ROLE OF THE CIRCULATION AT THE 150 MB LEVEL IN THE WINTER AND SUMMER MONSOON IN THE ASIAN AND AUSTRALIAN REGIONS

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ABSTRACT

The fluctuations of the winter and summer tropical monsoon circulation in the Asian and Australian regions were investigated by the examination of the circulation at the 150 mb level. The zonal wind data at the 150 mb level at Singapore for a 18 year period from 1961 to 1978 in the months from December to February were used to represent the fluctuation of the winter monsoon. In the case of the summer monsoon, the wind speed at the 150 mb level at 10°N from 40°E to 110°E was read from the analyzed maps and used to represent the fluctuations of the tropical easterly jet stream (TEJ) for a period from 1964 to 1977 in the months from June to August. The analysis is based on the comparison of the composite maps of the precipitation and monsoon circulation at the low level in the cases of the strong and weak circulation at the 150 mb level. The relationship to the other features of the general circulation of the atmosphere such as the Walker circulation and the circumpolar westerlies was also analyzed.

The results of the analysis have shown that in the case of the winter monsoon, the fluctuations of the monsoon are the part of the main fluctuations of the Walker circulation. This circulation consists of the zonal oscillation of the equatorial trough and the sea surface temperature in the equatorial Pacific. When the monsoon is strong, the regions of active double ITCZs with heavy precipitation are located near 140°E and the equatorial Pacific is cold. On the other hand, when the monsoon is weak, the region of active ITCZ shifts east to 170°E and the equatorial Pacific is warm.

In the case of summer monsoon, the poleward transport of heat by the circumpolar westerlies plays an important role. When the monsoon is strong, the circulation near 50°N is zonal. On the other hand, weak monsoon is associated with a blocking high to the north of the Caspian Sea and trough to east of Lake Baikal. The transport of the heat toward pole by the standing eddies increases in these years and this surpress the development of the South Asian high and the monsoon circulation. The Walker circulation was found to influence the monsoon circulation only during the years with a major El Niño when the TEJ was also unusually weak.

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ABSTRACT

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CHAPTER 1

INTRODUCTION

1-1 Purpose of the study

The region extending from west Africa (20°W) eastward through southern Asia into adjacent west Pacific (as far as 130°E) and the nothern coast of Australia experiences a seasonal reversal of wind direction called the 'Monsoon'. Simpson(1921) in his review of the monsoon has stated that the word monsoon itself was derived from the Arabic name for a season and this word was applied to the winds over the Arabian Sea which blow for approximately six months from the northeast and for six months from the southwest.

The recent developments of the rawinsonde network have shown us a three dimensional structure of the atmosphere. In the middle and high latitudes; poleward of the subtropical jet stream a Rossby regime dominated by the westerly waves at the 500 mb level and the travelling cyclones and anticyclones at the surface level prevails. On the other hand, in the tropical and subtropical latitudes; between the subtropical jet streams of both hemispheres, a Hadley regime dominated by rising warm air at the zone of the heavy rainfall and subsiding cooler air near the subtropical jet is observed.

In the oceanic regions, the seasonal migration of the Hadley regime is small. On the other hand, in the continental regions, because of the large annual range of the surface temperature, the seasonal migration of the Hadley regime is large enough to produce a seasonal reversal of wind direction. Figure 1 shows the vertical structure of the Asian summer monsoon. Below the 500 mb level, there are a low pressure in south Asia and the sub-

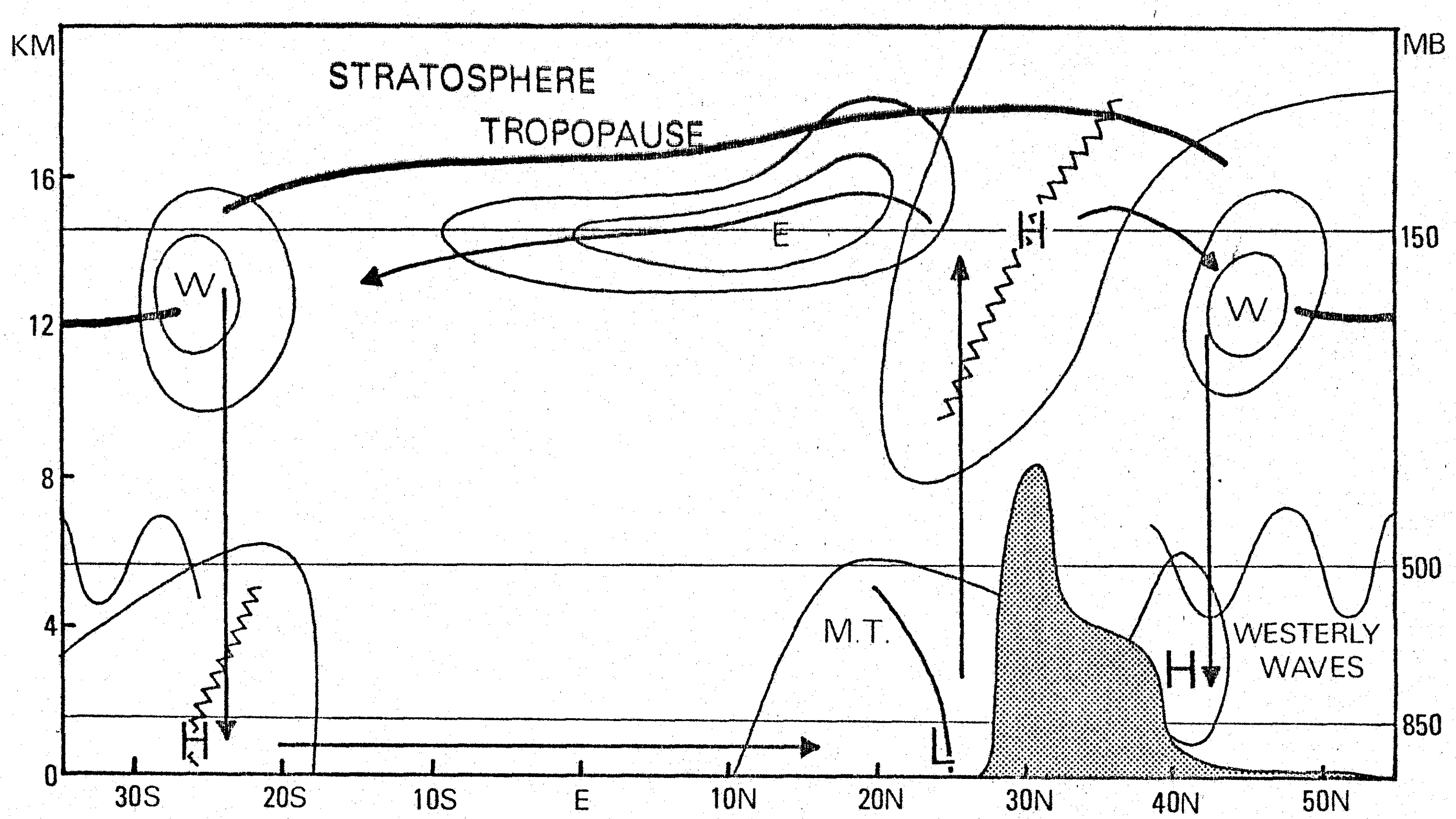

Fig. 1 Vertical structure of the Asian summer monsoon
(M.T.: Monsoon trough, E: Tropical easterly jet,
E: Subtropical westerly jet)

tropical anticyclones near 25°S. The low level currents shown by the arrow below the 850 mb level are the southeast trades to the south of the equator and the southwest monsoon to the north of the equator The rising warm moist air over southern Asia near the monsoon trough releases the latent heat of the fusion. For this reason, unlike the deserts of north Africa, the entire troposphere below the 150 mb level is warmer than the surrounding atmosphere.

Since the large atmospheric structure is approximately hydrostatic, the large thickness of the warm atmosphere results in the formation of the high pressure in the upper troposphere near the 150 mb level (South Asian high). Since the stratosphere is extremely stable, the rising warm air which constitutes the ascending branch of the Hadley cell could not penetrate deeply into stratosphere. Thus these air diverges from the South Asian high near the 150 mb level. The northward branch subsides just south of the subtropical jet stream located near 45°N and centered at the 200 mb level. The southward return flow to the southern hemisphere is deflected by the Coriolis force and is observed as a northeasterly flow near the 150 mb level. This flow is strong enough to form a tropical easterly jet. When the return flow reaches the subtropical jet stream in the southern hemisphere, it undergoes subsidence and maintains subtropical anticyclones near 25°S. Hence the Asian summer monsoon is a Hadley cell displaced northward to 25°N. Poleward of the subtropical jet stream, the Rossby regime dominated by the waves at the 500 mb level prevails.

The purpose of this research is to investigate the interannual fluctuations of the tropical winter and summer monsoons. The mechanism of these fluctuations and its relationship to the monsoon circulation at the low level are the main topic of the investigations. Because the monsoon circulation at the low level is subject to topographical distortion, the return flow at the 150 mb level was used to monitor the fluctuations of the tropical monsoon circulation.

1-2 Review of the recent research

From the previous section, it is clear that the monsoon circulation is a part of the general circulation of the atmosphere. For this reason, this review will be limited to the research works which have treated the tropical monsoon as a part of the general circulation of the atmosphere. Research works which have treated the synoptic components of the monsoon such as the monsoon trough in the Ganges valley of India will be treated as necessary in the later sections. Since most of the research works have analyzed the summer monsoon, this review will begin with a summary of the summer monsoon. This will be followed by a shorter review of the winter monsoon. Finally a summary of the numerical simulations of the monsoon circulation will be presented.

a) Summer monsoon

The first important paper utilizing the data from the rawinsonde network was completed by Yin (1949) who has analyzed the onset of the Indian summer monsoon in 1946. Using the data at the 5,000 feet, 10,000 feet, and 20,000 feet levels, which correspond to 1,500, 3,000 and 6,000 m respectively, he has concluded that the onset of the monsoon rainfall in India coincides with the shift of the subtropical jet to the north of the Himalaya. Thompson (1951) has analyzed the winter and summer monsoon circulation up to 10,000 feet in the southeast Asian and the west Pacific regions. This was the first comprehensive analysis of the monsoon circulation. Utilizing wind data as a criterion, he has identified several convergence zones in tropical southeast Asia. Some of these convergence zones such as the West Pacific convergence zone, continue to generate controversy with regards to their existences and locations.

