 21 - 30 September 1955, and reveals that precipitation is minimum in regions where superposition of the low-level continental anticyclone and the high-level subtropical ridge occurs. This aspect of the results is concordant with the findings of Reference [4].

It may also be seen from Figure 9 of Reference [6] (which consists of mean surface pressure and 500-mb charts for the period 8 - 17 September 1951 - 1957) that by 8 - 17 September the surface trough had already retreated equatorward to south of 20°N while the subtropical ridge at 500 mb still persisted near 25°N. Although the chosen situations in 1955 and 1957 are two discrete examples, they are found to represent the general trend in a more significant way.

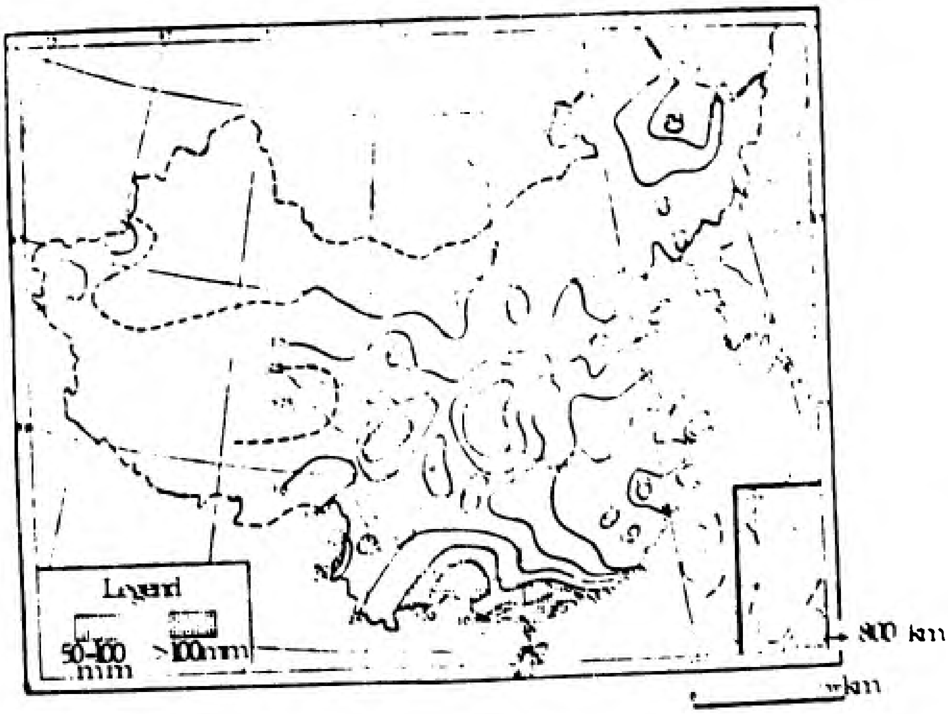


Figure 5

Distribution of the total amount of precipitation (mm) for the decade 21 - 30 September 1955.

During the transition from winter to summer, the rapid warming of the continent leads to the breakdown of the pressure field earlier at the surface than in upper levels. Thus the flow patterns are again not concordant at lower and upper levels. A very marked change in the zonal circulation generally occurs in early June over South China [3, 7, 8]. Before this sudden change, the characteristics of winter circulation in the basic flow persist at the upper levels [9], although the zonal speed may show two periods of weakening. What then will be the surface flow configurations at this time? It may be noted from the monthly mean charts for the various standard levels published by the U. S. Weather Bureau [10] that the major circulation system dominating the Eurasian Continent is an anticyclone centered south of Lake Baikal, which begins to weaken and move westward in March. During this month, the whole continental block is still under its influence. By April, further weakening and westward movement of the continental anticyclone is noted and cyclonic circulation appears over India and Indo-China. In May, the continental anticyclone becomes very weak, while a cyclonic circulation dominates Southeast Asia and the South China coast. The characteristics of the winter configuration however, still persist at 500 mb (see Figure 4 of Reference [6]), and the pattern does not break down until June. Thus during the transition from winter to summer, there is a certain time lag in the changes of circulation between the high and low levels. The change also occurs first at low levels but is more gradual in character. During March and April, continental thermal lows form over India and the Indo-China Peninsula with gradual intensification and development and they dominate a great part of Europe and Asia in June. Thus changes in the upper-air flow pattern lag behind those at the surface by two months over Southeast Asia and more than a month over South China.

Figures 6.1 and 6.2 are mean surface pressure and 500-mb contour charts for the period 21 - 30 May 1957. They may serve as an example to show the circulation changes at low and high levels during the transition from winter to summer. On the surface chart the continental low was centered over Northeast India with a trough of low pressure extending eastward across the Indo-China Peninsula to South China. At 500 mb (Figure 6.2), the subtropical high pressure belt was located near 15°N. A westerly flow prevailed to the north, and the only change during the period was a weakening in its intensity. In other words, the winter characteristics persisted in the circulation pattern, which did not break down until June. Thus the time lag between circulation changes at low and high levels during spring covers more than 45 days in the least.

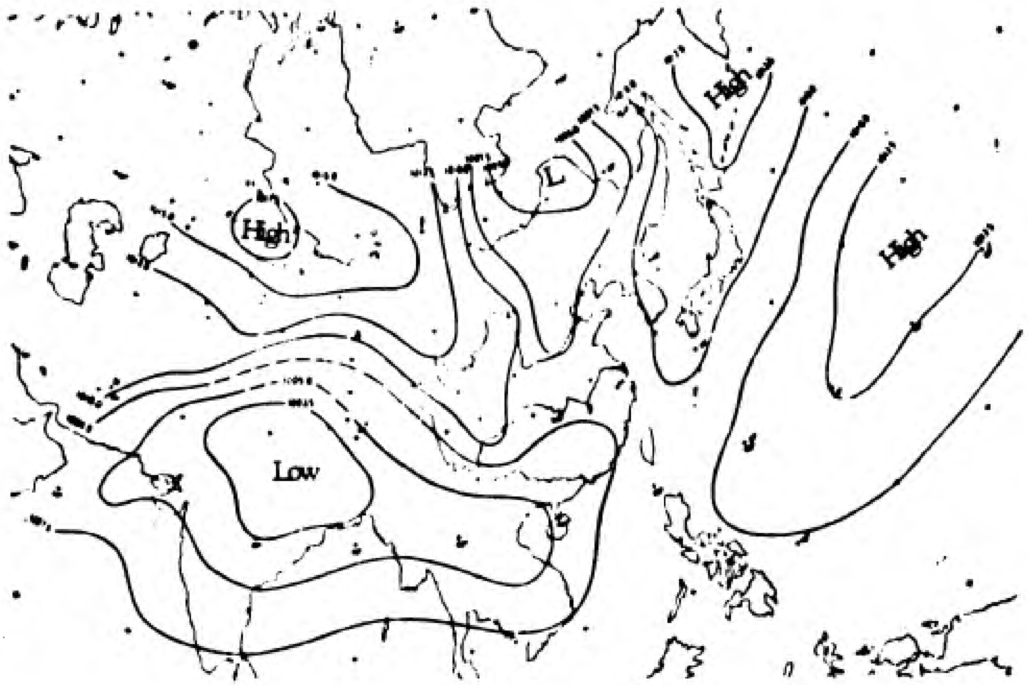


Figure 6.1

Mean surface pressure chart for 21 - 30 May 1957.

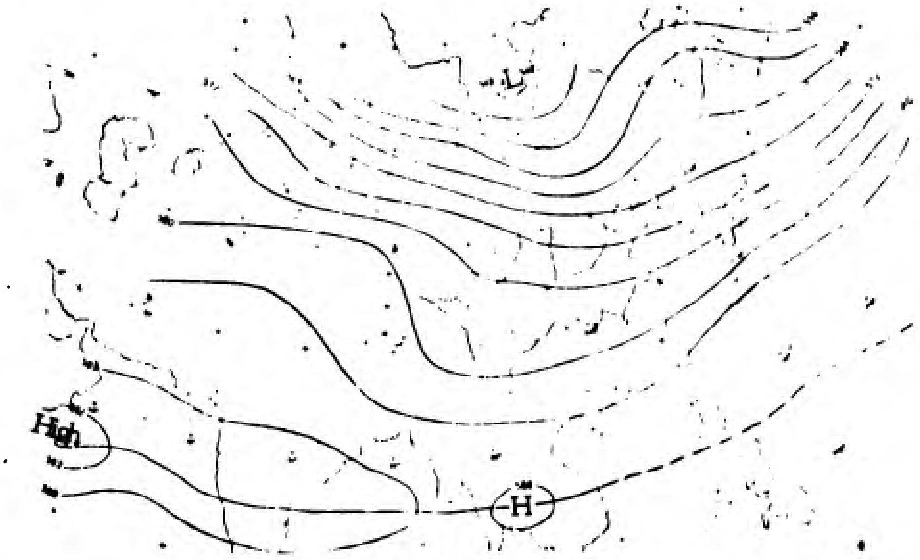


Figure 6.2

Mean 500-mb contour chart for 21 - 30 May 1957.

The above factual evidence shows that during the transition from winter to summer, changes in the circulation pattern first occur at the surface. Changes at low level begin during March and April and are gradual in nature, while those at high level occur later around June and are more abrupt in character.

III. THE BREAKDOWN OF NORMAL ANNUAL DISTRIBUTION OF THE PRESSURE FIELD IN THE LOWER TROPOSPHERE CAUSED BY THE INTENSE THERMAL LOW OF THE SUMMER MONSOON

The profile of the annual distribution of surface pressure over a continental region may be described in general terms as follows: In the surface layer, pressure is highest in winter and lowest in summer. The reversal of this distribution occurs at some level in the troposphere. Thus for all upper levels above this layer, pressure is lowest in winter and highest in summer. For example, the annual maximum pressure (at 850 mb) over North America occurs in July and August and the minimum in winter. However, the situation is different over some regions in Asia. The annual maximum (at 850 - 500 mb over these regions) does not occur in high summer but in September and October with a secondary maximum in April and May and a corresponding minimum in July and August. This phenomenon has been studied by previous investigators [4, 11, 12] and an explanation has been suggested by Lü [12]. However, limitations imposed by the sparseness of data, particularly upper-air observations, in the early days prevent a comprehensive review of its inherent characteristics. The present authors are of the opinion that the observed profile may be related to the pronounced monsoon circulation over East Asia. The regional distribution of this phenomenon, the governing factors involved and other related aspects are analyzed and discussed in the ensuing sections.

(a) Some Characteristics of the Pressure Distribution within the Troposphere over East Asia

Charts depicting the annual distribution of monthly mean contour heights at various isobaric surfaces over ten upper-air stations in China for the period 1953 - 1958 (Figure 7.1) and four in North America for the period 1946 - 1949 (Figure 7.2) and a cross section of the monthly mean 700-mb contour height in the latitude belt 25°N - 35°N for the different longitudes (Figure 8) were prepared. The following factual statements may be obtained from an analysis of these figures:

(i) In the middle and lower troposphere (mainly at 700 mb) the maximum of monthly mean contour heights occurs in September and October instead of July and August over certain regions in East Asia. A comparison between Figures 7.1 and 7.2 shows that the maximum values of monthly mean contour heights from 850 mb to 700 mb or higher levels occur in September and October over a majority of observing stations in China, while in North America the maxima are observed mainly in July over stations south of 30°N and in August over stations north of 30°N. Meanwhile, it may be seen from Figure 8 that the occurrence of maximum pressure in September and October is restricted to the region from 40°E to 140°E (see the thick full line in figure). Thus it may be said that the occurrence of maximum monthly mean pressure in the middle layer of the lower troposphere in autumn instead of summer is a regional feature of East Asia.

(ii) Over many regions in East Asia the profile of monthly mean contour heights at 850 - 500 mb is characterized by the existence of double maxima with the major one in autumn and the secondary in spring. Figures 7.1 and 7.2 show that over America the 850-mb profile exhibits only one maximum in the whole year, while over some stations in China, such as Chengtu, Nanning and Kunming, two or more maxima are observed, one in September or October and the other in April or May. This phenomenon can also be detected from Figure 8 (see the months with thick full lines) and is most pronounced over Southwest China. For example, the reflection of the double-maxima structure is always noticeable at 500 mb and may reach 300 mb on many occasions. It is less pronounced in Southeast and Northwest China. Thus the occurrence of double maxima is restricted to 700 mb over Foochow and Urumchi and this structure does not show up even at 850 mb over Northeast China. In Shenyang and Peking, the maxima of monthly mean contour heights occur in August at 700 mb and in July at 300 mb and there is only one maximum at each level throughout the whole year.

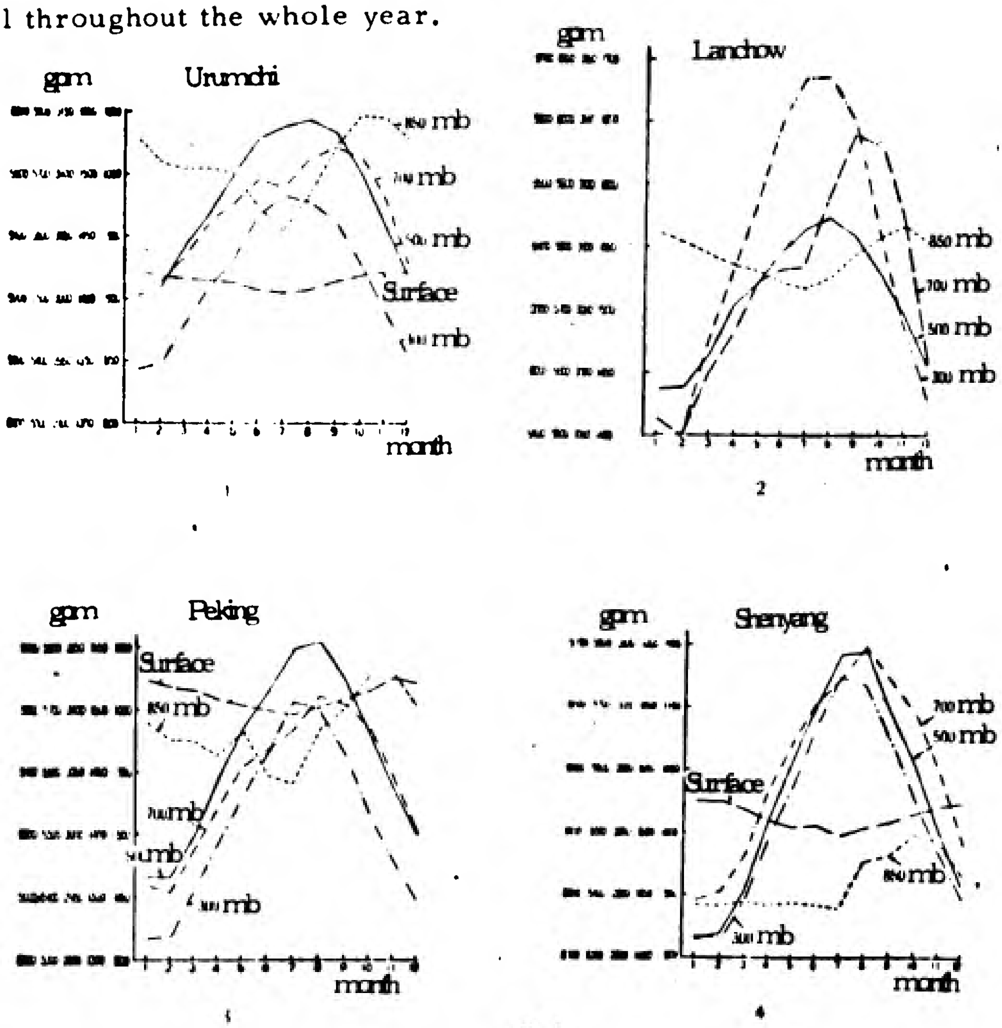


Figure 7.1

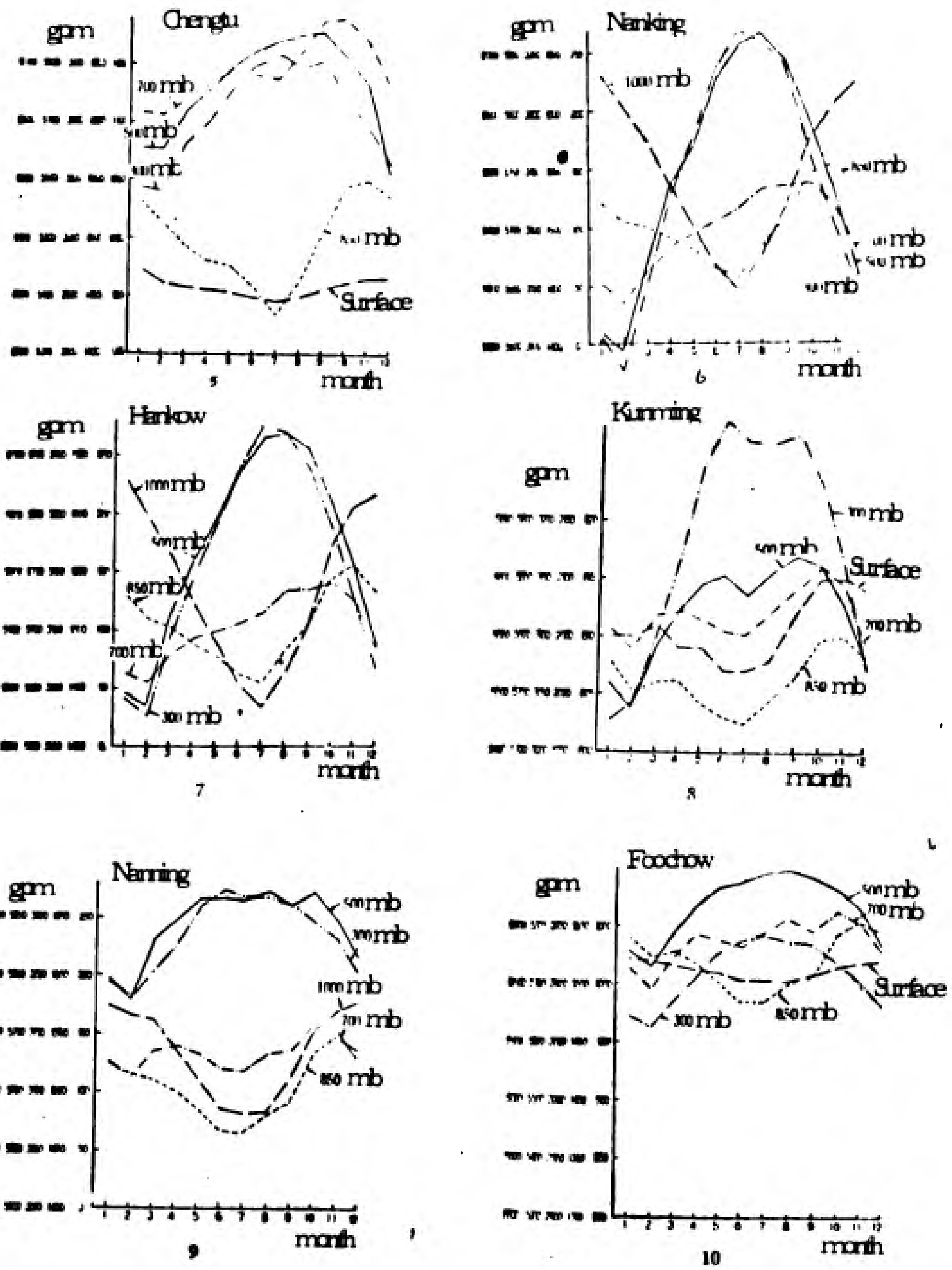
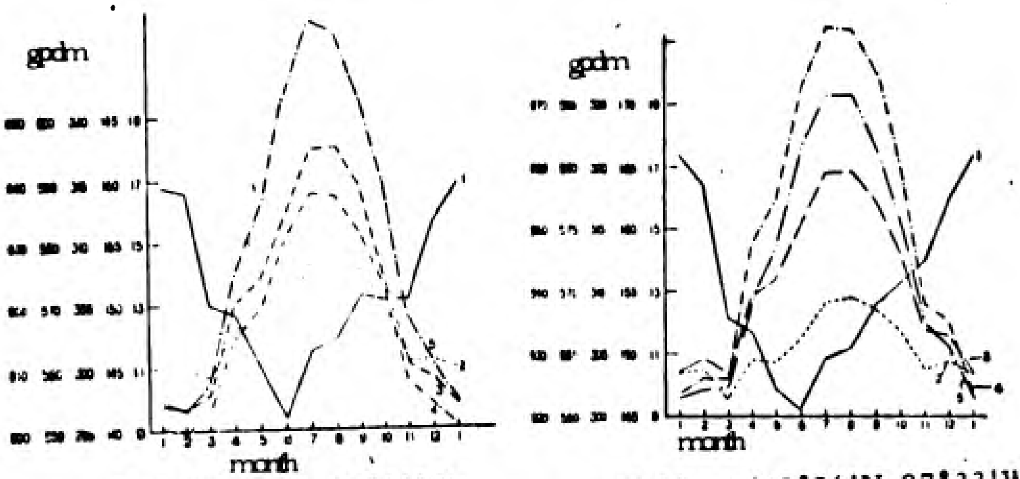


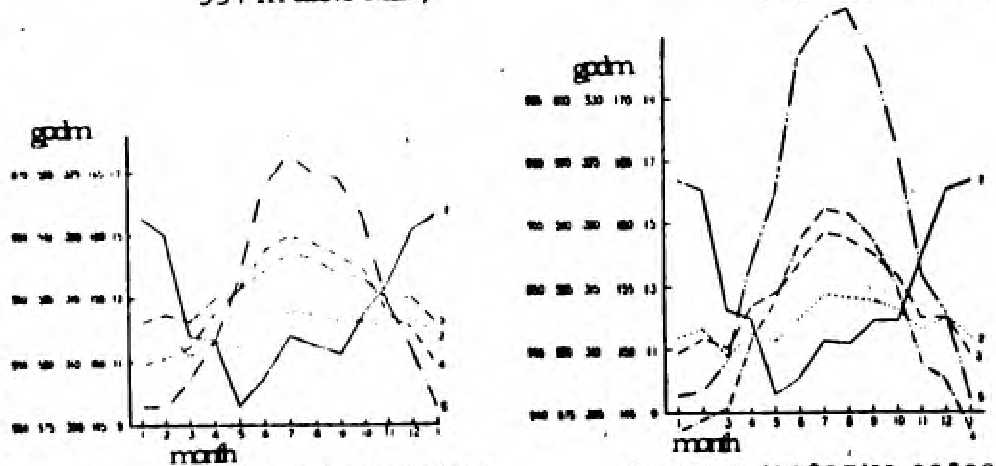
Figure 7.1 (Continued)

Annual distribution of monthly mean contour heights (1953-58) at standard pressure levels over various stations. [Ordinate: 1000-mb (surface), 850, 700, 500 and 300-mb contour heights in geopotential meters respectively (reading from right to left column by column.)]



Omaha (41°10'N, 95°56'W; 337 m above MSL)

Okalahoma (35°26'N, 97°33'W; 370 m above MSL)



Brownsville (26°00'N, 97°26'W; 17 m above MSL)

San Antonio (29°27'N, 98°28'W; 211 m above MSL)

Figure 7.2

Annual distribution of monthly mean contour heights at standard pressure levels over various stations.

[Ordinate: 1000-mb (surface), 850, 700, 500 and 300-mb contour heights in geopotential meters respectively (reading from right to left column by column.)]

1 - 1000 mb; 2 - 850 mb; 3 - 700 mb; 4 - 500 mb; 5 - 300 mb.

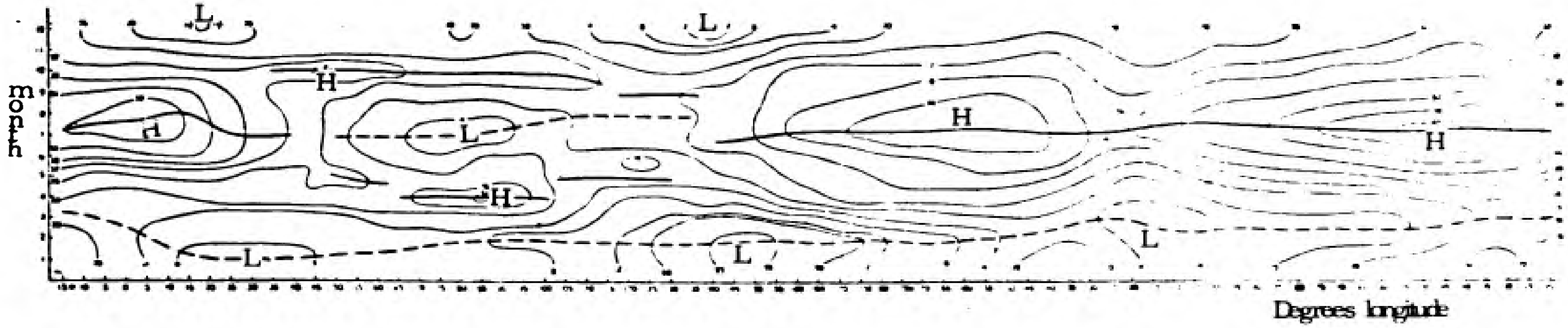


Figure 8

The distribution of monthly mean 700-mb contour height within the latitude belt 25°N - 35°N over a period of many years versus longitude.

(Thick full and broken lines depict axes of maximum and minimum values respectively.)

(b) The Characteristics of the Spatial Distribution and the Time of Occurrence of the Two Maxima and Two Minima

(i) Figure 9 shows the horizontal distribution of the occurrence of the two maxima and two minima at different isobaric surfaces over East Asia. It may be seen that the region of simultaneous occurrence of two maxima and two minima at both 850 and 700 mb is roughly bounded by 40°E - 140°E and 15°N - 40°N. On the other hand, the region for simultaneous occurrence at 850, 700 and 500 mb is much smaller and is roughly bounded by 45°E - 100°E and 25°N - 32°N. The two maxima usually occur in September or October and in April or May, while the two minima mainly prevail in high summer and deep winter.

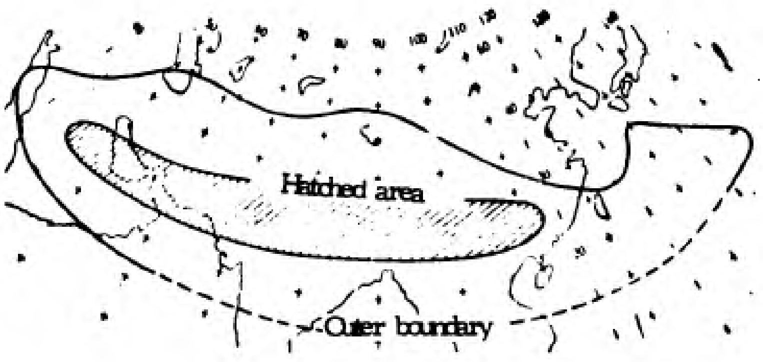


Figure 9

The horizontal distribution of the occurrence of the two maxima and two minima at different isobaric surfaces throughout the year.

(Outer boundary encloses region of simultaneous occurrence of two maxima and two minima at 850 and 700 mb. Hatched area depicts region for simultaneous occurrence at 850, 700 and 500 mb.)

(ii) The regional distribution of the time of occurrence of maximum monthly mean pressure at various levels in autumn: It may be seen from Figure 10 that in autumn the occurrence of maximum monthly mean pressure in the middle layer of the lower troposphere may be separated on a monthly basis by a line from Hainan Island to the Caspian Sea. Regions lying northeast of this line including a great part of China are characterized by the occurrence of the maximum in September, while those lying southwest of this line including India are characterized by the occurrence of the maximum in October. In examining the distributions at different levels along the vertical, the 30°N parallel may also serve as a rough boundary line. In this connection, maximum monthly mean pressure occurs only at 700 mb in regions north of 30°N, whereas it may occur at 700 and 500 mb in regions south of this parallel.

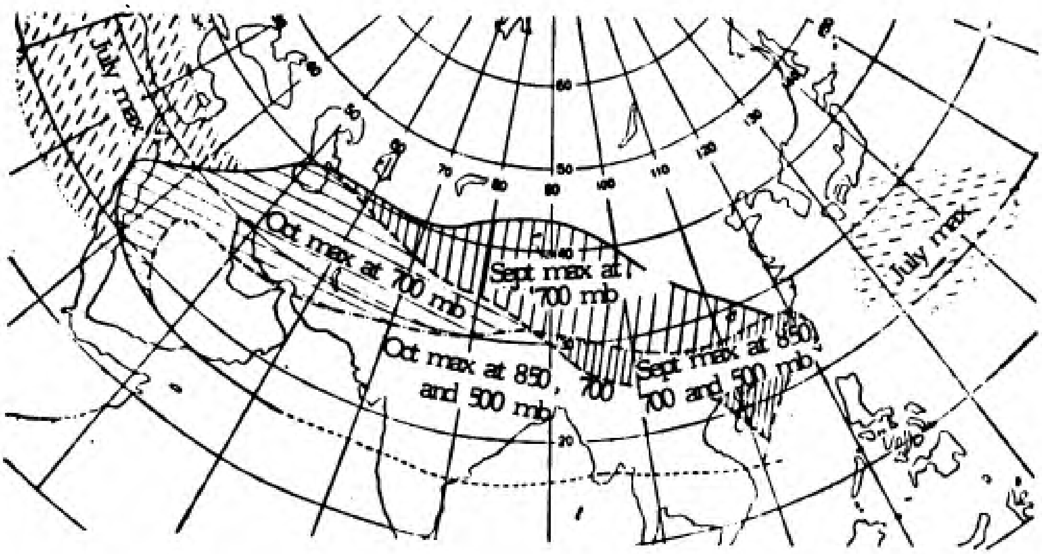
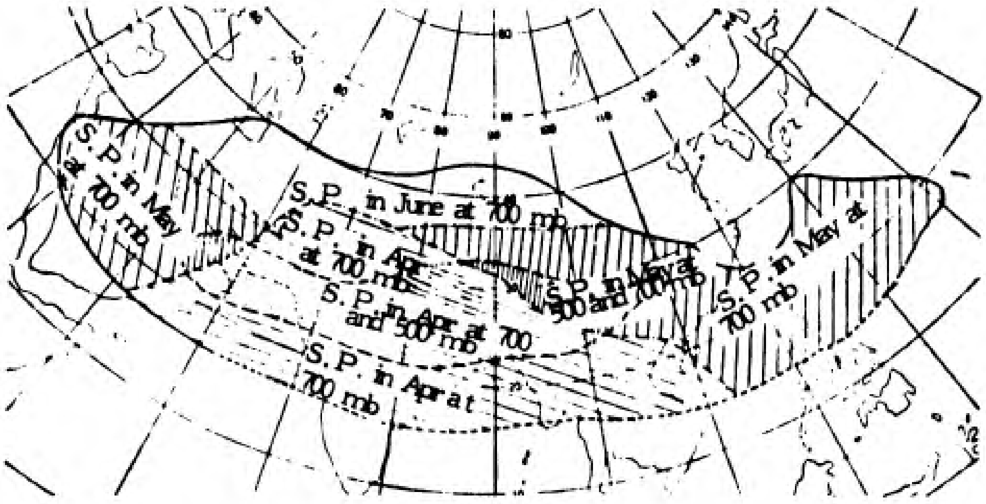


Figure 10

The regional distribution of the month of occurrence of maximum contour height at various isobaric levels.

(iii) Regional distribution of the month of occurrence of the secondary pressure peak at various levels in spring: In a similar manner as in Figure 10, a chart depicting the regional distribution of the month of occurrence of the secondary peak of the mean pressure profile in spring was prepared (Figure 11). It may be noted that the secondary peak occurs in April over India and Indo-China, in May over contiguous regions to the east and the west and in June over areas well to the north. The level at which the secondary peak occurs varies to some extent over different regions. During spring, it may occur at 700 and 500 mb in the region bounded approximately by 55°E - 110°E and 20°N - 30°N, and 700 mb alone outside this area. However, the secondary peak may also occur at 850 mb in spring in many places on the continent outside the above delineated region.



S. P. Secondary Peak

Figure 11

The regional distribution of the secondary pressure peak in spring at various isobaric levels.

(iv) Characteristics of the annual distribution of pressure at different altitudes over various stations: An analysis of the variation of the maximum monthly mean pressure with altitude from Figures 7.1 and 7.2 shows that over North America there is little difference in the time of occurrence of the maximum monthly mean pressure at the various levels except 1000 mb. The maximum generally occurs in July or August at the various upper levels, but in January at 1000 mb. However, a relatively high value also occurs in September over stations north of 40°N (such as Omaha). A much more complicated configuration is observed in China than in North America. Over each station, the month of occurrence of the maximum varies with height and the onset becomes earlier and earlier as we go upward. The general tendency is characterized by the occurrence of maximum in July at 300 mb, in August at 500 mb, and in December or January at 1000 mb. However, over stations like Kunming and Nanning the maximum occurs in September or October at 500 mb.

(c) Factors Governing the Exceptional Pressure Distribution over East Asia

We are of the opinion that the distribution of action centers (two highs and two lows) over the eastern Asiatic Continent during the year is essentially caused by the existence of exceptionally low pressure in July at all upper-air levels in the lower troposphere. Consequently, we must find out in the first place why the pressure is low in July at all upper levels in the lower troposphere over East Asia. Why is the pressure persistently low in July in some regions up to 10 km? To tackle this problem would require a study of the patterns at the various standard levels at different geographical locations. Now let us first consider the regions outside the full line in Figure 9. In general, two maxima and two minima are noted at 850 mb, with only a maximum above this level in July. We consider that this type of situation is essentially brought about by the heating effect of the underlying surface. As the air temperature rises after spring, the contour height from the lower level to the upper level falls gradually. This influence reaches 850 mb by July. Consequently, the resultant pressure distribution at this level is characterized by the existence of two maxima and two minima.

Secondly, we will take the hatched area of Figure 9, where two maxima and two minima can be detected at 500 mb or even at 300 mb. It is considered that the heating effect alone cannot adequately account for the observed configuration, although the phenomenon appears to be related to this effect. As far as geographical distribution is concerned, the configuration of this belt is similar to the equatorial trough and the Indian low. This implies that the particularly low pressure over this region in July may to some extent be related to the northward displacement of the equatorial trough to the region. Thus the observed low pressure in July may in some way be caused by factors of dynamical significance. However, complete domination of dynamical forces would result in a decrease of pressure with height. Since maximum pressure is observed in July at 300 mb or above, it is thought that the resultant pressure variation must be related to the heating effect. The air temperature is highest in July over this region, and the Tibetan Plateau further acts as a heat source in summer to warm up the air 1 - 2 km above it. Thus the fall in the contour height may probably be induced by the increase in air temperature.

Finally, we may turn our attention to the transitional zone between the hatched area and the region bounded by the full line in Figure 9. In July pressure in this zone is low at 700 mb with opposite tendency above it. In other words, maximum pressure occurs in July. This may be due to the fact that the zone under consideration forms a transitional belt between the aforesaid regions. In some provinces such as Chinghai, Kansu and Sinkiang, high temperature in July may result in a fall of 700-mb contour. In other provinces such as Fukien and Kiangsi, the transitional characteristics of the annual pressure variation is caused by the frequent but short visits of the equatorial trough together with the additional effect of surface heating.

Let us now discuss the reason why the secondary peak of the pressure profile over the various regions occurs in different months of spring. It is thought that this phenomenon is related to the two stages of development of the summer monsoon low, namely: the thermal stage and the thermal-dynamical stage. If pressure is high in April, May or June in the annual distribution, then it begins to fall in May, June or July respectively. For isobaric surfaces, this corresponds to a fall in contour height. Over South India a thermal low appears in March and becomes a closed system in April with further development and northward displacement in May. The effect of surface heating is relatively small in May and the layer of zero height change is low. Thus pressure tends to rise at all upper levels above this layer. From May onward, the continuing development of the thermal low over India raises the layer of zero height change to a level above 700 mb. In this way, the 700-mb contour height begins to fall. Thus the secondary peak occurs in April at 700 mb over regions south of 30°N in Southeast Asia. Over the region from North India to the Tibetan Plateau, the heating effect is particularly significant because of the proximity of the thermal low on the one hand and the rapid warming of the plateau in spring [13] on the other. As a result, the 500-mb contour height begins to fall in May. It should perhaps be borne in mind that the height of the plateau above mean sea level is in fact of the same order of magnitude as the 500-mb contour height so that the secondary peak may also occur in April at 500 mb over this region. The fall of contour height in June and July, particularly in June, may be closely related to the northward displacement of the equatorial trough. In this case, it is reasonable to expect a fall in contour height under the influence of the equatorial trough. Of course, in the first stage, heating effect also plays a part in bringing about the fall of contour height.

The third problem which confronts us is why the pressure is highest in September and October in the lower troposphere. This phenomenon may be accounted for by the time lag in circulation changes between the lower and the upper levels [4]. The equatorial trough and the subtropical ridge move southward after high summer. However, in September the subtropical ridge over East China lingers near 25°N. Thus the unusual pressure rise observed in this region after high summer may probably be due to the halt of the subtropical ridge on its return trip equatorward. Why then is the pressure higher in September or October than in April and May? This may probably be due to the higher temperatures in September and October than in April and May in the upper troposphere, although basically it is still related to the subtropical ridge. The reason is that westerlies prevail over the surface layer in April and May with the subtropical ridge lying well south of 20°N while in September and October the system lies just over the Asiatic Continent. In North America pressure and air temperature are higher in autumn than in spring, but there is no pressure peak in autumn because of the absence of low pressure values in July.

It may be seen from Figure 10 that the occurrence of the autumn pressure peak is related to the prevailing westerlies over East Asia, the establishment of the jet stream and the topographical influence of the Tibetan Plateau. The westerly circulation is established rather early to the east of the Tibetan Plateau. The subtropical anticyclone retreats from the mainland of China soon after the beginning of October with the establishment of the southern branch of the jet stream complex. As a result pressure is highest in September. However, the blocking action and the heating effect of the plateau may cause the subtropical ridge to persist at relatively high latitudes over the south of the plateau in October. Since the summer monsoon does not retreat completely from India before November which marks the recession of the subtropical ridge from the mainland of China south of the plateau, the pressure peak therefore occurs in October over the Tibetan Plateau and regions to its south.

We now come to the last problem about the earlier occurrence of the pressure peak at the lower levels. We know that in winter the continental anticyclone is most intense in December and January. Thus we are easily led to expect that the vertical extent of its influence is also highest in these two months. However, observational evidence does not lend support to this argument. It may be seen from Figure 7.1 that the anticyclone activities only reach 1000 mb in December and January. The pressure peak occurs in October or November at 850 mb and earlier at higher levels. This gives rise to another problem. When the pressure field of the winter monsoon is at its maximum intensity, why does the vertical extent of its influence not at its maximum? Our findings indicate that the winter monsoon attains its maximum intensity during the period when the upper westerlies are strongest. It may be seen from 500-mb mean contour charts (not shown) that troughs in the upper westerlies are deepest in December and January. Thus even though the pressure field of the winter monsoon is at its maximum in December and January, its effect is masked by the stronger influence of the westerlies aloft. Consequently high pressure is confined to the surface layer. It is well known that the surface pressure field of the winter monsoon begins to form in early September, while the southern branch of the upper westerlies does not become established after the second decade of October. Thus during October and November, although the surface pressure field has not yet attained its maximum intensity and may in fact be weaker than that in mid-winter, its circulation may become deeper in vertical extent because of the newly established upper westerlies. Consequently, pressure is higher in October and November than in deep winter in the lower troposphere. On the other hand, the subtropical ridge is situated at its northermost position in July and August and the heating effect of the continent in summer becomes more and more significant with increasing altitude so that high pressure occurs in summer at very high levels. It may be said that although the monsoon is a low-level phenomenon, its vertical extent is dependent on the intensity of the planetary wind belt at upper levels in addition to its own pressure field. If the monsoon flow is taken to be most pronounced when the vertical extent of its circulation is deepest, then the observed pressure field in October or November instead of January should be used for discussion.

Summing up, we note that the monsoon flow is particularly pronounced over the Eurasian Continent. The location of the equatorial trough at its northermost position makes the annual change of the pressure field more complicated over East Asia than elsewhere. Firstly, the heating effect is sufficiently pronounced to maintain a deep and large thermal low over the continent in July and the vertical extent of this circulation reaches a great height. As a result, the annual distribution of pressure over regions south of 45°N in the eastern Asiatic Continent is usually characterized by the existence of two maxima and two minima at 850 mb. At higher levels where the influence of the thermal low could not reach, the annual pressure distribution is found to be normal. Secondly, the northward displacement of the equatorial trough to the latitude regime of 25°N - 30°N over the eastern Asiatic Continent enhances the vertical extent of the abnormal characteristics in the annual pressure distribution. Thirdly, the intense heating effect of the Tibetan Plateau tends to cause the thermal low in its vicinity to become deeper in its vertical extent. Consequently, the annual pressure distribution at 700 mb is found to be abnormal near the plateau. Finally, differences in topography, latitudes, and the intensity of pre-summer pressure systems would induce variations in the occurrence of abnormal pressure distribution with respect to time and space. Of course, the effect of these abnormal characteristics on weather and climate is a more important problem. As far as China is concerned, the appearance of the secondary peak in the pressure profile in April and May at upper levels in low latitudes or the development of the surface heat low is favorable for an earlier onset of the summer monsoon or rainy season over South China and the northward retreat of the polar front. On the other hand, the appearance of a thermal low over North China suppresses the development of vortices and prevents the southward advance of the polar front, leading to an intensification of the spring drought. The occurrence of upper-level pressure maximum in September and October may strengthen the fine weather spell of high autumn over Central and South China, resulting in still less rainfall in the relatively dry autumn.

IV. MONSOONAL INFLUENCE ON THE FLOW PATTERN OVER CHINA IN THE LOWER TROPOSPHERE

It may be seen from the daily weather maps in each winter month that the flow pattern and the pressure field are usually more complicated in the lower troposphere (850 and 700 mb) than at higher levels. We are of the opinion that this is due to the intense monsoon which transforms the usual complicated pattern at the very low levels and the simple ones at high levels into a configuration characterized by simplicity at low and high levels but complicity in the intermediate layer. The height of this transitional layer varies to some extent with respect to the different combinations of intensities in the monsoon flow and the upper westerlies and is at 700 mb in general. However, during the final stage of development of the monsoonal anticyclone when the high pressure belt breaks up into small cells with further weakening, the transitional characteristics could then be detected at 850 mb.

The structure of the flow pattern on 4 December 1958 at the various levels may now be used to exemplify the monsoonal influence on the flow field. Figures 12.1 - 12.4 depict the surface pressure, 850-, 700- and 500-mb flow fields respectively for 0000 hour GMT 4 December 1958. This is a good example of the complete life-history of the development of a monsoonal anticyclone. The anticyclone is centered over the Yellow River basin with the polar front lying over the adjacent seas in Southeast China. The winter monsoon exerts its influence as far south as the coastal regions of South China. Because of the high intensity of the monsoonal anticyclone, its pronounced circulation can be detected at 850 mb (Figure 12.2). The northwest monsoon dominates north of the Hwai Ho basin with prevailing northeast monsoon to the south. This feature is concordant with the pattern shown by monthly mean flow charts [5]. The configuration of the flow field, however, undergoes significant changes as we go upward. Not only is the direction of the basic current at high levels different from that at 850 mb, but significant differences in the flow distribution are also found. The pattern at 700 mb is much more complicated than that at the surface and 850 mb. A convergence belt appears over the coastal regions of South China with another one lying just west of the meandering zone of the Yellow River basin. A small anticyclonic circulation is centered to the northwest of the Yunnan-Kweichow Plateau, with a trough extending from Southwest Kiangsu across Anhwei and the basins of the two lakes (Poyang Hu and Tungting Hu) to Kwangsi. In addition, in regions south of the Yangtze, winds are lighter at 700 mb than 850 mb. In general, westerly or northwesterly currents prevail over the majority of regions in China, except some parts in the south where southwesterlies are observed. The distribution of pressure systems is not concordant at the various levels (charts omitted). The flow below 850 mb is controlled by a single high cell while a westerly circulation prevails at 500 mb with a main trough over the east coast of China. The circulation at 700 mb is different from that of the lower and upper levels, and is characterized by a trough of low pressure over Southeast China, a small anticyclone over the Yunnan-Kweichow Plateau, a weak high cell over the meandering zone of the Yellow River basin and a trough to the west of it.

It may be said that the distribution described above is rather the rule than the exception. An analysis of the synoptic situations of the winter months in 1957 and 1958 (six months) shows that when a cold high appears at the surface this distribution occurs simultaneously on 73% of the occasions. In other words, it is climatologically significant.

It is generally thought that under the influence of the underlying surface the tropospheric flow pattern is relatively complicated in the lowest layer but becomes simpler with increasing altitude. However, according to the foregoing analysis, the observed circulation over China does not lend support to this conjecture. Under the influence of the winter monsoon, complicated flow pattern in the troposphere does not occur in the lowest layer, but in a transitional layer between the monsoon circulation and the planetary wind belt aloft. When the monsoon is prevailing, the flow pattern exhibits simplicity at low and high levels and complicity in the intermediate layer.

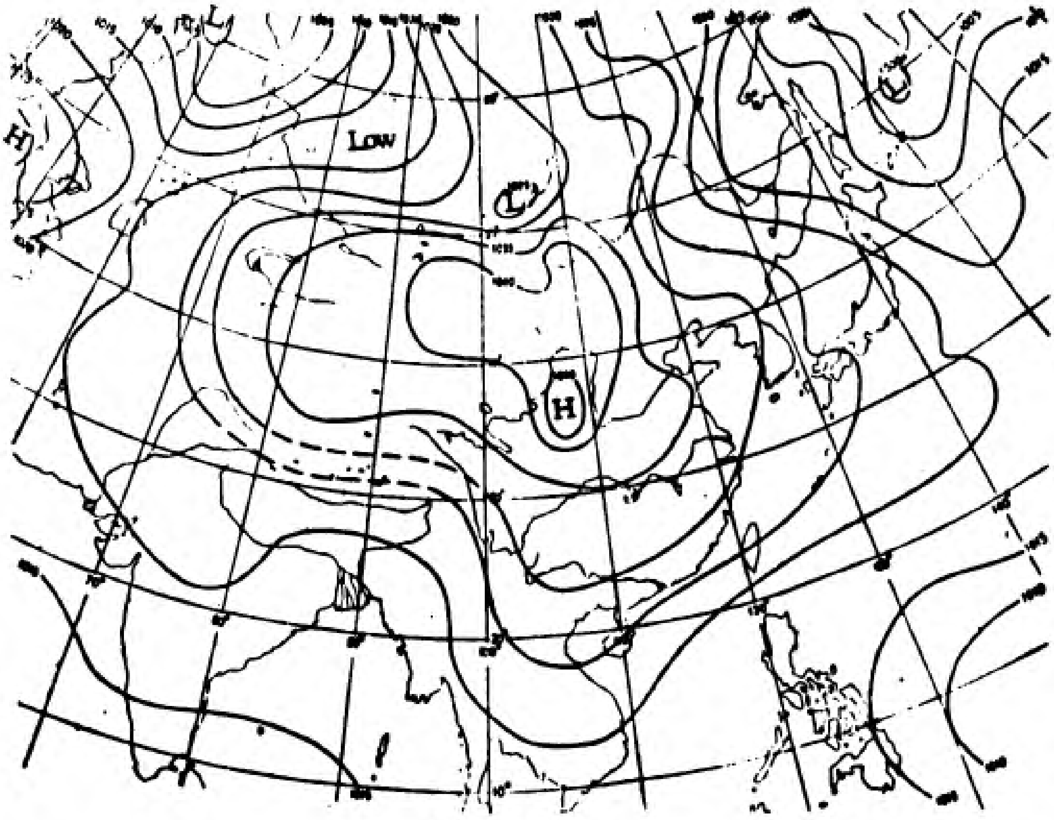


Figure 12.1

Surface pressure distribution at 0000 hour GMT, 4 December 1958.

Figure 12. 2

850-mb streamlines at 0000 hour GMT, 4 December 1958.

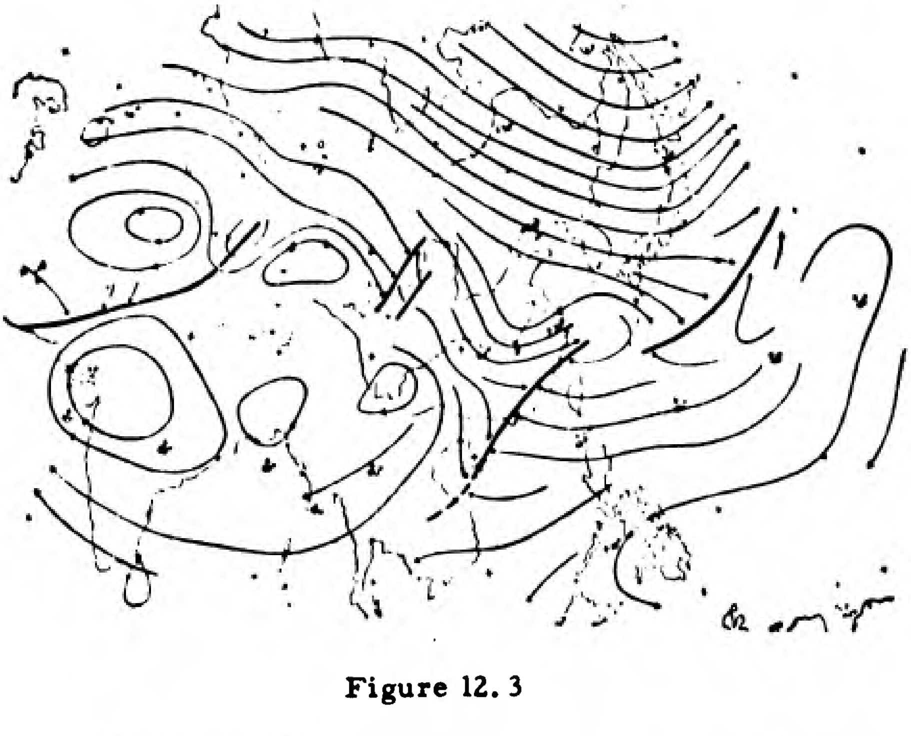


Figure 12. 3

700-mb streamlines at 0000 hour GMT, 4 December 1958.

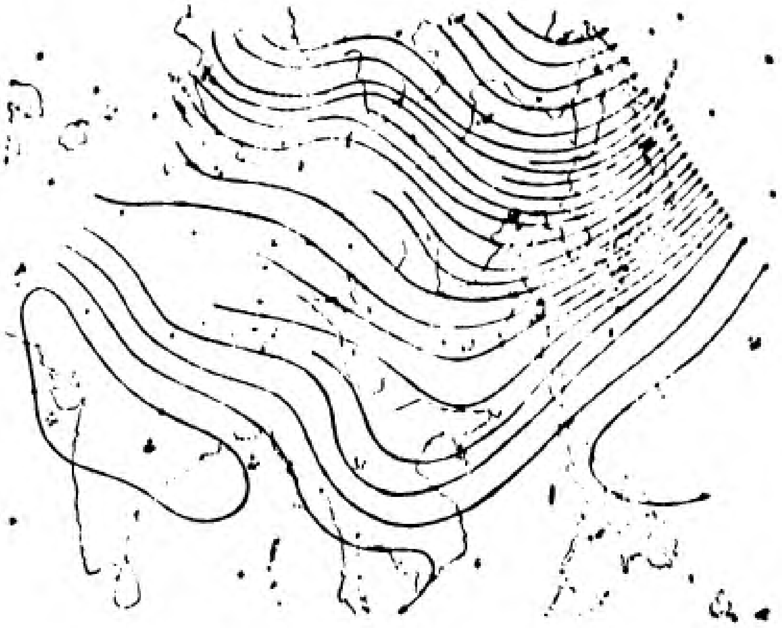


Figure 12.4

500-mb streamlines at 0000 hr GMT 4 December 1958.

V. MONSOONAL INFLUENCE ON THE TROPOSPHERIC TEMPERATURE FIELD OVER CHINA

The pronounced monsoons over East Asia tend to strengthen the meridional temperature advection in general. In winter the intense monsoon flow reaches very low latitudes. Thus the meridional temperature gradient is greater at low than high latitudes over China in winter. As southerly winds prevail during the summer monsoon, the meridional temperature gradient is greater at high than low latitudes. In other words, the configuration in summer is opposite to that in winter. This phenomenon can be seen clearly from Figures 13.1 and 13.2, which depict the meridional distribution of mean 1000 - 700 mb thickness at spatial intervals of 5° latitude for January and July respectively [10]. This may be taken as equivalent to the variation of mean temperature gradient (degrees/5° latitude) between 1000 and 700 mb. A comparison of the three curves in each figure shows that in January the maximum mean temperature gradient lies at 40°N - 45°N for the entire northern hemisphere, 50°N - 55°N over North America and 25°N - 30°N over the eastern Asiatic Continent, which is 15° and 25° latitude south of the corresponding positions for the entire northern hemisphere and North America respectively. The latter value in the difference of latitude seems to be very large. Meanwhile, it may be seen that on a global basis, the maximum temperature gradient occurs over North America in the latitude regime north of 30°N - 35°N and over the eastern Asiatic Continent south of 30°N - 35°N. On the other hand, in July the maximum temperature gradient over North America, the eastern Asiatic Continent and the entire northern hemisphere all occurs at 45°N - 50°N. However, the characteristics of their distributions are significantly different. In regions south of 45°N - 50°N the temperature gradient is smaller over the eastern Asiatic Continent than North America and the mean value for the entire northern hemisphere. However, in regions north of 45°N - 50°N, the temperature gradient over East Asia becomes greatest and is greater than that over North America and the mean value for the entire northern hemisphere.

In order to assess the stability of the above phenomenon, the meridional distribution of differences in thickness values (broken lines in Figure 13) over the mainland of China (105°E - 120°E) was obtained for January and July from 1000 - 500 mb thickness data for the years 1954 - 1957. In a similar manner, this additional information shows that in winter the meridional temperature gradient becomes greater as we go from high to low latitudes with the maximum value at 25°N - 30°N. The situation is reversed in summer, during which the maximum temperature gradient occurs at the northern boundary of the summer monsoon. For a better understanding of the monsoonal influence on the distribution of the temperature field, let us review the variation of temperature gradient with height. In this connection a comparison of Tables 1 and 2 shows that over a great part of China the temperature gradient is small at low level and large at the intermediate high levels, while over North America it is large in the lowest layer and decreases upward. This feature is found to be most outstanding at 850 mb in a comparison between East Asia and North America, and shows that the monsoon is more pronounced over East Asia, resulting in differences in the profiles of the temperature field over these two continents.

Table 3 shows the annual range of mean temperature at the various standard levels for the years 1953 - 1957 over Foochow, Nanking, Peking and other stations in China versus stations in North America near 25°N, 30°N and 40°N along 80°W. Since the monsoons are more intense over East Asia, the annual range of temperature is much greater over East Asia than North America at all levels at the same latitude. Thus we may say that the intense monsoon over East Asia strengthens the meridional advection of temperature in this region. As the winter monsoon penetrates down to very low latitudes, the maximum horizontal temperature gradient over China should be larger in the south than the north in winter. The configuration is opposite in summer, during which the maximum horizontal temperature gradient occurs at the northern edge of the summer monsoon because the flow is from low to high latitudes. Furthermore, since significant changes occur between the summer and winter monsoons, the annual temperature range is also largest in regions under the influence of the summer and winter monsoons.

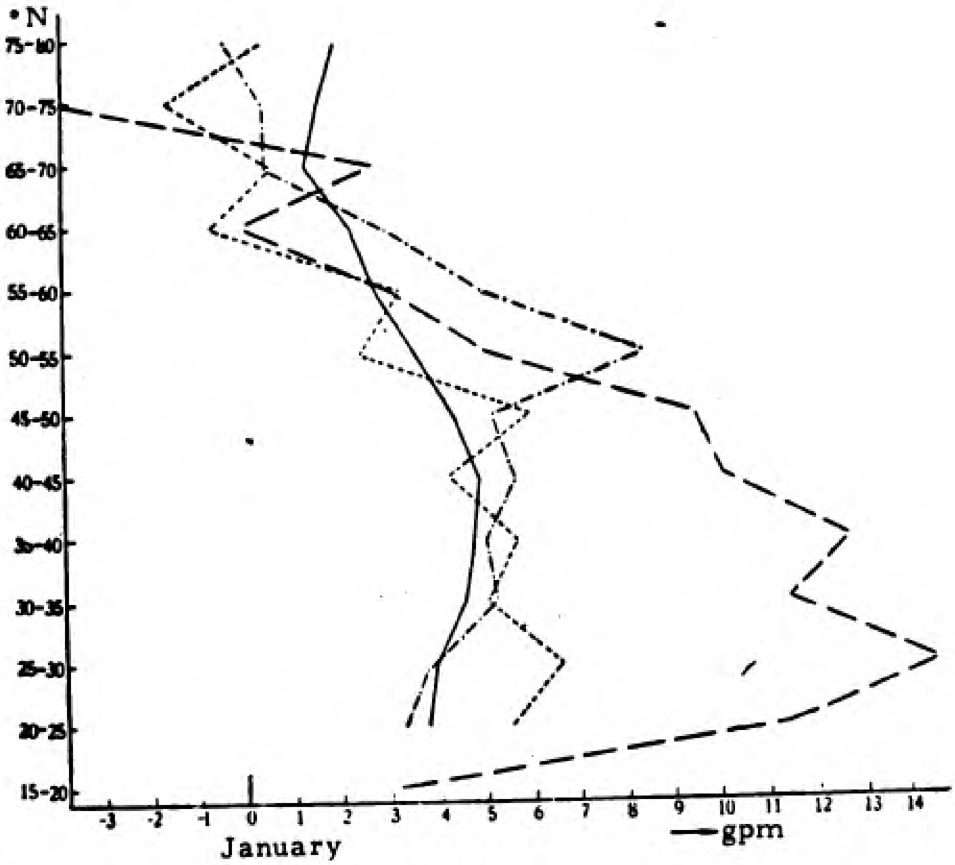


Figure 13.1

Meridional distribution of mean 1000 - 700 mb thickness
at intervals of 5° latitude in January.

—— Northern hemisphere
---- Eastern Asiatic Continent
—·—· North America
——— 1000 - 500 mb thickness over
Eastern Asiatic Continent.

All values averaged along latitude circle.

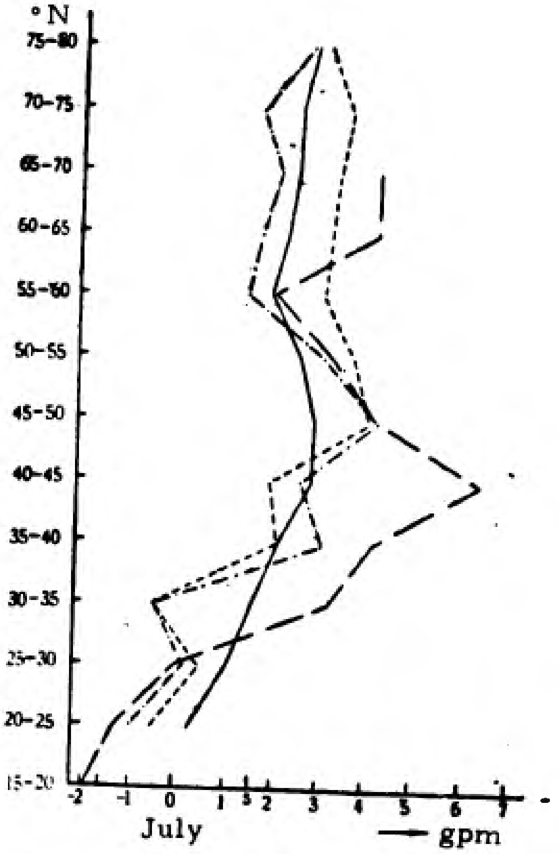


Figure 13.2

Meridional distribution of mean 1000 - 700 mb thickness at intervals of 5° latitude in July.

—— Northern hemisphere
---- Eastern Asiatic Continent
—·—· North America
— · — 1000 - 500 mb thickness over Eastern Asiatic Continent.

All values averaged along latitude circle.

TABLE 1

The mean south-north temperature difference at each standard pressure level over various stations in China for January and July 1953 - 1957

Station Temperature Difference (°C) Height (mb)	Nanning to Chihchiang	Chihchiang to Hankow	Hankow to Chengchow	Chengchow to Taiyuan	Taiyuan to Peking	Peking to Silinhot
		January				
900	9.4	1.0	2.9	4.1	0.9	12.9
850	8.3	1.5	3.5	3.3	2.6	6.9
800	7.2	2.2	4.6	3.3	3.0	4.4
700	6.4	3.6	5.4	3.9	3.0	2.7
600	5.2	4.4	5.5	3.8	2.7	3.0
500	3.5	6.1	5.5	3.3	4.0	-0.5
400	3.3	6.1	6.1	3.4	2.6	1.7
300	0.5	7.5	6.4	3.3	2.4	0.2
200	0.8	3.3	0.6	3.0	-0.4	1.1
		July				
900	-0.1	0.2	-1.1	3.0	0.0	2.2
850	0.0	-0.2	-0.3	1.1	1.7	0.1
800	-0.3	0.3	-0.4	3.8	-1.3	0.2
700	-0.2	-0.2	1.0	1.2	1.1	1.5
600	-0.6	0.4	1.4	2.0	1.1	1.4
500	-0.7	0.0	1.2	1.7	1.6	1.8
400	-1.4	0.5	0.7	2.5	2.0	2.0
300	-2.0	0.7	0.0	2.5	2.6	0.2
200	-2.6	1.0	-0.6	-0.8	0.8	0.4

TABLE 2

Temperature difference along 80°W at intervals of 5° latitude at the various standard pressure levels for January and July 1953 - 1957

Latitude / Temperature Difference (°C) / Height (mb)	25–20	30–25	35–30	40–35	45–40	50–45	55–50	60–55	25–20	30–25	35–30	40–35	45–40	50–45	55–50	60–55
	January								July							
850	2.4	3.3	5.1	6.9	6.2	7.4	4.4	2.1	0.1	-0.1	-0.2	1.7	2.9	2.6	2.7	2.1
700	1.7	3.2	4.9	6.4	5.4	5.8	4.1	2.6	0.6	0.0	-0.2	1.5	2.4	3.0	2.1	1.7
500	2.5	3.1	4.0	5.5	5.0	4.7	3.4	2.9	0.2	-0.3	0.1	1.8	2.3	3.0	2.3	1.5
300	2.7	2.4	2.8	3.6	2.7	2.6	0.8	1.2	0.5	-0.6	0.3	1.7	2.8	2.4	2.2	1.8
200*	1.0	1.0	0.0	-1.5	-2.0	1.0	1.8	2.2	-0.5	0.0	-0.7	-0.6	-0.7	-2.5	-1.1	-0.9

* Less than 5 years of data.

TABLE 3

Annual temperature ranges at various stations in China versus those at corresponding latitudes in America

(Mean of January and July 1953 - 1957)

Station Temperature Difference (°C) Height (mb)	Foochow	25°N	Nanking	30°N	Chengchow	35°N	Peking	40°N	Changchun	45°N
850	17.1	5.4	24.5	8.8	25.4	14.1	29.5	11.9	33.0	22.6
700	11.5	2.6	17.6	5.8	21.5	10.9	26.0	15.8	28.6	18.8
500	8.4	3.5	17.4	6.9	20.2	10.8	24.2	14.5	27.3	17.2
300	4.3	4.8	15.7	7.8	20.9	10.3	21.5	13.2	20.3	12.1

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MONSOON REGIONS AND REGIONAL CLIMATES IN CHINA*

by

Kao Yu-hsie, Hsu Shu-ying, Kuo Chi-yunn and Chang Min-lie

Due to the vast extent of China, the planetary circulation varies markedly from place to place during each season depending on topography and the distance from sea. The general concept that China is a monsoon region is thus a vague and misleading one. In order to examine the common characteristics as well as the unique peculiarities associated with the monsoons, an attempt has been made to classify the whole country into various monsoon regions in terms of the seasonal variations in the pressure, wind and humidity fields and other meteorological elements such as precipitation and cloudiness. The climates of these regions are then described.

I. MONSOON REGIONS IN CHINA

(a) The Boundary between the Summer and Winter Monsoons

Because of the differences in topography, prevailing wind regime and position with respect to the planetary circulation, the degree of significance and the steadiness of the monsoon phenomenon are different over different regions and exert an important influence on the local climates. During the winter half year, the whole of the East Asiatic Continent is dominated by the winter monsoon whose southern limit is designated by the average position of the polar front in January (Line 1 in Figure 1). The

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* Manuscript completed in December 1956.

front lies across the southwestern Pacific and the South China Sea with the northeast trades or the equatorial easterlies to the south where the distribution of monthly temperature and rainfall is relatively uniform with small variability. Figures 2.1, 2.2 and 2.3 depict the mean monthly rainfall, cloudiness and relative humidity for four stations north and south of 15°N in the Philippines, and north and south of 7°N over the Malaya Peninsula.

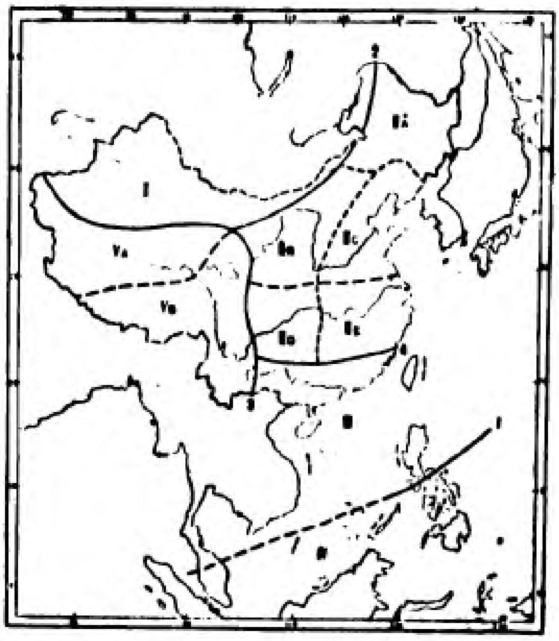

Figure 1

Climatic classification of monsoon regions over China.

It may be seen from these figures that the variations of the meteorological elements throughout the year are small south of Line 1. The region north of this line is affected by both the winter monsoon and the equatorial easterlies, and the climate is therefore dry and cold with little cloud and low humidity in winter but cloudy, wet and humid with a marked rise in temperature during summer. Thus the position of the polar front during deep winter not only represents the southern limit of the winter monsoon but also serves as a boundary between regions of tropical and equatorial monsoon climates.

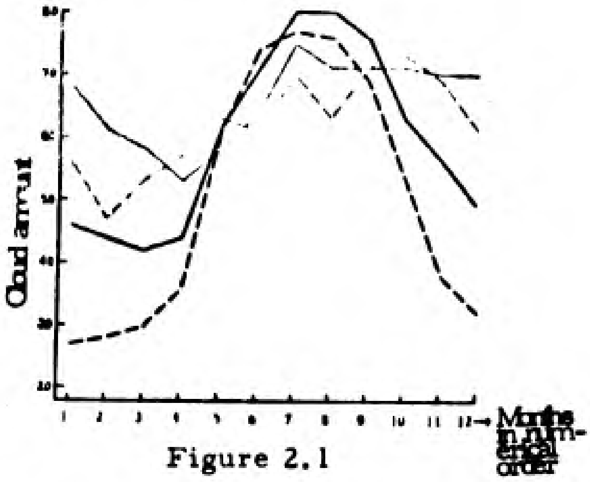

Figure 2.1

Annual distribution of monthly mean cloud amount at various stations.

(Legend: Thick solid line represents the mean for Baguio, Vigan and two other stations in the Philippines north of 15°N. Thin solid line represents the mean for Legaspi and three other stations in the Philippines south of 15°N. Thick broken line is the mean of four stations in Thailand north of 7°N. Thin broken line is the mean of four stations in Malaya south of 7°N.)

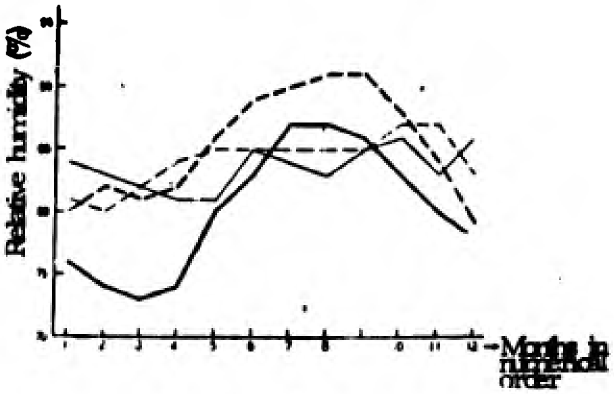

Figure 2.2

Monthly mean relative humidity at the various stations.

(Legend: same as in Figure 2.1.)

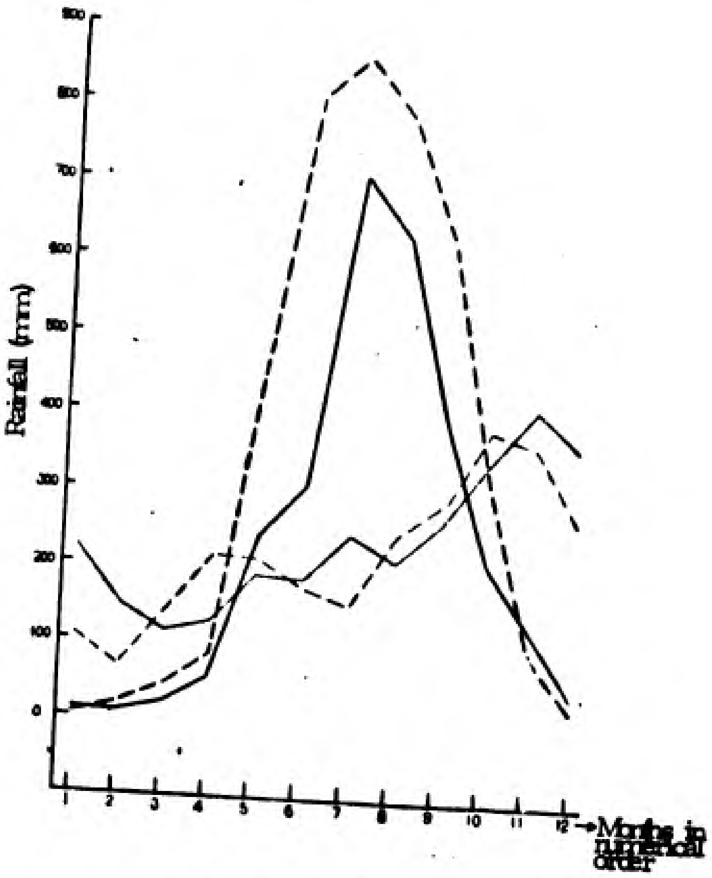

Figure 2.3

Annual distribution of monthly mean rainfall at the various stations.
(Legend: same as in Figure 2.1.)

The northernmost position of the polar front in the high-summer months of July and August is indicated by Line 2 in Figure 1. Again, this line depicts the limit of the summer monsoon as all meteorological elements to the south exhibit variations with a period of one year. Winter is dry and cold with northerly winds while summer is wet and hot with southerly airstreams. In regions to the north or northwest of the polar front, the prevailing winds are northerly throughout most of the year and the influence of the monsoons from the oceans is rarely noticeable even in high summer. Consequently, the climate is dry with little rain during both seasons and summer is particularly hot with exceedingly low humidity. The average position of the polar front in summer is therefore a boundary between relatively dry and wet climates and also separates the temperate regions from the subtropical zones.

(b) The Western Limit of the Winter Monsoon

The Tibetan Plateau is situated more than 3000 meters above sea level and is affected by the planetary circulation during both winter and summer. Conditions over the plateau are thus markedly different from those over the lowlands to the east, particularly in the neighboring regions east of the 1.5 - 3.0 km contours (i.e., the west and southwest parts of Szechwan Province and the southeast part of Kansu Province). Overcast and dull weather with a high frequency of fog is prevalent over these regions in winter while generally clear skies are observed over the plateau. The difference in weather is obviously due to the presence of the orographic front shown as Line 3 in Figure 1, whose southeastern portion is usually known as the Kunming Quasi-stationary Front. This system reaches a height of about 3 km over Hoshi (western side of the Yellow River), between 1.5 and 3.0 km along the eastern edge of the Tibetan Plateau and between 1.5 and 2.0 km in the south. It serves as a dividing line between regions of different monsoon climates near the plateau.