The works during the decade of the late 1940s and early 1950s were based on the primitive radiosonde and pilot baloon data which covered only the lower half of the troposphere. The subsequent advances in observational technology have extended the observation to the upper troposphere. Utilizing these new data, a series of important discoveries were made. Koteswaram (1958) has analyzed the upper tropospheric circulation during the summer of 1955 over southern Asia and adjacent north Africa and discovered the tropical easterly jet stream. The level of strongest winds is located at the 150-100 mb level at the latitude of 15°N over southern Asia. He has suggested that there should be a close connection between fluctuations in the upper easterly current and fluctuations in the low level monsoon. This problem is the main topic of the current investigation. The latitude of the core of the jet stream was found to exist near 10°N in later investigations using the new data from southern India and the Indian ocean. Mason and Anderson (1963) have used IGY data and analyzed the high level anticyclone at the 100 mb level over southern Asia. They have suggested that there are considerable year to year fluctuations in the position and intensity of this anticyclone.

Flohn (1964) has done an extensive study of the nature of the tropical easterly jet. This jet is located in the layer 200-100 mb level from 5°N to 20°N with a core of maximum wind at 150 mb level near the southern tip of the Indian peninsula. From a mean climatological map of the jet, he suggested the existence of an ageostrophic cross circulation. In the entrance region (80°-150°E) there is a thermally direct cell with uplift to the north of the jet axis and subsidence to the south. In the exit region (20°W-70°E) there is an indirect cell with subsidence to the north of the jet axis and uplift to the south. Subbaramayya and Ramanadham (1966) have made a three dimensional diagram of the summer monsoon in Asia which illustrates the tropical easterly jet as a return flow of the Asian summer monsoon to the southern hemisphere.

The first study of the dynamical processes which maintain the tropical easterly jet was carried out by Keshavamurti (1968). He has used a mean wind chart at the 200 mb and 700 mb levels and computed an angular momentum balance of the mean climatological circulation in southern Asia and Indian Ocean north of the Equator. It was shown that the mean meridional circulation (which is direct circulation) maintains the tropical easterly jet.

The importance of the zonal asymmetry of the tropical general circulation was shown by Krishnamurti (1971). The velocity potential at the 200 mb level and the temperature at the 300 mb level over the tropical atmosphere between 45°N and 25°S during a period from June to August 1967 has clearly shown the existence of a thermally direct east-west circulation associated with the monsoon. This circulation has ascending warm air over Burma and descending cold air over the oceanic tropics. The intensity of this circulation was found to be comparable to that of Hadley type circulation. This study has shown that the classical model of the Hadley cell of the tropical troposphere is unrealistic. The monsoon circulation is a major component of the tropical general circulation.

At this point two important research work on the tropical climatology were completed. The first work is a comprehensive review of the regional and synoptic climatology of the tropical monsoon (Ramage, 1971). The second work is an atlas of the general circulation of the tropical atmosphere (Newell et al.,1972). This work gives an excellent view of the vertical structure of the tropical atmosphere which is fundamentaly different from that of the middle latitudes. These research works have essentially completed the investigation of the mean climatological conditions of the Asian summer monsoon.

In the recent years, with the approach of the Monsoon Experiment (MONEX) the emphasis of the research has shifted to the fluctuations of the monsoon system. Krishnamurti and Bhalme (1976) have analyzed the daily observational records in the years 1957,1962 and 1967. Nine elements of the monsoon system were analyzed. These are: Monsoon trough of northern India at the sea level, Mascarene high at the sea level, Somali jet, Tibetan high at the 200 mb level, tropical easterly jet, monsoon cloud cover using satellite brightness data in 1967, monsoon rainfall in central India, moist and dry static energy in central India. The spectral analysis of these nine elemént was conducted which has shown a quasi-biweekly period oscillation of about 14 $\pm$ 2 days for all of the nine elements. The possible reasons for these oscillations are currently under investigation.

Kanamitsu and Krishnamurti (1978) have compared two contrasting years of the tropical circulation in the summer. The summer circulations with normal rainfall in central India in 1967 were compared with widespread drought in the tropical area in 1972.

The circulation in 1967 is the same as Krishnamurti (1971) with well developed east-west circulation with ascending warm air over Burma and descending cold air over the oceanic tropics. On the other hand, in the drought year of 1972, the tropical circulation in the upper troposphere was displaced some 10° of latitude to southeast over the Asian monsoon region. In addition, the tropical easterly jet and east-west circulation in the Pacific area were much weaker than the normals. Hence the weaker monsoon circulation were partially offset by intense Hadley cell in the central Pacific Ocean.

b) Winter monsoon

In contrast to the summer monsoon, there are only few researches on the circulations of the tropical monsoon in the winter. Ramage (1968) analyzed the winter monsoon circulation and associated precipitation during two Januaries of 1963 and 1964. In 1963, strong easterly winds were observed near Singapore and precipitation was heavy in North Borneo and light in New Guinea. On the other hand, in January of 1964, opposite conditions were observed with the weak easterlies near Singapore.

Krishnamurti et al. (1973) have computed the velocity potential similar to that of Krishnamurti (1971) in the northern winter of 1969. The ascending branch of the east-west circulation was located over the Indonesian region where the winter monsoon is most intense while the descending branch is located in the central Pacific. The Hadley cell which is zonally symmetric prevails in the other areas of the tropics.

Murakami and Unninayar (1977) have analyzed the tropical general circulation during the northern winter from December 1970 through February 1971. This study has shown that the strong winter monsoon is associated with a large eddy kinetic energy at the 200 mb level while the contrasting situation with weak monsoon occurs when the eddy kinetic energy is small. They have suggested that the nature of the winter circulation is different from one winter to another and even within the same winter.

c) Numerical simulation of the monsoon

The numerical models of the general circulation of the atmosphere are still in primitive stages of development. For this reason, only two of the most recent researches will be presented here. Hahn and Manabe (1975) have used the 11 level GFDL numerical model of the atmospheric circulation and conducted a simulation of the Asian summer monsoon with emphasis on the role of the Himalaya mountains. The simulation with the mountains produced a south Asian low pressure slightly to the east of the climatological position. The simulated atmosphere is dynamically too active in the western Pacific, while in the northern Bay of Bengal and northern India it is relatively inactive. When the mountains were removed from the model, the Tibetan high shifts east and is located to the east of Taiwan and the Indian monsoon circulation is weak. Thus it was concluded that the presence of the Himalaya mountains enhance the Asian monsoon circulation.

Washington (1976) has used the NCAR model and simulated the winter monsoon circulation over Asia and Africa. Although the seasonal reversal of the Somali jet is well simulated, the computed circulation near the East China Sea is unrealistic and in fact resembles summer circulation.

d) Current scientific problems of the monsoon research

The current status of monsoon researches can be summarized as follows: With regard to the climatological mean structure of the large scale aspects of the monsoon in the Asian region, the research is nearly completed. On the other hand, the analysis of the interannual fluctuations is hampered by the lack of three dimensional data of high quality using satellite, aircraft, and rawinsonde data covering at least 10 years or more. The numerical simulation of the monsoon circulation by the general circulation model have just started. The difference between the actual and simulated monsoon circulation is far larger than the interannual fluctuations of this circulation. For this reason, it will be some time in the future that the realistic researches of the simulation of the year to year fluctuations can be undertaken.

The monsoon experiment (MONEX) is currently being carried out to obtain data of high quality for one monsoon year (winter 1978-1979 and summer 1979). This will probably yield many valuable discoveries concerning the short period oscillations of the monsoon circulation. However, many years of these type of data are necessary in order that any dynamical study of the interannual fluctuations of the monsoon might be carried out.

1-3 Method of the study

a) Data source and period of analysis

The method of the current study is closely related to the scientific problems of the monsoon research which were summarized at the end of the last section. Although data of high quality are not readily available for many consecutive years, the quality of the available rawinsonde data is sufficient for a synoptic climatological study if the density of the observation network and wind data up to the 100 mb level are adequate.

An inspection of the data in the "Monthly Climatic Data for the World"published regulary by National Oceanic and Atmospheric Administration has shown that in the case of the summer monsoon a 14 year period from 1964 to 1977 can be used for the research. In the case of the winter monsoon, an 18 year period from 1961 to 1978 was found to be useful. This publication also contains monthly data at the surface level such as the pressure, temperature, vapor pressure at the sea level, and monthly total precipitations for about 400 to 500 stations in the world.

Figure 2 shows the locations of the upper air stations used for the study of the winter monsoon. There are approximately 60 stations in the domain of the current research. The stations with open circles have only 2 to 3 years of data. Figure 3 shows the locations of the precipitation stations used for the study of the winter monsoon. Major gaps in the data are located in the People's Republic of China and Indonesia. There are approximately 120 stations depending on the years of data.

Figure 4 shows the locations of the upper air stations used for the study of the summer monsoon. There are approximately 80 to 100 stations in the domain of the current research. The stations with open circles have only 2 to 3 years of data. Figure 5 shows the locations of the precipitation stations used for the study of the summer monsoon. Major gaps in the continental regions are again located in the People's Republic of China and Indonesia. There are approximately 200 stations in the domain of the current research.