It is evident that the monsoon phenomenon over the region bounded by the three lines shown in Figure 1 is manifested by the interplay of the polar continental and tropical maritime air masses and the variations of all climatic and synoptic elements are associated with a period of one year. In places outside this region, such as the southern part of the Tibetan Plateau, the phenomenon may take an entirely different form. Thus some places are characterized by a complete absence of monsoons while in others, the prevailing wind is not "stable" enough and its seasonal variation is not well marked.

(c) The Northern Limit of the Southwest Monsoon

Although the mainland of China is affected by only one homogeneous monsoon airstream in winter, conditions are very different in summer when both the southeast monsoon and the southwest monsoon come into play. Since these two airstreams originate from different sources, a differentiation is often necessary. During the high-summer months when the polar front moves to its northernmost position, the equatorial convergence line which forms the foremost part of the southwest monsoon also shifts toward its northernmost limit (Line 4 in Figure 1). Thus some regions on the mainland such as South China and the coastal belts in the southeast are open to the winter monsoon in the cold season and the southeast and southwest monsoons in summer. The position of the convergence line at this time of the year forms the boundary between these two monsoon airstreams. In terms of climatic conditions, the region south of the polar front in high summer may be classified into a humid climatic zone north of Line 4 and a wet one to its south.

In principle, the mainland of China may be divided into five monsoon regions by means of the average positions of the polar front in deep winter and high summer, the orographic front in winter and the equatorial convergence zone in high summer (see Figure 1). It can also be classified latitudinally into the Westerly Wind Belt (I), the Subtropical Monsoon Region (II), the Tropical Monsoon Region (III), the Equatorial Monsoon Region (IV) and the Plateau Monsoon Region (V). In general, the transition of monsoons is not apparent in the Westerly Wind Belt. In the Subtropical Monsoon Region, the winter monsoon alternates with the southeast monsoon. In the Tropical Monsoon Region, the winter, southeast and southwest monsoons all come into play. The Equatorial Monsoon Region is controlled by the northeast trades and the southwest monsoon while over the plateau, monsoons are only caused by the seasonal displacement of the planetary westerly wind belt and the easterlies.

(d) Delineation of Major Monsoon Boundaries

Despite the fact that we have made use of frontal discontinuity in the classification of the monsoon regions, little has been achieved in accurately determining the mean positions of these systems on a monthly basis. Although numerous investigations have been made by workers in this country and abroad, no satisfactory and conclusive results have so far been obtained. In the present study, an attempt has been made to delineate the mean positions of the various frontal systems by means of all available data collected during 1955 - 1958.

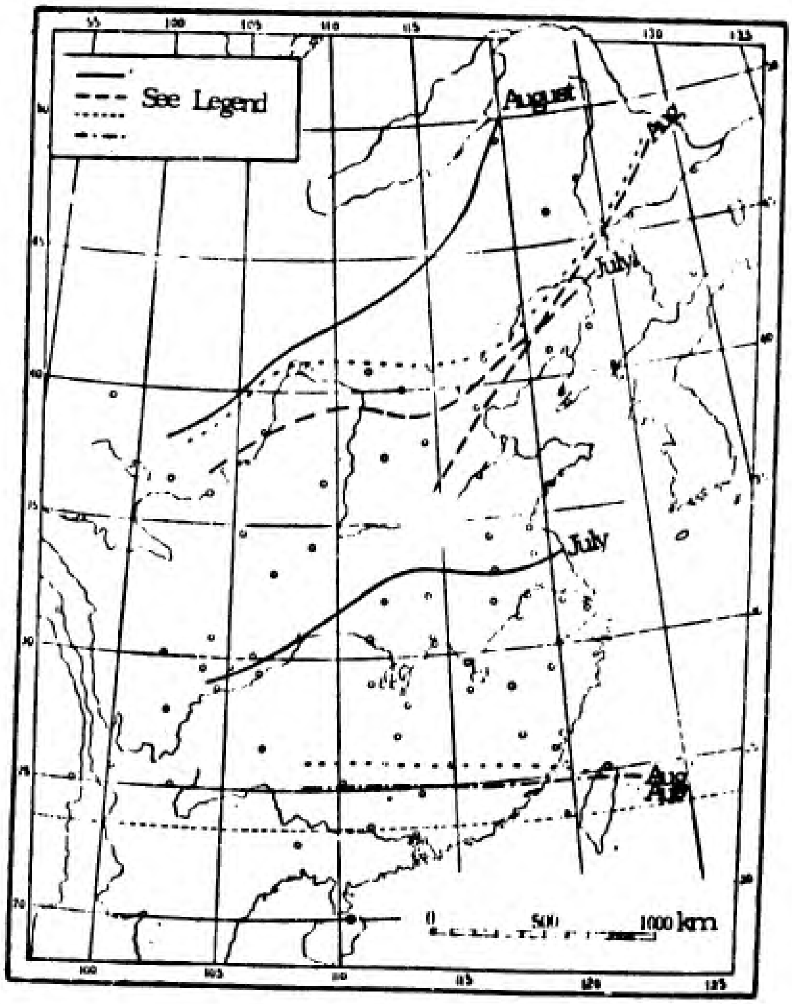

Figure 3

Frequency distribution of the polar front during high summer.

(Legend: Solid line represents axis of maximum frequency in July and August.
Dashed line represents axis of the second maximum frequency in July and August.
Dotted line is mean position of sea-level trough in August.
Dashed and dotted line is axis of streamline convergence in August.)

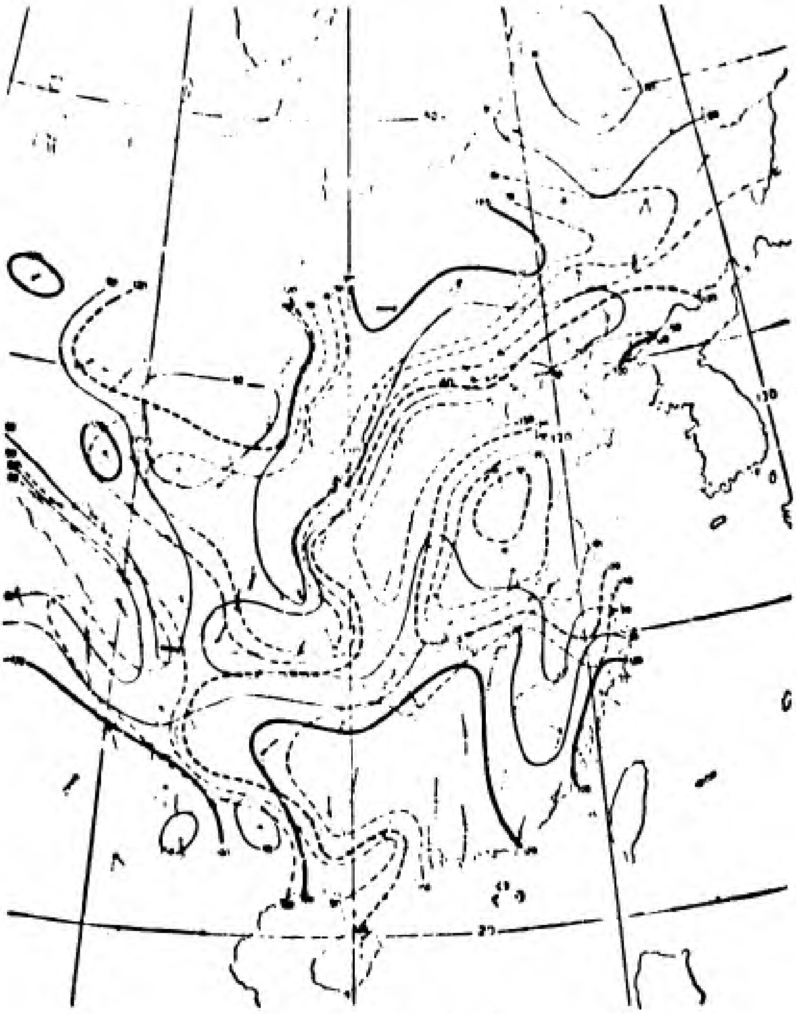

Figure 4

Direction and steadiness of resultant winds in January and July.

(Legend: Solid arrows indicate wind direction for January and broken arrows for July. Broken lines are isogons of resultant winds and solid lines are isolines of wind steadiness.)

Positions of the polar front and equatorial convergence zone (equatorial front) in high summer: Figure 3 depicts the distribution of the frequency of daily occurrence of the polar front over each 2½° square for the months of July and August 1955 - 1958. The region of maximum frequency (shown by thick solid line) extends from the western part of the Greater Khingan Mountains southwestward across the great bend of the Yellow River to the southeast of Chiuchuan. A secondary maximum is found in the belt stretching from Heilungkiang southwestward across Tsitsihar to the region between Chengchow and the eastern coast in Hopeh (broken line) and may be due to the "stationary front" during the period from 20 August to the end of the month. The mean sea-level pressure chart for August shows that the region of maximum frontal frequency coincides with the position of the mean trough lying across North and Northeast China (see dotted line in Figure 3). This indicates that the trough system is not entirely thermal in nature and that the pattern presented in Figure 3, though based on only 4 years of observations, is of definite climatological significance.

Figure 4 shows the distribution of the resultant wind and its degree of constancy for January and July. Solid arrows depict the wind direction for January while dotted arrows that for July. By examining the seasonal change of the resultant wind, it is possible to define regions for which the change is either well-marked or insignificant. The boundary separating these regions naturally fluctuates from month to month but is in fair agreement with the belt of maximum frequency of frontal occurrence. A study of the flow patterns at the various levels prepared by Hsu et al [1] shows that the convergence line over North China in August (Figure 5) lies close to the zone of maximum frontal frequency. It is well known from synoptic experience that a heavy rain belt is usually located within a frontal zone (surface or upper level) and the mean position of this belt as determined from pentad rainfall observations may thus be used to represent the position of the polar front. Figure 6 shows

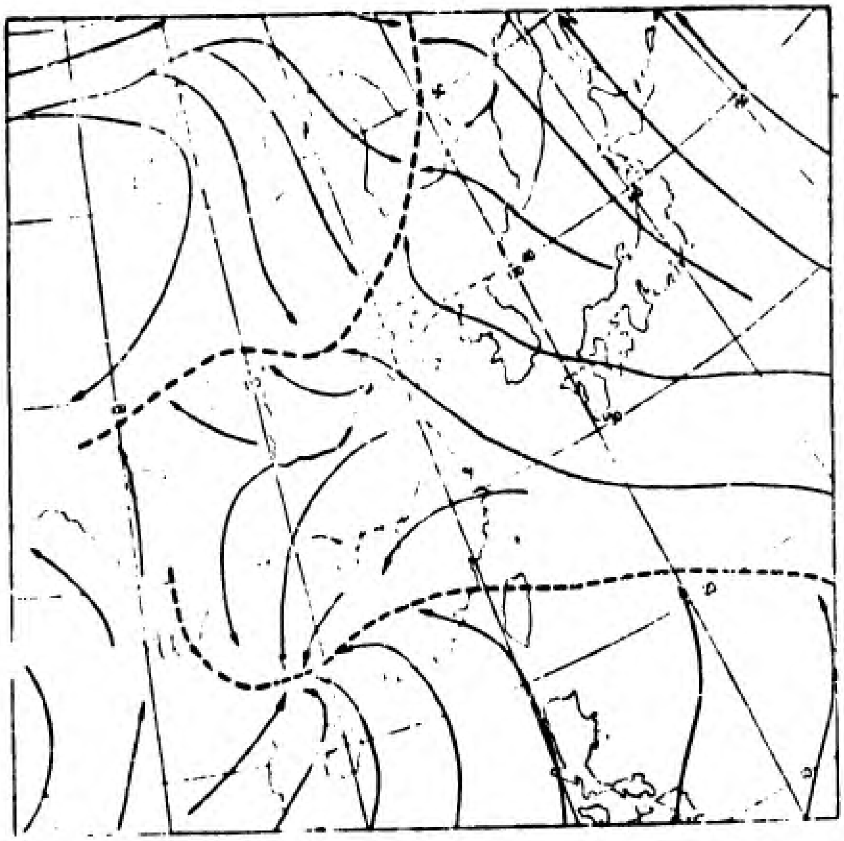

Figure 5

Mean surface flow pattern in August.

such heavy rain belts over China during the period 10 July to 2 September and clearly reveals the existence of two separate rain areas, one over North China and the other over South China. During the third pentad of August (9th to 13th), the rain belts are located at their northernmost positions. The northern rain belt is associated with the polar front while the southern one is considered to be due to typhoons and the equatorial trough. However, the observed east-west orientation of the pentad profile of the heavy rain belt together with its systematic northward displacement cannot simply be attributed to rainfall from typhoons. The observed pattern can only be accounted for by the seasonal poleward movement of the low pressure systems associated with the equatorial trough or front. The poleward boundaries of these two rain belts can therefore be taken to represent the northern limits of the polar and the equatorial fronts. Similar results would be obtained by the other methods described previously.

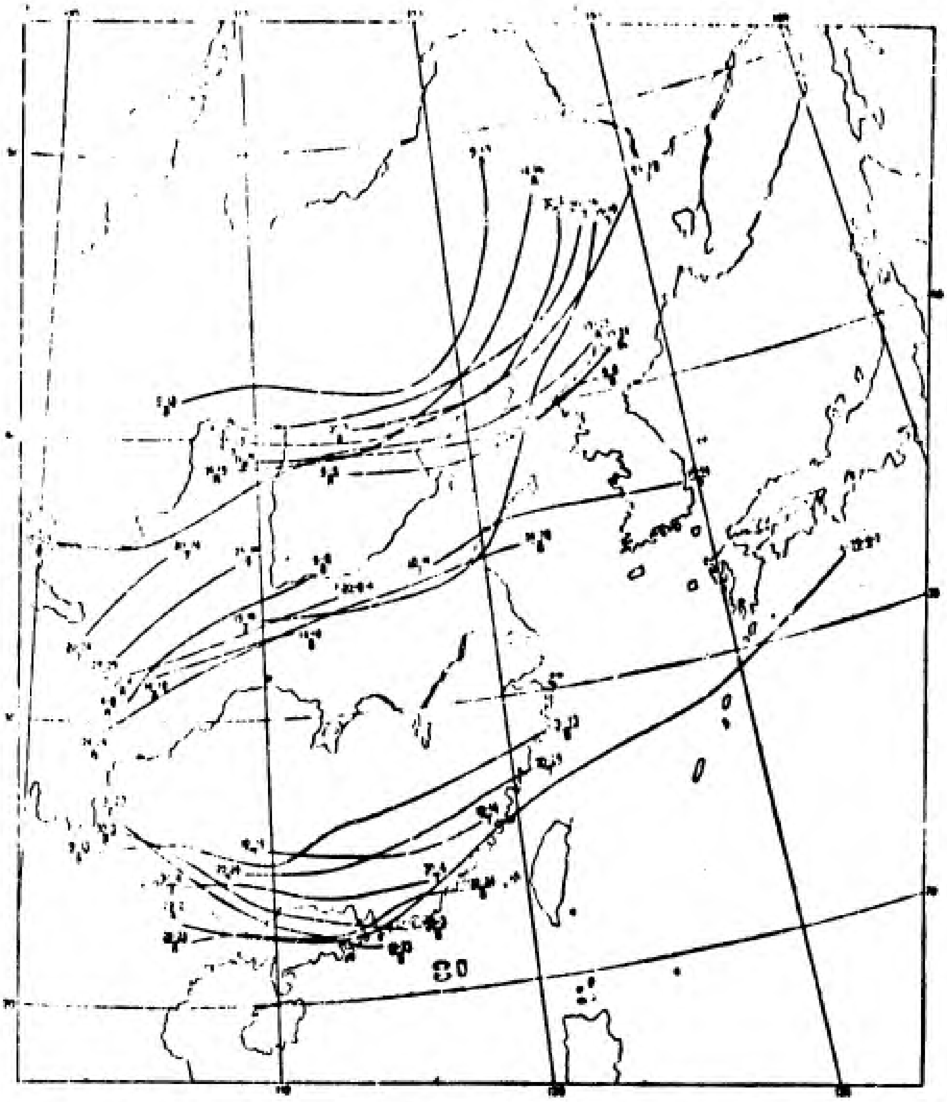

Figure 6

Positions of the line of mean pentad rainfall maximum over China during the period 10 July - 2 September.

The average position of the polar front during high summer is relatively difficult to determine. Meteorologists have also brought forward different concepts concerning the structure of the equatorial front over China, and the positioning of this system varies considerably among the various workers [3 - 5]. The existence of divergent views is probably due to the fact that the equatorial front is difficult to identify on daily synoptic charts because of lack of organized convergence at the upper and lower levels and contrast in meteorological elements across the front, resulting in the absence of significant weather. The analysis is even more formidable when the front has moved into China. In July, the system is not too far from the coast and identification is still possible. In the present study, the mean position of the front in August is determined by a close examination of the positions of the trough and pentad rain belt together with the annual variation of the frequency of wind direction over South China.

Figures 7.1 and 7.2 indicate the annual variation of surface pressure with latitude along 115°E and 120°E in August during the period 1951 - 1957. It is seen that the trough over South China lingers around 25°N along 115°E and fluctuates between 20° and 30°N along 120°E. The system stays north of 25°N in 1953 - 1957, north of 30°N in 1957 and at 20° - 25°N in 1951 and 1952. The corresponding profile for 110°E (chart omitted) shows a larger latitudinal variation with the trough located more frequently to the south. The above results are contrary to synoptic experience and also discordant with the position of the mean sea-level trough determined from long-period records. However, it is noted in Figure 3 that the dotted line taken from

Reference [5] agrees closely with the dot-dash line extracted from Reference [1]. This suggests that the mean position of the trough of low pressure over China should be located near 25°N in August.

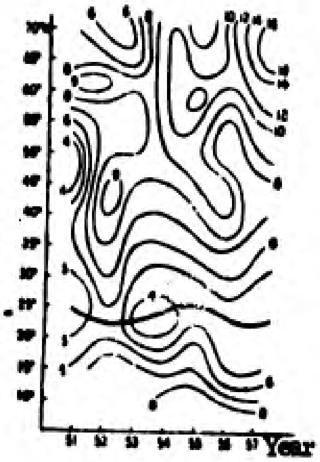

Figure 7.1

Annual variation of surface pressure at various latitudes along the 115°E meridian in August.

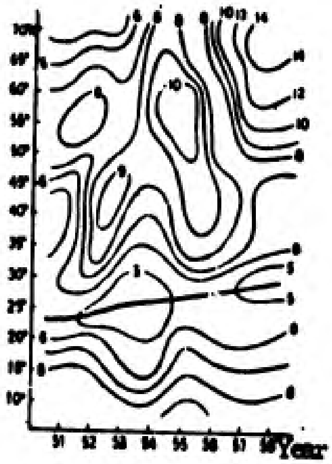

Figure 7.2

Annual variation of surface pressure at various latitudes along the 120°E meridian in August.

Since the equatorial front forms the boundary between the southeast and southwest monsoons, it may be thought that these monsoon regions can easily be identified by means of the prevailing wind direction. However, a study of prevailing wind charts presented by various workers [5, 6] shows that the surface wind over South China is either southerly or southeasterly. Maps of resultant winds, on the other hand, give southerly and southwesterly over the region. Thus both sets of maps fail to bring out the existence of the monsoon discussed above. The discrepancy may be due to the following reasons: (i) The selection of the calender month as a time unit for computing averages may introduce certain artificial distortion. (ii) There are advantages as well as disadvantages in the use of resultant or prevailing winds. Prevailing winds are sometimes difficult to determine when there is a wide scatter in the frequency distribution while resultant winds tend to be much less representative if the wind speed shows a large variation. (iii) Topography may give rise to complications in the flow pattern. A discussion on the annual variation of wind direction over several coastal stations during the period 1924 - 1936 is now given below.

The monthly frequency of occurrence of southwesterlies and the winter prevailing winds (northerlies, northwesterlies and northeasterlies) for 8 stations is presented in Figure 8. With the exception of the Chiungshan and Sanshui area and the vicinity of Amoy and Tungting Tao (Island), where the increase of southwesterlies is less marked than the southeasterlies, all the coastal stations south of Foochow experience a significant increase in the frequency of the southwesterlies after May with a corresponding decrease in the winter monsoon. Thus southwesterlies are more predominant than

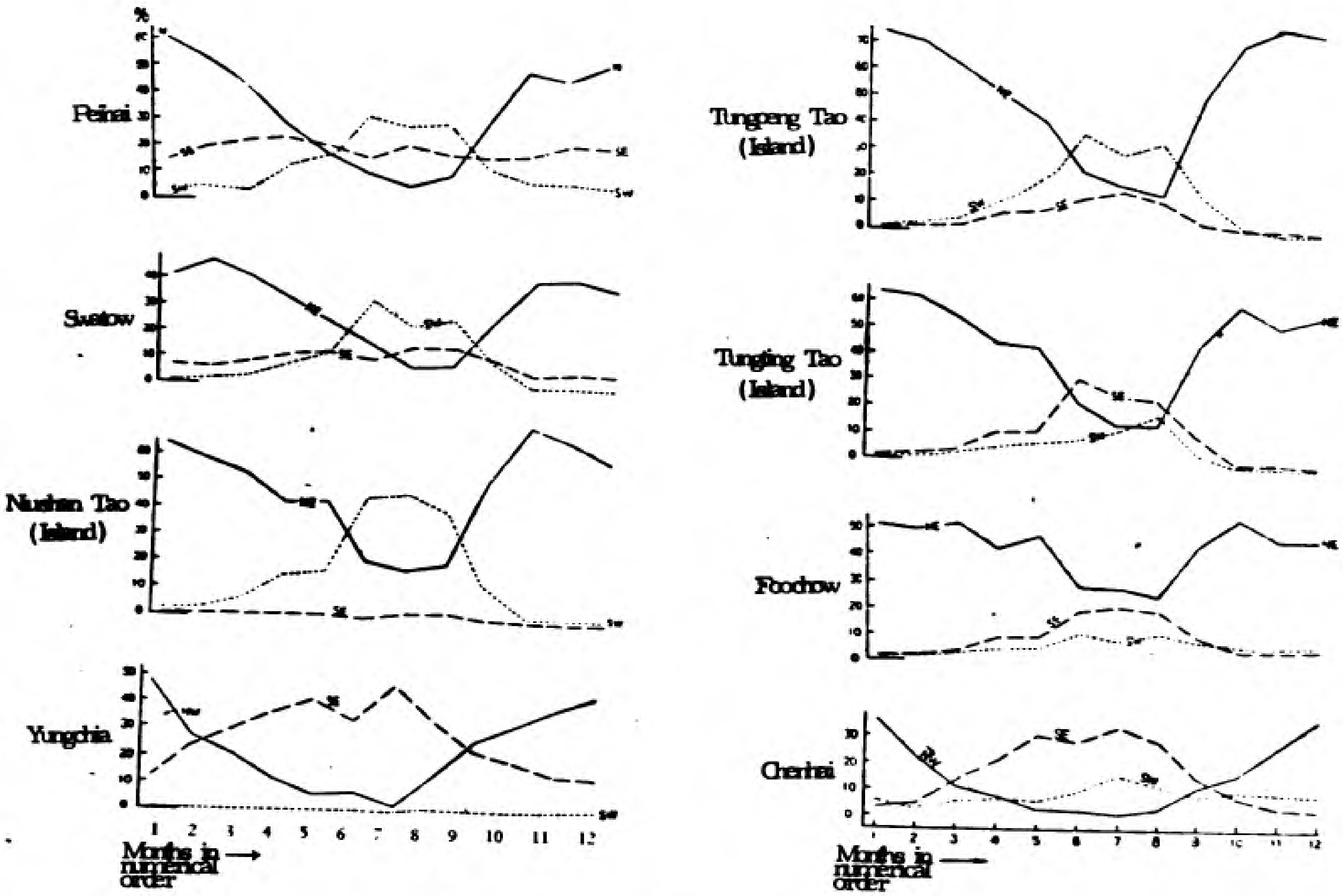

Figure 8

Variations of the frequency of occurrence of southeast, southwest and the winter prevailing winds (north, northeast and northwest) at the various coastal stations.

southeasterlies in summer. Conditions are, however, reversed north of Foochow and the frequency of southeast winds is considerably greater than that of the southwesterlies in places like Yungchia and Chenhai. An examination of the ratio of the frequency of occurrence of these two air-streams shows that Foochow lies in the transition zone with southeasterlies prevailing in the north and southwesterlies in the south. It is unfortunate that due to the sparse network of observing stations inland and the lack of long-period records from available stations, it has not been possible to carry out a similar comparison meridionally to determine the position of the equatorial trough. It is hoped that this goal will be achieved when more upper-air observations become available so that a final adjustment may be made.

In order to locate the fronts as precisely as possible, the distribution patterns of the difference in relative humidity between winter (January and February) and summer (July and August) near the surface and at the various upper levels have been used as a final correction factor. Since other air mass properties such as specific humidity and the various temperature parameters are characterized by higher values in summer, their annual range is incapable of depicting the boundary between different air masses. Relative humidity, on the other hand, is particularly useful in this respect. Although the specific humidity of a polar air mass is slightly higher in summer than in winter, the relative humidity shows an opposite configuration because temperatures are much higher in summer. In the desert regions of Inner Mongolia and Sinkiang, where the moisture supply is cut off and the air temperature is extremely high in summer, the relative humidity is much lower than in winter. In areas affected by maritime air masses which are heavily ladened with moisture, the effect of high temperatures is not so significant and the relative humidity is then higher in summer. Consequently, the use of the above method will enable us to define regions which are normally affected by northerly or southerly airstreams in summer. The zero isopleth depicts very approximately the position of a front separating two different monsoon regions.

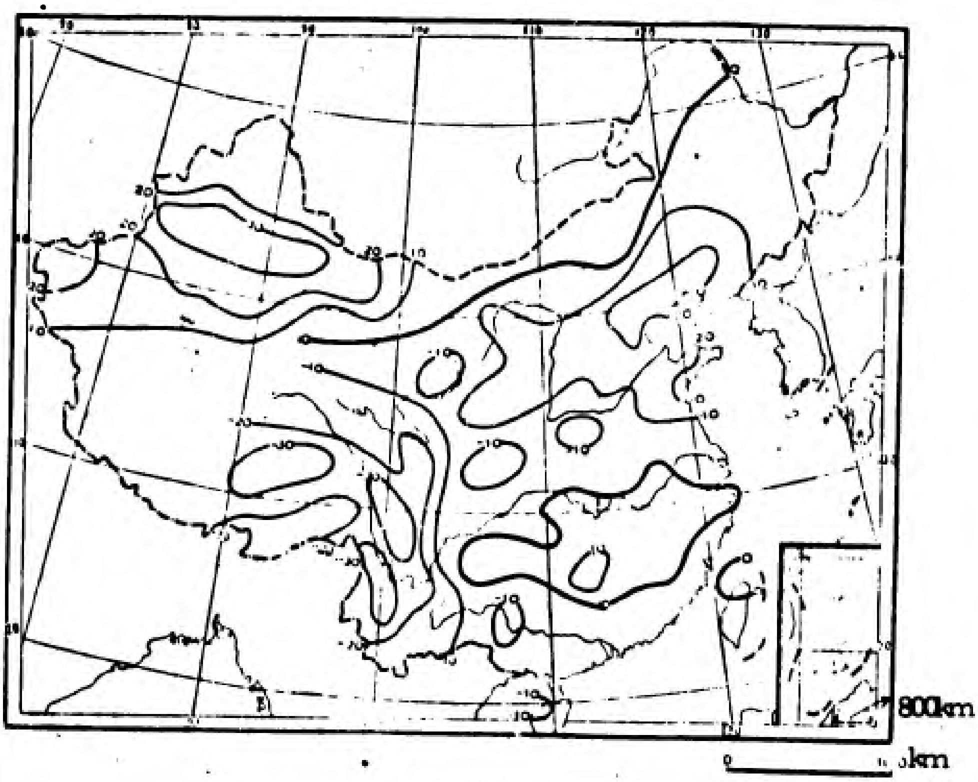

Figure 9

Difference in surface relative humidity between winter and summer (values of January and February minus values of July and August).

Figure 9 represents such an attempt and shows that the zero isopleth runs from the upper Nen Chiang (River) in the northeast across the Khingan

Mountains and Yin Shan to Kaotai and Chang-I in Kansu. Inner Mongolia, Sinkiang and a part of Northeast China lie in a positive region northwest of this isopleth and is associated with higher relative humidity in winter than mid-summer. The difference reaches 44% in the Tarim Basin. With the exception of the central regions south of the Yangtze, negative values are found in all places southeast of this line where the relative humidity is higher in summer. The zero isopleth is therefore the probable northern limit of the summer polar front.

A second positive region occurs south of the Yangtze at 25° - 30°N. The lower summer humidity may be due to the fact that the lake districts are mainly influenced by the subtropical ridge in July and August so that the weather is generally fine and hot with very little rain, resulting in the highest temperatures over the country and consequently a reduction in relative humidity. Conditions are, however, different south of 25°N where the high values in summer indicate that both Kwangtung and Kwangsi are under the influence of different air masses or pressure systems. The zero isopleth along the 25°N is probably most representative of the northernmost position of the equatorial front.

Figure 10 depicts the typical profiles of the vertical distribution of the difference in relative humidity between January and July for 44 stations including Peking. (The length of records from which the means were computed is 10 years for 10 stations, 3 years for 17 stations and 1 year for the remaining 17 stations.) Positive values indicate that the relative humidity is higher in January. These profiles can be classified into five types. The first is the southwest monsoon type which is characterized by higher values at all levels in July (Line I in the figure). The curves for Tengchung, Hsichang, Kunming, Lhasa, Heiho, Kantse, Chakhan Usu, Chamdo (or Changtu) and Nanning all belong to this type. The second type (Line II in the figure) shows opposite characteristics to the first with higher relative humidity in January from surface to upper levels. Typical examples are given by the profiles of Silinhot, Hailar, B. Sunid (Undor Temple), Laochunmiao, Golmo, Mangyai and Kashgar. Type 3 is associated with the southeast monsoon (Line III in the figure) with negative values at the lower levels and positive above. This category includes most stations between 30° and 40°N such as Peking, Taiyuan, Chengchow, Sian, Nanking and Shanghai. The fourth type is a transitional one between the southeast and southwest monsoons and the typical vertical distribution of relative humidity is given by the curves of Foochow, Swatow, Haikou, Chihchiang, Kweiyang and Chengtu. The general distribution is similar to that of the southwest monsoon with predominantly negative values at the lower and upper levels but positive ones in the middle layer, the latter feature being characteristic of the Type 3 pattern. Type 5 is given by the profiles of Hankow, Sian, Chengchow, Ichang and Kanchow and is associated with the southeast monsoon (Line V in the figure). Because of the high summer temperatures in the surface layer, the relative humidity is lower in July than in January, resulting in an opposite distribution of that of Type 4. Thus negative differences are observed at the middle levels with positive ones above and below.

From a study of the relative humidity differences at the various levels, three distinct boundaries can be established. The boundary between the southeast and winter monsoons runs from the south of Hailar, B. Sunid and

Silinhot across the north of Chakan Usu and Heiho to the Himalayas at about 30°N. The boundary between the southeast and southwest monsoons lies to the south of Foochow, Chihchiang and Kweiyang. The above findings agree closely with the results discussed earlier. The third boundary is located within the region of the southeast monsoon.

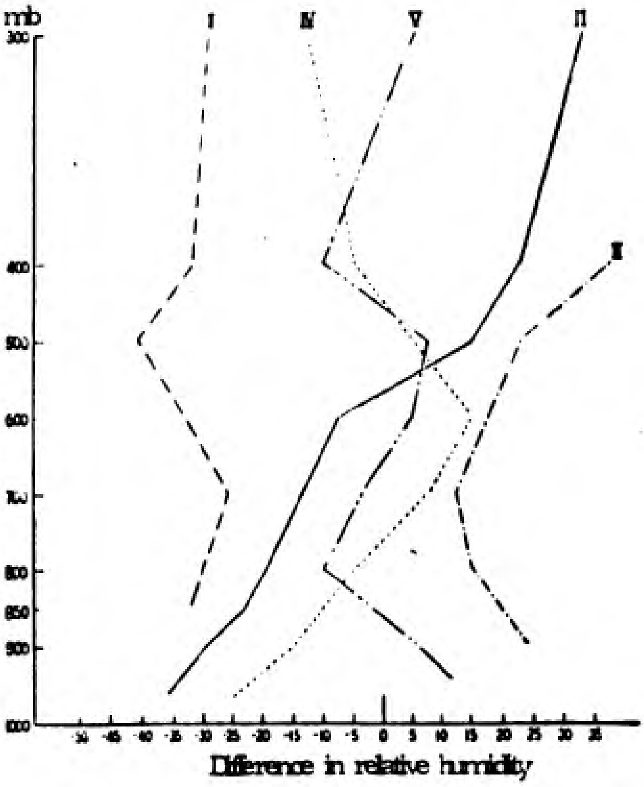

Figure 10

Vertical distribution of the difference in relative humidity between January and July (January values minus July values).

II. SIGNIFICANCE AND STABILITY OF THE MONSOONS OVER THE VARIOUS REGIONS IN CHINA

Because of the difference in geographical setting and the degree of continentality, the influence of the monsoons varies considerably from place to place even within the same monsoon region and it appears necessary to devise some form of index to evaluate this influence.

Due to the vast extent of China, aperiodic or random travelling synoptic systems may exert unequal influence on the weather at different locations, which may further be modified by local topography. Thus, the methods of Hann [7], Conrad [8] and Khromov [9] for the determination of monsoon indices all suffer from certain limitations although they have their individual points of merit. In an attempt to remove these shortcomings, the present authors have derived a new index which is obtained by analyzing the differences in the frequency of winds of the same directions between January and July. The maximum positive and negative differences are first computed from which the corresponding frequencies of winds from opposite directions are subtracted. If the values after the subtraction are not exceeded by any differences from other wind directions, then the sum of their absolute values is taken as the monsoon index. The angular distance between these two directions can also be used to indicate the degree of predominance of the monsoons.

In regions of complicated topography where travelling pressure systems are frequent, the above index is considered more representative than that derived by Conrad because the influence of these factors has been removed in theory.

Indices due to Conrad, Hann and Khromov have been computed from the same set of observations and compared with that proposed by the present authors. (Distribution charts for the Conrad index and ours are presented in Figures 11.1 and 11.2; charts for the other indices are omitted.) The results show that the angular displacement of monsoon winds or the "monsoon angle" given by other methods is generally very small in most regions of China implying that the monsoon phenomenon should not be significant. This is obviously contrary to observations and the discrepancy lies in the fact that the indices from these methods are not suitable for application in regions of complicated topography.

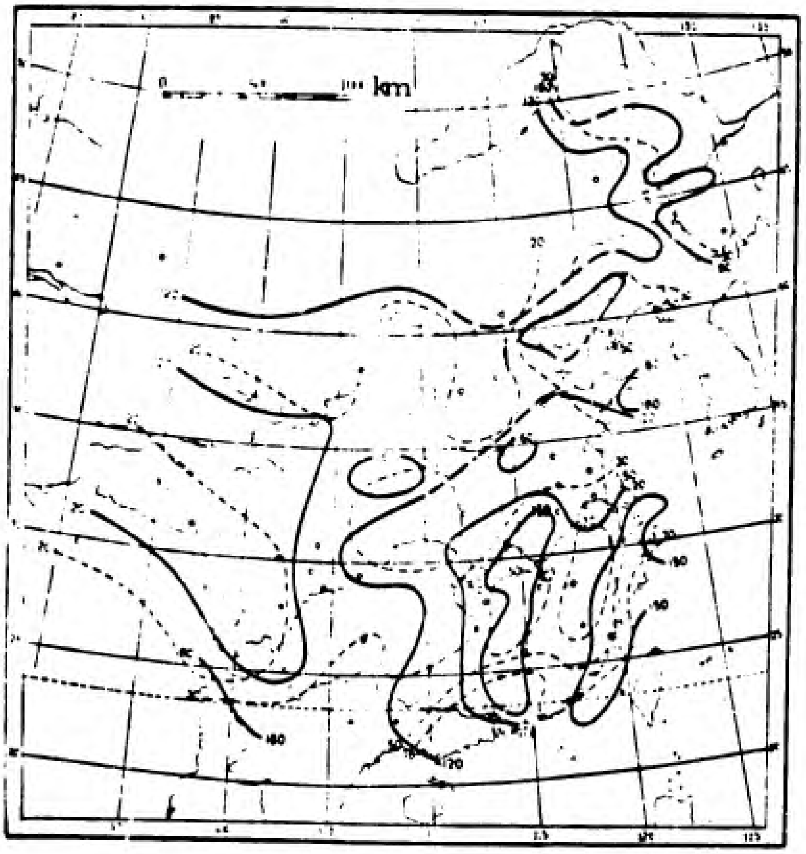

Figure 11.1

Distribution of the Conrad monsoon index.

(Legend: Solid lines are isogons of "monsoon angle". Broken lines are isopleths of the monsoon index.)

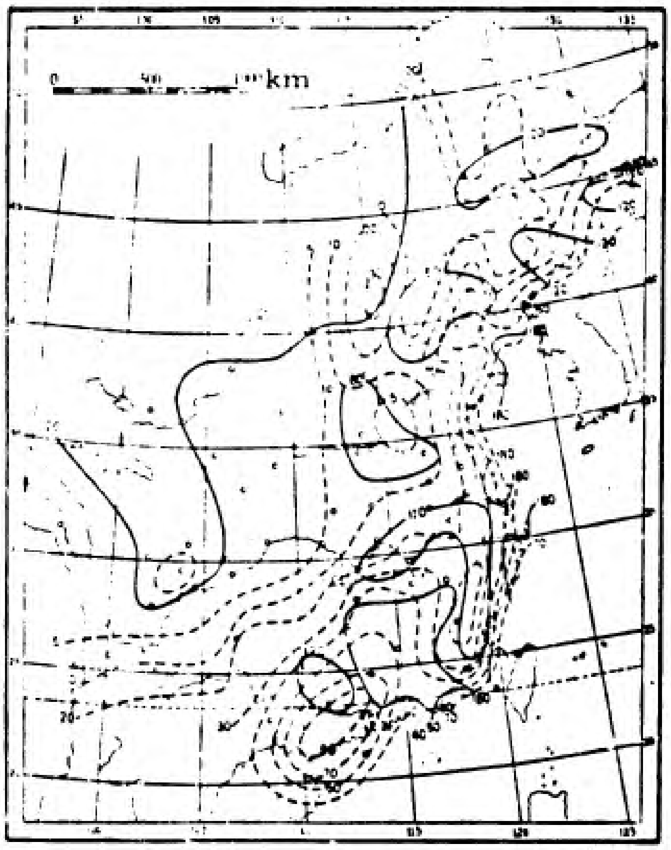

Figure 11. 2

Distribution of the monsoon index proposed in this study.
(Legend: Solid lines are isogons of "monsoon angle". Broken lines are isopleths of the monsoon index.)

A comparison between Figures 11. 1 and 11. 2 reveals that with the exception of the eastern edge and the northwestern part of the Tibetan Plateau, monsoons are observed over all other regions of the highlands. Figure 11. 2 clearly indicates that the monsoon angle is greater than 120° over these regions. The monsoon index is, however, generally smaller than that of Hann and Khromov and negative values are also found in some places. Figure 11.1 shows that the monsoon angle is greater than 120° in most parts of the chart area except the region extending from the Yangtze - Hwai Ho basin across East Honan and West Hunan to the border line along Szechwan and Kwangsi. The small values over this region are probably due to the effect of migratory pressure systems. This phenomenon is not found in Figure 11.2. Instead, the following interesting features are noted: (i) The 120° isogon lies very close to the northernmost periphery of the summer monsoon over the northwestern region. (ii) The monsoon index decreases systematically toward inland as is expected from the influence of land-sea distribution. (iii) The smallest monsoon index occurs west of 110°E. This is in agreement with synoptic experience since the region of small index values is generally known as the "stagnant" region with a high frequency of calm winds in winter and also a wide scatter in the distribution of wind direction. In summer, cold air also frequently spreads into the region behind a minor upper-air trough between 110° and 115°E and gives rise to a large variation in wind direction.

The stability and significance of the monsoon phenomenon can best be illustrated by the magnitude of the monsoon index and the monsoon angle given in Figure 11.2. Monsoons are generally more stable and significant in the east than in the west and also more pronounced over Southeast and South China than over Central, North and Northeast China. The phenomenon is, however, less pronounced over the Yellow River - Hwai Ho region and the lower basin of the Liao Ho where the monsoon angle is relatively large.

By considering the distribution of these two parameters as well as the variations of the relative humidity field and the isopleths of angular change greater than 135° in the resultant wind field for January and July, we may subdivide the southeast monsoon region into five smaller regions (see Figure 1).

III. CLIMATOLOGICAL SIGNIFICANCE IN THE CLASSIFICATION OF MONSOON REGIONS AND THE DELINEATION OF MONSOON BOUNDARIES

(a) Classification of Monsoon Regions

By using the mean positions of the polar and equatorial fronts during deep winter and high summer, we have classified the mainland of China into five major monsoon regions, namely, the Westerly Wind Belt (I), the Subtropical Monsoon Region (II), the Tropical Monsoon Region (III), the Equatorial Monsoon Region (IV) and the Tibetan Plateau Monsoon Region (V). The Subtropical Monsoon Region is most extensive with a multitude of complex topographical features and a significant spatial variation in the planetary circulation. The stability of the monsoon phenomenon therefore varies considerably within the region which may further be subdivided into five smaller regions in terms of the variations of the monsoon parameters between January and July. These are the Northeast Region, the North China Region, the Lower and Middle Yangtze Basin, the River-bend Region (i.e., the region near the great bend of the Yellow River) and the Szechwan Basin as shown in Figure 1.

(b) Climatological Characteristics of Monsoon Regions

The reason for choosing the mean positions of fronts as boundaries of monsoon regions is too plain to need emphasis. In the region bounded by the two seasonal limits of the polar front, the climate is dry in winter and rainy in summer. Outside this region [e.g., near to the Equatorial Region (IV)], the contrast becomes less marked. Regions to the south are essentially controlled by the equatorial airstream and rarely affected by continental air masses in winter so that the rainfall is higher during this season. In the high-latitude region (I), the polar airstream dominates throughout the year and dry weather prevails in all seasons with exceptionally high temperatures in summer. In other regions, conditions are generally dry and cold in winter and moist and hot in summer with rainfall concentrated in the hot months. However, the pattern varies from place to place due to the differences in topographical effects and the intensity of the planetary circulation aloft.

Tropical Monsoon Region (III): This region is bounded by the deep-winter polar front in the south and the high-summer equatorial trough in the north. The northern boundary marks the southern limit of high pressure centers which move eastward into the sea in January and the region is thus situated at the southern periphery of the polar cold anticyclone in deep winter with relatively steady east to northeast winds and is little affected by migratory systems. In summer, south to southwest winds prevail south of the equatorial trough. Consequently, the monsoon phenomenon is most well-marked over this region. Four transitions of the monsoons take place each year; the change from the winter to the southeast monsoons occurs in May and June and that from the southeast to the southwest monsoons in July and August with two similar changes in the reversed order from summer to winter. The transitional periods are, however, shorter in autumn than in spring, which give rise to two peaks in the annual distribution of rainfall, one in May and June and another in August and September. The first peak is due to the polar front while the second is associated with the equatorial trough. Although a relatively dry period occurs between these two rain spells, the amount of rainfall recorded during this period is considerable and the situation is in no way similar to the summer droughts in the Yangtze basin, which may be disastrous. The weather in this region is also drier than in the north and south in winter because of the prevalence of monsoon winds near the surface and the presence of the subtropical ridge at higher levels over certain parts of the region.

Subtropical Monsoon Region (II): The region lies between the polar front in high summer and the equatorial trough in July and August and is only affected by two monsoons during the year. As the region covers a vast extent with a wide variation in topography, continentality and the influence of planetary circulation, a further sub-division is necessary:

Central China Region (II_D, II_E): The northern boundary is given by the mean path of eastward moving cold anticyclones in January while the southern one is marked by the southernmost limit of these tracks. The upper-level subtropical ridge fluctuates within these two lines which therefore contain the mean axis of wind-shift from easterly to westerly or vice versa. Cyclonic waves and migratory high pressure systems generally pass through this region into the neighboring seas thus reducing the steadiness of the monsoon winds. The region may be separated into two sub-regions (II_D and II_E) by considering the meridional track of cold air in winter and the eastern limit of cold air intrusion in summer (cold air may travel along 110° - 115°E behind an upper-air trough into Szechwan and Kweichow). Monsoons are more pronounced in the east than in the west.

The climatological characteristics of the two sub-regions can easily be deduced from the general discussions presented above. II_D is a "stagnant" sub-region in winter when wind changes are highly irregular. In summer, the frequent invasion of cold air also gives rise to variable winds and the monsoon phenomenon is therefore insignificant. Random migratory systems are frequent in the sub-region II_E and although monsoons are slightly more pronounced here, they are not so well-marked in comparison with other regions to the north and south. On the other hand, winter precipitation is generally high over this sub-region due to the presence of migratory synoptic systems and the upper jet stream. In summer, rain is deficient under the influence of the subtropical ridge and droughts often result. Since cold surges are more frequent in the Szechwan basin than in Central China, the rainy season begins in June and July over II_D and in July and August over II_E. The temperature rises to a maximum during the dry spell of July and August.

North China Region (II_B, II_C): The region is bounded by the high-summer polar front in the northwest and the northernmost limit of the tracks of migratory anticyclones in deep winter in the northeast. Its southern border forms the northern boundary of the Central China Region. Eastward moving systems are often observed to the southeast of the northern boundary but are extremely rare northeast of the line. Tropical easterlies prevail to the south of the region which therefore comes under the influence of the westerlies north of the subtropical ridge. Since the westerly wind regime dominates the region throughout the year with a large number of migratory synoptic systems, the monsoon phenomenon is not so pronounced as in Central and South China. The region may also be divided into two sub-regions II_B and II_C in terms of the effect of land-sea distribution and the activities of cold air in high summer.

The climate of II_C is exceptionally dry and cold in winter due to its location at high latitudes and the influence of the monsoon. The increase in rainfall in spring is insignificant because the summer monsoon rarely reaches this sub-region during the period. The rainy season sets in during the full establishment of the summer monsoon around the first decade of July when the polar front has reached North China and the heaviest falls are confined to the two months of July and August. The scarcity of rain in spring and autumn often gives rise to severe droughts.

Conditions in II_B are similar to those of II_D. As the invasion of cold air is more frequent in summer, the onset of the rainy season in this sub-region is delayed by one month compared with II_C. The highest temperature is usually observed before the rainy season.

Northeast Region (II_A): The high-summer polar front forms the western boundary of the region while the northern limit of the tracks of eastward moving cold anticyclones in deep winter marks its southwestern frontier. The northeastern and southeastern borders are delineated by the national boundary. Northwest monsoons are prevalent over this region in winter while cyclonic activities are frequent in summer. Since there is a significant difference in the properties of pressure systems affecting the region in January and July, the monsoon phenomenon is more pronounced than over the North China Region. The activities of cyclones during April - September give rise to considerable amounts of rainfall in spring and autumn but the annual distribution is more uniform than in the North China Region.

Tibetan Plateau Monsoon Region (V): Due to the sparseness of observation stations which are mostly situated in the eastern part of the Tibetan Plateau, it has not been possible to carry out a comprehensive study of the monsoon phenomenon over the region. However, from an examination of the rainfall distribution and the seasonal variation of the planetary circulation, it is almost certain that monsoons do exist over the Tibetan Plateau although they may be somewhat different from those in the plain region. An analysis of the upper-air records for 1957 - 1958 reveals that the monsoon airstreams over the northwest part of the plateau are completely different from those over the southeast. In general, the climatological characteristics in the former region resemble those over Inner Mongolia and Sinkiang while conditions in the latter are little different from those over India. It is still not yet possible to ascertain whether the southeast and southwest monsoons prevail over the southeastern part of the plateau and the best that can be achieved at present is to divide the plateau into two climatic regions. The characteristic features associated with each region will be discussed in a separate paper [10].

Summing up, we may say that the monsoon phenomenon is more pronounced in the east than in the west of China while the Tibetan Plateau forms a separate region of its own. Monsoons are also more prominent and complex in the south than in the north. Since the summer monsoon spreads northward from the south, the rainy season sets in earlier over South China. The Szechwan basin is more frequently affected by cold surges than eastern China during summer so that its rainy season is correspondingly delayed. The onset of the winter monsoon is, however, earlier in eastern China and this results in an early termination of its rainy season. In relating monsoons to droughts over China, it may be said that the rainy season generally begins during the full establishment of the summer monsoon and consequently conditions in spring are relatively dry over most parts of the country. The summer monsoon reaches South China in March and is followed closely by the rainy season in April. Hence spring droughts are relatively unimportant over this region but become most serious in North and Northwest China. Summer droughts on the other hand are severest over Central China, slightly less so in South China and least important in North China where rainfall reaches its peak during this season. In autumn, droughts are severer in the east than in the west.

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THE ADVANCE AND RETREAT OF MONSOONS IN EAST ASIA IN RELATION TO THE SEASONAL VARIATIONS OF THE WESTERLY CIRCULATIONS*

by

Hsu Shu-ying

In recent years discussions on the advance and retreat of monsoons by meteorologists in China and other countries have often related their activities to the seasonal variation of the general circulation. For example, in his studies on the onset of the summer monsoon over India and Burma, Yin [1] has pointed out in 1949 that the outburst of the Indian monsoon is related to the dissolution of the southern branch of the jet stream complex over the Far East and the rapid retrogression of the mean upper-trough from its quasi-stationary position at 90°E to 80°E. The long-wave pattern undergoes a distinct rearrangement characterized by the appearance of a mean trough to replace the pre-existing mean ridge over Central Siberia. On the other hand, Chakravortty [2] has discussed the onset of the Indian monsoon in terms of westerly disturbances, while the relationship between the southwest monsoon over India and the jet stream was dealt with by Ramaswamy [3]. In 1952 Yeh et al [4] have pointed out that there are two abrupt changes in the general circulation, one in June and the other in October, which are accompanied by marked changes in the circulation pattern over East Asia, in the movement of pressure systems and in the distribution of meteorological elements. They have also pointed out that in early June of 1949 the onset of the southwest monsoon over India coincided with the disappearance of the southern branch of the jet stream. Thus it is obvious that

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* Manuscript completed in September 1959.

the onset of the summer monsoon over India is mutually related to the seasonal variation of the upper westerly circulation.

In 1956 Liu et al [5] have attempted a classification of the natural seasons over China in terms of the weakening, dissolution and reappearance of the westerly jet stream at 500 mb. Yeh, Dao and Li [6] have examined the onset of the mei-yü over China and Japan in relation to the abrupt northward advancement of the easterly and westerly wind belts in June. They studied the onset of the summer monsoon over India and Calcutta in relation to the sudden poleward displacement of the intertropical convergence zone and also analyzed the relationship between the abrupt southward displacement of the westerly and easterly wind belts and the equatorward recession of the southwest monsoon over India and of the intertropical convergence zone.

Of course, their points of view differ from one another to some extent. For example, the results of Liu et al reveal that although the westerlies weaken toward the end of May or at the beginning of June, they persist and do not disappear until mid-July. This means that the upper westerlies only vary in intensity in early June over regions south of the Yangtze, but a radical change in the circulation pattern does not take place until July.

However, from the analysis of upper-air data over some stations, Dao with Chen [7] and Sutcliffe [8] have reported significant changes in the cir- tion pattern at 200 mb in early June which are shown by the replacement rlies by easterlies. The present author is of the opinion that the t discordance in their findings is probably due to the difference in de, longitude and altitude of the regions examined by these investigators.

Since the westerly wind belt retreats northward in phase with the seasonal change from winter to summer, the weakening of westerlies and the appearance of easterlies occur earlier at low latitudes than at high latitudes.

In a discussion on the variation of the pressure field in the surface layer, Kuo and Kao [9] have recently reported that the breakdown of the pressure field of the winter monsoon is first observed at the surface and then at upper-levels. They have noted that in mid-May the continental anticyclone over East Asia has begun to give way to the heat low but the winter circulation still prevails at the upper-levels. In autumn the low-level pressure field of the summer monsoon begins to change in early September. Although the flow configuration at upper-levels begins to undergo fluctuations in early September, the winter circulation pattern does not become established until mid-October. These phenomena seem to suggest that the seasonal change of the general circulation sets in earlier at low levels than at high levels. Furthermore, in practically all cases, a sudden change at the surface will excite a corresponding change at the upper level and vice versa. The differences only lie in the degree of variation.

According to studies on the onset of the summer monsoon over different regions in China, the summer monsoon begins to affect South China in early March and Central China in mid-April. It becomes prevalent in mid-June, attains its maximum intensity in mid-July and begins to retreat from North and Central China in early September and from South China after mid-October. Since China is so vast as to cover a latitudinal range of 49°, it is virtually impossible for the summer monsoon to set in simultaneously over the entire mainland. This also applies to the occurrence of the prevailing and the peak periods. However, the various phases are interrelated to some extent. Thus, at a particular time, the summer monsoon may begin to set in at some place, becomes established in another and may attain its maximum intensity elsewhere. In general a relatively significant march of both the winter and summer monsoon is observed annually in early March, mid-April, mid-June, mid-July, early September and mid-October.

I. ADVANCE AND RETREAT OF MONSOONS IN RELATION TO THE EVOLUTION OF THE PLANETARY WESTERLY CIRCULATIONS

From the above discussion we note that the onset of the southwest monsoon over India is closely related to the abrupt dissolution of the southern branch of the westerly jet. The activities of the monsoons over East Asia in relation to the evolution of the planetary westerly circulation are depicted by Figure 1, in which the full lines denote the average pressure difference for the period 1951 - 1957 between land (110° - 120°E) and sea (135° - 145°E) at 25°N and 30°N. This entity may be used to represent the meridional component of the prevailing wind speed in the surface layer. The broken lines depict the mean 500-mb meridional height difference over the region from 110° to 115°E at 25°N and 30°N for the same period. This parameter is used to portray the variation of the zonal wind speed at 500 mb.

As these curves illustrate the variation of the atmospheric processes with time, we may begin each phase of our discussion from the times of abrupt rise (or fall) and at points where the meridional or zonal wind component becomes zero with a change in sign. It may be noted from Figure 1 that the

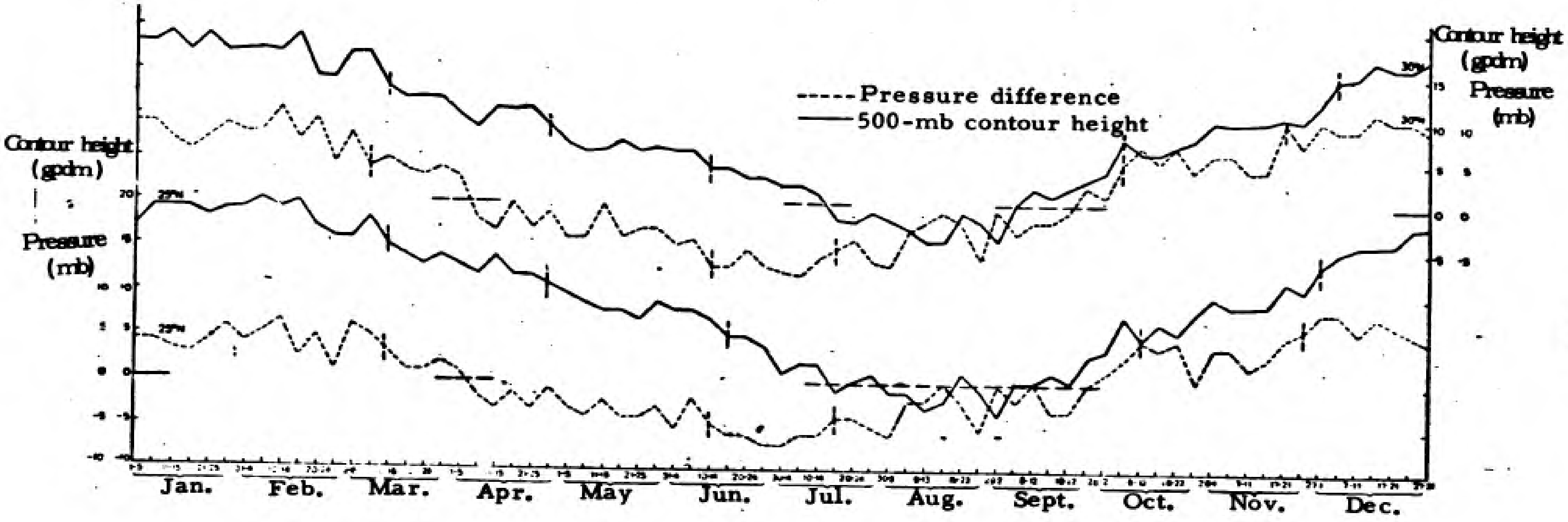

Figure 1

Pentad mean values of land-sea pressure difference and 500-mb meridional height difference over the region from 115° to 120°E at 25° and 30°N.

times of occurrence of the rise or fall in the values of both components are basically in phase. This shows that variations in the monsoon flow in the surface layer as well as the advance and retreat of the monsoons are closely related to the seasonal change of the upper westerly circulation.

(a) "Quantitative" and "Qualitative" Changes in the Planetary Westerly Circulation

In order to gain a better insight into the variations of the zonal circulation, we have prepared 500-mb time-sections of 5-day means of geostrophic westerly wind at various latitudes along 90°E (4 years of data only), over the region 115° - 120°E for the period 1953 - 1958 and 140° - 145°E for 1951 - 1957. The long period time-sections are shown as Figures 2.1 to 2.6 in this paper together with those for 1956 (others omitted). The westerlies along 90°E portray the upper-air cross-section over the central part of the Tibetan Plateau. It may be seen from Figures 2.1 and 2.2 that two distinct jets co-exist north and south of the Tibetan Plateau (500-mb westerlies are reckoned as strong when the mean zonal speed $\geq$ 10 m/sec). The southern jet lies near 22°N and the northern one near 42°N. The former is found to be stronger and more stable in its position than the latter. A significant weakening of the westerlies occurs in the last decade of February and an abrupt decrease of intensity toward the end of April or early May. By the last decade of June, westerlies disappear completely south of 20°N and are replaced by easterlies. In early July the wind flow in the region between 20°N and 35°N turns into easterlies within the lapse of a pentad. Figure 2.2 shows that in 1956 the wind flow in the region south of 30°N had already become easterlies by the end of May with a core of strong easterlies near 27°N by mid-July. However, in 1957 and 1958, easterlies did not appear in this region until the end of June or early July. This shows that the flow configurations are characterized by large inter-annual variations. When changes are taking place in the southern jet, the northern jet usually weakens. The latter is highly unstable in position, and the fluctuation is particularly marked in individual years. Figure 2.2 shows that the northern jet is mobile in character and is apparently a migratory feature from high latitudes. In early June when the wind speed of the southern westerly jet falls light, the average intensity of the northern jet also falls to below 10 m/sec abruptly. After mid-September (early October in 1956), when the subtropical ridge begins to retreat southward, the northern branch of the westerlies intensifies slightly. The southern branch becomes established in mid-October with its intensity gradually increasing. Thus the configuration of the northern and southern branches of the westerlies re-appears. By early December, a further intensification of the southern branch gives rise to the maximum speed of the westerly flow in winter.

The cross-sections of the mean geostrophic westerly wind over East China 115° - 120°E (Figures 2.3 and 2.4) are slightly different from those along 90°E. The zonal westerly wind speed is strongest during the winter months (early December to mid-March). In general, it is about 30 m/sec, which is stronger than the upstream flow near 90°E. In mid-March, the wind speed drops off abruptly to 25 m/sec. By the end of April, a further weakening of the westerlies brings the speed down to below 20 m/sec together with a slight northward retreat of the whole westerly wind belt. By mid-June the breakdown of the westerlies is marked by the appearance of easterlies south of 20°N and finally the prevailing wind flow south of 30°N becomes easterly by mid-July. On average the onset of easterlies occurs at approximately the same time along this meridian as further upstreams, although the easterlies reach higher latitudes in the latter region. After mid-July, westerlies still dominate regions north of 30°N, but they have become very weak with wind speeds of less than 10 m/sec on most occasions. By early September, the westerlies intensify again at 40°N with a core of maximum winds to the north. The westerly wind belt then advances southward and reaches 35°N by the end of September, but the intensification takes place only very slowly. Toward the end of October, the position of the westerly wind belt becomes steady. In early December it advances further southward to 27°N with some intensification and subsequently remains stationary near its deep winter position. The above sequence of events is often more noticeable in individual years. For example, the variations of the intensities and the positions of both the northern and southern branches of the westerlies are much more well-marked in 1956 (Figure 2.4) than shown on the average patterns.

Figures 2.3 and 2.4 show that over the South China coast (south of 20°N) the westerlies are replaced by easterlies in mid-June but re-appear toward the end of September or mid-October. Thus the easterly circulation persists for almost four months. However, at 30°N over East China the easterlies appear in mid-July and are replaced by westerlies in September. In other words the easterlies dominate East China for one and a half months only.

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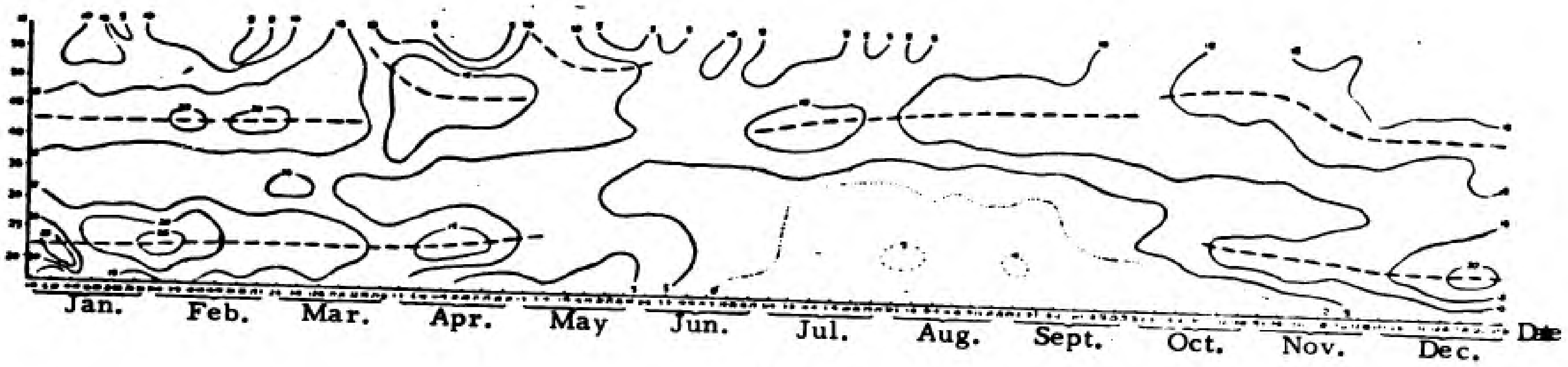

Figure 2.1

Cross-section of pentad mean 500-mb geostrophic zonal wind component over the latitude range 15° - 60°N along 90°E for the period 1955 - 1958.

(Full and broken lines are isotachs of west and east winds in m/sec respectively. Thick broken lines depict axes of strong westerlies.)

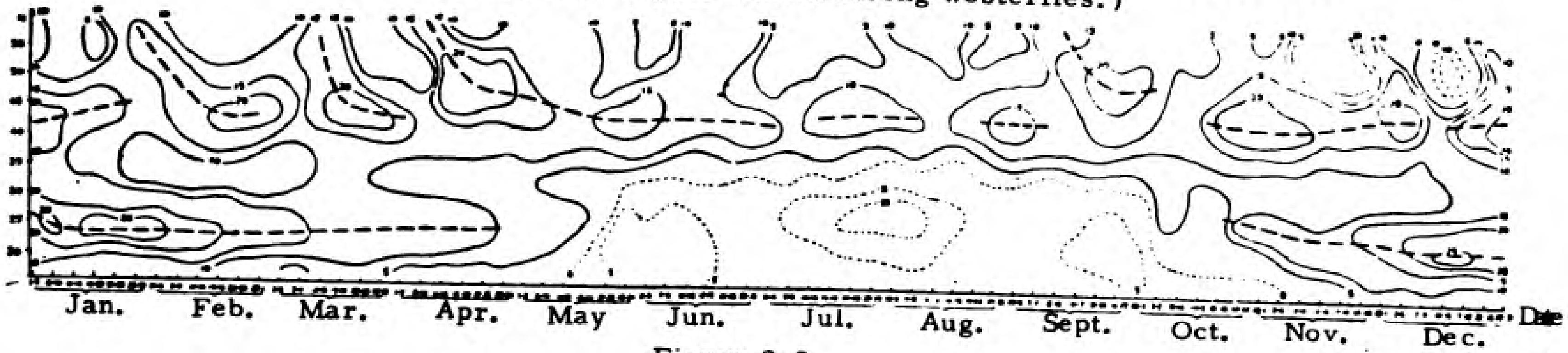

Figure 2.2

Time-section of geostrophic zonal wind component over the latitude range 15° - 60°N along 90°E for 1956.

(Legend same as in Figure 2.1.)

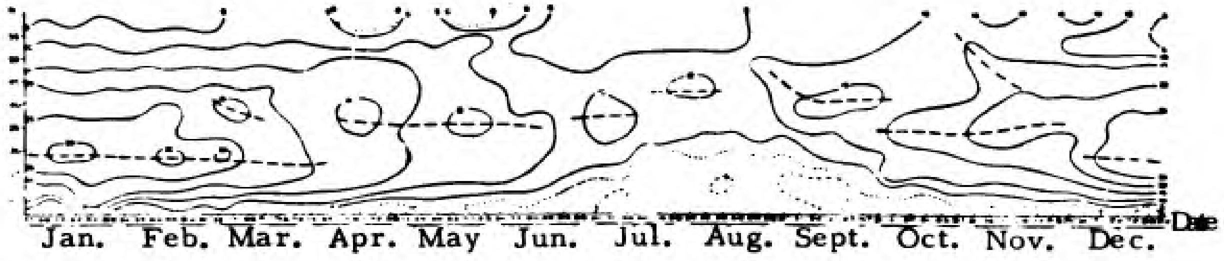

Figure 2.3

Cross-section of long-period pentad mean of 500-mb geostrophic zonal wind component over the region 15° - 16°N, 115° - 120°E.

(Legend same as in Figure 2.1.)

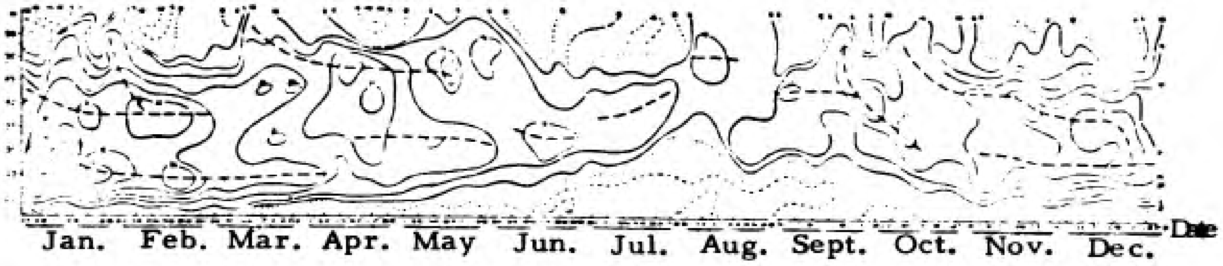

Figure 2.4

Cross-section of pentad mean of 500-mb geostrophic zonal wind component over the region 50° - 60°N, 115° - 120°E for 1956.

(Legend same as in Figure 2.1.)

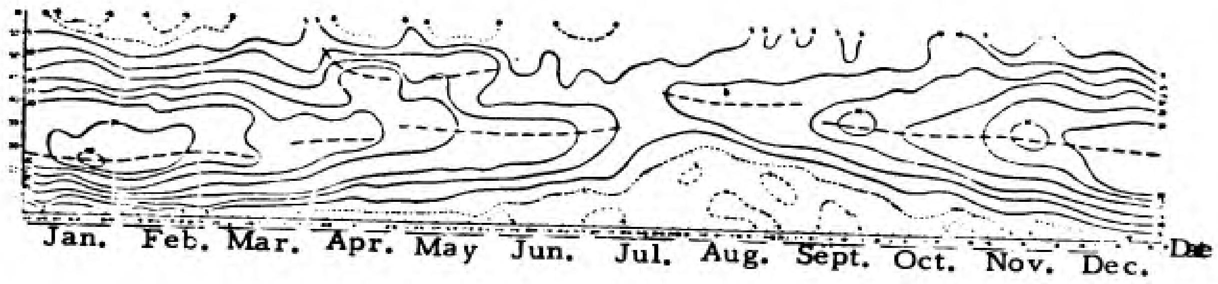

Figure 2. 5

Cross-section of long-period pentad mean of 500-mb geostrophic zonal wind component over the region 15° - 60°N, 140° - 145°E.

(Legend same as in Figure 2.1.)

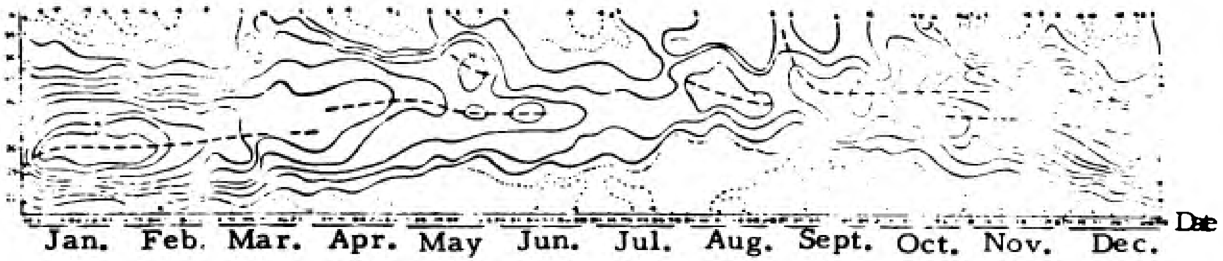

Figure 2. 6

Cross-section of pentad mean of 500-mb geostrophic zonal wind component over the region 15° - 60°N, 140° -145°E for 1956.

(Legend same as in Figure 2.1.)

It may be noted from Figures 2.5 and 2.6 that the variation of geostrophic westerly wind with time over the region 140° - 145°E is not so marked as that over the region depicted by Figures 2.3 and 2.4. The westerlies in the former region are stronger than those in the latter and the position of the strong westerlies is located at higher latitudes with little change in position. The winter jet stream is usually located near 27°N and begins to weaken and move northward to 30°N toward the end of February or in early March. By mid-April, a further weakening of the westerlies is accompanied by a northward retreat to about 35°N. The wind speed decreases most significantly in mid-July. In early August a belt of strong westerlies appears at 42°N, which gradually moves southward. A significant strengthening of the westerlies occurs in early September and is followed by a further intensification in early November. This can be clearly seen from Figure 2.6. In this region, the easterlies appear at low latitudes earlier and disappear later than upstreams. In the three longitude belts under consideration, the easterlies reach the highest latitude near 90°E and the lowest along the east Asiatic coast with those in the region 140° - 145°E in between. Both the northern and the southern branches of the westerlies are very steady in the upstream region. The northern branch is very unsteady in the middle zone and the two branches usually merge into one westerly airstream in the downstream region.