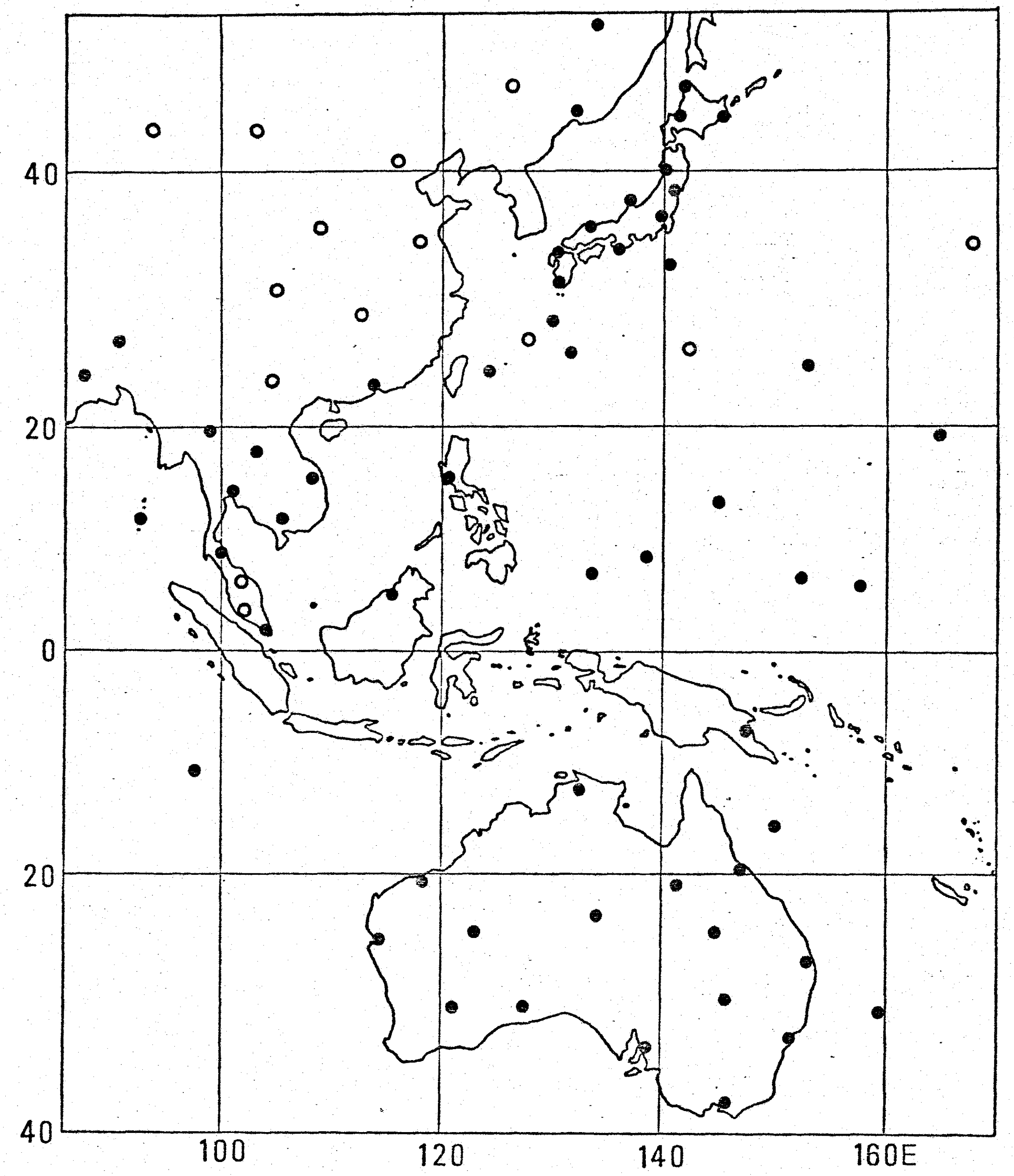

Fig. 2 Aerological stations used for the study of the winter monsoon

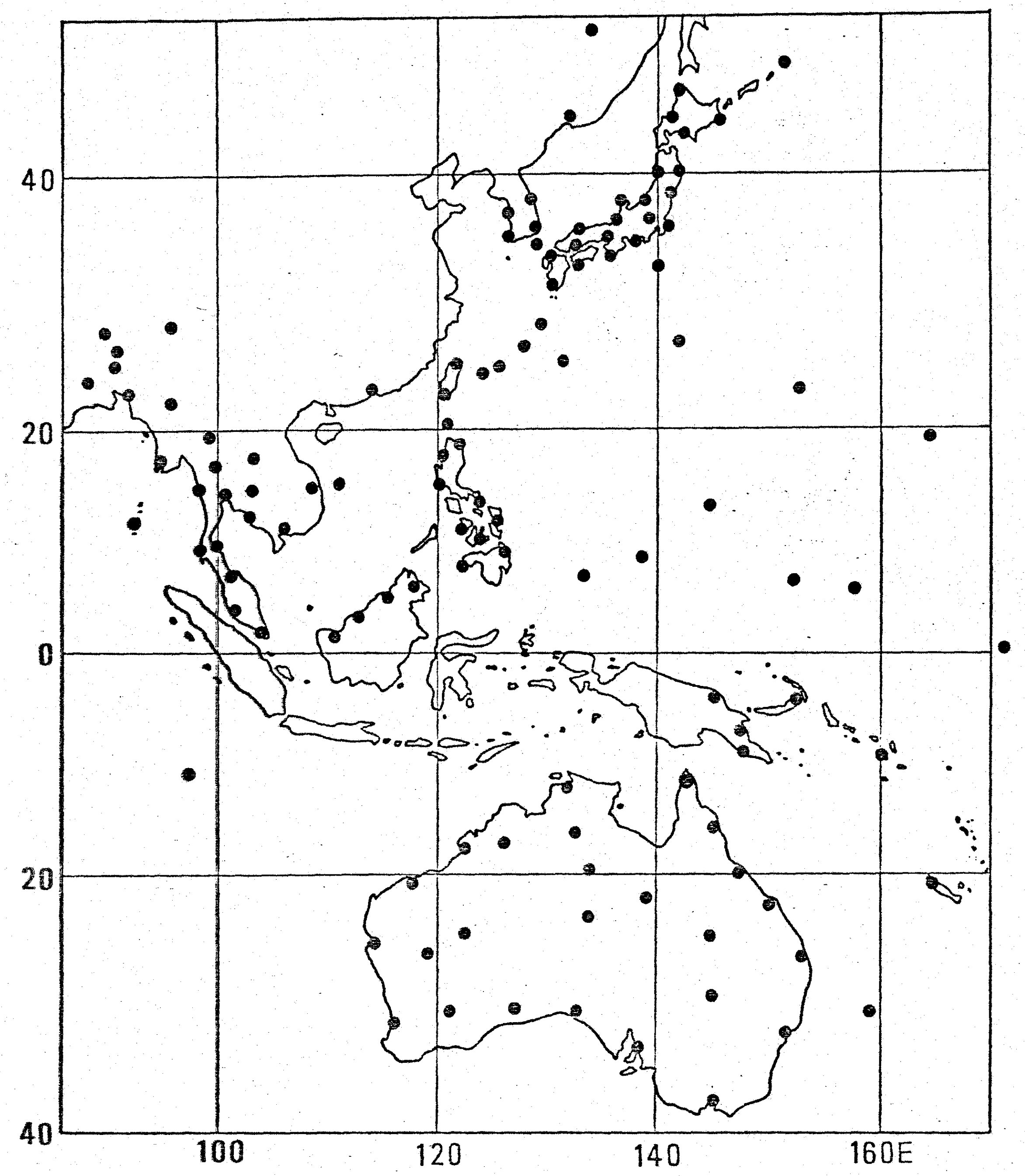

Fig. 3 Precipitation stations used for the study of the **winter** monsoon

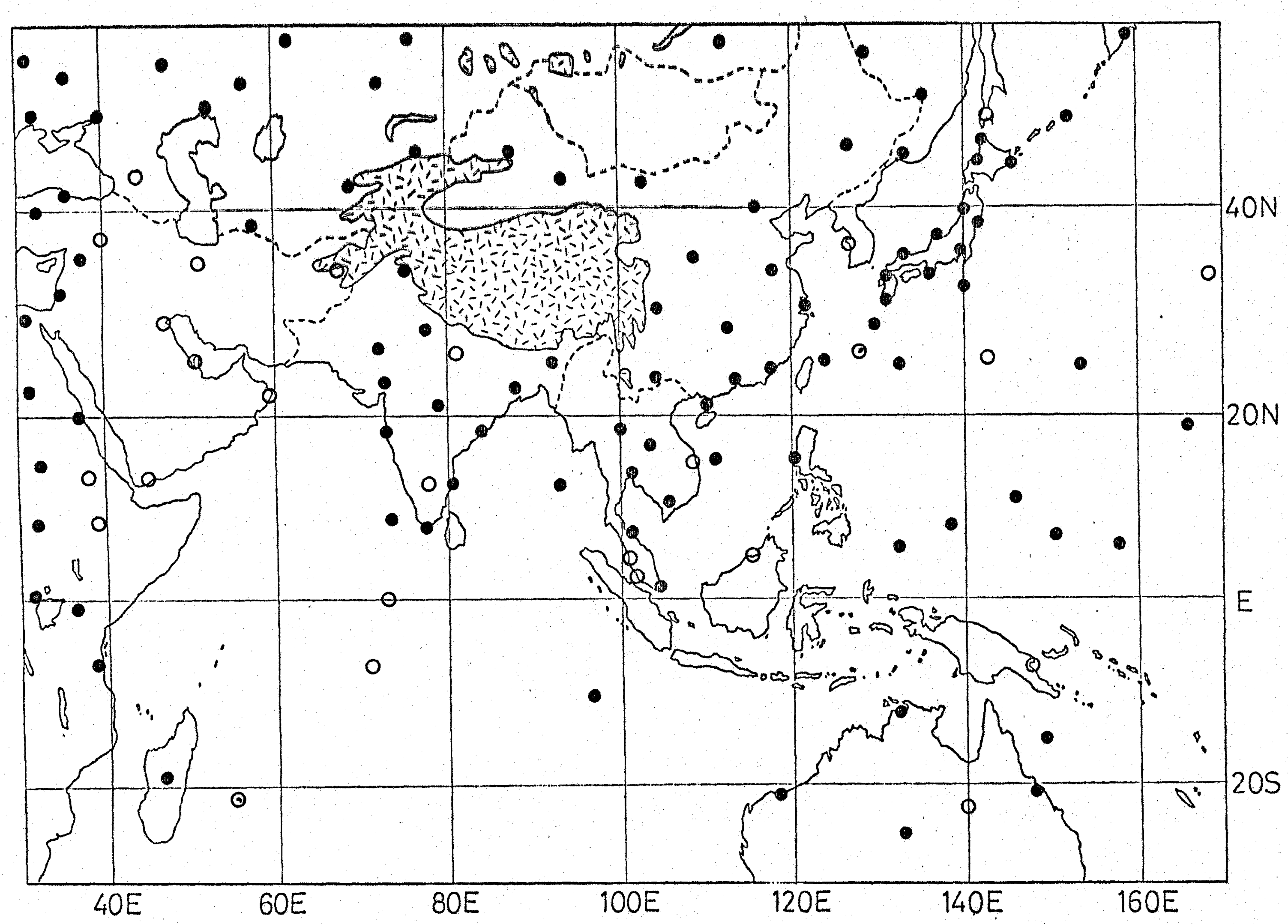

Fig. 4 Aerological stations used for the study of the summer monsoon

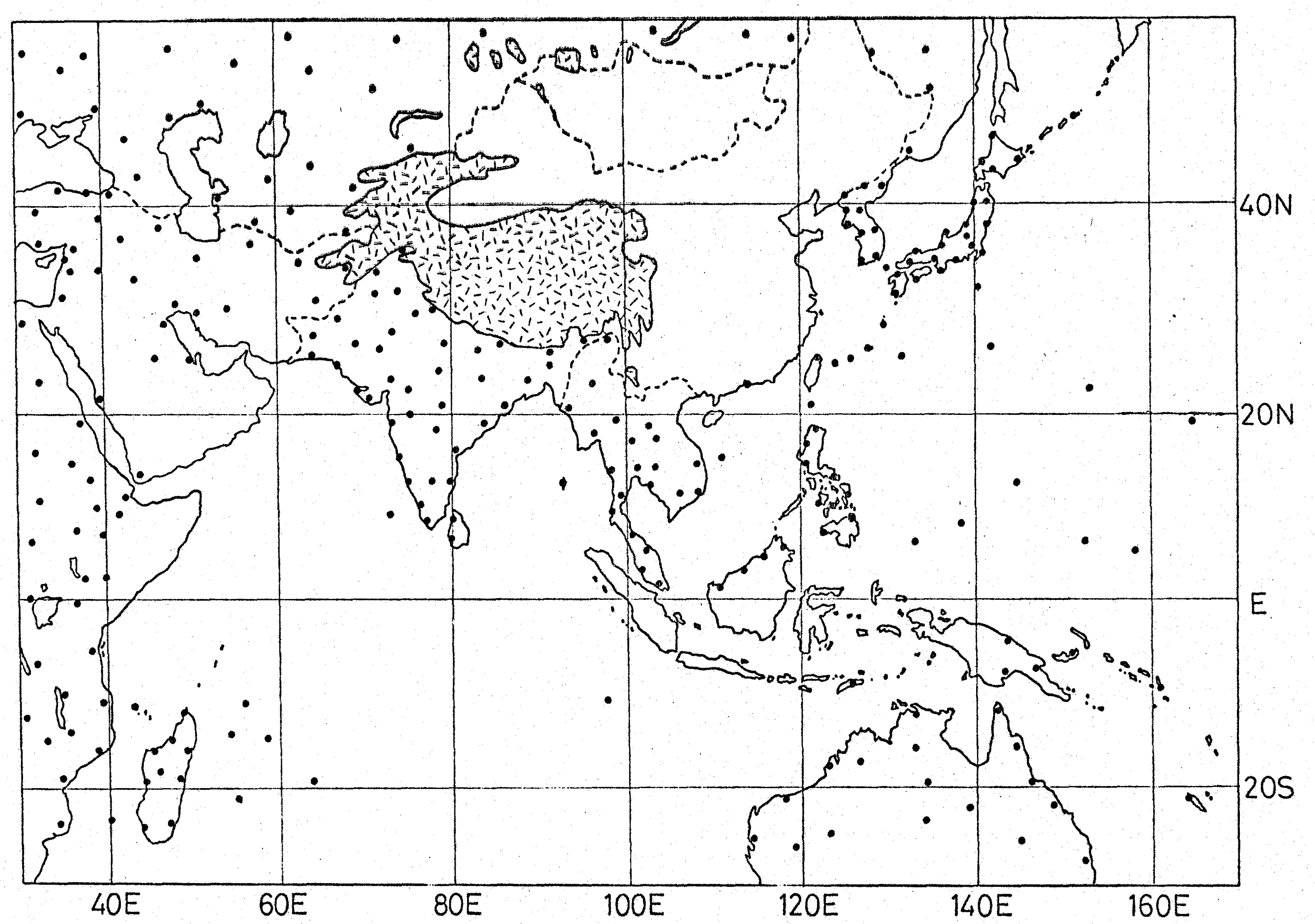

Fig. 5 Precipitation stations used for the study of the summer monsoon

Although most of the data are from the "Monthly Climatic Data for the World", the upper air data for the People's Republic of China for the study of the summer monsoon are obtained from the "Daily Weather Map" published by the Japan Meteorological Agency. In addition, the data for Chichijima, Minamitorishima (Marcus Island), and Clark air force base (Philippines), Osan air force base (Republic of Korea) are obtained from the "Aerological Data of Japan" published by the Japan Meteorological Agency.

b) Method of analysis

In every year of the research period (1961 to 1978 in the case of the winter monsoon), the monthly synoptic maps of the total precipitation, the circulation at the 850 mb level, and the circulation at the 150 mb level were drawn. The circulation at the 150 mb level was used to represent the intensity of the monsoon circulation for a given month. A review of Figure 1 shows that this circulation at the high level is a return flow of the monsoon circulation at the low level.

On the other hand, the circulation at the 850 mb level was analyzed to investigate the response of the monsoon circulation at the low level to the fluctuations of the large scale monsoon represented by the wind at the 150 mb level. The location of the ITCZ, the activities of typhoons, and Baiu fronts may be coupled with the fluctuations of the monsoon. The 850 mb level was chosen instead of pressure at the sea level, because it was felt that the sufficient depth of the current at the low level was required to be considered as a major monsoon circulation at the low level.

Finally, the analysis of the monthly total precipitation has two important purposes. The first obvious reason is the relationship of drought and flood to the monsoon circulation. These fluctuation has a major impact on the food production in Asia (Tanaka, 1978). The second reason is the role of the latent heat release in the maintenance of the monsoon circulation. The direct heating of the surface of the earth by insolation is not sufficient to initiate the monsoon circulation. The best example is the case of the Sahara desert where the surface temperature is highest in the world during the months of July and August, but the tropospheric temperature at 5 to 10 km high is colder than the region of ITCZ located near 10°N (Tanaka, 1976).