It should be pointed out here that for the several cases under examination changes in the speed of the westerly flow are relatively significant and systematic in regions south of 35°N, while those at higher latitudes are more erratic. Figures 2.3 and 2.4 show that changes in the flow configuration of the southern and northern branches are occasionally out of phase. This is indicated by an intensification of the northern branch when the southern one weakens and vice versa. However, variations in mid-June, mid-July and early September are associated with development on a large scale. In June, changes in the flow pattern are most significant over West China with a very abrupt northward excursion of the air stream but are insignificant in the region 140° - 145°E. There is a general weakening of the westerlies in the three regions under examination. Changes in July are generally more marked and rapid, covering a wide range in latitude. Figure 2. 5 shows that the disappearance of westerlies at 35°N is followed by an abrupt intrusion of easterlies. The prevailing easterlies are mainly restricted to the south of 20°N in early July, but rapidly extend to 32. 5°N during the second and third pentads (5 - 14th). The dissolution of the westerlies and the abrupt northward advancement of easterlies shown in Figures 2. 3 and 2. 4 are, however, not so marked as at 90°E. The strong upper westerlies begin to move southward over China and the southern part of the Western Pacific in early September but do not extend southward over the Tibetan Plateau until October. The intensification of westerlies in October together with the southward retreat of easterlies marks a major change in the circulation pattern. From the foregoing discussion, we may ascertain that there are several changes in the upper-westerly circulations over East Asia in the course of the year. Significant changes in the speed of the westerly current occur in early March, mid-April, mid-October and early December, while the position of the strong westerlies undergoes marked displacement in April and mid-October. In general, there are three radical changes in the properties of the upper-westerly circulation over the region south of 35°N in East Asia and they occur in mid-June, mid-July and early September. The change is most marked in mid-July, less conspicuous in mid-June and relatively weak in early September.

An analysis of the cross-sections of the zonal component of westerlies at 500 mb for the period 1953 - 1958 (charts omitted) shows that generally the onset dates of the various circulation patterns were more marked and better defined for the various individual years than on the average. However, exceptional cases were also noted. For example, in the westerly wind cross-sections for the region 115° - 120°E, a significant weakening of the westerlies occurred during March in five years out of six. In 1954, the only year of exception, weakening of the westerlies was not observed and the position of strong westerlies was found to move southward. In April, the zonal component of the westerly wind underwent marked weakening near the average position of the jet stream in all years. The position of the strong westerlies was found to shift northward by 5°- 10° latitudes in five years out of six but little change was noted in this respect in 1954. In June, a weakening of the westerlies was observed in five years out of six (except in 1957). In addition, the weakening was associated with a poleward displacement of the westerly wind belt in four years, but in 1953 and 1958 the strong westerlies dissolved without displacement. In the six-year period under examination, abrupt northward advance of low-latitude easterlies occurred in 1954, 1956 and 1958, the phenomenon being most pronounced in 1956. A gradual poleward invasion was noted in the other three years. In July, there was a general decrease in the speed of the westerlies in all six years, while a large-amplitude northward excursion of the easterlies was noted in four years. In 1957, the poleward intrusion of the easterlies was gradual in nature and the northernmost latitude reached was relatively low (about 27°N). The easterlies in 1954, on the other hand, spread to about 27°N by the end of June and underwent little change afterward. In September, the westerlies intensified in general, but only exceeded 10 m/sec in four years only. In 1954 and 1957, the westerlies formed in August at middle latitudes and moved southward. In October there was a general intensification of the westerlies and the southern branch of the westerlies re-assumed its winter position. However, there was little change in the intensity of 1953 and 1957. In early December the westerlies intensified very significantly at low latitudes and remained steady near 27°N. The intensification of the westerlies was insignificant in 1955.

Special features were also observed in the other two cross-sections (90°E and 145° - 150°E), but they will not be described in the present paper for the sake of brevity.

(b) Variations of Action Centers in the Surface Layer in Relation to the Advance and Retreat of Monsoons

The monsoon phenomenon is most pronounced in the surface layer and its evolution is closely related to the variations of action centers at sea level. Since the foregoing discussion indicates that the onset dates of the various phases of changes in the upper-westerly circulation are concordant with the advance and retreat of the monsoons, it may be inferred that when changes occur in the upper-air circulation there will likewise be significant development in the surface action centers. This point may be verified by an inspection of the variations in surface land-sea pressure difference and those in 500-mb meridional height difference as shown in Figure 1. A further discussion of the subject will be given in the present section. As Reference [11] is the only available source of information in this connection, the main theme of our discussion must be restricted to the evolution of the formation and dissolution of action centers north of 30°N.

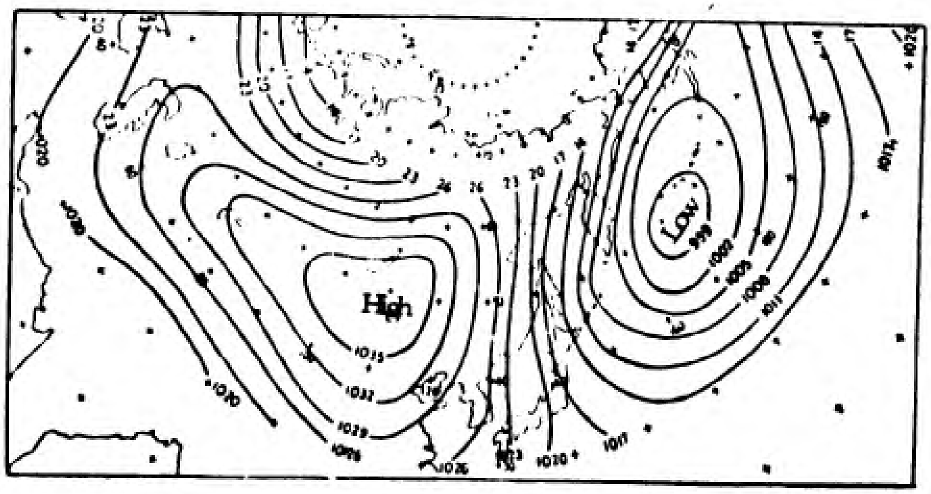

Figure 3.1

Normal surface pressure chart for 21 - 25 January
(Unit: mb).

During the period from about the second pentad of December (7 - 12th) to next March, there is little change in the position of the Mongolian High and the Aleutian Low. Figure 3.1 shows the normal mean sea level pressure for 21 - 25 January and can be taken as representative for the period under consideration. It may be noted that both the continental high and the oceanic low reach their maximum intensity, resulting in a maximum pressure gradient between these systems. The winter monsoon is also most intense.

In March, there is a relatively large change in the pressure field between land and sea. This may be seen from normal pentad charts given in Figure 3.2 (12 - 16 March and 17 - 21 March):

(i) The Mongolian High remains steady near 50°N, 100°E from January onward, moves westward to 50°N, 85°E in the fourth pentad of March (17th - 21st) and rarely retreats to the east again afterward. Following the above displacement, the coverage of the innermost close isobar of 1026 mb near the center decreases in size very rapidly. Prior to this development the central intensity falls only 6 mb from 1035 to 1029 mb in one and a half months from late January to the third pentad of March (12 - 16th).

(ii) Within the lapse of a pentad the intensity of the Aleutian Low weakens by 3 mb with the appearance of two centers, one laying east of the deep winter position (50°N, 170°E) and the other over the southern part of the Alaska Peninsula, which then become common features on the charts.

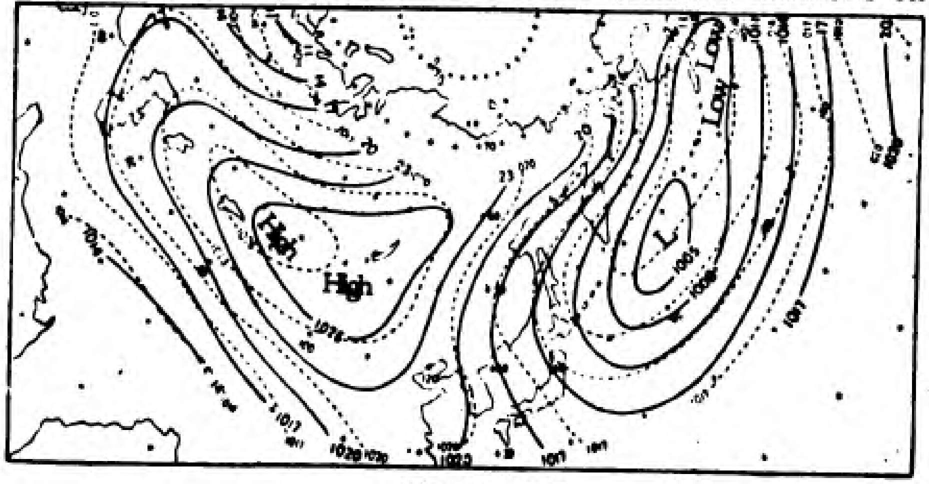

Figure 3.2

Normal surface pressure chart for 12 - 16 and 17 - 21 March.
(Full lines represent isobars for 12 - 16 March and broken lines for 17 - 21 March.)

(iii) The 500-mb trough along the coast weakens considerably and the contours become more zonally oriented.

(iv) Following the above development, a thermal low appears over India and the summer monsoon begins to affect South China. In short, a major change in the general circulation takes place during the first fortnight of March. If strong westerlies prevail before mid-March, then weak westerly circulation may occur relatively frequently thereafter.

In the course of 25 days from about mid-March to 6 - 10 April, the Siberian High weakens by 3 mb from 1026 to 1023 mb, while the intensity of the Aleutian Low undergoes little variation with its central position between 50°N, 160°E* and 54°N, 160°W. The co-existence of two centers within the low complex occurs occasionally. After the second pentad of April (6 - 10th), another development in the surface pressure field is shown by the appearance of an area of low pressure with close isobars near the east coast of Russia with the Aleutian Low displaced eastward to the Alaska Peninsula. From then onward, low pressure systems do not occur in the seat of the winter Aleutian Low. The mean pressure distribution for 16 - 20 April is typical of the circulation pattern for the period under consideration. Figure 3.3 shows that the continental anticyclone is displaced westward by 10° longitude from 85°E to 75°E and weakens by 6 mb from 1023 to 1017 mb in 10 days. However, the intensity of the continental anticyclone remains more or less the same during the 15 days before this occurs. Because of the appearance of a low pressure area over the east coast of Russia, a relatively

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* Translator's Note: 100°E in the original text appears to be a misprint.

small high cell appears over Kamchatka between the two depressions. This is accompanied by a slight intensification of the Aleutian Low instead of weakening.

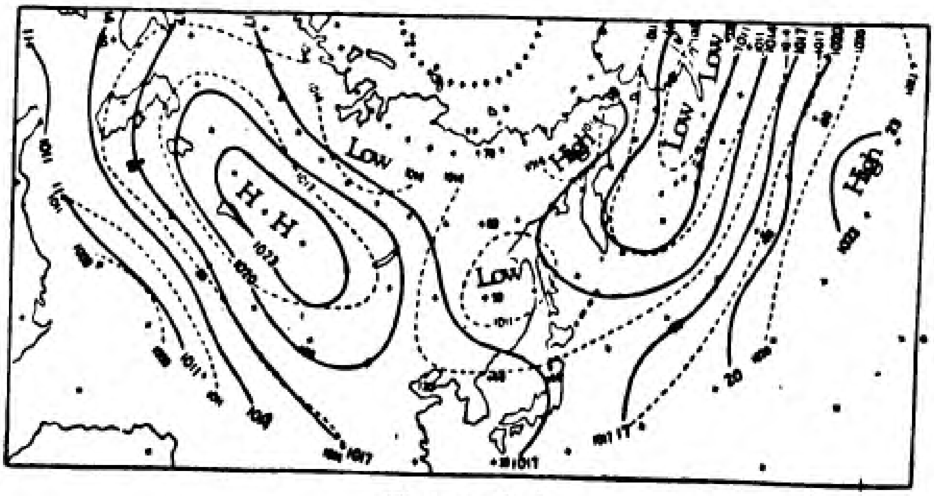

Figure 3.3

Normal surface pressure chart for 1 - 5 and 16 - 20 April.
(Full lines represent isobars for 1 - 5 April and
broken lines for 16 - 20 April.)

During the period from mid-April to the end of May, there is little change in the surface pressure pattern with the only possible exception of the westward movement of an area of low pressure from the east coast of Russia to Northeast China in early May. Due to the westward displacement of this depression and the persistence of the Aleutian Low, the anticyclone over the sea of Okhotsk deepens considerably, which is accompanied by a slight weakening of the continental anticyclone in April.

The continental anticyclone and the Aleutian Low weaken rapidly at the end of May, and are barely detectable by mid-June (Figure 3.4 for 9 - 14th and 15 - 19th). At this time the 500-mb trough along the coast dissolves completely.

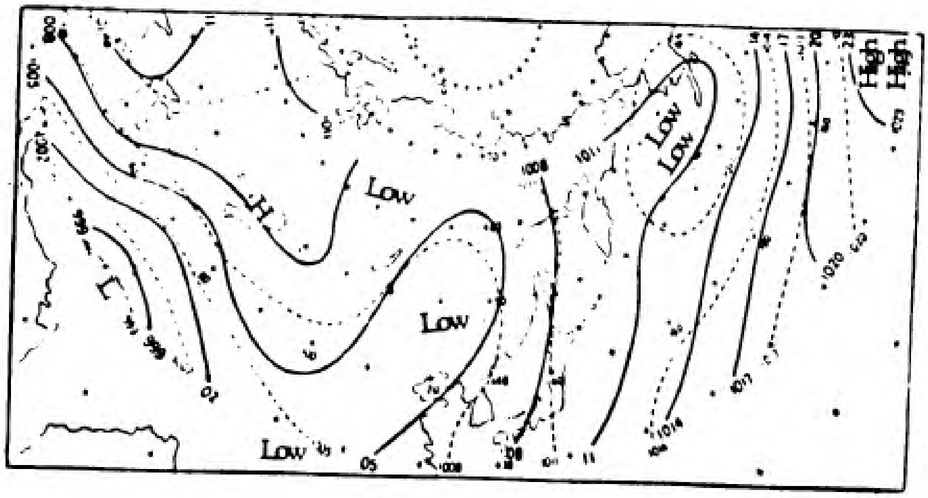

Figure 3.4

Normal surface pressure chart for 9 - 14 and 15 - 19 June.
(Broken lines represent isobars for 9 - 14 June and
full lines for 15 - 19 June.)

The remnants of the "Northeast Depression", the Aleutian Low and the continental anticyclone can still be identified after mid-June but disappear completely by the second pentad of July (5 - 9th) as seen in Figure 3.5 for 5 - 9 July and 25 - 29 July. At this time the Pacific Anticyclone and the Indian Low intensify considerably and extend northward and westward respectively. This is concordant with the establishment of the intense summer monsoon over China with a maximum poleward extent.

From mid-July to the second decade of August, the Pacific High and the Indian Low assume the major role. However, it is unfortunate that the evolutionary processes of these systems cannot be clearly described with the limited size of the charts available. Consequently, it has not been possible to achieve a comprehensive understanding on the evolution of the action centers over East Asia during the advance and retreat of the monsoons. After 20 August, the continental anticyclone, the "Northeast Depression" and the "Bering Sea" Low begin to take shape. The situation becomes very similar to that of mid-April.

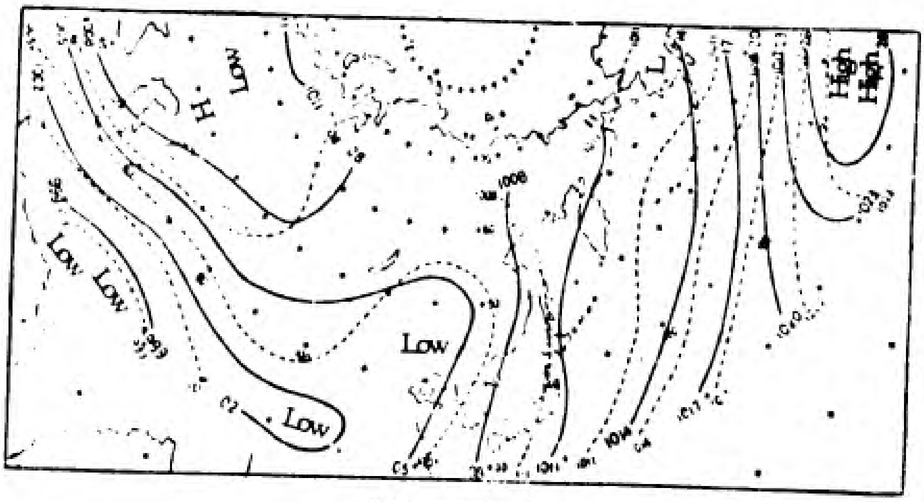

Figure 3.5

Normal surface pressure chart for 5 - 9 and 25 - 29 July.
(Broken lines represent isobars for 5 - 9 July and
full lines for 25 - 29 July.)

In early September the continental anticyclone forms rapidly and remains stationary in its deep winter position. Its central pressure rises by 6 mb in 5 - 10 days during the first decade of September. Before this period the system takes one and a half months from mid-July to the end of August to intensify by 6 mb from 1005 to 1011 mb. Meanwhile an area of low pressure with close isobars develops over the Aleutian Islands, but remains weak in intensity. Figure 3.6 is a pentad mean chart for 29 August - 2 September and 8 - 12 September, which is representative of the situations before and after the change in circulation under discussion. A comparison of the two situations on the chart shows that the intensities of the Indian Low and the Pacific High decrease very rapidly, but the isobaric configuration is basically characterized by the existence of a meridional component in the flow before the end of August. However it begins to assume a zonal orientation in early September and marks the onset of the active winter monsoon.

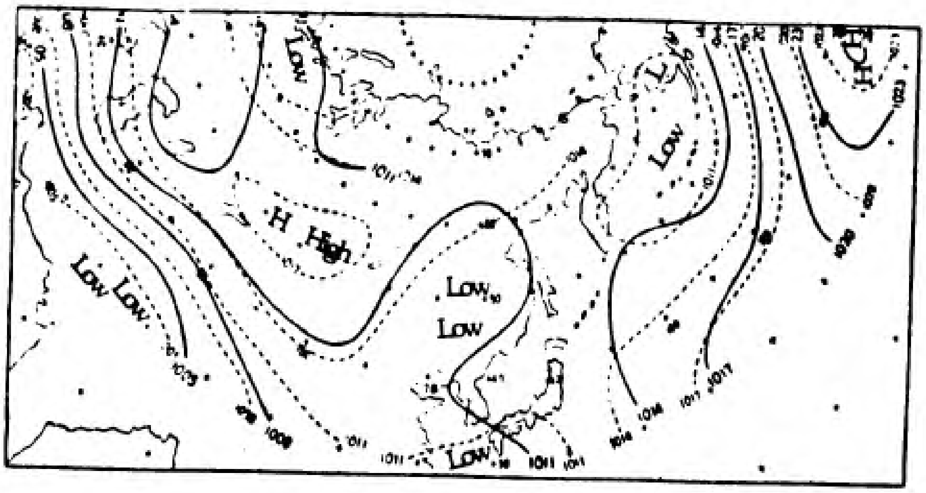

Figure 3.6

Normal surface pressure chart for 29 August - 2 September and 8 - 12 September.
(Full lines represent isobars for 29 August - 2 September and broken lines for 8 - 12 September.)

The pressure pattern and the intensity of the action centers prevailing during the period 8 - 12 September are found to persist until the end of the month. In late September or early October, the continental anticyclone and the Aleutian Low also intensify rapidly. During the second fortnight of October, both action centers deepen by 6 mb. By 8 - 12 October the flow configuration basically resembles the winter pattern with the exception that the center of the Aleutian Low lies some 35° longitude east of its deep winter position (Figure 3.7). From then onward, the Siberian Anticyclone remains in situ but deepens and becomes more extensive until the end of November. However, the system does not attain its deep winter intensity until early December.

The formation, evolution and dissolution of action centers occur in mid-June and early September. This is in phase with the seasonal variations of the upper westerlies and the embedded long-waves and also the adjustment processes in the atmosphere on a large or even global scale. The annual variations of the Atlantic Anticyclone, the North American High and the Iceland Low have, in fact, been examined to confirm this point. For example, significant variations of the intensity of the Iceland Low are also noted in early March, mid-April, early June, September and December. However, the formation, dissolution and variations of action centers in the western hemisphere are generally not so marked as in the eastern hemisphere. It is clearly shown that the major trough over the coastal region of China weakens significantly in March and April, degenerates rapidly in June, dissolves completely in July, re-appears in September, intensifies in October and becomes fully established again in December. This sequence of development reveals that the march and the evolution of the monsoons are explicitly related to changes in the upper planetary circulation.

Although the pentad mean charts used do not contain information for regions south of 30°N, it is known that changes in the activities of the two major low-latitude action centers also occur at about the same time as described in the foregoing sections. However, variations in the activities of these action centers are not so abrupt and well-marked as those at high latitudes.

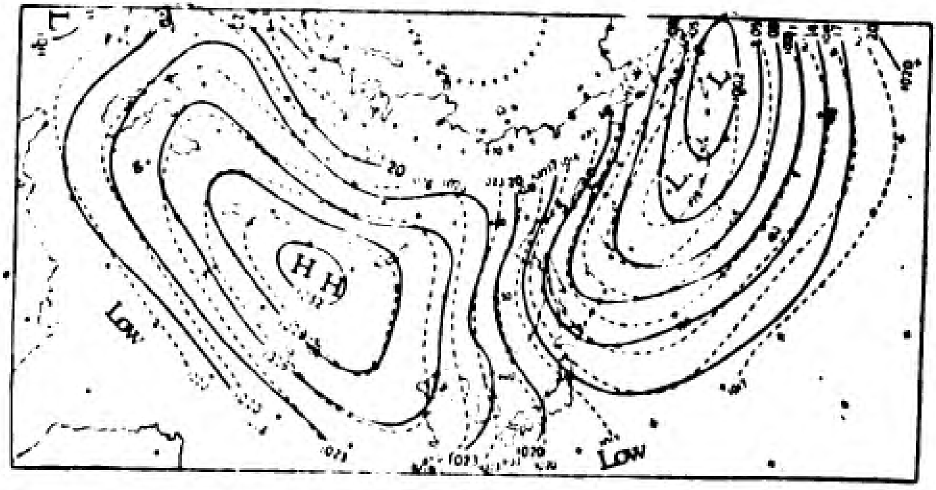

Figure 3.7

Normal surface pressure chart for 22 - 26 November and 2 - 6 December.

(Full lines represent isobars for 22 - 26 November and broken lines for 2 - 6 December.)

Summing up, we note that during a relatively significant advance or retreat of the monsoons over East Asia, distinct variations occur in the intensity, the distribution and the positions of the action centers, namely, the Mongolian High (or the Indian Low), the Aleutian Low (or the Pacific High), the major trough over the east coast and the upper westerly circulation.

II. DISCUSSION OF THE PROBLEM

The effect of the distribution of land and sea can only give rise to a gradual advance and retreat of the monsoons. Since the onset and ending of the monsoons are marked by several particularly distinct stages, the present author is of the opinion that the observed phenomena are related to both "quantitative" variations (variations in the intensity of the westerlies) and "qualitative" variations (displacement of the planetary wind belt) of the upper-air westerly circulation. The present state of our knowledge does not enable us to conclude that the heating effect due to the distribution of land and sea plays no part in causing large "quantitative" variations of the planetary circulation but we may infer that these variations are the result of large-scale development or changes on a global basis. Thus, a discussion on the advance and retreat of the monsoons should be coordinated with the changes of the upper-air planetary flow. Of course, the existing land-sea distribution plays a part in modifying the variations of the planetary or hemispheric flow. In other words, in studying the march of the monsoons, attention must also be drawn to the regional characteristics of the seasonal variation of the general circulation. When "quantitative" and "qualitative" changes of the westerly circulation occur, the advance or retreat of the monsoons over the mainland of China is characterized by a large horizontal amplitude. Three typical examples of "qualitative" changes in the westerly circulation are presented for discussion in the ensuing sections.

Many writers [4, 6, 7] have pointed out that a large-scale abrupt change of the general circulation occurs in June and is characterized by the dissolution of westerlies and the appearance of easterlies in East Asia or over South China and the Tibetan Plateau. This is accompanied by a significant weakening of the strong westerlies over East and North China and represents a period of radical change of monsoon activities. The pressure gradient between land and sea shows that surface northerlies are replaced by southerlies over South China to accord with the march of the pronounced summer monsoon to its maximum intensity. At this time, the rain season begins over Central China as the summer monsoon extends northward and becomes fully established in this region. Precipitation also increases over North China to mark the onset of the summer monsoon with a notable fall in the pressure difference between land and sea. It may be seen from the 500-mb pentad mean charts over East Asia for the period 1951 - 1957 that the subtropical ridge moves rapidly from 15°N to 20° - 25°N in June and the westerlies over the South China coast are replaced by easterlies. The dissolution of the jet stream and a substantial weakening of the westerlies also bring about a re-arrangement of the long-wave systems over the northern hemisphere. Both the Aleutian Low and the Mongolian High dissipate (absence of close isobars) from the surface with only their remnants on the subsequent pentad charts. Following the dissolution of the major upper-air trough over the coast, the rain belt over China advances to the Yangtze basin within the next one or two pentads. Figure 4 [12] shows the configuration of the 556-gpdm contour at 500 mb over the northern hemisphere for the period May - October. An examination of the development from May to June reveals that a major trough is found near 40°E and another along the east Asiatic coast near 125°E in May. As the coastal major trough moves away eastward in June, its seat is occupied by a ridge and the re-arrangement is accompanied by the dissipation of the trough near 40°E and the formation of another major trough near 90°E. This evolution undoubtedly represents a large-scale change in the circulation pattern. However, the variation in the circulation over the region south of 25°N is a qualitative one since this is reflected by the replacement of westerlies by the easterlies to mark the domination of the subtropical anticyclone.

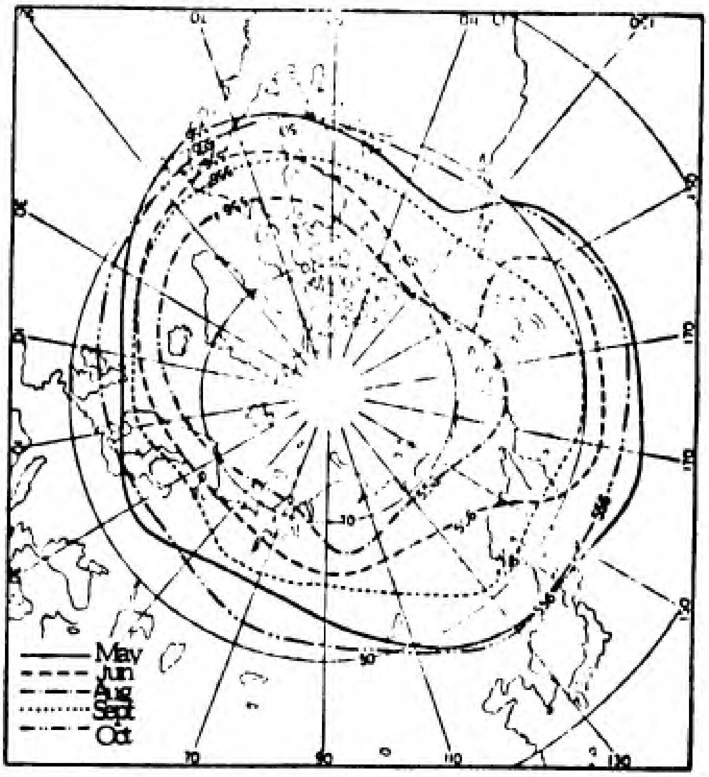

Figure 4

Distribution of the 556-gpdm contour at 500 mb over the northern hemisphere for May, June, August, September and October.

The westerly wind belt moves rapidly northward to its northernmost position in mid-July, and this is accompanied by the abrupt poleward advance of the subtropical anticyclone. Thus the Yangtze basin and the regions south of it come under the control of the subtropical ridge with easterly circulation aloft, and the weather becomes fine, sunny and stable. However, westerly flow still dominates the Hwai Ho basin and the regions to its north although its intensity continues to decrease. By this time, the remnants of the Mongolian High and the Aleutian Low as a legacy from June disappear completely. From mid-July to 20 August, the continental heat low attains its maximum growth while the Pacific Anticyclone becomes fully developed and is located in its northernmost position. During the period the pressure gradient between land and sea attains a maximum to mark the peak of the summer monsoon and south China is not only invaded by the pronounced southeast winds of the monsoon but is also affected by the southwesterlies from time to time. The equatorial convergence zone moves poleward to its northernmost position in August. This is the time when the summer monsoon reaches its peak intensity over Central China where hot and dry weather prevails, and begins to become pronounced at higher latitudes in North China. Thus the rain belt advances from the Yangtze basin to north of the meandering region of the Yellow River and its arrival marks the onset of the rainy season over North China and the Northeast Provinces. The summer season is also said to commence over China in terms of the natural seasons described by Liu [5]. The above sequence of events is a consequence of the second "qualitative" change in the westerly circulation over regions south of 35°N.

The pronounced summer monsoon circulation persists up to the end of August, and by early September the upper-air flow begins to prepare its way for the advent of winter, which is associated with an intensification and southward displacement of the westerly wind belt as a whole and the replacement of easterlies by westerlies over the Yangtze basin and the neighboring regions. As the cold anticyclone in the surface layer moves rapidly southward, a major part of the mainland of China comes under the influence of the winter monsoon. The development marks the third "qualitative" change in the circulation pattern, and the Mongolian High, the Aleutian Low and the coastal upper-air major trough become established rapidly after this phase.

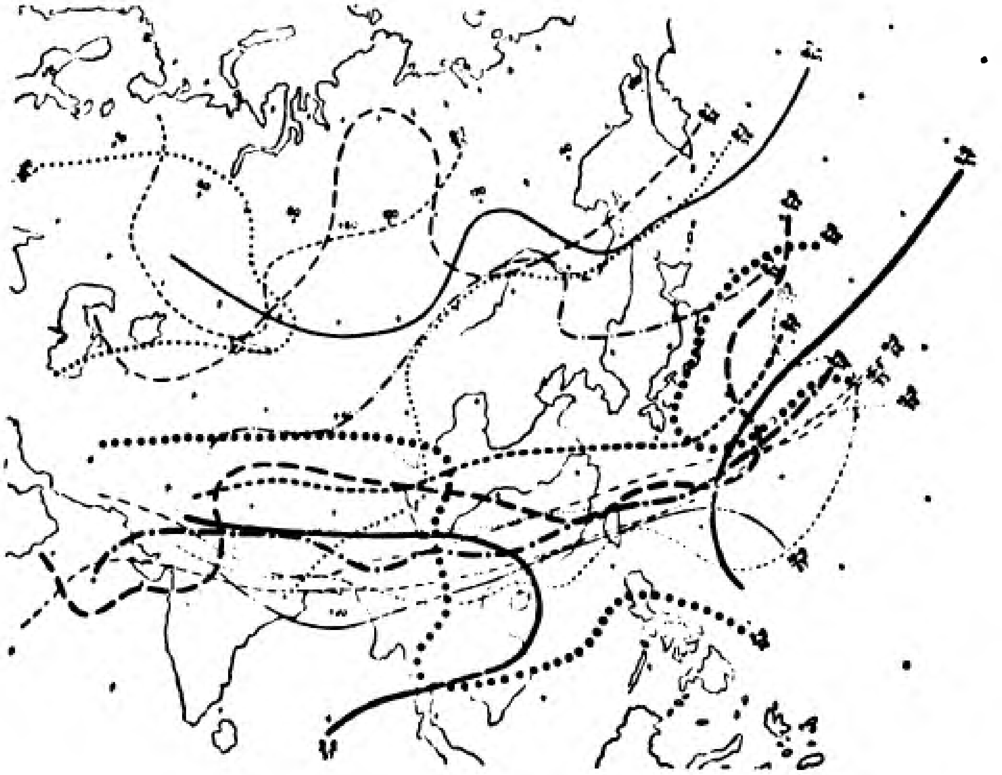

Figure 5

Zero-isopleth of monthly isallohypse at 500 mb for July - October for the period 1953 - 1957.

(Legend: Positive and negative anomalies are located south and north of the zero-isopleth respectively. The denominator of the fraction at the end of each isopleth denotes the year in abbreviation and the numerator defines the months for which the difference in monthly mean contour heights was taken; e.g., $\frac{9-8}{57}$ means $\frac{\text{September - August}}{1957}$.)

-.-.- 1953; ——— 1954 --- 1955; ... 1956; —— 1957.

An examination of the geographical distribution of 700-mb troughs and ridges (charts omitted) summarized by Klein and Winston [13] reveals abrupt changes during the transition from summer to winter. For example, the maximum frequency of occurrence of troughs lies in the middle- and high-latitude regions near 110°E over the East Asiatic Continent, while the occurrence of ridges is relatively frequent over the east coast of the mainland. However, the mean maximum frequency of occurrence of troughs and ridges shows a major change in September. Troughs appear over the coast which has previously been the seat of ridges, while the major trough over the East Asiatic Continent is replaced by a ridge from the west. These changes agree with the variations of the configuration of the 556-gpdm contour at 500 mb from August to September as shown in Figure 4. Although similar features are also observed from October to November, the differences are not so marked as those from August to September. Thus early September may appropriately be taken as the onset of a period of "qualitative" change in the general circulation.

It may be seen from charts depicting the monthly variation of 500-mb contour heights that a positive height anomaly occurs over the Eurasian Continent after March with a negative one from the beginning of August. Figure 5 shows the configuration of the zero-isopleth of the monthly variation of 500-mb contour heights from July to October for the period 1953 - 1957. Positive and negative height anomalies lies south and north of the zero-isopleth respectively. The denominator of the fraction at the end of each isopleth denotes the year and the numerator the months in numerical representation over which the height difference is computed. The figure shows that a negative height anomaly begins to appear in early August over the Eurasian Continent, mainly over the northern part of the block (Siberia and Outer Mongolia) with a positive height anomaly over China. By September, the zero-isopleth moves rapidly southward, reaching the middle and lower basins of the Yangtze or regions to the south of it. (The thick full lines in Figure 5 represent the zero-isopleths of height changes computed by subtracting the mean monthly contour height of August from that of September.) This indicates that the maximum variation of the circulation pattern over middle- and high-latitude regions in China occurs in September, which is a month of major transformation in the circulation pattern over these regions. However, changes at low latitudes are not significant at this time of the year. In terms of the advance and retreat of the monsoons, the period represents a phase of radical change in the flow pattern because the winter monsoon begins to extend its influence over East Asia in September and spreads rapidly southward. The re-appearance of the winter monsoon and its subsequent intensification are also closely related to the abrupt changes of the circulation over middle and high latitudes during this period.

In the light of the foregoing discussion, the author is of the opinion that the advance and retreat of the monsoons are not only related to the formation, intensity, position and dissipation of the four action centers over East Asia and the variations in their properties, but are also affected by the several distinct changes in the upper-air westerly circulation.

ACKNOWLEDGMENT

The author wishes to thank Kao Yu-hsie for his valuable guidance and constant encouragement during the preparation of this paper.

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THE ONSET AND TERMINATION OF THE RAINY SEASON IN RELATION TO THE ADVANCE AND RETREAT OF THE MONSOONS IN EAST ASIA

by

Kao Yu-hsie and Hsu Shu-ying

From a study of the seasonal displacement of the mean pentad rain belt [1] (which is the axis of isohyet maximum of mean pentad rainfall), we have defined the dates of onset and termination of the rainy season over the various regions in China. Their relationship with the advance and retreat of the monsoons is also examined. Due to the lack of homogeneity in the period of data analyzed, certain inconsistency in the movement of the rain belt is observed. However, the general tendency of the displacement of this system is closely related to the activities of the monsoons.

I. DATES OF ONSET AND TERMINATION OF THE RAINY SEASON

The rainy season sets in earlier in South than in North China and the difference in the time of onset amounts to almost three months. The season also commences about one month earlier in the east than in the west. The northward displacement of the rain belt is characterized by two "jumps" and three "halts". The jumps are associated with the abrupt seasonal change of the westerly circulation from winter to summer while the halts give rise to the rainy season in South and North China and the mei-yü period over the Yangtze basin (Figure 1). After the first jump of

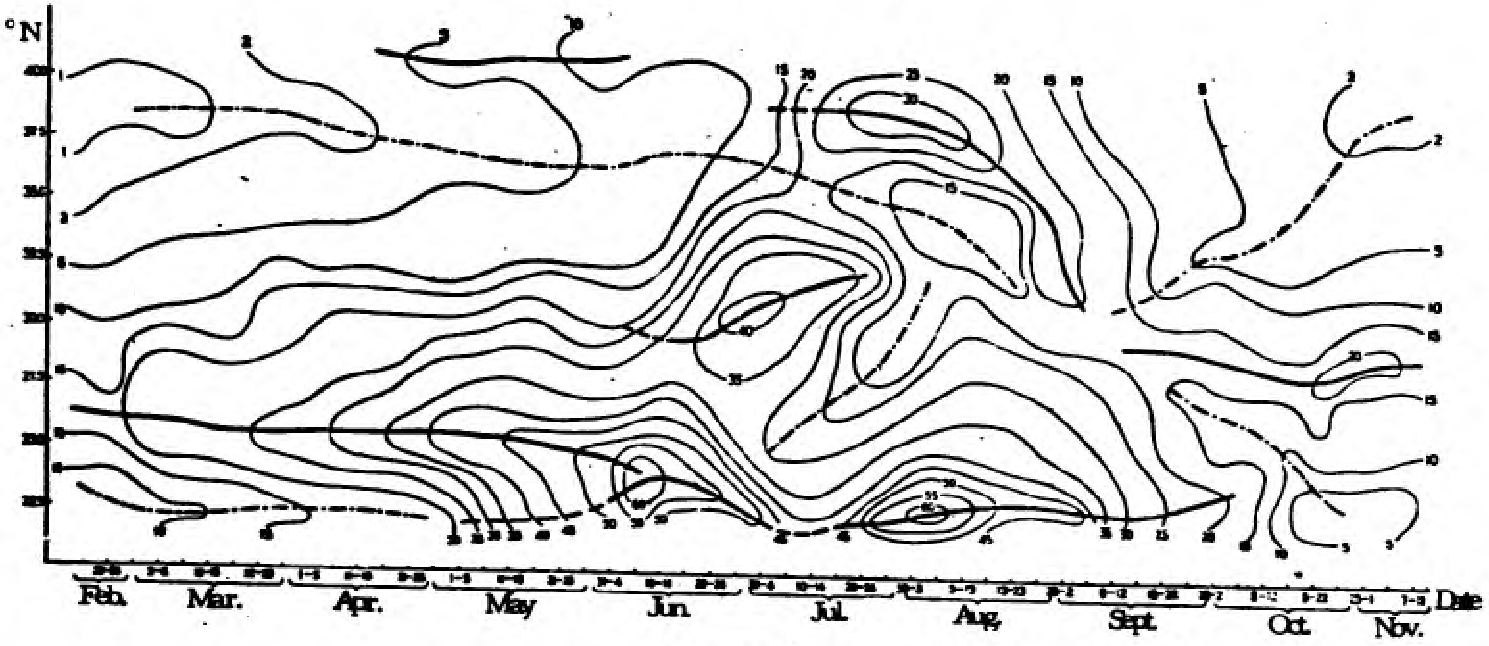

Figure 1

Distribution of mean pentad rainfall at various latitudes along 110° - 115°E and 120°E.

(Legend: Solid lines are isohyets in mm; thick solid lines are axes of maximum rainfall.
Thick broken lines mark regions of rapid increase in rainfall and dash-dot lines represent regions of little rain.)

the rain belt from South China to the Yangtze basin during the last decade of June, a second heavy rain belt appears over the South China coast, which also moves northward with the first and reaches its northernmost position near the Nan Ling Ranges around mid-August. The rainy season terminates about 20 days earlier in North than in South China and lasts longer in East China and the Szechwan basin than in Central China.

(a) Characteristic Features Associated with the Seasonal Displacement of the Pentad Rain Belt

During the period from mid-October to mid-May, the heavy rain belt fluctuates between the north of the Nan Ling Ranges and the south of Yangtze. (The thick line in Figure 1 represents the maximum rain belt during the various periods.) This may be due to the persistence of the upper jet stream over the region. When a second heavy rain belt appears near the South China coast in mid-May, the northern winter heavy rain belt shifts toward the south. (The thick line in the south of the figure indicates a rapid increase of rainfall.) During a later stage, the rain belts merge and begin to move north as shown in Figure 2.1 which is based on the mean pentad rainfall computed from long-period records. The heavy rain belt is designated by the line joining places of maximum rainfall although heavy rain does not necessarily occur along a line. Before the third pentad of June (10th to 14th), the rain belt lingers over the south of the Nan Ling Ranges and the first northward jump takes place in the fourth pentad (15th to 19th). During the next five days, the rain belt

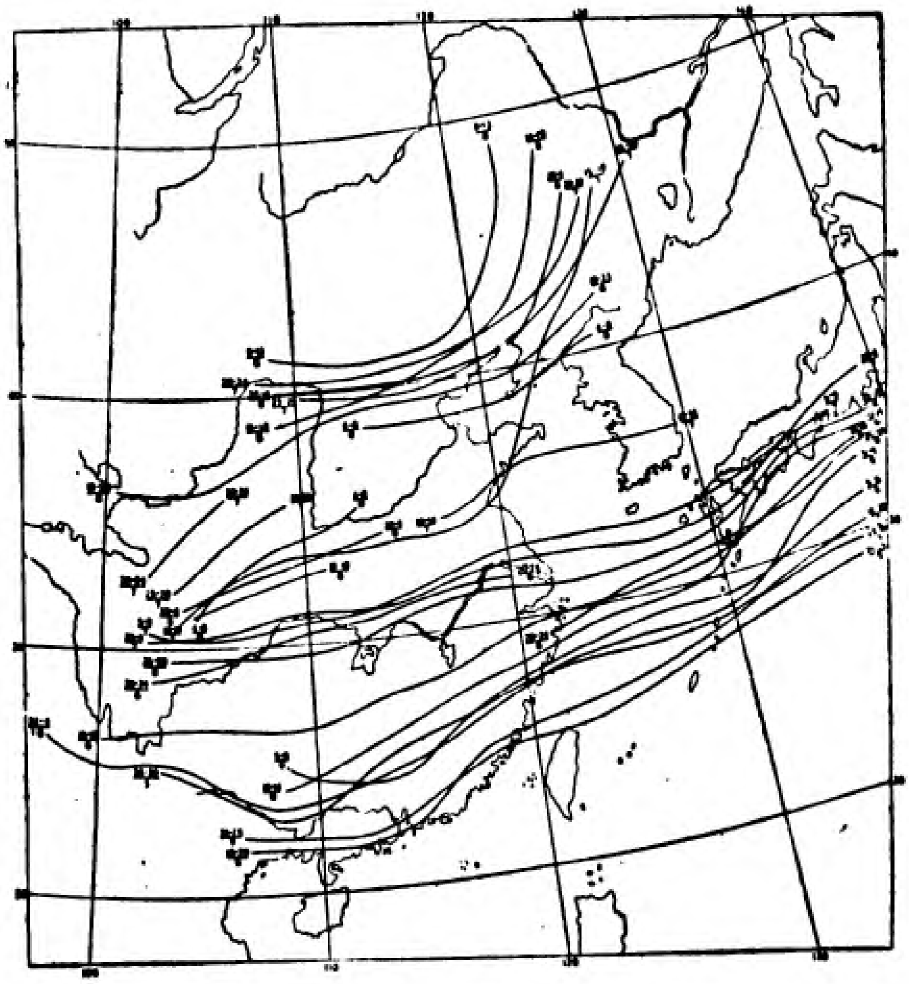

Figure 2.1

Northward displacement of the mean pentad heavy rain belt. (Numerator at the end of each line denotes the days and the denominator the month.)

moves across the lake basins to the Yangtze with its eastern edge passing over the south of Japan. From the last decade of June to the first decade of July, the belt persists near the banks of the Yangtze and over Japan. After the third pentad of July (10th to 14th), another jump is observed and the system passes rapidly over the Hwai Ho and the lower Yellow River basins to the River-bend region and Northeast China. By mid-August, the rain belt reaches its northernmost position and begins to travel southward. The belt moves across the Yellow River basin during the sixth pentad of August (24th to 28th) and the rainy season over North China comes to an end. The system arrives at the South China coast by early September and the rainy season terminates over the whole country as it continues to move farther away from the mainland during the following fifteen days. The rapid movement of the rain belt from the end of August to early September is known as the "third abrupt change". The rainfall after this period becomes more meridionally distributed with more rain over East China and the Szechwan basin than over Central China.

If the arrival of the belt of heavy pentad rainfall is taken to represent the onset of the rainy season (the period of rapid increase in rainfall excluded), then the rainy season starts around mid-May over the South China coast while the mei-yü begins to affect Japan in mid-June. The onset of the rainy season occurs over the Yunnan-Kweichow Plateau, South Kwangsi, Nan Ling and the border region between Fukien and Chekiang in early June, over the Yangtze during the fourth pentad of June (15th to 19th), over the Yellow River basin and part of the central plain after the second pentad of July and over North China and the region north of the Yellow River during the third pentad of July (10th to 14th). It takes five pentads for the rain belt to move 350 km from the South China coast to Nan Ling with an average speed of 70 km per pentad. Although the Yangtze basin is a little more than 280 km from Nan Ling, the onset of rainy season differs only by one pentad. The rain belt at the northern bank of the Yangtze first travels about 30 km per pentad, but accelerates to 300 km per pentad between the north of Yangtze and the northern part of North China. In high summer, the onset of the rainy season is delayed by one pentad for a separation of every 50 km over the stretch from North China to the northernmost limit of the rain belt. Thus three "stagnant periods" are noted from mid-May to early June, from mid-June to early July and from mid-July to mid-August during which the northward displacement of the rain belt is only very gradual in character. The three jumps occur during 10 - 20 June, 10 - 20 July and between 16 August and early September, the last being associated with the southward retreat of the rain belt.

(b) Duration of the Rainy Season

The duration of the rainy season over the various regions of China may be determined by examining the movement of the rain belt shown in Figure 2.1. The mei-yü period lasts for about one month from early June to early July in Japan while in China, it represents the first stage of the rainy season and lasts for 25 days from mid-May to early June in the south and a little more than 20 days from mid-June to early July over the Yangtze basin. The duration of the rainy season is about 40 days from mid-July to late August in North China and north of the Yellow River, about two and a half months from mid-June to late August in the neighborhood of the Yangtze basin and about two and a half to three months in the coastal region of South China. If summer is taken to be the period between the northward and southward passages of the heavy rain belt, then the season lasts for more than 80 days from early June to early September in regions south of the Nan Ling Ranges, about 50 days from early

July to the end of August over the Yangtze basin and only about 30 days from mid-July to mid-August in North China. The above definition of the summer season is concordant with the findings of Liu [2] on the duration of the natural summer season and also with our early criterion for determining the peak period of the summer monsoon in terms of monsoon activities.

If we include the period of rapidly increasing rainfall ahead of the rain belt as shown by the broken line in Figure 1 or the period associated with the jumps of the rain belt as the rainy season, then its duration will be correspondingly increased to 160 days from late April to mid-October over coastal China, nearly 90 days from the first decade of June to early September over the Yangtze basin and about 50 days from early July to the end of August in North China. This agrees better with common experience in defining the rainy season over China.

If the period separating the two major rain maxima at each latitude is defined as a dry spell (see Figure 1), then this singularity is most marked south of 27°N and lasts for nearly a whole month from the last decade of June to the last decade of July. However, the degree of "dryness" during this spell decreases southward so that the rainfall over the coastal region of South China is considerably higher than that near Nan Ling. Between 27° and 32°N, the dry period persists for more than one month from the last decade of July to the middle or end of August. The dry spell becomes insignificant near 35°N and is not observed north of this latitude.

Figure 1 also reveals several other features. The heavy rain belt moves from north to south toward the South China coast in May but a rapid increase in rainfall is already observed over the region at the end of April. Two abrupt northward movements of the rain belts are noted in mid-June and mid-July but as the belt reaches the Yangtze after its first jump, a second heavy rain belt forms along the South China coast (shown by the dotted line in Figure 2.2). The rainy season terminates at the end of August over the northernmost region in North China as the heavy rain belt begins to retreat southward in the middle of the month. The belt shows two rapid displacements during its southward journey; it takes only two decades to move from North China to the Hwai Ho basin and another one to reach the South China coast. After the first decade of September, the only significant heavy rain belt is found over this southern region.

The rainy season begins earlier in the east than in the west but terminates earlier in Central China than in regions east and west of it. The dates of the onset of the rainy season over the various regions in East Asia are given in Table 1.

TABLE 1

The dates of onset of the rainy season over the various regions in East Asia. (Dates for the western regions taken from Reference [3].)

Date °E / °N	75°	80°	85°	90°	95°	100°	105°	110°	115°	120°	125°	130°	135°	140°
22°.5	10—14/6	10—14/6	10—14/6	5—9/6	30/5—4/6	21—25/5	21—25/5	16—20/5	6—10/5	6—10/5	—	—	—	—
25°.0	15—19/6	10—14/6	10—14/6	5—9/6	30/5—4/6	30/5—4/6	30/5—4/6	10—14/6	5—9/6	16—20/6	—	—	—	—
30°.0	30/6—4/7	15—19/6	10—14/6	30/5—4/6	—	5—9/6	25—29/6	20—24/6	15—19/6	15—19/6	15—19/6	10—14/6	—	—
35°.0	—	—	—	—	—	—	20—24/7	20—24/7	15—19/7	4—10/7	4—10/7	4—10/7	30/6—4/7	25—29/6
40°.0	—	—	—	—	—	—	—	20—24/7	20—24/7	20—24/7	15—19/7	—	—	—
45°.0	—	—	—	—	—	—	—	—	—	9—12/8	15—19/7	15—19/7	—	—

Remarks: 10-14/6 given in the table means 10-14 June.

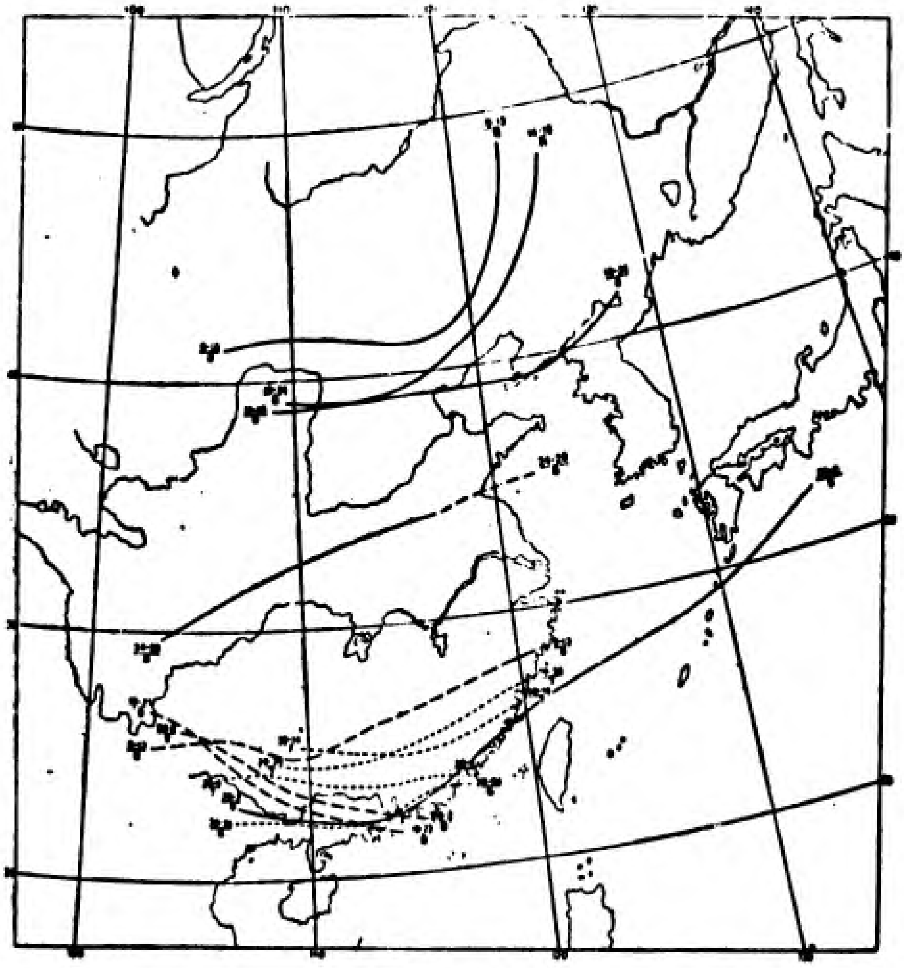

Figure 2. 2

The advancement and retreat of the mean pentad heavy rain belts.

(Legend: Solid lines are the positions of the heavy rain belt during its southward movement between August and early September; dotted lines are the positions of the second heavy rain belt over South China during its northward advancement and broken lines are the positions during its southward retreat. Numerator at the end of each line gives the days and the denominator the month.)

It is clearly seen from Table 1 that the march of the rainy season is more systematic in the region bounded by 22. 5°N and 35°N. The onset of the season is earlier in the more eastern longitudes. Thus rain begins earlier along the 120°E meridian than over the western coast of the Indian Peninsula by one month and a similar difference is noted for regions in Japan near 35°N and South Kansu. However, as the plateau region at the above latitudes in China is characterized by topographical discontinuities the onset of the rainy season is earlier over the eastern side of the highlands. An example is given by the Yunnan Plateau where the season commences earlier by 10 days than over South Hunan and half a month than over the western coast of India. The gradual retardation of the ending of the rainy season to the west of the Central China region has already been discussed in a previous study [4].

The 105°E meridian may therefore be taken as a boundary which separates regions with different rainfall regimes. In the western region which comprises Southwest China, the rainy season is similar to that over India. In the eastern region, the rainy season is determined by the seasonal position of the polar front east of the Tibetan Plateau and the activities of the southeast monsoon. It should be pointed out that the rainy season over the Tibetan Plateau is controlled by the southwest monsoon and its onset and termination are therefore in phase with those of India. Rain first sets in over the southeastern part of the plateau and most places of India when the summer circulation pattern becomes established after the breakdown of the upper winter planetary easterly over East Asia and the disappearance of the westerly jet stream over the southern part of the continent, and the southwesterlies also advance rapidly northward. As the Indian Low moves westward, rain begins to affect the western part of the plateau. However, to the east, the seasonal displacement of the polar front remains the main factor in controlling the march of the rainy season. Thus, the middle and lower Yangtze basins and part of Japan continue to be influenced by the upper westerly wind belt after the dissolution of the jet stream over southern Asia. The polar front still frequently moves south to the neighborhood of the Nan Ling Ranges during the time resulting in a delay in the onset of the rainy season in comparison with regions further west (near 95° - 105°E). The rapid southward movement of the polar front over the eastern plain in China at the end of summer, on the other hand, leads to an early termination of the rainy season over Central China. The late ending of the season over West and East China is due to the blocking effect of the plateau to the west and the influence of the ocean to the east.

It may further be noted that the rainy season over Shensi, South Kansu and North Szechwan including parts of Hupeh and Honan not only sets in later than in the eastern provinces but their dates of onset are also more erratic (see Figure 2.1). It is a synoptic experience that the rain does not arrive at the Szechwan region until the mei-yü terminates over the Yangtze basin and the rain belt moves to North China. This is due to the fact that the region is situated near the mean upper-air trough at 110° - 115°E in high summer and is frequently affected by the intrusion of cold air. Yeh and others [5] have described the significance of this phenomenon in their study on cold-air activities in China.

Figure 2.1 also indicates that the northern limit of the rain belt is located near 42°N over North China. When the rain belt reaches this position between 9 and 13 August, the second pentad heavy rain belt associated with the equatorial trough also arrives at its northernmost limit extending from Wenchow across

Chian and Kweilin to Kunming. If the two rain belts are considered to be associated with the southwest monsoon and the monsoon from the Pacific, then these positions may be taken to represent the northern limits of their influence.

II. THE ONSET AND TERMINATION OF THE RAINY SEASON IN RELATION TO THE ADVANCE AND RETREAT OF THE MONSOONS

The march of the monsoons over China has been studied by Tu and Hwang [6] more than 10 years ago in terms of variations of the wet-bulb temperature. After the liberation of China, upper-air observations have increased considerably and they have been used by many investigators in studying the onset and termination of the mei-yü in relation to the outburst of the Indian summer monsoon. In our earlier study [7], we have determined the dates of the advance and retreat of the monsoons by means of a composite analysis of the five-day changes of the land-sea gradients of the various meteorological elements and variations in the upper-air westerly circulation. A comparison of the dates of onset and termination of the rainy season with those of advance and retreat of the monsoons obtained by the various authors reveals several interesting features which will be discussed below.

We shall first compare our dates of onset of the summer monsoon over East China with those determined by Tu [6]. It is found that: (a) the dates by Tu are generally earlier by one month; (b) the northern limits of the Pacific and equatorial air masses determined by Tu are more to the north of the corresponding rain belts presented by us; (c) the dates of the retreat of the monsoons are in fair agreement although the dates determined by us are, in general, slightly later. The two characteristic jumps of the monsoon during its northward advancement and the earlier onset of the monsoon in the eastern region are well recognized in both studies though a slight difference in the time of occurrence is noted.

Table 2 shows that the rainy season sets in after the full establishment of the monsoon. The time lag as shown by Rows I and III in the table is 20 days for South China and 10 days for Central and North China. The termination of the rainy season, on the other hand, is generally earlier than the retreat of the monsoon and the difference in time is two weeks for North China, 20 days for Central China and 25 days for South China. Thus the lag between the ending of the rainy season and the retreat of the monsoon increases toward the south. However, if the periods of rapid increase (decrease) in rainfall ahead (behind) the rain belt are included as the rainy season, then the results shown in Rows II and III are in close agreement with each other.

The difference between the dates given in Rows I and III is probably due to the fact that the various phases of monsoon activity are only defined by large-scale variations in the flow pattern and pressure systems supplemented by local changes in meteorological elements and that the transitional periods of abrupt developments have not been isolated; whereas in determining the dates of onset of the rainy season, the period of rapid change in rainfall has been excluded. Since the period of steady rainfall in the heavy rain belt is always preceded by a finite interval with rapid increase in rainfall, the rainy season determined by the movement of the heavy rain belt in fact refers more appropriately to the period between the beginning and ending of the period of steady rainfall. As an example, the mei-yü over the Yangtze basin sets in during the first decade of June but does not increase to a steady rate of fall until the second decade of the month. Thus, it is purely a matter of definition that gives rise to the observed differences.

TABLE 2

Comparison of dates of onset and termination of the rainy season with the advance and retreat of the monsoons

	Dates of onset and termination		
	South China	Central China	North China
I. Rainy season (as defined by the dates of the two passages of the heavy rain belt)	16/5- 2/9	20/6-24/8	20/7-19/8
II. Rainy season (transitional period included)	30/4-21/9	9/6- 5/9	10/7-28/8
III. Pronounced summer monsoon	26/4-27/9	10/6-12/9	10/7- 2/9

III. THE ONSET AND TERMINATION OF THE RAINY SEASON IN RELATION TO THE SEASONAL VARIATION OF THE WESTERLY CIRCULATION

In discussing the problems on the general circulation during June and October, meteorologists in China often emphasized the relationship between the onset of the mei-yü and the abrupt change in the flow pattern of June [8]. It is therefore obvious that the rainy season and the displacement of the rain belt over China are greatly dependent on the evolution of the general circulation.

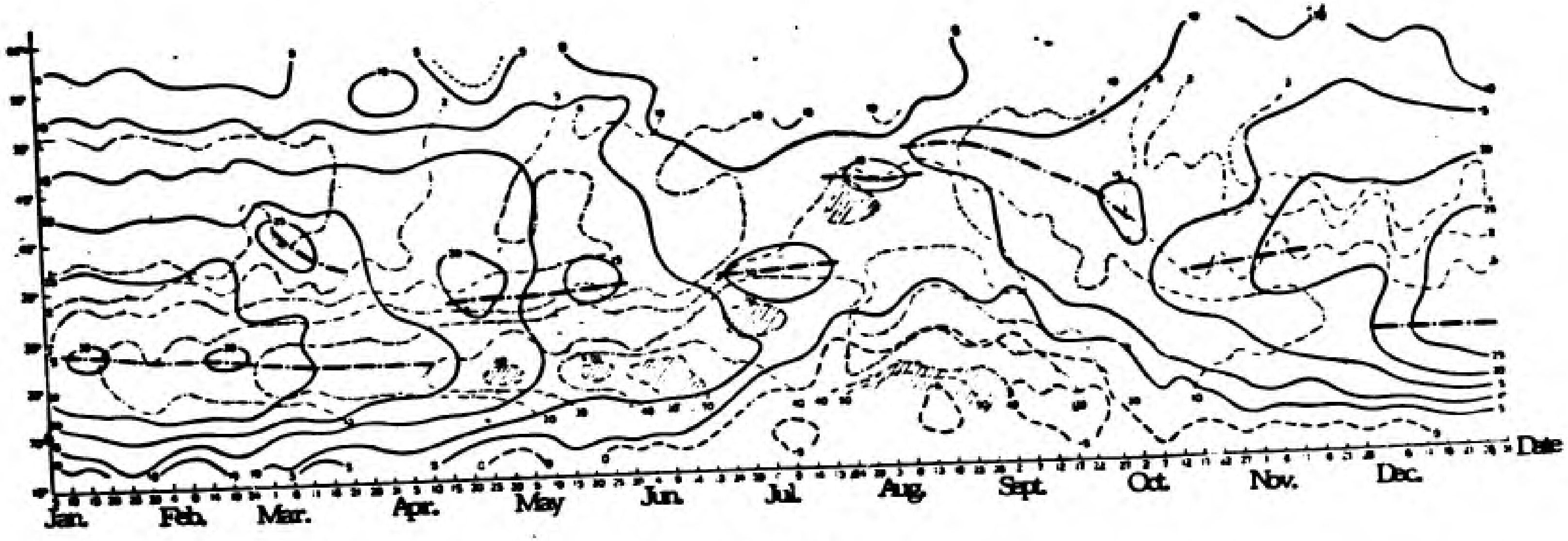

Figure 3.1

Distribution of rainfall and mean 500-mb zonal component of the geostrophic wind (1953 - 1957) at various latitudes along 115° - 120°E.

(Legend: Solid lines are isotachs of zonal wind in m/sec; broken lines are isotachs of easterlies in m/sec. Thick dot-dash lines represent belts of strong westerly wind; thin dot-dash lines are isohyets in mm and shaded areas represent regions of heavy rainfall.)

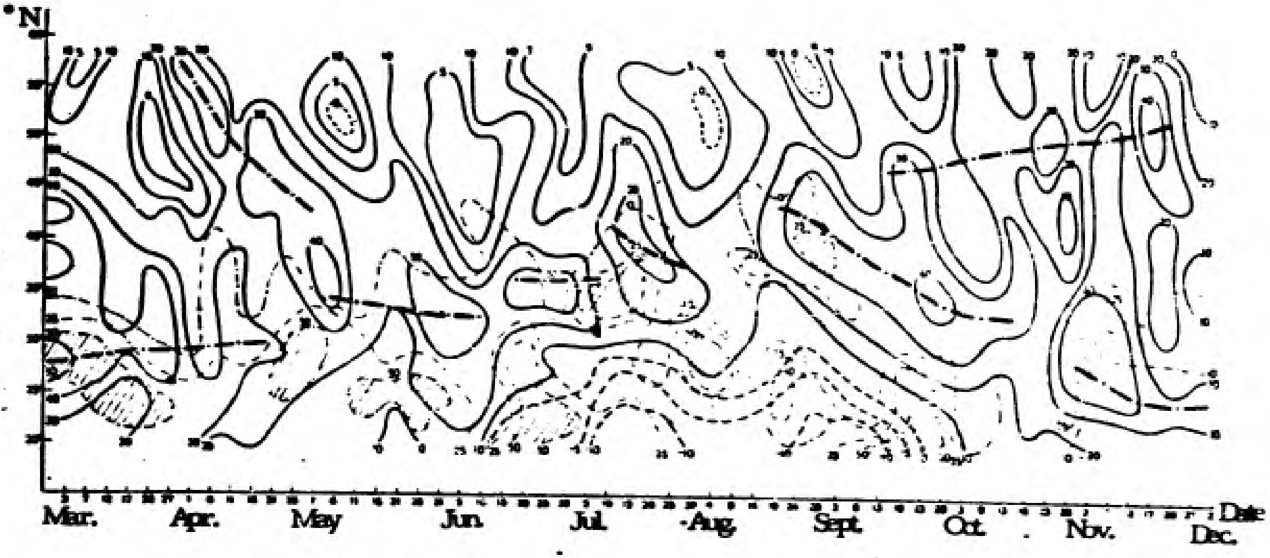

Figure 3.2

Distribution of rainfall and the 300-mb zonal geostrophic wind at various latitudes along 115° - 120°E for 1957.

(Legend same as in Figure 3.1.)

The movements of the rain belt is characterized by three slow displacements and three rapid advancements. Two of the jumps take place during their northward trip while the remaining one occurs during its southward journey. The march of the rain belt is in phase with the advancement and retreat of the monsoons so that the commencement of the stage of their full development also marks the arrival of the rainy season.

Studies on the seasonal variation of the westerly circulation in relation to monsoon activities [9] indicate that the three rapid advancements of the monsoons are associated with the three jumps of the rain belt in June, July and September. The first jump from South to Central China occurs in early June when the southern branch of the westerly jet stream weakens or dissipates over China. The second from the Yangtze basin to North China in mid-July is accompanied by the weakening and poleward displacement of the northern branch of the westerly jet. The tropical easterlies, the surface Pacific High and the upper subtropical ridge near 20°N also moves rapidly north to the Yangtze basin. When the westerlies reappear over this region in early September, the rain belt retreats southward and moves to South China in 20 days. Thereafter, the system becomes meridionally oriented and persists over the region of the two lakes for a little more than a month and finally dissipates with the intensification and full establishment of the upper westerly wind circulation by mid-October.

Figure 3.1 depicts the 5-day mean zonal component of the geostrophic wind at 500 mb at the various latitudes along 115° - 120°E (solid line) and the seasonal variations of the mean pentad rainfall (dotted line) and the heavy rain belt (shaded area) along 110° - 120°E. It is seen that the jet core is located near 27.5°N from December to April with a persistent rain belt below. The upper westerlies moves about 5° latitude to the north during late April and this is accompanied by an increase in rainfall with a slight southward displacement of the heavy rain belt, which reaches the coastal region in mid-May. Easterlies appear south of 25°N in mid-June and is associated with the weakening of the westerlies to below 10 m/sec. However, strong west winds still persist in places near 35°N. At this time, the heavy rain belt advances abruptly northward to the southern edge of the jet core in accord with the first jump from South China to the Yangtze basin and the spreading of the tropical easterlies toward the South China coast. The second rapid advancement of the easterlies occur in mid-July when the strong westerlies shift toward higher latitudes and the rain belt shortly reaches North China. The intensification and southward displacement of the westerly winds in early September are accompanied by the southward migration of the easterlies and the heavy rain belt. The speed of movement of the latter is, however, much greater than that of the westerlies so that its east-west orientation soon becomes insignificant, resulting in widespread rain over Szechwan, Yunnan, Kweichow and the southeast coastal region. The easterlies retreat completely from the mainland by October when the westerly jet core intensifies and becomes stable near 32°N. Rainfall at this time of the year is much less intense than in summer but the precipitation belt persists with little movement near 30°N. The further intensification of the westerly jet near 27°N causes the rain belt to become stable near the jet core.

It should be mentioned that the geostrophic wind presented in Figure 3.1 is computed from data for the period 1951 - 1957 while the pentad rainfall is extracted from records of 10 years or more before 1954. Despite the heterogeneity in the basic data, these two parameters are shown to be closely related. Figure 3.2 represents a similar cross-section for the year 1957 except that the zonal component of the geostrophic wind is computed from 300-mb data. The relationship between the heavy rain belt and the position and intensity of the west wind circulation is clearly shown in the figure. It may be added that the development of the second belt of heavy rain over the South China coast is associated with the appearance of the easterlies at 300 mb. This feature is also noted in Figure 3.1. Thus we may say that while precipitation over China is mainly observed in frontal zones in the westerly wind belt, strong upper easterlies are also an important factor for production of rain over South China.

A more detailed comparison between Figures 3.1 and 3.2 reveals that the positions of the rain belt and the strong westerlies are generally in phase during the winter and summer half years (December to April and June to August) but are out of phase in the months of April, May and September. Although the latitudinal displacements of the westerly jet are accompanied by marked changes in the position of the rain belt, the movements of the two systems (and perhaps the frontal zone) are completely uncorrelated. Thus the heavy rain belt remains almost stationary during the first weakening of the southern branch of the westerlies after April and the system even moves southward to the South China coast in May. In early September, the winter monsoon begins to spread southward and the upper westerlies also shift in the same direction to the northern bank of the Yangtze. However, the east-west oriented heavy rain belt is only found further south over the South China region, which is a feature difficult to explain. On the other hand, during the period October - April, the polar front is often located over the ocean far away from the mainland of China and the upper-air frontal zone becomes the controlling factor for production of rain over the whole country. Consequently, the rain belt coincides with the zone of strong upper winds. During the transitional periods, the polar front in the lower troposphere slowly migrates toward the north as a result of the invasion of the summer monsoon and constitutes a major rain-bearing system. Since the effect of the upper westerly jet greatly diminishes after its first weakening in May, the front is even more important over South China where the mean position of the system is located during this period. Thus, the heavy rain belt is always found to follow the polar front. Similarly, at the beginning of summer, the rain belt over Central and North China is also controlled by the polar front which is associated with strong upper westerlies. At this time, the southern branch of the west wind belt is not discernible. By early September, the westerlies begin to strengthen but the polar front rapidly sweeps across Southeast China and the lateral separation between the two gives rise to a more irregular distribution of rainfall over the country.

IV. CONCLUDING REMARKS

Early attempts to determine the onset and ending of the rainy season and the movement of the rain belt have been made with the aid of monthly mean rainfall data and the results have not proved conclusive. In the present study, we have utilized the mean pentad rainfall to define the heavy rain belt which is found to show systematic latitudinal displacements and is closely related to the advance and retreat of the monsoons. Hence the movement of this system has been used to determine the onset and termination of the rainy season over the various regions.

The movement of the rain belt over China is characterized by three slow displacements and three rapid advancements. The mean pentad heavy rain belt first appears over the coastal region of South China in mid-May when it begins to move northward. By mid-June, the belt reaches the north of the Nan Ling Ranges and the border between Fukien and Chekiang. The first rapid movement or jump occurs in the second decade of June when the rain belt spreads rapidly northward across the region of the two lakes. Heavy rain arrives at the Yangtze basin in mid-June and persists over the region for more than 20 days. The rain belt advances rapidly for the second time about the second decade of July and sweeps across the Hwai Ho and the lower Yellow River basins to the region near or north of 40°N. By mid-August, the axis reaches its northernmost position and begins to retreat southward. It stays north of 38°N before the last decade of the month but moves rapidly southward thereafter, and returns to the coastal region of South China in less than two and a half months. If the period between the two successive passages of the rain belt is taken to be the rainy season (the transitional period of increasing rainfall is excluded, otherwise the onset of the season has to be advanced by about 10 days), then this season lasts for about four months from mid-May to mid-September in South China, more than two and a half months from mid-June to the end of August in the Yangtze region and a little more than one month from mid-July to the last decade of August in the North China region north of 39°N. These intervals all coincide with the peak periods of the summer monsoon over the various regions.

The seasonal variation of the pentad rain belt shows that: (a) the march of the rainy season is significantly related to the outburst of the summer monsoon and the onset and termination of the season closely agree with the dates of the full establishment and retreat of the monsoon; (b) the three rapid displacements of the rain belt are all accompanied by "radical changes" in the general circulation (including the monsoon flow); (c) the onset and termination of the mei-yü are dependent on the slow displacements of the pentad rain belt, and (d) over East Asia, the movement of the rain belt, the advance and retreat of the monsoons and the seasonal variation of the general circulation are all characterized by an interplay of gradual and rapid changes.

In examining the movement of the rain belt over China in relation to the outburst of the southwest monsoon over India, it is noted that the onset of the rainy season is earlier in the east than in the west at the same latitude.