The fluctuations of the circulation at the 150 mb level were monitored by the following method. In the case of winter monsoon, the wind at the 150 mb level at Singapore was used to represent the monsoon circulation. In the case of the summer monsoon, the intensity of the tropical easterly jet stream (TEJ) was measured at every ten degrees of longitude from 40°E to 110°E at 10°N. The average of these eight values was used to represent the monsoon circulation.

The structures of the circulation of the strong and weak monsoon were compared by construction of the two sets of the composite maps. In the case of the winter monsoon, the composite maps of 6 Januaries with strong monsoon circulation and 6 Januaries with weak monsoon circulation were made. The difference between the two composite maps was then analyzed (strong minus weak monsoon). This map of the difference clearly shows the change in the circulation regime associated with the change in the intensity of monsoon circulation. Similar procedure was followed in the case of the summer monsoon. The results of this investigation are discussed in Chapters 3 and 6.

Finally, the relationship to the general circulation was investigated. Three important features of the general circulation were chosen to analyze the relationship. These are the circumpolar westerlies in the northern hemisphere at the 500 mb level, the Walker circulation, and the east-west circulation in the case of summer monsoon. The results of this survey are discussed in Chapters 4 and 7.

CHAPTER 2

GENERAL DESCRIPTION OF THE WINTER MONSOON

2-1 Climatology

In this section the mean climatological conditions during the peak of the winter monsoon are discussed. The seasonal change to the winter monsoon regime is described in the next section. Figure 6 shows the mean precipitation in January. Major regions of heavy precipitation over 200 mm are located within 10° latitude of the equator. These are associated with the location of the ITCZ. A closer inspection of the equatorial rainbelt shows the tendencies for formation of the two ITCZs located in Micronesia and northern Australia respectively. Two dry regions associated with the subtropical high are located near 20°N and 30°S respectively. In the extratropical regions, there is a belt of heavy precipitaion over the Kuroshio current where activities of the polar fronts are enhanced by continuous supplies of heat and moisture of warm ocean. In Japan, there are regions of heavy orographic precipitation in Hokuriku and a dry rainshadow on the Pacific side caused by an persistant northwesterly monsoon from cold dry Siberia.

Figure 7 shows the mean geopotential height at the 850 mb level in January. The Siberian high which is strong at the surface level has diminished its intensity and is located near the eastern edge of Tibetan plateau. The subtropical high is located near 20°N and 30°S respectively and coincides with the zone of the low precipitation. The hatched line over northern Australia is a southern ITCZ which separates the equatorial westerlies and the southeast trade winds. The northern ITCZ which is less conspicuous is located just north of the equator. The trough over southern

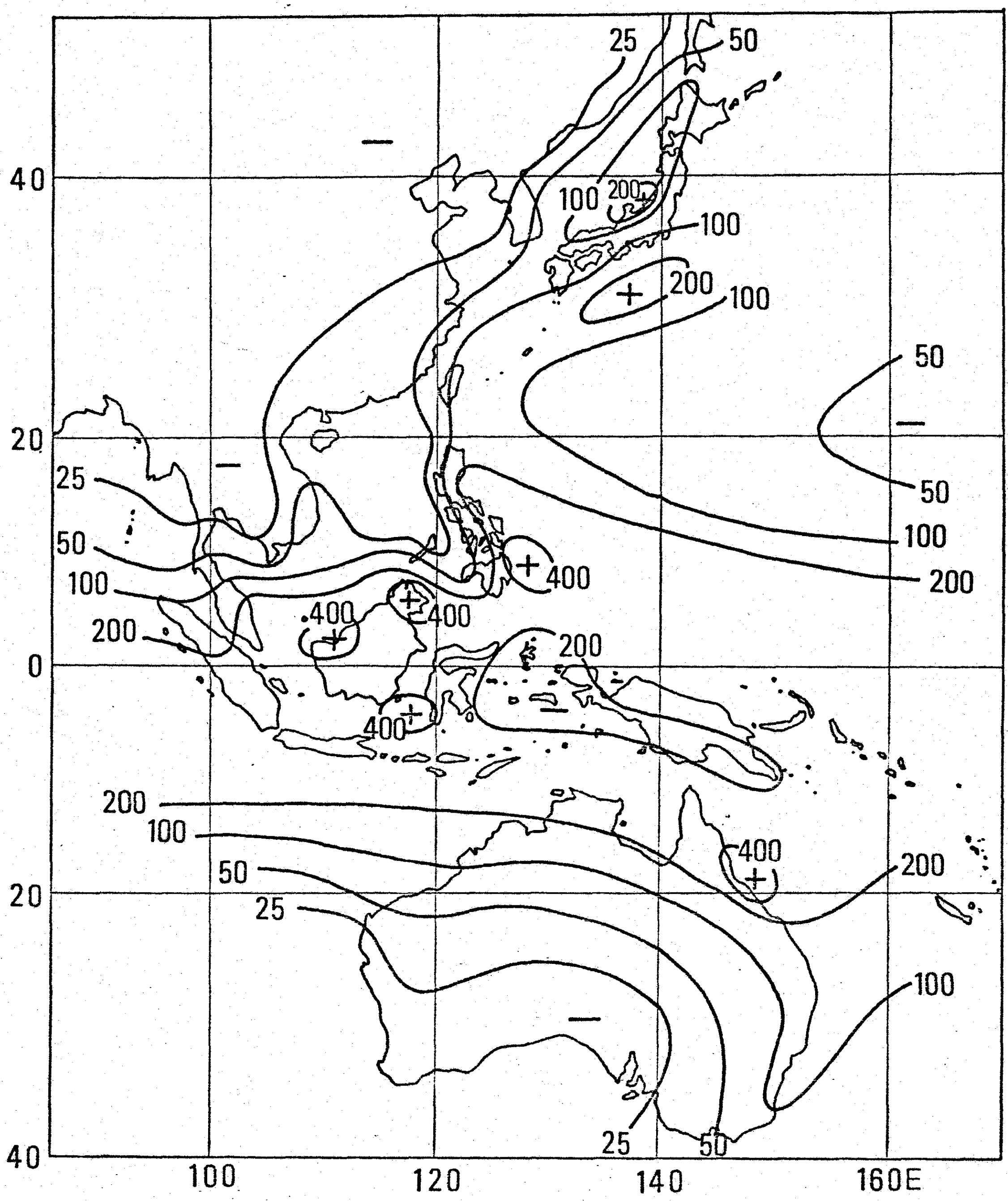

Fig. 6 Precipitation in January (1951-1960 average)

(unit: mm)

(data adopted from World Weather Records, 1967)

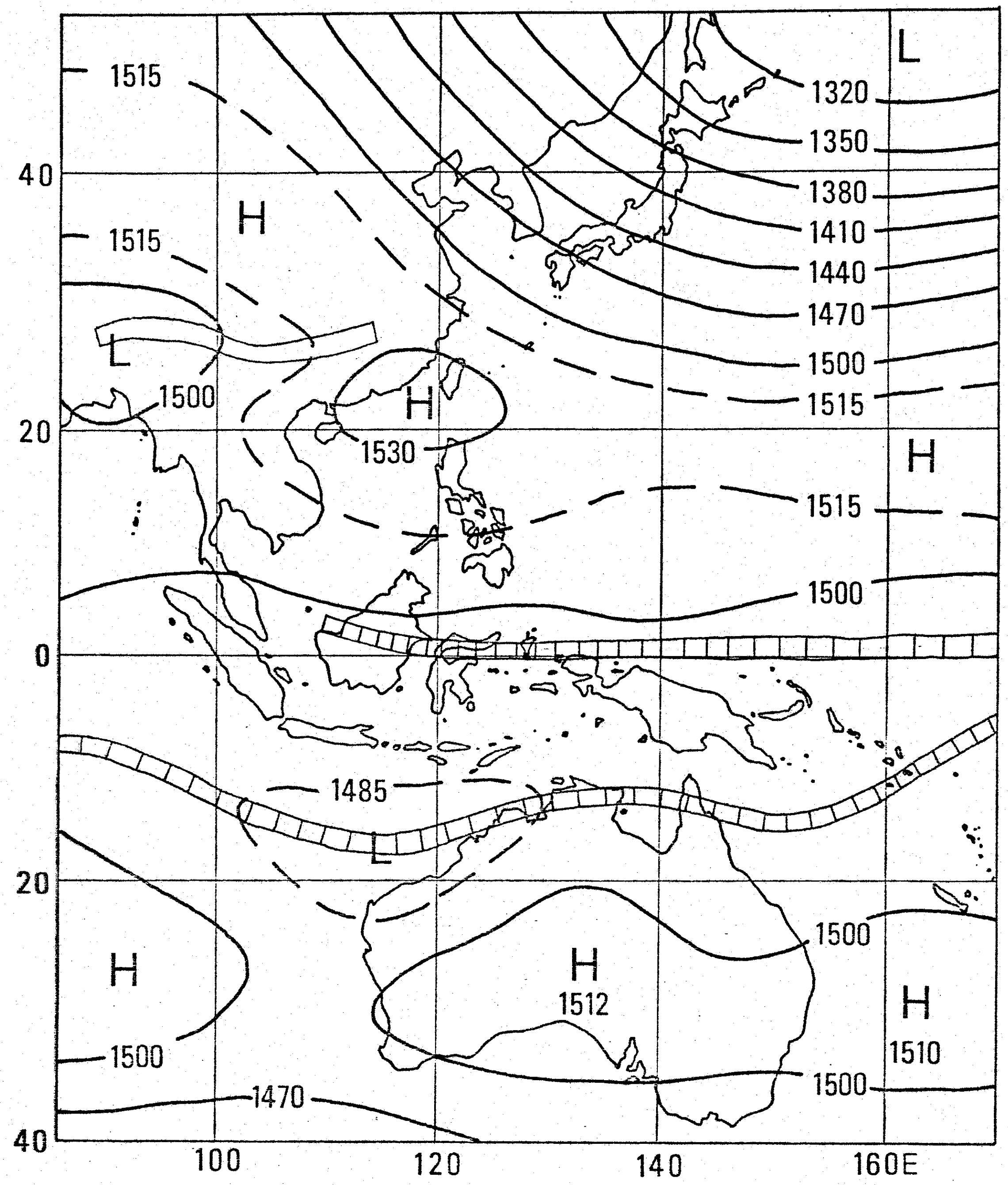

Fig. 7 Geopotential height at the 850 mb level in January
(1961-1975 average) (unit: geopotential meters)

China is called the Kunming frontal zone. This is a frontal zone at the 850 mb level separating warm dry air of the subtropical high and dense cold air of the Siberian high. Chu (1963) has analyzed the frontal zone at the surface level in this region. However, the direction of the frontal zone in his figure is different from the east-west direction shown in Figure 7.

The important point in this figure is the fact that the northeast monsoon is a low level branch of the Hadley cell originating in the subtropical high. Thus the northwesterly wind near Japan and the tropical northeast monsoon is a distinct wind system of the general circulation of the atmosphere.

Figure 8 shows the mean zonal wind at the 850 mb level in the winter. The cold northwesterly flow from Siberia gradually changes its direction and becomes a westerly flow of more than 10 m/sec to the east of Japan. The Kunming frontal zone is well marked as a boundary between the dry southwesterly wind and a weak northeasterly wind from the Siberian high.

The northeast wind to the south of the subtropical high is called the northeast monsoon to the west of about 130°E and northeast trades to the east of this longitude. The equatorial westerlies which are called the "West monsoon" in Java prevail between the equator and approximately 15°S. South of the southern ITCZ, the southeast trade wind occupies most of Australia. In southern Australia, the middle latitude westerlies are observed to the south of the subtropical high.

Figure 9 shows the mean zonal wind at the 150 mb level in the winter. There are three major wind systems. The most intense wind is the subtropical jet in the northern hemisphere with maximum wind speed of over 60 m/sec near Japan. A similar but less intense jet is located in southern Australia with a wind

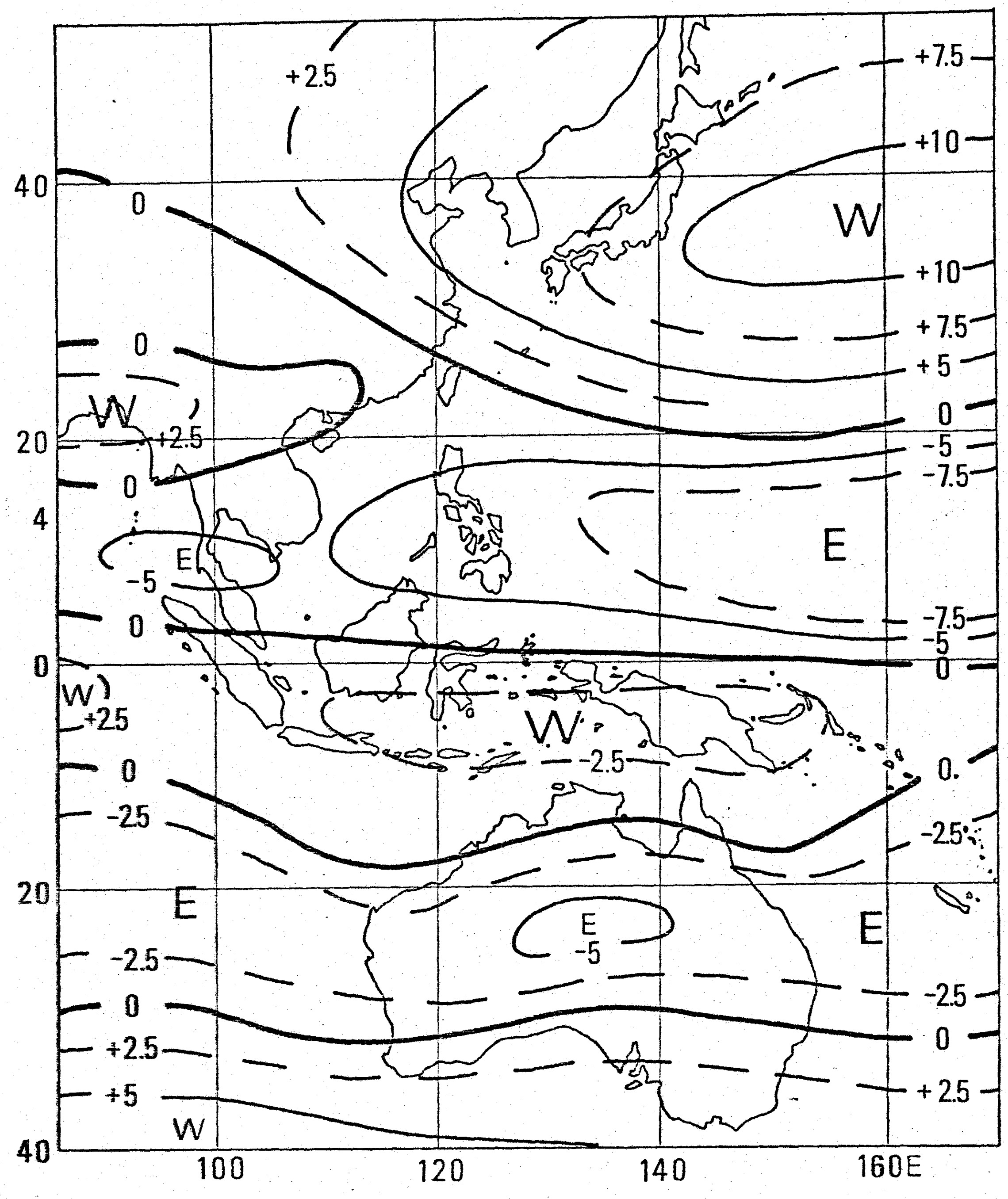

Fig. 8 Mean zonal wind at the 850 mb level in winter (data adopted from Newell et al. 1972) (unit: m/sec)