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MONSOONS IN EAST ASIA IN RELATION TO SYNOPTIC SEASONS*

by

Hsu Shu-ying and Kao Yu-hsie

The "synoptic seasons of China" has been a subject for discussion by many meteorologists since 1955. In most of these studies, the investigators tried to determine the dates of onset and termination of each climatological phase in relation to the seasonal variation of the general circulation or the westerly circulation. Kao [1] has discussed the relationship between the dates of onset and ending of the mei-yü period over the Yangtze basin and the variation of the flow configurations in the middle and the lower layers of the troposphere. Yeh and Chu [2] have pointed out that the development of the "Northeast Depression" in March and April and the replacement of the upper-air trough over Lake Baikal by a ridge may be used to mark the onset of spring and autumn. Later, Liu et al [3] have attempted to define the mei-yü period over China as well as the natural synoptic season of summer mainly in terms of the establishment, dissolution and variations in the intensity of the strong westerlies at 500 mb over South Asia. Although these writers tend to restrict their discussion to the characteristics of the synoptic seasons and have not delved into their relationship with the monsoons, they have presented some good methods for investigating this problem.

On the other hand, some writers have also dealt with the activities of the summer monsoon as a subsidiary issue of their discussions on the onset and ending of certain climatological phenomena. In a collaborative study on the "Mei-yü over China", Dao et al [4] have pointed out that not only is the

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* Manuscript completed in September 1959.

onset of the mei-yü over China in phase with the establishment of the summer monsoon over Calcutta in India, but its ending is also closely related to the dissolution of the jet stream in the upper westerlies or the appearance of easterlies over Tateno in Japan. Chen [5] has noted this phenomenon and particularly emphasized that the mei-yü period is closely related to the seasonal variations of the general circulation over Asia. The study of Kao [6] on the "Clear Weather in High Autumn" reveals that autumn over China begins in early September and ends in mid-October, lasting about 45 days; whereas in an investigation on "Precipitation in Autumn over China" [7] it has been shown that the onset and cessation of the autumn rain over West China are closely related to the appearance of the westerly circulation and the establishment of the jet stream. It is also connected with the retreat of the summer monsoon.

Yeh et al [8] have related the outburst and recession of the southwest monsoon over India and the mei-yü to the abrupt changes of the general circulation in June and October. The above studies serve as valuable references for our investigation on the monsoon problem.

I. MONSOON AND SYNOPTIC SEASONS

It may be noted from Reference [9] that significant seasonal variations occur in the position and intensity of the four action centers over the Eurasian Continent and the Pacific. These changes are found to exert a pronounced influence on the advance and retreat of the summer monsoon over China. Although the position and the central intensity of the action centers undergo periodic variations in the course of the year, a stable pattern is, however, often observed at certain particular stages. This pattern is related to some flow configuration of the general circulation and large-scale synoptic processes, and is naturally associated with distinct climatological features. Multanovskii [10] gives the same view and is of the opinion that the large-scale synoptic processes operate steadily throughout a natural synoptic season during which both the influence of action centers and their interrelationship undergo little change. A change in the synoptic season is undoubtedly accompanied by a relatively long period of significant variations in the flow field over a region in the form of the breakdown of one or more action centers or formation of new action centers. This is concordant with the results of References [9] and [11] on the advance and retreat of the monsoons in relation to the activities of certain atmospheric action centers. Thus monsoons also exhibit seasonal variations.

The demarcation of the seasons is a very important problem in climatological studies as well as in long-range forecasting. The classification of the 24 "Chhi" (phenophases) of the Chinese agrometeorological calendar is the outcome of phenological studies and farming empiricism. Chang [12] has made use of phenological concepts to define winter in terms of a mean temperature of less than 10°C, summer greater than 22°C and the transitional seasons (spring and autumn) in the range of 10 - 22°C. This forms an objective and comprehensive system of climatological classification. The threshold temperature of 10°C possesses a particular significant meaning in respect of the growth of farm products. The agricultural productivity of each region is greatly influenced by the length of the period during which the air temperature is greater than 10°C. In addition to heating, water supply is also an important factor for the growth of farm products. Thus lack of water results in droughts which are known to be detrimental to crops. The advance and retreat of the monsoons bring about distinct variations in moisture supply and heating, resulting in the characteristic variations of climatological and meteorological features with time. These features form distinct seasonal patterns.

The seasonal variations of weather and climate over different regions in China are thus determined to a great extent by the activities associated with the advance and retreat of the monsoons. Since the monsoons form part of the general circulation, the flow pattern and the position of pressure systems together with the various meteorological and climatological elements undergo significant changes from season to season. The degree of change also varies with time and space. It is therefore difficult to formulate a system to account for the seasonal and spatial variations of meteorological and climatological elements for the vast area of China, which spans 49° of latitude and 63° of longitude with a multitude of complex topographical features. In the present study, the climate of East China is classified into seven synoptic seasons in accordance with the several stages of the advance and retreat of the monsoons and the evolutionary characteristics of the action centers and the westerly circulation. The characteristic distribution of temperature and precipitation in each season is then described.

II. THE SEVEN NATURAL SYNOPTIC SEASONS

(a) Deep Winter

The winter monsoon prevails over the entire mainland of China with a maximum intensity during the period from early December to early March.

The monsoon winds are strong and steady with little change. During this period, the center of the Mongolian High as shown by the pentad mean pressure (chart omitted) moves about within the region bounded by 45° - 55°N and 100° - 105°E, while its monthly mean position remains steady at 50°N, 100°E. The position of the Aleutian Low as shown by the pentad mean pressure (charts omitted) varies within the region of 45° - 50°N and 165° - 180°E with its stable monthly position at 50°N, 170°E. At this time, the other two action centers at low latitudes over the Asian region from 60°E to 180°E, namely the Pacific High and the Indian Low, are completely absent. The westerly circulation becomes strongest and the isotach maximum of 30 m/sec of the 500-mb westerlies over China reaches its southernmost position at about 27°N, while the major trough remains quasi-stationary near the Asiatic coast at about 140°E

Figure 1 shows the mean 500-mb contours (broken lines) and the sea-level isobars (full lines) during the period 16 - 25 January for the years 1954 - 1957, which represent the typical synoptic configuration for deep winter. The Mongolian High appears to be the only dominant pressure system over the mainland of China. The continental cold high and the stable winter monsoon only break down for short periods during the development of wave disturbances and the passage of upper troughs. The eastward movement of these systems tends to induce a fresh cold outbreak and the intrusion of a new cold high into China, resulting in the re-establishment of the winter monsoon. When the surface cyclone is absorbed into the Aleutian Low, the cold front and the continental anticyclone behind it spread southward into low latitudes. From then onward they remain quasi-stationary for some time and the duration depends on the

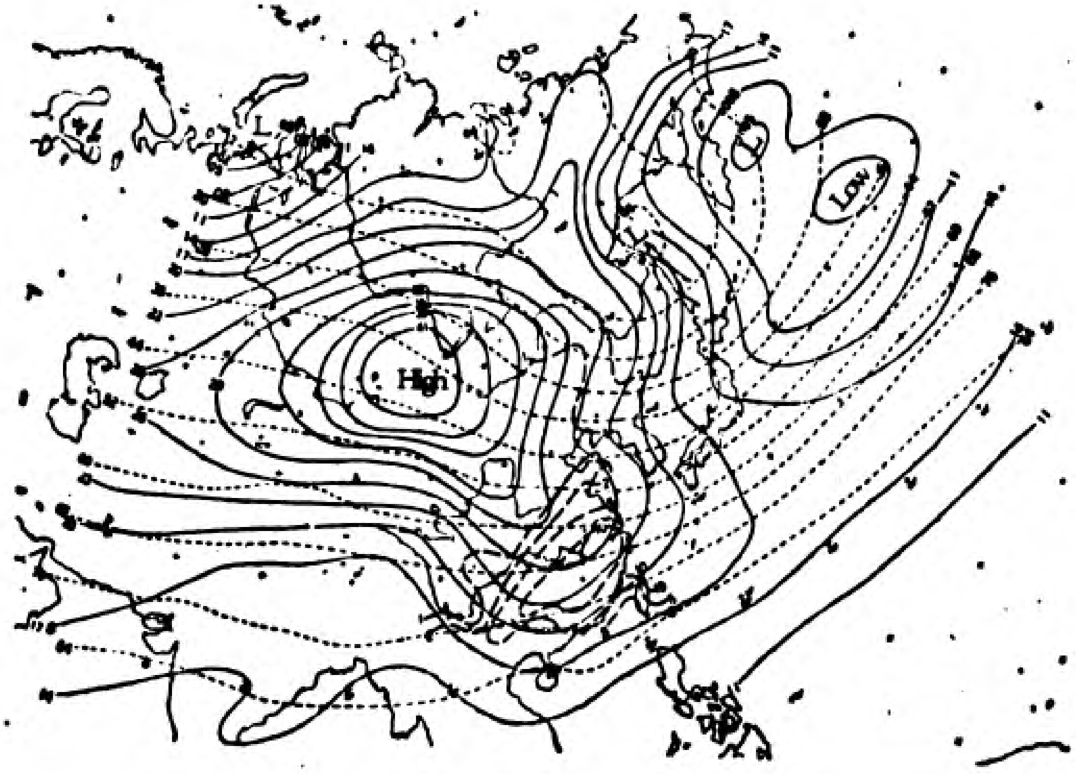

Figure 1

Mean 500-mb contours and mean surface isobars for the period 16 - 25 January.

[Legend: ——— surface isobars (unit: mb add 1000 mb);
----- 500-mb contours (unit: gpdm add 500 gpdm);
———— isohyets.]

intensity of the intrusive cold high and the extent of its southward excursion. In general, the greater the intensity of this anticyclone, the farther is its southward spreading and the longer is the duration of the quasi-stationary period. Consequently, before the completion of one cycle of a monsoon surge induced by the development of a cyclone, the formation of another one or two cyclones and their eastward displacement may trigger a fresh cold outbreak to spread southward without causing a breakdown of the wind and pressure fields of the pre-existing monsoon. In a series of cold outbreaks, the continental cold anticyclone intensifies and spreads southward to very low latitudes, resulting in a relatively long stable period. Thus the synoptic processes of deep winter are characterized by an intense and stable winter monsoon and a long stable period. An examination on the distribution of the pentad mean temperature and the pentad rain belt reveals that the rain belt (see thick broken lines in Figure 1) lingers around 27°N while the 10°C isotherm [13] persists at about 25°N after early December (Figure 2).

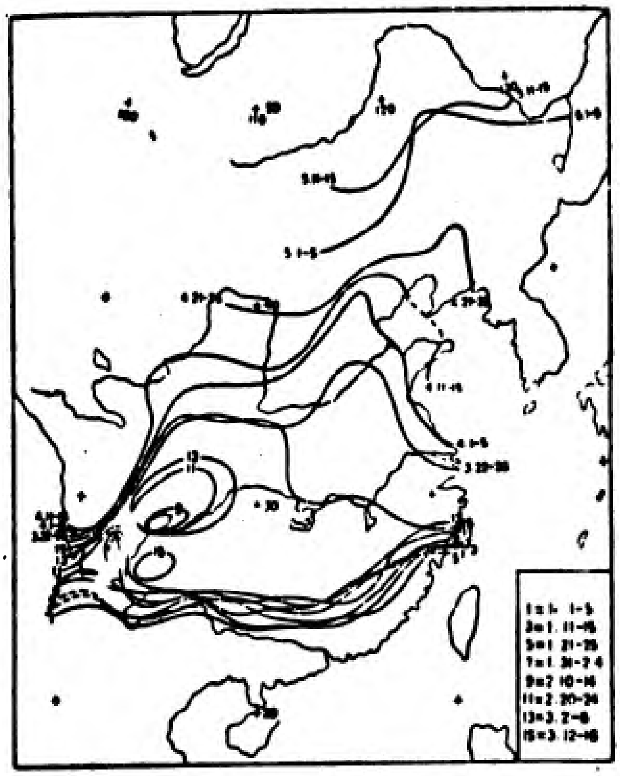

Figure 2

Position of the 10°C isotherm of pentad mean temperature.

(Numerals at the extremities of each isopleth denote the pentad number from 1 - 15 and the corresponding dates are shown in the inset. In regions north of 30°N, the number before the decimal denotes month in numerical form and the days are given after the decimal.)

(b) Late Winter

Late winter covers a period of about 45 days from early March to mid-April. The significant weakening of the winter monsoon in early March is accompanied by the onset of the summer monsoon over South China. The beginning of late winter is marked by a distinct weakening of the Mongolian High, whose center retreats westward by 15° longitude from 100° to 85°E within two to three pentads. From then onward, it rarely occurs in its deep winter position. A marked weakening of the Aleutian Low also takes place in early March, but its central position undergoes little change. After the onset of late winter, two low centers often form in the northern part of the North Pacific. One persists in the position of the original Aleutian Low, while the other is located over the southern part of the Alaska Peninsula. Following the first weakening of the action centers at high latitude, the Indian Low and the Pacific High begin to show "glimpses" of their individual identity (Figure 3). By mid-March the existence of the Indian Low is shown by the appearance of a closed isobar of 1011 mb, while the gradual westward extension of the Pacific Ridge is accompanied by the first significant weakening of the strong upper westerlies over the mainland of China. In general the 500-mb westerlies decrease from 30 m/sec to 20 m/sec [11] with little change in position. The quasi-stationary major trough over the coast becomes less meridionally oriented (broken lines in Figure 3). Its intensity diminishes but there is little change in position.

The above analysis thus shows that deep winter is characterized by high index circulation followed by a low index one after early March.

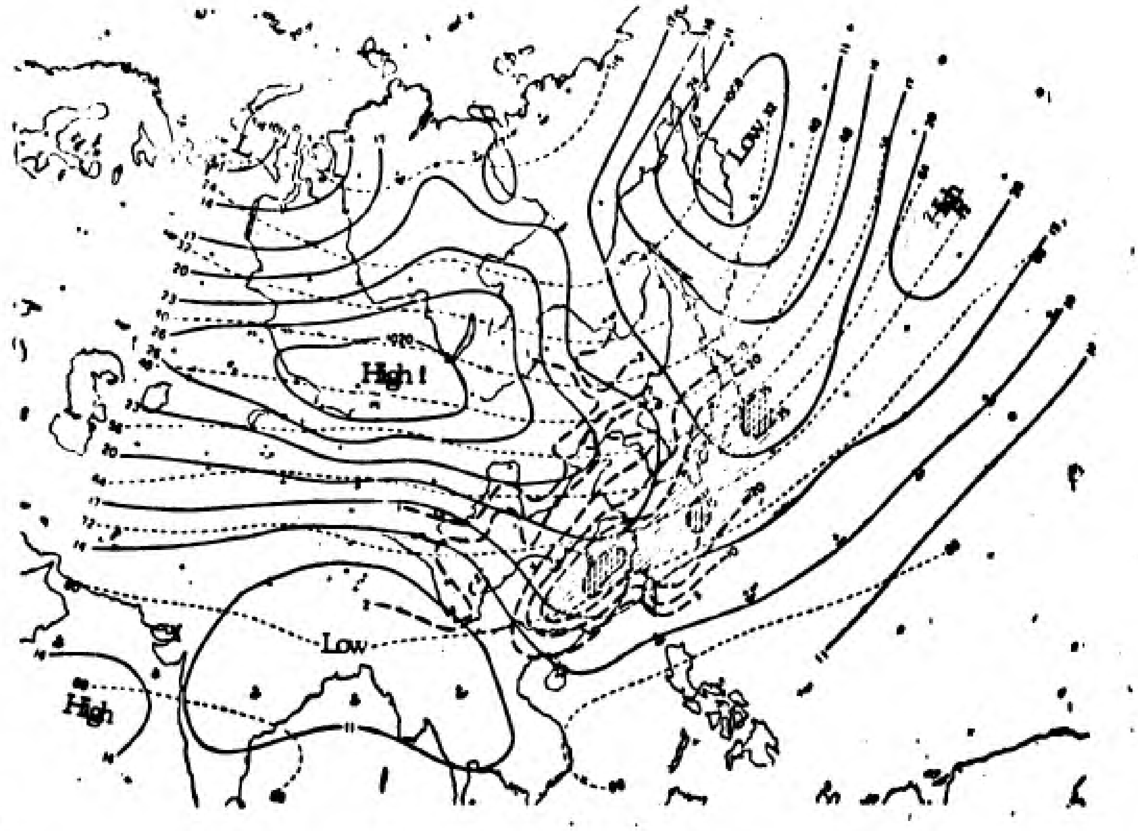

Figure 3

Mean 500-mb contours and mean surface isobars
for the period 17 - 26 March.
(Legend same as in Figure 1.)

It has been pointed out by Namias [14] that a distinct weakening of circulation index occurs during the period from late February to early March. This phenomenon has also been discussed recently by Chen [15]. We are of the opinion that the occurrence of low index circulation during this period not only marks the weakening of the winter monsoon but also acts as a signal to indicate more frequent occurrence of low index circulation from early March onward.

A comparison between Figures 1 and 3 shows that there is little change in the upper-air circulation except a notable weakening of the westerly wind speed and the intensity of the trough. Apart from some weakening of the two surface action centers at high latitudes, the principal change is characterized by the existence of the Pacific High and the Indian Low in the surface layer. Their appearance signifies that the summer monsoon begins to affect the mainland of China and that the seasonal change in the pressure field over East Asia takes place earlier near the surface than at the upper levels. This phenomenon is particularly significant over India. The breakdown of the pressure field of the winter monsoon over the region is marked by the appearance of a low pressure system and this occurs when the westerly circulation though significantly weaker still predominates in the upper air as in deep winter.

If we say that the weather is stable in deep winter, then the onset of late winter may be said to be characterized by changeable weather. Although there are no radical changes in the synoptic process between deep and late winter, low pressure systems occur much more frequently over the mainland in the latter period.

The 10°C isotherm of mean temperature moves rapidly from Nan Ling to the Yangtze basin after the 13th pentad (2 - 6 March) at the beginning of the season, reaches 35°N by the 17th pentad (22 - 26 March) and remains almost stationary subsequently (Figure 2). Table 1 indicates that the difference between maximum and minimum temperature is larger during this period than in deep winter. The pentad total rainfall also increases during late winter. For example, the amount of rainfall as shown in Figure 3 is about twice that in the previous period. This accords with the results in Table 1 for Foochow, Lungchi, Nanning and Lungchow. However, there is little change in the pentad position of the belt of maximum rainfall.

(c) Spring

Spring lasts about 60 days from mid-April to mid-June. The change in mid-April marks the end of winter and the beginning of spring. As the winter monsoon further weakens in mid-April, summer monsoon prevails over South China to mark the onset of the rain season. At this time, Central China begins to come under the influence of the summer monsoon and rainfall increases. At the beginning of spring, the Mongolian High shifts westward to 75°E with a weakening of its intensity by 6 mb in 10 days while the rapid eastward displacement of the Aleutian Low by 30° longitude to 160°W is associated with the appearance of the so-called "Northeast Depression" [17] over the coastal region of Russia between these two action centers. Yeh and Chu [2] have used the occurrence of this low pressure center as an indicator to mark the onset of spring. We concur with this criterion. At this time, a high cell appears over the Sea of Okhotsk between these two low pressure centers and is accompanied by the establishment of an area of low pressure at low latitude over the region from the Bay of Bengal to Burma. The Pacific High intensifies and extends westward (Figure 4). By mid-April, the westerly wind speed weakens further and the strong westerly wind belt migrates northward by 5° latitude from its winter position of 27.5°N. Since the two high-latitude action centers move in opposite direction away from each other, their separation distance increases to 60° longitude. This development together with the two periods of weakening of the upper westerlies gives rise to a re-adjustment of the wave pattern and configuration of troughs and ridges over East Asia in mid-April and the formation of another upper trough system over the coastal region of Russia within the major trough zone along the Asiatic coast.

Variations in the atmospheric action centers as well as the intensity of the westerlies and the configuration of troughs and ridges in the westerly wind belt in spring are greatly different from those in winter, although there are no basic differences in properties of these systems. The flow configuration in winter is controlled by the Mongolian High and the Aleutian Low, while the synoptic pattern in spring is characterized by the co-existence of four action centers. In addition to this, two seasonal pressure systems are observed at high latitudes over East Asia, namely, the "Northeast Depression" and the anticyclone over the Sea of Okhotsk. Their appearance marks the onset of spring while their breakdown serves as an indicator of the beginning of high summer.

A comparison between Figures 3 and 4 reveals that although the major trough along the coast fills up considerably and moves toward the coastal region of Russia, there is still little change in the westerly circulation at middle and low latitudes. However, marked differences are noted in the distribution of the surface pressure field. They are characterized by significant weakening of the two high-latitude action centers and the establishment of two low-latitude ones. The interaction among these systems and the variations of their intensities and positions are the special features observed in spring. As far as the mainland of China is concerned, the continental cold anticyclone is the dominating feature in winter, although it is occasionally replaced by low pressure systems. After mid-April, the cold anticyclone may spread southward and linger over the mainland of China for a day or two before it moves eastward into the adjacent seas. Under such circumstances a low pressure area develops over the mainland (the "Northwest Trough" or the "Southwest Trough" forms, or both develop simultaneously). It moves eastward in a day or two after its formation and thus triggers another southward migration of the cold anticyclone. A comparison of the evolution of the synoptic patterns shows that both the breakdown of the winter pressure field and the establishment of the summer one occur earlier at the surface than at the upper levels.

Following the onset of spring in mid-April, the 20°C isotherm [13] of pentad mean temperature moves rapidly northward from the northern part of Kwangtung across the Yangtze around 11 - 15 April to 35°N during 11 - 15 May. This isotherm travels 12° latitude northward in 5 pentads and remains almost stationary thereafter. From the second decade of May to the first decade of June, the 20°C isotherm advances only 5° latitude northward in one month. It may be seen from Figure 2 that the 10°C isotherm migrates rapidly from 40°N to 48°N during the period from mid-April to early May but moves 4 - 5° latitude poleward in 20 days before this period, causing a rapid rise of the air temperature over the whole of China.

A comparison between Figures 3 and 4 indicates relatively large differences in the distribution of the pentad precipitation. After the first decade of May the rainfall over the coastal region of South China is greater than that over the lake basins in Hunan and Hupeh. Thus the heavy rain belt as shown by pentad values is located well south of its winter position, implying that the polar front has reached the coastal region of South China in early May. This suggests that precipitation over Central China is no longer closely related to the upper frontal zone and that the factors controlling the rainfall in spring are different from those in winter.

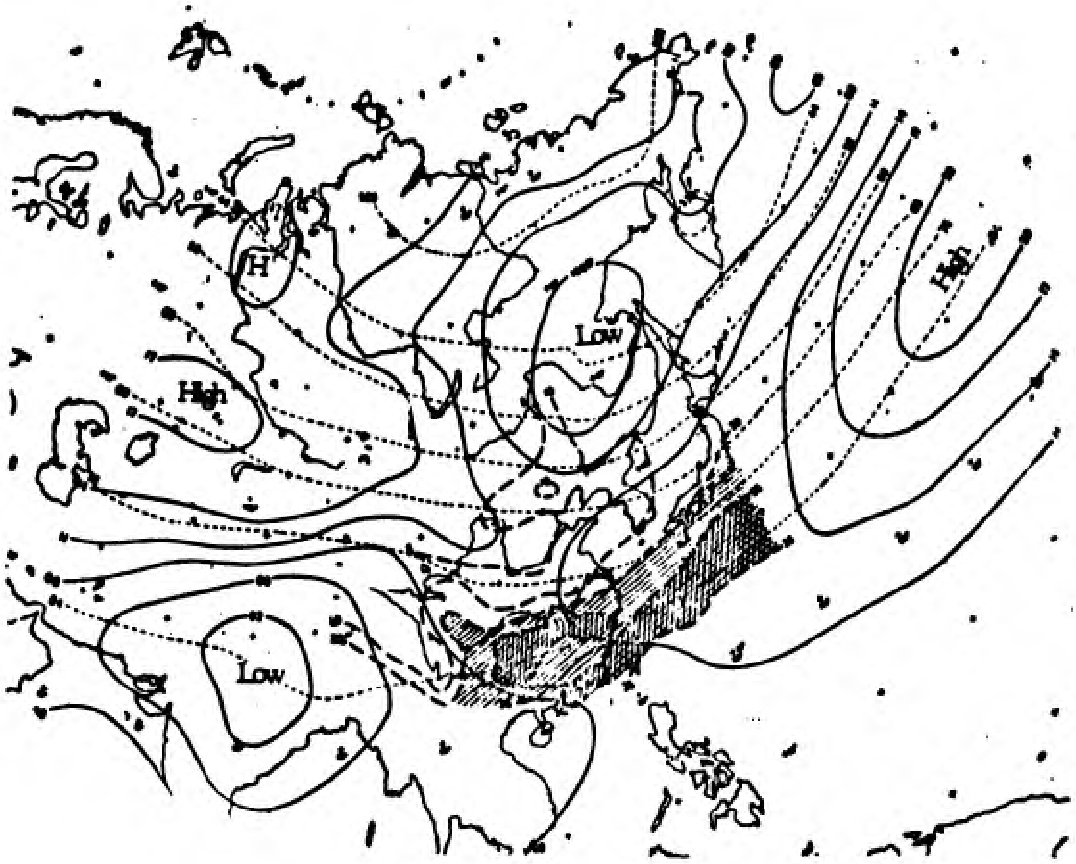

Figure 4

Mean 500-mb contours and mean surface isobars for the period 11 - 20 May.

(Legend same as in Figure 1.)

TABLE 1

Distribution of temperature and precipitation at different seasons

Date		Ulan-hot	Pao-tow	Pek-ing	Sian	Kai-feng	Han-kow	Yu-yang	Chang-sha	Chih-chiang	Foo-chow	Lung-chi	Can-ton	Nan-ning
2/12—1/3	$\bar{T}$	−14.2	−10.3	−3.2	0.7	1.1	5.5	5.1	5.9	6.3	12.0	14.8	14.6	14.3
	T'	− 9.2	− 6.1	0.1	1.1	3.5	7.4	8.2	7.6	8.9	14.0	17.0	16.5	16.3
	T''	−19.6	−13.3	−4.9	−4.8	−1.8	3.5	2.8	2.9	4.5	9.8	12.7	13.1	12.2
	γ	0.58	0.38	0.28	0.33	0.29	0.22	0.3	0.26	0.24	0.21	0.24	0.19	0.23
	R	7.1	4.8	10.4	23.5	31.4	121.7	86.5	228.3	128.9	140.3	154.6	138.4	109.4
	R/n	0.4	0.3	0.58	1.3	1.7	6.8	21.8	13.3	7.2	7.8	8.6	7.7	6.1
2/3—10/4	$\bar{T}$	−1.8	1.3	6.6	9.6	9.2	11.3	10.7	11.8	12.1	13.9	16.9	17.9	17.9
	T'	5.5	5.5	11.4	13.0	13.4	15.2	14.0	16.2	15.5	16.6	19.1	20.3	22.1
	T''	−10.0	−4.7	1.5	6.0	4.5	8.5	6.9	8.1	7.7	12.0	14.6	15.2	15.1
	γ	1.94	1.28	1.24	0.87	1.1	0.84	0.89	1.01	0.97	0.57	0.56	0.64	0.88
	R	6.5	11.7	9.3	39.6	22.0	129.3	105.4	197.2	129.9	192.9	126.8	168.8	98.1
	R/n	0.8	1.4	1.16	5.0	2.7	16.2	13.1	24.6	16.2	23.1	15.8	21.1	12.2
11/4—9/6	$\bar{T}$	12.4	13.6	19.1	19.5	19.9	20.8	19.1	20.9	20.6	21.7	23.3	24.7	25.0
	T'	17.8	18.8	23.7	24.8	26.1	24.9	23.4	24.6	25.0	24.8	25.7	27.0	27.5
	T''	4.3	5.5	13.4	13.2	12.2	16.6	13.9	15.4	15.5	17.3	19.2	20.7	21.7
	γ	13.3	11.1	0.85	0.97	1.16	0.15	0.79	0.77	0.79	0.63	0.54	0.53	0.48
	R	71.3	48.9	71.9	87.8	71.8	274.1	316.2	419.0	336.2	378.0	416.5	491.7	313.9
	R/n	5.9	4.1	6.0	7.3	6.0	23.1	29.6	34.8	28.0	31.6	34.6	41.0	26.2
10/6—14/7	$\bar{T}$	22.2	22.0	25.6	26.7	26.5	26.7	24.4	27.2	26.6	27.2	27.8	28.0	28.1
	T'	24.5	23.0	26.2	27.9	27.8	27.9	25.7	29.4	28.1	28.6	28.6	28.0	29.0
	T''	19.7	21.2	24.5	25.7	25.6	25.6	23.1	24.8	25.0	24.9	26.3	26.9	27.7
	γ	0.68	0.29	0.24	0.31	0.31	0.33	0.37	0.66	0.44	0.53	0.33	0.29	0.18
	R	110.9	56.2	33.7	93.6	120.0	311.3	229.7	252.9	220.9	240.1	316.7	279.5	203.7
	R/n	15.9	8.0	4.8	13.4	17.2	44.5	32.8	36.2	31.4	34.3	45.0	40.0	29.2
15/7—2/9	$\bar{T}$	22.2	21.3	25.5	26.7	26.5	28.8	25.7	29.2	27.8	28.3	28.9	28.3	28.2
	T'	24.9	24.1	26.9	28.7	28.3	29.9	26.9	29.9	28.5	28.4	29.4	28.6	28.8
	T''	19.2	18.1	22.9	23.3	23.7	26.8	24.7	27.6	26.1	27.9	28.5	27.8	27.9
	γ	0.57	0.60	0.40	0.54	0.46	0.31	0.22	0.23	0.24	0.05	0.09	0.08	0.09
	R	149.3	129.4	301.2	140.2	195.1	153.4	289.9	198.7	185.4	251.6	316.6	410.9	308.6
	R/n	14.9	12.9	30.1	14.0	19.5	15.3	29.0	19.9	18.5	25.2	31.7	41.1	30.9
3/9—22/10	$\bar{T}$	10.6	12.2	17.3	17.8	19.3	21.7	20.0	22.1	21.6	24.7	25.9	25.9	26.0
	T'	15.9	16.9	22.4	21.9	23.2	26.4	23.4	26.7	26.0	27.8	28.3	27.9	28.0
	T''	3.9	6.5	12.1	13.4	15.0	17.4	15.2	18.1	17.3	21.6	22.8	23.4	22.8
	γ	1.20	10.4	1.0	0.85	0.82	0.90	0.82	0.88	0.87	0.62	0.55	0.45	0.52
	R	45.6	44.1	60.5	158.1	53.3	110.9	210.8	140.8	147.5	146.3	269.7	193.2	191.0
	R/n	4.6	4.4	6.1	15.8	5.3	11.1	21.1	14.1	14.8	14.6	27.0	19.3	19.1
23/10—1/12	$\bar{T}$	−4.5	−0.7	5.2	7.8	9.3	12.8	11.7	13.1	13.1	18.7	20.5	20.6	20.2
	T'	2.8	4.3	10.4	11.0	13.7	16.8	14.3	16.6	15.7	21.4	22.0	22.9	23.0
	T''	−10.9	−7.1	−0.2	2.8	4.2	8.5	8.0	8.4	8.8	16.3	18.1	18.2	17.5
	γ	1.71	1.42	1.32	1.0	1.19	1.03	0.79	1.02	0.86	0.64	0.49	0.59	0.69
	R	3.6	8.9	14.8	33.4	33.0	85.7	103.9	108.3	88.0	39.0	41.7	56.5	41.0
	R/n	0.5	1.1	1.85	4.2	3.8	10.7	13.0	13.5	11.6	4.9	5.1	7.1	5.1

(Legend: $\bar{T}$ = Pentad mean temperature; T' = Pentad maximum temperature; T'' = Pentad minimum temperature; γ = Pentad temperature variation $(T'_{max} - T''_{min})/n$; R = Total rainfall; R/n = Pentad mean rainfall; n = number of pentads.)

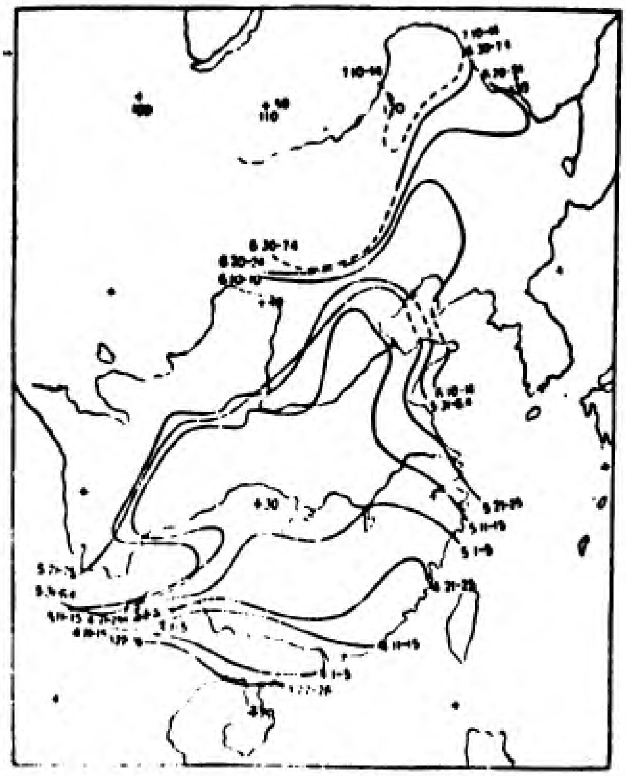

Figure 5

Position of the 20°C isotherm of pentad mean temperature. (Number before the decimal point denotes the month in numerical order and the days are given after the decimal, e.g., 6.10 - 14 means June 10 - 14.)

(d) Early Summer

Early summer lasts about one and a half months from early June to late July. The change of the flow pattern in mid-June marks the end of spring and the onset of summer. The beginning of this season is characterized by the full establishment of the summer monsoon over South China, resulting in a slight decrease of precipitation. This is accompanied by the occurrence of relatively dry spells over the hilly regions in Southeast China and in the vicinity of Nan Ling. Meanwhile, the summer monsoon reaches Central China to mark the onset of the mei-yü period and its influence extends to North China. On the other hand, the activities of the action centers are shown by the breakdown of the Mongolian High and the Aleutian Low, leaving only their remnants in early summer. The situation is significantly different from that in mid-March, when both the Mongolian High and the Aleutian Low, though weaker and notably displaced from their normal positions, maintain their dominating influence. The Pacific High and the Indian Low also make their first appearance at this time, and become the controlling systems in June when the action centers of mid-March are only barely detectable. Thus mid-March is taken as late winter and June early summer. Another feature of interest is the formation and establishment of an anticyclone over the Sea of Okhotsk. The synoptic characteristics during its life cycle often act as an indicator to mark the onset and termination of the mei-yü period. At 500 mb, the evolution process is shown by (i) the replacement of the quasi-stationary trough along the coast by a ridge, and the ridge at 80° - 90°E by a trough (see Figure 6); (ii) the presence of a trough over the mainland from Northeast to Central China; (iii) significant northward displacement of the subtropical ridge from 15°N to 22°N accompanied by the appearance of easterly winds south of 22°N at 500 mb.

It is obvious that the change in the circulation pattern at both the lower and upper levels in June [11] is a major one. This is evident by a comparison of Figures 5 and 6. Changes in the upper westerly circulation and wavelength and position of the planetary waves at this time of the year are extremely important seasonal variations of the flow pattern. Although the planetary circulation weakens occasionally before mid-June, there are no radical changes in the wavelength and the position of the long-wave troughs and ridges. In other words, the flow pattern is still typical of the winter monsoon. An adjustment of the wavelength of the planetary waves occurs in mid-June when the number of long waves changes to four. After this, the circulation pattern is typical of the summer season. The above aspect has been studied by many investigators [4, 8, 17]. Our opinion is that mid-June should mark the beginning of summer and the end of winter, if the seasonal changes of the general circulation are considered.