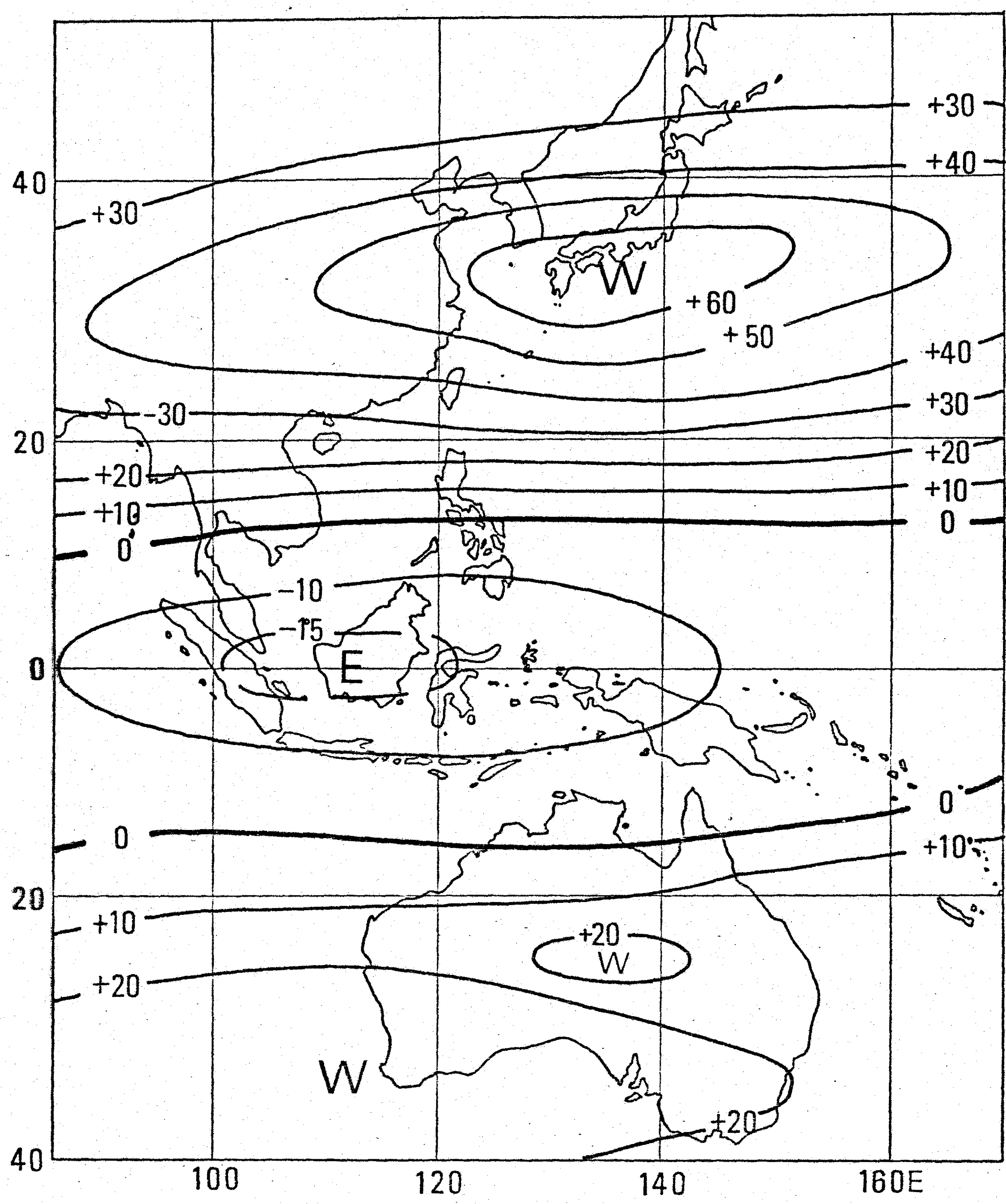

Fig. 9 Mean zonal wind at the 150 mb level in winter
(data adopted from Newell et al. 1972)
(unit: m/sec)

speed of 20 m/sec. Both of these westerly jets are located poleward of the subtropical high of the respective hemispheres. The easterly wind observed near the equator has a maximum speed of 15 m/sec near Singapore. This is the upper level branch of the Hadley cell whose ascending branch is located over the northern Australian ITCZ. In the current study, the year to year fluctuations of this wind measured at Singapore are used as an indicator of the strength of the tropical winter monsoon.

2-2 Regional characteristics of the monsoon regime.

In this section the seasonal change of the circulation toward the winter monsoon regime is discussed. The monthly climatological precipitation (1931-1960: if available) and the monthly wind at the 850 mb level for a 10 year period from 1968 to 1977 were used to investigate the climatology of the seasonal change of the monsoon regime. The data covered all months of the year, but the regions of analysis are restricted to these with a pronounced wind shift of at least 120° between the two monsoons of the late autumn or winter and of the summer. A 10 year period was chosen because this is the only period when the data of steadiness factor, which is defined as a ratio of the scalar wind speed to the vector wind speed, are available.

In all the tropical monsoon regions, the passage of the ITCZ is characterized by minimum values of steadiness factor for both summer and winter monsoons. In the transition month when the ITCZ passes over the station, the steadiness factor drops to the values of 30 to 50 %. For this reason, the tropical monsoon winds were defined to have at least 60 % steadiness factor.

Figure 10 shows the southward migration of the ITCZ at the 850 mb level. In contrast to the northward migration, the retreat of the ITCZ to its winter position is slow. This is a result of the warm sea which occupies larger areas in the southern latitudes. The winter monsoon regime is fully established in December when the southern ITCZ develops to the south of New Guinea. The double ITCZ regime persists until near the end of March.

The slow progress of ITCZ is accompanied by heavy precipitation. Regions to the east of 100°E generally observe their wettest period in the year shortly after the southward passage of the ITCZ (Figure 11). On the east coast of Indochina the maximum rainfall is observed one month after the passage of ITCZ but before the peak of the northeast monsoon. In the Philippines it occurs little more than one month later primary due to the influence of the Kuroshio current. In addition, the east coast of the Indian peninsula also observe its wettest period in late autumn.

The double zones of heavy precipitation which characterize the winter conditions begin to develop during December and by January, the southern rainbelt in northern Australia is well developed. The southward migration of heavy rain in eastern Japan is associated with typhoons and seasonal shifts of the location of the polar fronts.

Figure 12 illustrates the onset of the winter monsoon at the 850 mb level defined by the first month with more than 60 % steadiness factor of the northeast wind to the north of the equator. In the southern hemisphere, the monsoon begins when

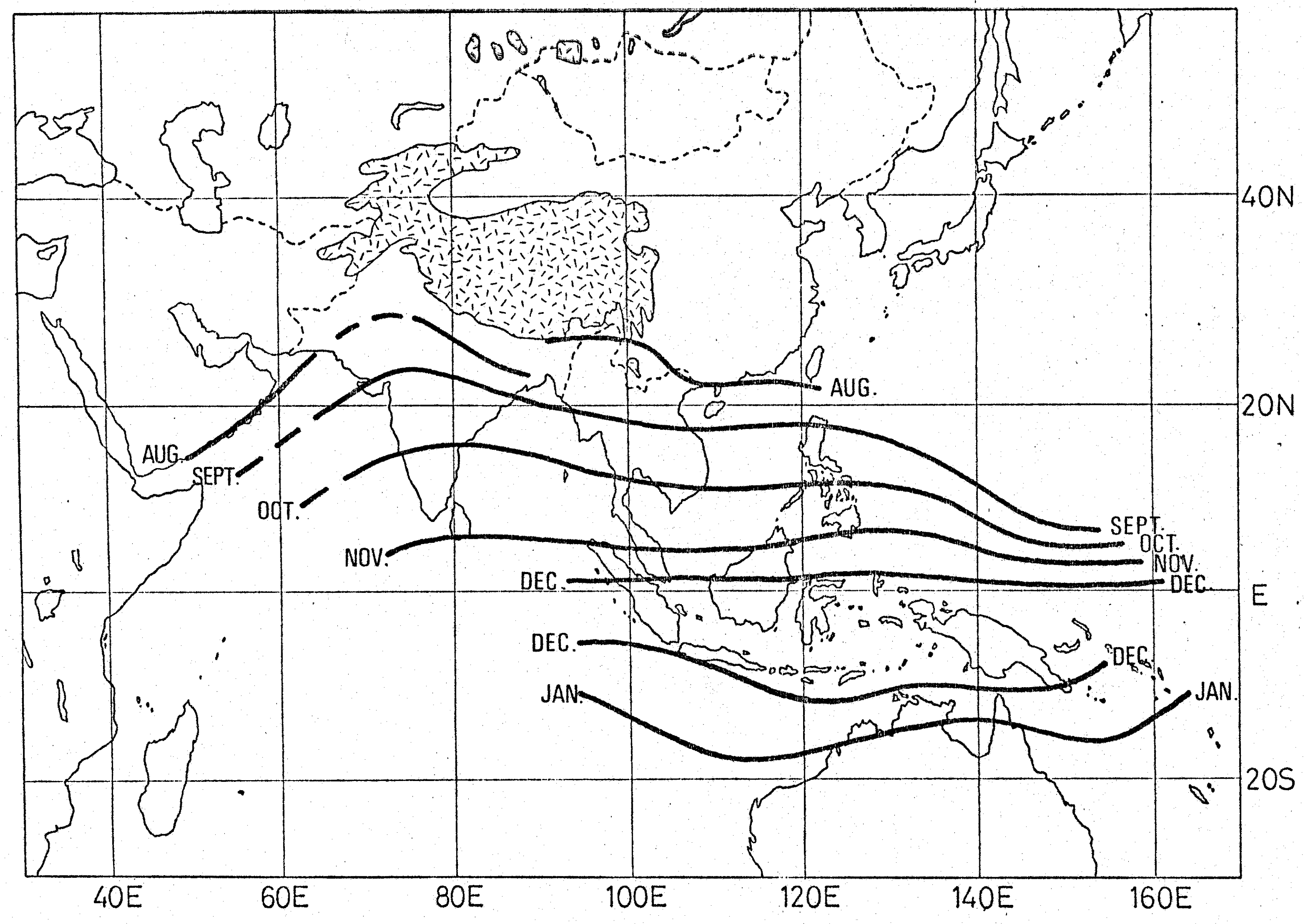

Fig. 10 Southward migration of the ITCZ at the 850 mb level from August to January