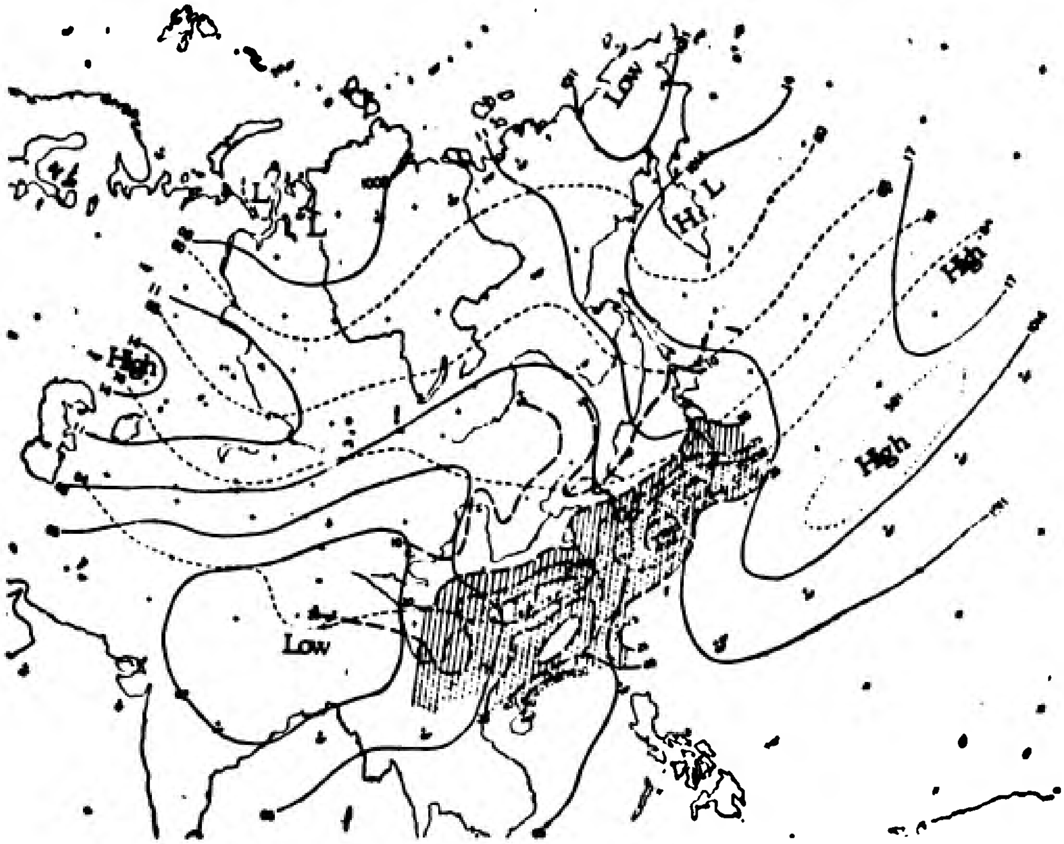

Figure 6

Mean 500-mb contours and mean surface isobars for the period 20 - 29 June.

(Legend same as in Figure 1.)

An examination of the 26°C isotherm [13] shows that this isopleth moves slowly from the northern part of the Leichow Peninsula to 25°N from early May to 10 - 14 June (Figure 7). The rate of northward advance of 5° latitude in 35 days is less than 1° latitude per pentad. However, it migrates as much as 15° latitude in the ten-day period from 10 - 14 June to the last decade of the month. This indicates the magnitude of the colossal change that can occur in June. A relatively stable period is noted in the last decade of June but abrupt changes begin to occur again in mid-July. After the onset of summer, the rate of increase of pentad (mean) temperature becomes much smaller than that in spring and is generally about 1/2 to 1/3 of that during the transitional periods (see Table 1).

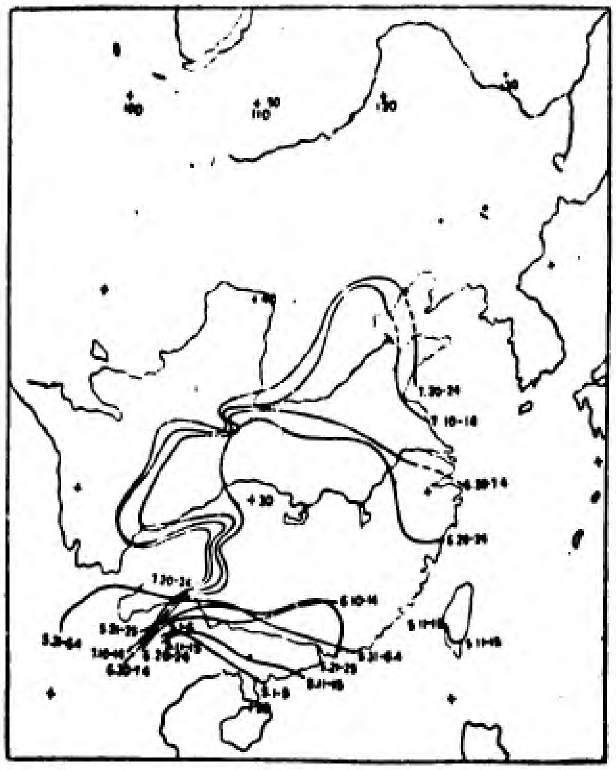

Figure 7

Position of the 26°C isotherm of pentad mean temperature.
(Legend same as in Figure 5.)

Following the circulation changes in mid-June, the "pentad rain belt" or the belt of rain shown by observations of pentad rainfall, which lingers in the vicinity of Nan Ling before 5 - 9 June becomes a region of heavy rain and moves rapidly to the Yangtze basin along with the polar front during the period 10 - 19 June. For the next 20 days the rain belt stays near the northern bank of the Yangtze to give mei-yü weather over the region. After the poleward displacement of this rain belt to the northern bank of the Yangtze, a secondary heavy-rain belt develops along the South China coast under the influence of the equatorial trough (Figure 6). The change in mid-June not only represents the establishment of the summer circulation but also brings the mainland of China to come under the influence of tropical synoptic systems. After the replacement of the flat ridge near 90°E by a trough in summer, the upper-air circulation becomes unfavorable for the maintenance of the anticyclone near the Lake of Baikal. Thus any subsequent intrusion of cold air into China can no longer take place in the form of a cold anticyclone as in the previous periods. In the majority of cases, the cold air spreads southward as a cold tongue or ridge. Meanwhile, the trough over Northeast China gradually moves westward from the beginning of spring and becomes stationary near 110° - 115°E by early summer. This indicates that the track of the cold polar air and the affected region tend to shift westward day by day, and the size of the region also decreases gradually. In fact, the cold air passes southward just to the east of the Tibetan Plateau into the Szechwan-Shensi region after the onset of summer and then spreads eastward to greet the arrival of the summer monsoon in the Yangtze basin, resulting in the characteristic weather of early summer - the mei-yü.

(e) High Summer

High summer lasts about 50 days from mid-July to early September. The change in mid-July is reflected by the prominent march of the monsoons and marks the onset of high summer. At this time, the summer monsoon is most pronounced over Central China and the mei-yü season comes to an end to be followed by a dry spell. In North China, the summer monsoon prevails and the rain season begins. In South China which comes under the influence of the equatorial trough rainfall increases and the dry spell ends. An analysis of the variations of action centers shows that with the complete dissolution of the remnants of the two high-latitude systems, the two low-latitude centers form the only controlling system with the Pacific High undergoing a marked displacement toward the northwest. Due to the presence of the upper ridge, the easterly wind begins to attain its maximum prominence in the upper-air circulation over the mainland of China south of 35°N.

Figure 8 depicts the mean high summer synoptic configuration for the decade 20 - 29 July. A comparison between Figures 6 and 8 indicates that a more marked development of the Pacific High and the Indian Low is shown at the surface. During this period, the Pacific High which is particularly active moves 10° - 15° latitude northward and covers the entire Northern Pacific. The upper-air circulation is characterized by a poleward displacement of the subtropical anticyclone to about 30°N with a further weakening of the high-latitude ridges and troughs. An examination of Figures 1 and 8 reveals contrasting features in the distribution of the surface pressure patterns, which are most significant south of 35°N. At 500 mb, the seasonal reversal of easterly and westerly winds is very distinct over this region.

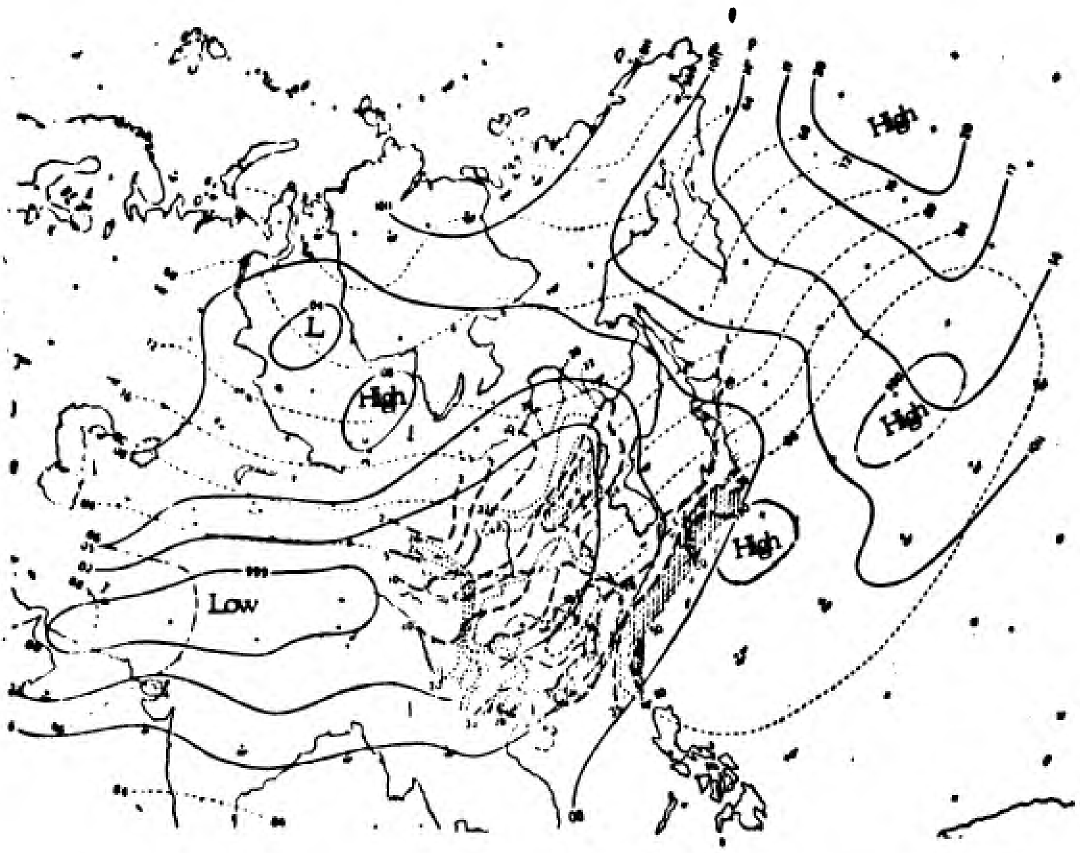

Figure 8

Mean 500-mb contours and mean surface isobars for the period 20 - 29 July.
(Legend: .... isotherm; the rest is the same as in Figure 1.)

A marked change in the circulation patterns also occurs in mid-July over China south of 35°N. (In fact, this change affects the whole of East Asia.) Despite the dissolution of strong westerlies over the southern part of Asia, the mainland of China is basically under the influence of the westerly circulation in early summer except the coastal region of South China. Because of the further weakening of the circulation from mid-July onward, the subtropical anticyclone rapidly moves northward from about 20°N to 25° - 35°N (see Figures 2.3 - 2.6 in Reference [11]), giving rise to the various changes: (i) The tropical easterlies replace the westerlies in regions south of 30°N, representing the seasonal reversal of the westerly and easterly wind flows [11]. (ii) The mei-yü period ends in the Yangtze basin (and over Japan), while the "pentad rain belt" moves rapidly poleward along with the polar front to North China to herald the onset of the rainy season [18]. This marks a prominent northward advancement of the summer monsoon. (iii) Under the influence of the subtropical ridge, the weather becomes hot and sunny over the Yangtze basin with a decrease in precipitation, and the dry spell begins. However, conditions are different over South China where the dry spell ends and another wet spell develops. (iv) The tropical synoptic systems exert their maximum influence on the mainland of China, and the equatorial trough moves to about 25°N.

Thus we may say that while the synoptic processes over China are basically controlled by the westerly wind belt in summer, the situation is much more complicated and is characterized by the existence of the westerly wind belt, the subtropical anticyclone and the tropical easterly wind belt over different regions. From the monsoon viewpoint it may be said that while the winter weather is controlled solely by the winter monsoon, the summer weather is affected by the northerly, the southeast and the southwest monsoons. Hence the synoptic processes in summer are much more complicated than those in winter.

Table 1 shows that both the pentad temperature difference in the various regions and the meridional temperature difference over China attain their lowest values during high summer. The highest temperature occurs over Central China (Figure 8). The regional distribution of rainfall shows that the pentad mean precipitation decreases significantly over Central China and increases markedly over North and South China.

(f) Autumn

Autumn lasts some 50 days from early September to mid-October. The change in the flow pattern in early September marks the end of summer and the beginning of autumn. The march of the monsoons is reflected by the rapid southward extension of the winter monsoon, which dominates almost the whole of the mainland of China. An analysis of the variations of the action centers indicates that the establishment of the Mongolian High is accompanied by rapid intensification while the Aleutian Low re-appears but its position is more to the east than at other times of the year. The Indian Low undergoes a significant reduction in size with the southward excursion of the continental cold anticyclone. Because of the heating effect of the underlying ocean, the subtropical anticyclone still stays at relatively high latitudes with little change in position and intensity, but lies somewhere to the southeast of its high summer position. Changes in the upper-air circulation are characterized by the abrupt dissolution of the trough near 90°E and the rapid establishment of a new trough along the coast somewhat west of its deep winter position.

Figure 9 depicts the mean synoptic configuration for the period 8 - 17 September. If this figure is compared with the mean spring pattern (Figure 3), the following interesting features may be noted: (i) There is a close resemblance in the distribution of the surface pressure pattern. The Aleutian Low and the "Northeast Depression" not only appear in the two seasons but also lie in approximately the same position. If the appearance of the "Northeast Depression" over China is used as an indicator to mark the onset of spring, then its re-appearance in early September is a manifestation of the onset of autumn. Because of the various abrupt changes that take place in autumn, the significance of the re-appearance of this system may have been overlooked by previous investigators. (ii) The high cell between the Aleutian Low and the "Northeast Depression" is more pronounced in autumn than in spring. (iii) The 500-mb autumn trough along the coast begins to take shape while the trough associated with the "Northeast Depression" remains stable as in summer. (iv) Figures 8 and 9 reveal that the breakdown of the pressure field of the summer monsoon occurs first at the low levels and then extends upward to the upper levels of the troposphere.

The change in early September signifies a major re-adjustment in the circulation pattern on a scale which is sufficiently large to affect the entire East Asia. This is witnessed by the breakdown of the summer flow configuration and the establishment of the winter circulation pattern. Although we have not delved deeply into the circulation changes in the northern hemisphere beyond the stretch of East Asia, we feel that the change in early September must be associated with the process of re-adjustment of planetary waves over the whole of the northern hemisphere. It may be seen from Figure 4 of Reference [11] which depicts the variation of the 556-gpdm contour at 500 mb for the months of August, September and October that the positions of troughs over the northern hemisphere vary considerably from August to September but remain generally steady from September to October. Studies on the variations of the general circulation in October have engaged the attention of many investigators, but we are of the opinion that developments in early September are more representative of the seasonal transition from summer to winter. In less than one month after early September, the winter monsoon spreads southward from North China to South China, whereas the poleward advance of the summer monsoon takes more than three months to cover the same area. Because of the fast southward intrusion of the winter monsoon, the air temperature falls very rapidly. Consequently, in some places the temperature variation at this time of the year is greater than that in spring (Table 1). The 10°C isotherm [13] first appears over Northeast China in early September and moves rapidly southward (Figure 10), and the pentad mean rainfall is also less than that in spring. This is the characteristic phenomenon of the clear weather spell. However, the rainy season has not yet come to an end over Southwest China where the southwest monsoon still prevails.

Although the equatorial trough retreats rapidly from the mainland, tropical systems remain dominant along the coast of South China. For example, the frequency of occurrence of typhoons in the southwestern Pacific does not decrease after an intense winter monsoon. A marked fall in the number of typhoons crossing the coast [19] is noted in early September though the frequency of occurrence is still much higher than that in April and May. These observations indicate that in autumn South China remains under the influence of tropical pressure systems and upper easterlies.

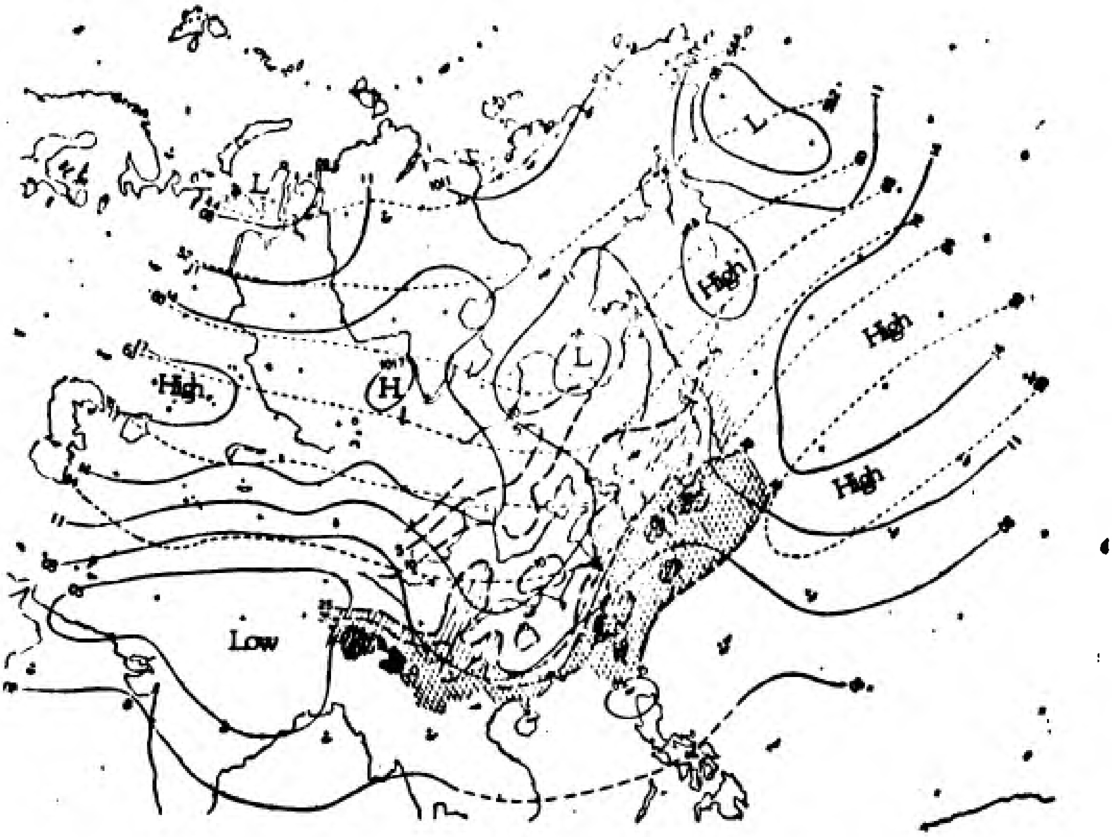

Figure 9

Mean 500-mb contours and mean surface isobars for the period 8 - 17 September.
(Legend same as in Figure 1.)

(g) Early Winter

Early winter lasts about 50 days from mid-October to early December and is characterized by the complete recession of the summer monsoon from the mainland of China. This is followed by a stable monsoon of deep winter at its maximum intensity. At the beginning of the season (Figure 11) a deepening of the Aleutian Low is observed with its center lying some 25°

longitude east of its deep winter position. The Mongolian High also intensifies and is centered about 15° longitude west of its deep winter position. Both the Indian Low and the Pacific High have receded from the mainland of China by this time.

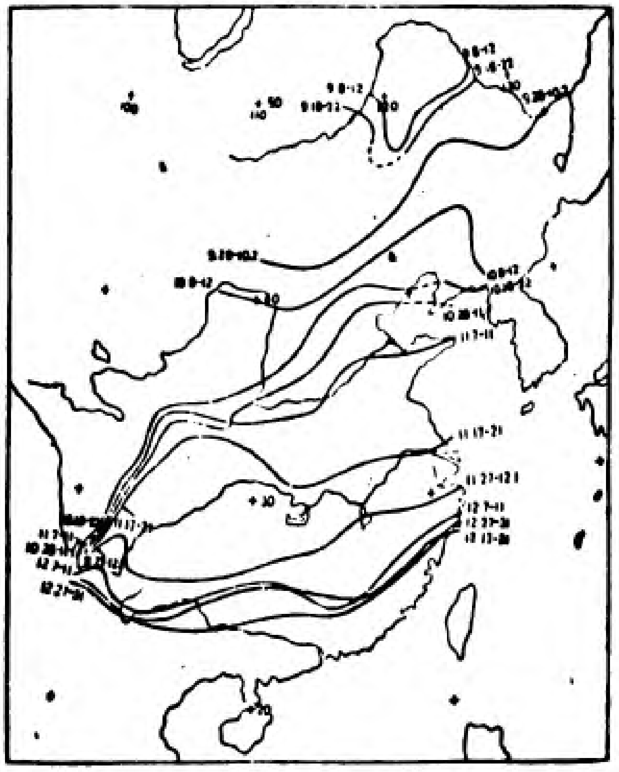

Figure 10

Dates of the southward displacement of the 10°C isotherm of pentad mean temperature.
(Legend same as in Figure 5.)

There is little change in the general circulation at the upper levels except the deepening of the permanent major trough along the coast. The jet stream over China intensifies and moves southward in early December, and becomes steady in its deep winter position at 27°N in the first decade of December.

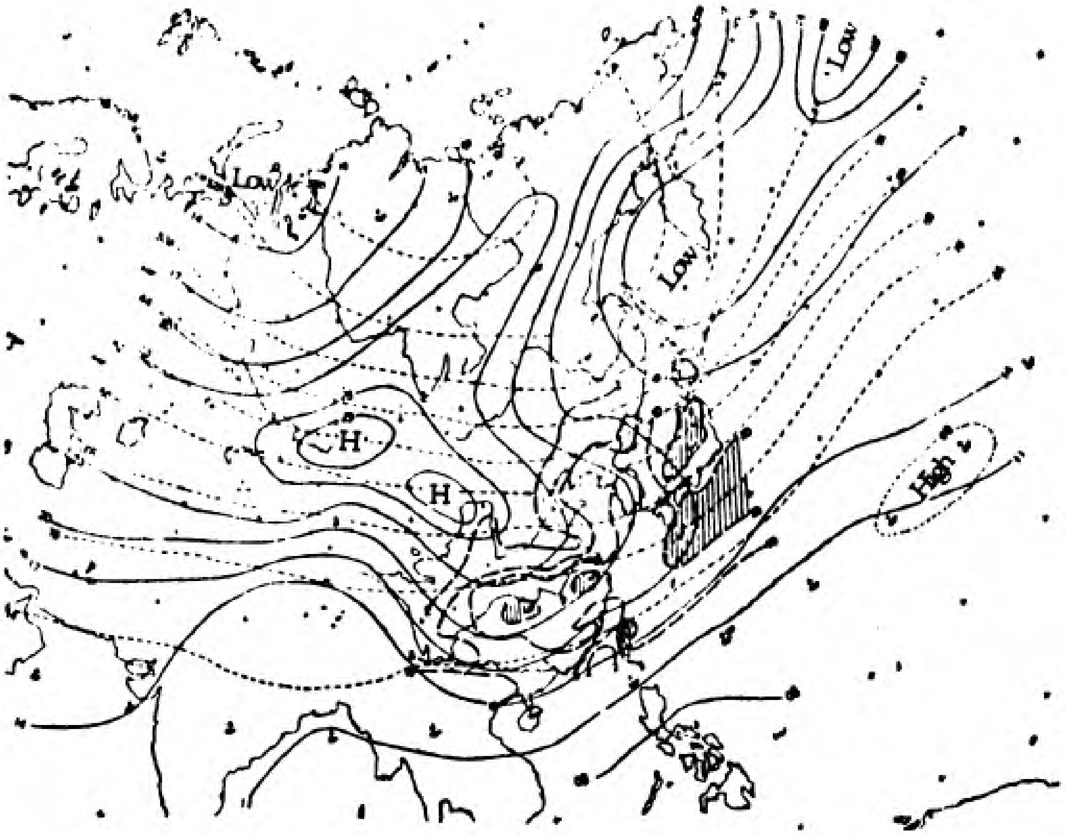

Figure 11

Mean 500-mb contours and mean surface isobars
for the period 2 - 11 November.
(Legend same as in Figure 1.)

III. CONCLUSIVE REMARKS

The march of the monsoons over China may be divided into several distinct stages. The onset of each stage is manifested by abrupt changes in the various meteorological elements and is followed by a definite relatively stable period. Hence each stage may be taken as a synoptic season by itself. A change of synoptic season is often associated by the breakdown of one or two of the existing action centers in the surface layer and accompanied by the formation of new ones. The Indian Low and the Pacific High first appear in early March. The appearance of the "Northeast Depression" together with the establishment of the Indian Low and the Pacific High marks the onset of spring. The beginning of early summer is associated with the formation of a high cell over the Sea of Okhotsk and the dissipation of the Mongolian High and the Aleutian Low while the dissolution of these systems signifies the onset of high summer. The re-establishment of the Mongolian High, the Aleutian Low and the "Northeast Depression" features the commencement of autumn. The breakdown of the Indian Low and the Pacific High signals the approach of early winter, while the onset of deep winter is designated by the filling of the Indian Low and the recession of the Pacific High from China with the Mongolian High and the Aleutian Low in their stationary positions. During the march of the monsoons, the upper-air general circulation over East Asia also undergoes very large variations and transformations. Significant weakening of the westerly circulation and the major trough along the coast occurs in early March. A further weakening of these systems accompanied by the appearance of the "Northeast Depression" takes place in April. The major trough along the coast disappears in June, while another deep trough forms along 90°E. The summer flow pattern is fully developed by mid-July. The major trough re-appears along the coast in early September and this is accompanied by the establishment of westerly circulation north of 25°N. By mid-October, strong upper westerlies prevail over the southern part of Asia and the major trough along the coast deepens and becomes fully established. In early December the westerlies intensify further over the southern part of Asia and remain steady at the latitude of its southernmost position.

The weather conditions over the mainland of China change with each synoptic season. The winter monsoon is intense in deep winter and the pressure field remains stable with little variation. The weather becomes changeable in late winter and highly variable in spring due to the interaction of the four action centers. The influence of tropical and subtropical weather systems begins to come into play in early summer and becomes most pronounced in high summer. In contrast, the effect of the systems in the westerly wind belt is minimum at this time of the year. The four action centers re-appear in autumn, but the weather is more stable than that in spring. Meanwhile the tropical and subtropical synoptic systems still play an important role on the weather over South China. The influence of the tropical pressure systems ends in early winter. It may further be noted that the trajectories of the cold anticyclone and the cold air stream over the mainland of China exhibit definite changes at the onset of each synoptic season. In winter, the meridional tracks are located to the extreme east, while zonal trajectories toward the adjacent seas lie along the southernmost frontier. Following the establishment of the "Northeast Upper Trough" and its subsequent westward displacement as the season advances, the trajectories of the cold anticyclone or the cold air stream shift toward the west and the north. In high summer, the cold air moves southward mainly along the eastern edge of the Tibetan Plateau. The continental anticyclone enters the adjacent seas at the northernmost latitude in autumn, after which time the point of entry gradually shifts southward.

Large variations in pentad rainfall and pentad temperature also occur at the onset of each synoptic season. Thus we are of the opinion that as the synoptic season advances abrupt changes in the pressure system occur first at the surface, bringing about marked variations in weather and climate, but a relatively stable period is reached subsequently. Significant changes in the upper-air flow pattern and the westerly wind circulation are also observed concurrently.

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CONCLUDING REMARKS

The following is a summary of the preliminary findings presented in the previous articles:

I. FORMATION OF THE MONSOONS OVER EAST ASIA

The monsoon phenomenon is considered to be caused by the combined effect of the distribution of land and sea, the planetary circulation and topography. The relative importance of these factors varies with both time and space. For East Asia, the differential heating over land and sea constitutes the primary cause of the monsoons. The seasonal displacement of the planetary wind belt may also be important but it only plays a secondary role over this region. Likewise, the influence of topography is limited to the modification of the regional characteristics of the monsoons and certainly cannot be taken as the main factor for the occurrence of the monsoon phenomenon. The above view is well substantiated by the following observational evidence.

In the first place, it is found from the analyses of the planetary wind belt over the two hemispheres and the variations of the annual pressure and land-sea temperature difference that the planetary circulation is completely broken up at the lower levels over East Asia. Thus monsoons cannot be accounted for solely by the seasonal displacement of the planetary wind belt.

Secondly, investigations on the seasonal variations of pressure and the difference in heating between land and sea, and the effect of land-sea distribution on the low-level pressure field reveal that the formation of the monsoons is fundamentally the result of differential heating over land and sea.

Finally, from studies of the mechanical and thermal effects of the Tibetan Plateau, it is noted that the presence of this huge land mass produces a significant influence on the monsoon climates over China and renders the monsoon phenomenon even more complicated for analysis.

II. CHARACTERISTICS OF MONSOON CLIMATES OVER CHINA

In examining the distribution of pressure, the seasonal variation of the general circulation and the temperature and wind fields at the various levels over East Asia in relation to the monsoon climates over China, the following features may be noted:

(a) Breakdown of the Low-level Planetary Circulation by the Monsoons: The exceptional intense monsoons over East Asia give rise to a complete breakdown of the normal pattern of the low-level planetary circulation. The height of the monsoons extends to 800 - 700 mb in winter and 700 - 500 mb in summer and the planetary flow can only be detected above the monsoons.

(b) Time Lag in the Seasonal Transition of the Flow Pattern between the Upper and the Lower Levels: The intense monsoons exert a particularly strong influence on the pressure field near the surface, resulting in a time lag in the seasonal transition of the flow pattern between the upper and the lower levels. Thus when the winter pressure pattern becomes established near the surface during autumn, summer-type pattern still prevails at the upper levels. Similarly when the continental heat low develops over India and spreads slowly across China during April, no marked change is observed in the upper-air planetary circulation until June.

(c) Disruption of the Normal Low-level Pressure Field by the Intense Summer Monsoon: The pressure distribution over a continent is generally characterized by a maximum in winter and a minimum in summer. However, above a certain level, the opposite configuration is observed and a winter minimum with a summer maximum is noted. This feature is found over North America but in certain regions over East Asia, conditions are found to be different. Over these regions, pressure is highest in the autumn months of September and October instead of summer in the layer between 850 and 500 mb with a secondary maximum in April and May. The pressure value in the high summer months of July and August is on the other hand the second lowest of the year.

(d) Influence of the Monsoons on the Low-level Tropospheric Flow Field over China: The influence of the monsoons is most pronounced in the transitional layer between the monsoons and the planetary circulation. Thus during the period of prevailing monsoons over China, a complex flow configuration is found sandwiched between two simple flow patterns at the lower and upper levels. The height of the transitional layer varies with the season and also with the intensity of the monsoons and the planetary circulation.

(e) Influence of the Monsoons on the Temperature Field in the Troposphere over China: Monsoons over East Asia tend to increase the north-south temperature gradient. In winter, the intense winter monsoon reaches relatively low latitudes and the meridional temperature gradient is larger in the south than in the north. The opposite is found in summer when the southerly monsoon winds give rise to a tighter gradient in the north.

All the features discussed above are important factors in controlling the weather and climate of China. It should be pointed out here that in routine synoptic analysis, the 700-mb level may not be high enough to reveal the variations of the planetary circulation over the country in winter. In summer, the configuration of the upper-level planetary circulation is hardly apparent at 500 mb, particularly over regions south of 30°N.

From the study of the seasonal variation of the prevailing airstreams and the extent of their influence, the vast territory of China has been classified into five climatic regions, namely the Westerly Wind Belt, the Subtropical Monsoon Region, the Tropical Monsoon Region, the Equatorial Monsoon Region and the Tibetan Plateau Monsoon Region. The Subtropical Monsoon Region covers a wide range of latitude with a multitude of complex topographical features so that it may further be sub-divided into five smaller regions in accordance with the intensity and stability of the monsoons and also the difference in relative humidity between the two months of January and July, namely, the Northeast Region, the North China Region, the Lower and Middle Yangtze Basin, the River-bend Region and the Szechwan Basin. A description of the climatic conditions in relation to the activities of the monsoons has also been made for each region.

III. ADVANCE AND RETREAT OF MONSOONS

The advance and retreat of monsoons over the various regions have been discussed in terms of the evolutionary processes of the four major action centers over East Asia (i.e., the Mongolian High, the Aleutian Low, the Indian Low and the Pacific High) and the seasonal variation of the pentad mean land-sea pressure gradient at the various levels. The onset of the rainy season in relation to the advance and retreat of the monsoons and the seasonal variation of the upper westerly wind circulation has also been examined.

It is noted that the march of the summer and winter monsoons over China takes place in seven distinct stages. Each phase of activities necessarily gives rise to a corresponding change in weather and other meteorological elements, particularly rainfall. Thus a large rain belt first appears over the coastal area of South China in mid-May and moves systematically northward. By early June, it reaches the border between Fukien and Chekiang Provinces to the north of the Nan Ling Ranges. During mid-June, the rain belt shows a marked "jump" and rapidly sweeps northward across the low-lying basins of Tungting Hu (Lake) and Poyang Hu (Lake) to the Yangtze. A second "jump" is noted in mid-July when the rain belt moves across Hwai Ho to the middle and lower valleys of the Yellow River. The rain belt reaches its northernmost position in mid-August and by the end of the month or early September, it begins to retreat rapidly southward again. The latitudinal displacement of the rain belt can therefore be used to accurately mark the onset or termination of the monsoons.

The latitudinal movement of the pentad rain belt takes the form of "jumps" alternating with more gradual displacements. The three jumps observed during each cycle are all in phase with the advance or retreat of the monsoons. The onset and termination of the rainy season in South and Northeast China and the period of mei-yü over the Yangtze - Hwai Ho region are related to the "slow-moving" phase of the rain belt.

The advance and retreat of the monsoons are also found to depend on the position, intensity and behavior of the surface action centers. The Mongolian High and the Aleutian Low diminish in intensity during early March while the Indian Low and the Pacific Anticyclone begin to appear on the surface chart during this period. By mid-April, a further weakening of the Mongolian High and the Aleutian Low is observed, which is accompanied by the full establishment of the Indian Low and the Pacific High. The two high-latitude centers dissipate in mid-June and the two low-latitude systems become the dominant features over East Asia, which reach their peak intensity in mid-July. By September, both the Mongolian High and the Aleutian Low re-appear on the chart and assume control over the whole region after mid-October.

A significant advance or retreat of the monsoons is generally associated with a corresponding change in the position and intensity of the westerly circulation which can normally be traced in the 500-mb pattern. Variations in the westerlies likewise affect the horizontal extent of the low-level monsoons.

The advance or retreat of the monsoons may either be abrupt or gradual in character. Rapid development is noted in mid-June, mid-July and early September when a "qualitative" change takes place in the high-level westerlies in the form of a marked displacement of the wind belt in the upper troposphere. A complete reversal of the wind regime (i.e., a change from easterlies to westerlies or vice versa) is also observed near 20°N in June, to the south of

30°N in mid-July and in early September. Hence, the seasonal variation of the planetary circulation should be carefully examined when preparing long-range forecasts on the advance or retreat of the monsoons.

Finally, the advance and retreat of the monsoons have been related to the march of the natural seasons over China. It is noted that each stage of the monsoon activities is associated with a relatively stable period of homogeneous climatological characteristics. From the study of the variations of these characteristics and the evolutionary processes of the action centers and the upper circulation, it has been possible to divide the year into seven natural seasons over East China, namely, deep winter, late winter, spring, early summer, high summer, autumn and early winter.

The conclusions presented above can only be considered as preliminary. Further research is needed to study the annual variation of the configuration and intensity of the monsoon pressure field by means of adequate upper-air and oceanic data. The influence of the monsoons on the occurrence of floods and droughts also requires to be examined. The physical processes responsible for the formation and development of the monsoons and their evolution should be thoroughly investigated in order that a better understanding of the monsoon phenomenon can be achieved. The results may then provide useful guidance for the selection of appropriate predictors for preparing medium- and long-range forecasts.

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