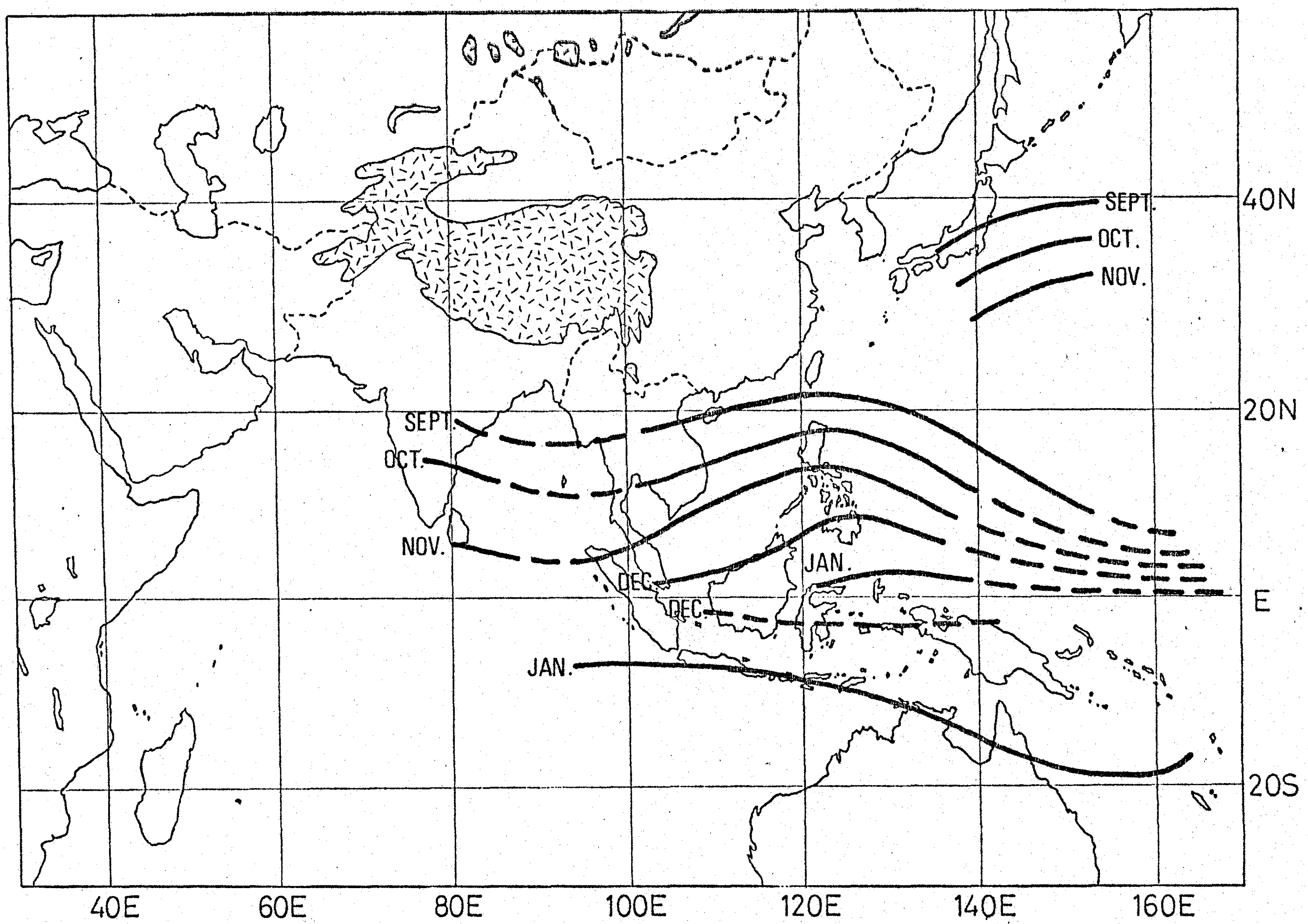

Fig. 11 Southward migration of rainfall maximum from September to January

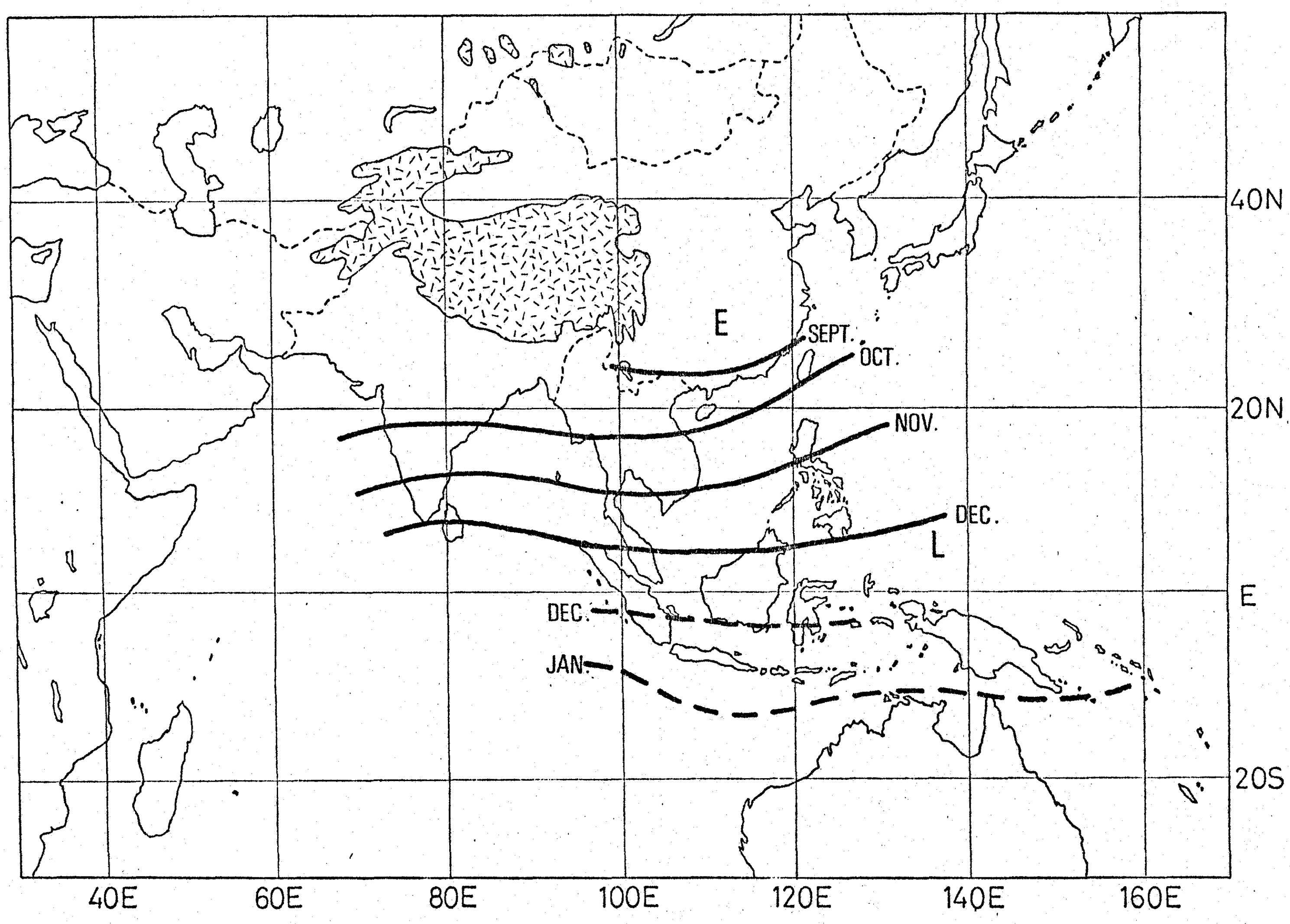

Fig. 12 Southward migration of onset of the winter monsoon at the 850 mb level from September to January

the northwest wind is first observed. At Hong Kong, which is located near the northern limit of the monsoon region, the northeast monsoon begins as early as September. The southward migration of the development of the winter monsoon is accompanied by a decrease in the precipitation in the respective regions to the north of the equator. The dashed lines to the south of the equator indicate the beginning of the northwest monsoon.

Figure 13 shows the month with peak intensity of the northeast monsoon. To the north of 15°N, the peak occurs about 15 days after the beginning of the monsoon. In the lower latitudes, the peak occurs one month after the beginning of the monsoon. The most interesting aspect of this map is the occurrence of the peak of the northeast monsoon during the months of October and November in the regions to the north of 12°N. This fact clearly shows that the northeast monsoon at the 850 mb level is associated with the outflow from the subtropical high and constitutes a part of the Hadley cell.

Finally Figure 14 shows the southward migration of the southern edge of the subtropical high at the 850 mb level. This coincides with the end of the northeast monsoon defined by 60 % steadiness factor. The regions bounded by two hatched lines located about 10°N and 3°N are the area where the northeast monsoon prevails during the winter months from December to March. In the outer fringe of the tropics, the 850 mb level wind became light and variable during the winter.

The seasonal changes to the winter monsoon regime shown by Figures 10 to 14 have illustrated that the slow southward migration of the Hadley cell is responsible for the slow pace

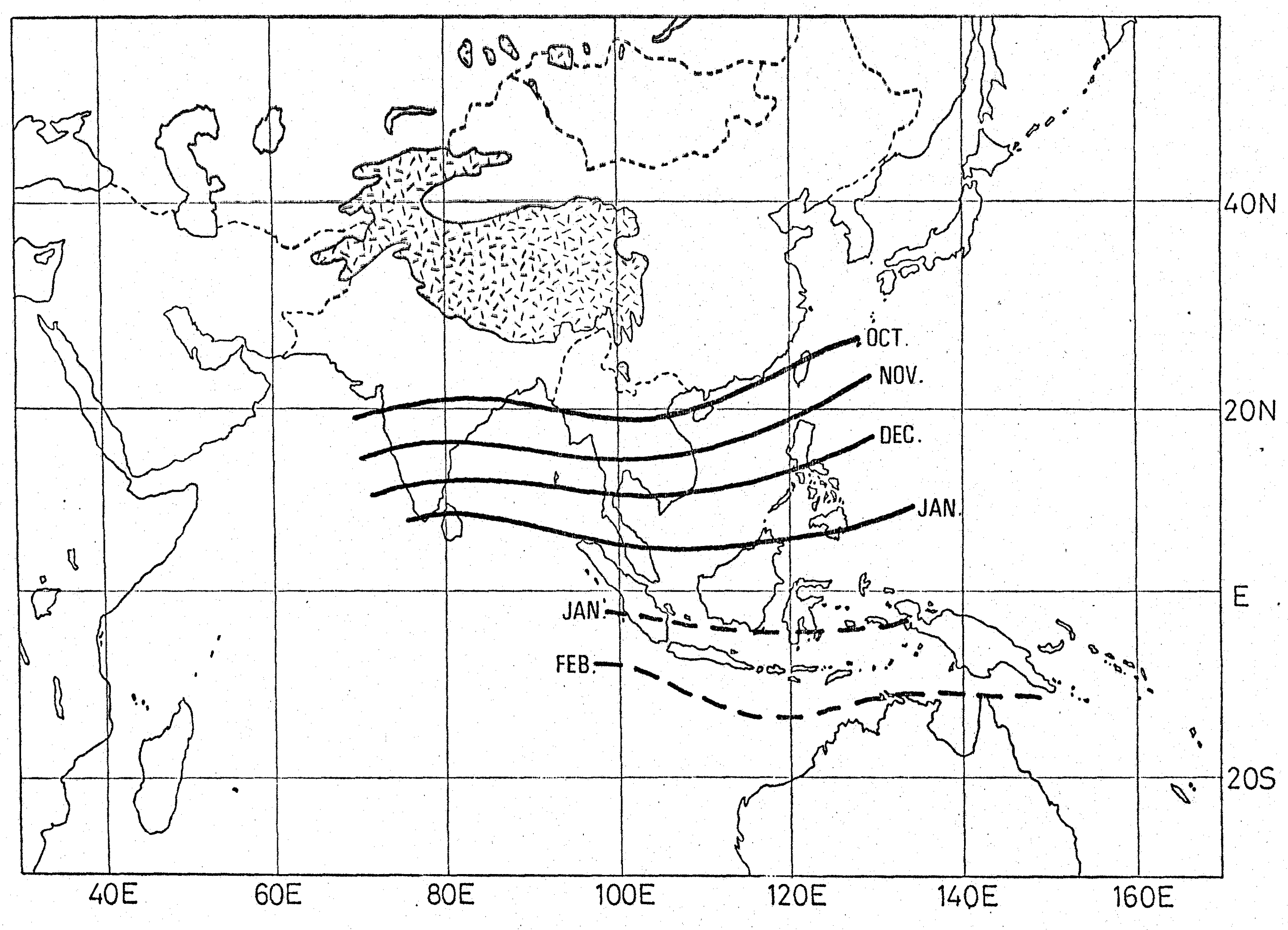

Fig. 13 Southward migration of the peak of the winter monsoon from October to February

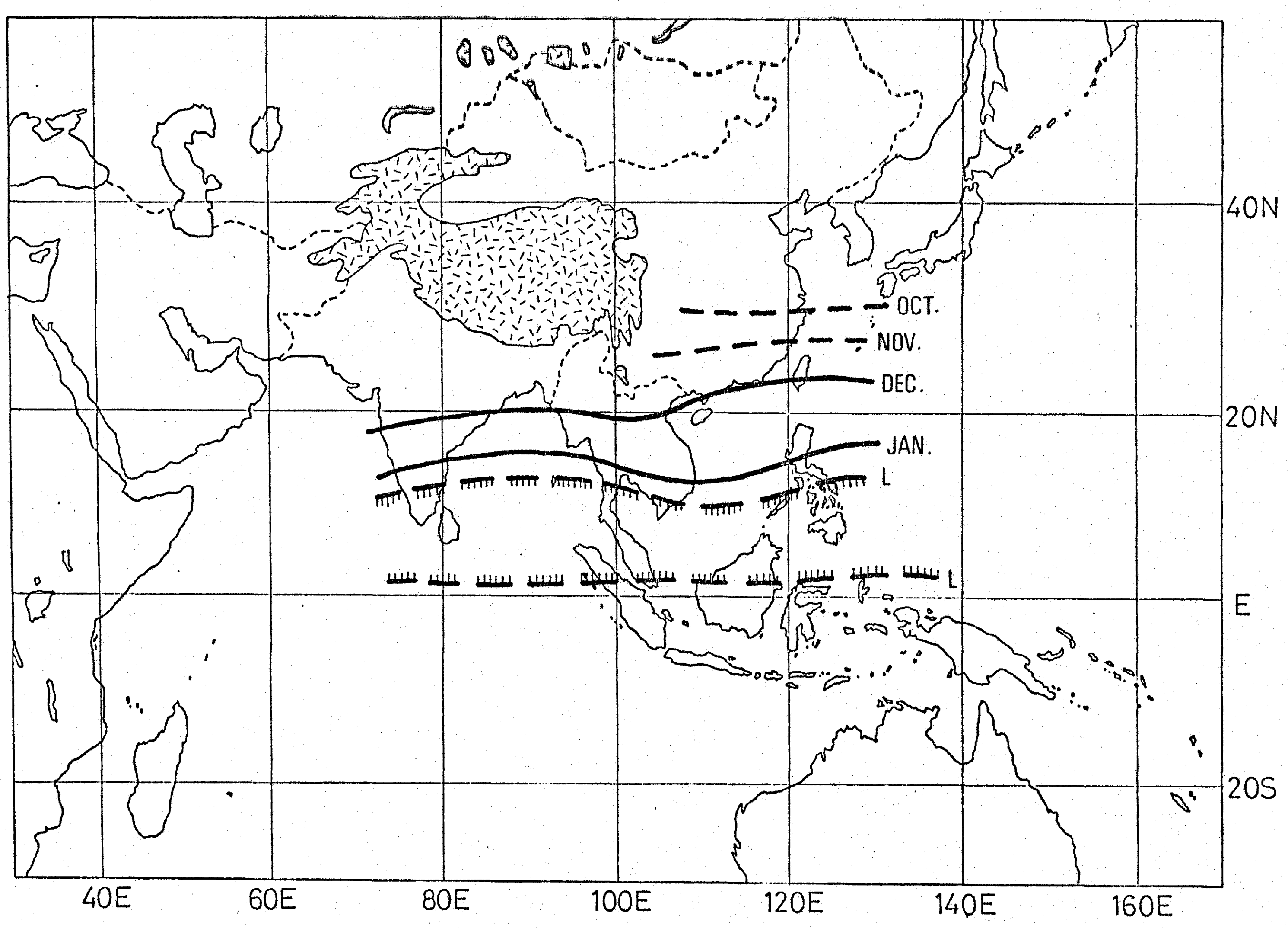

Fig. 14 Southward migration of the subtropical high at the 850 mb level from October to January ; L indicates the region of the winter monsoon during January

of the change to the winter monsoon regime. In given regions to the south of 20°N, the following sequences are observed: (i) The passage of the ITCZ accompanied by a shift of the wind to the northeast. (ii) The peak of the rainfall shortly after this passage. (iii) The beginning of the steady northeast monsoon and a slow decrease of rainfall. (iv) The peak of the northeast monsoon and a beginning of a dry season. And (v) the end of the steady notheast monsoon to the north of the 13°N. The only small exception to this sequence in the east of 100°E is in central Thailand, where the peak of rainfall occurs slightly before the passage of the ITCZ.

The precipitation on the east coast of Indochina, the east coast of the Philippines and the east coast of the Malay peninsula has been traditionally considered as a orographic precipitation. However the above sequence also applies to these regions and the peak of the rainfall occurs just before the increase in the steadiness of the northeast monsoon. For this reason, the proximity of the ITCZ is the necessary condition for the heavy precipitation,as far as the monthly total precipitation is concerned. Without the instabilities and the disturbance favorable for the precipitation, the orography can not produce the heavy precipitation.

CHAPTER 3

ROLE OF THE CIRCULATION AT THE 150 MB LEVEL DURING THE WINTER MONSOON

3-1 Upper level wind at Singapore

The fluctuations of the winter monsoon are discussed in this chapter with emphasis on the fluctuations of the tropical easterly wind at the 150 mb level. Figure 9 shows that the easterly wind at the 150 mb level in the winter is strongest near Singapore. Figure 15 shows the time-height cross section of the zonal wind observed at Singapore in January. Strong easterly wind is observed near the 150 mb level. A three to four year cycles are observed with strong wind in 1962,1963,1967, 1971 and 1974. Alternating zonal wind at the 50 mb level is a Quasi-Biennial oscillation of the zonal wind in the tropical stratosphere. Below the 600 mb level, equatorial westerlies are observed with the exceptions of 1964,1970 and 1973.

Table 1 shows the zonal wind at the 150 mb level at Singapore during three winter months from December to February in the period from January 1961 to February 1978. The interannual fluctuations of the zonal wind are largest in January with a standard deviation of 7.8 m/sec. For this reason, the emphasis of this study is on the month of January. With a notable exception of 1978, all the Januaries which observed strong easterly wind are the winters with strong easterly wind. Hence, month to month persistence of the intensity of the winter monsoon is high.

Table 2 summarizes the monthly variation of the winds at the 150 mb level at Singapore. The zonal component of the wind

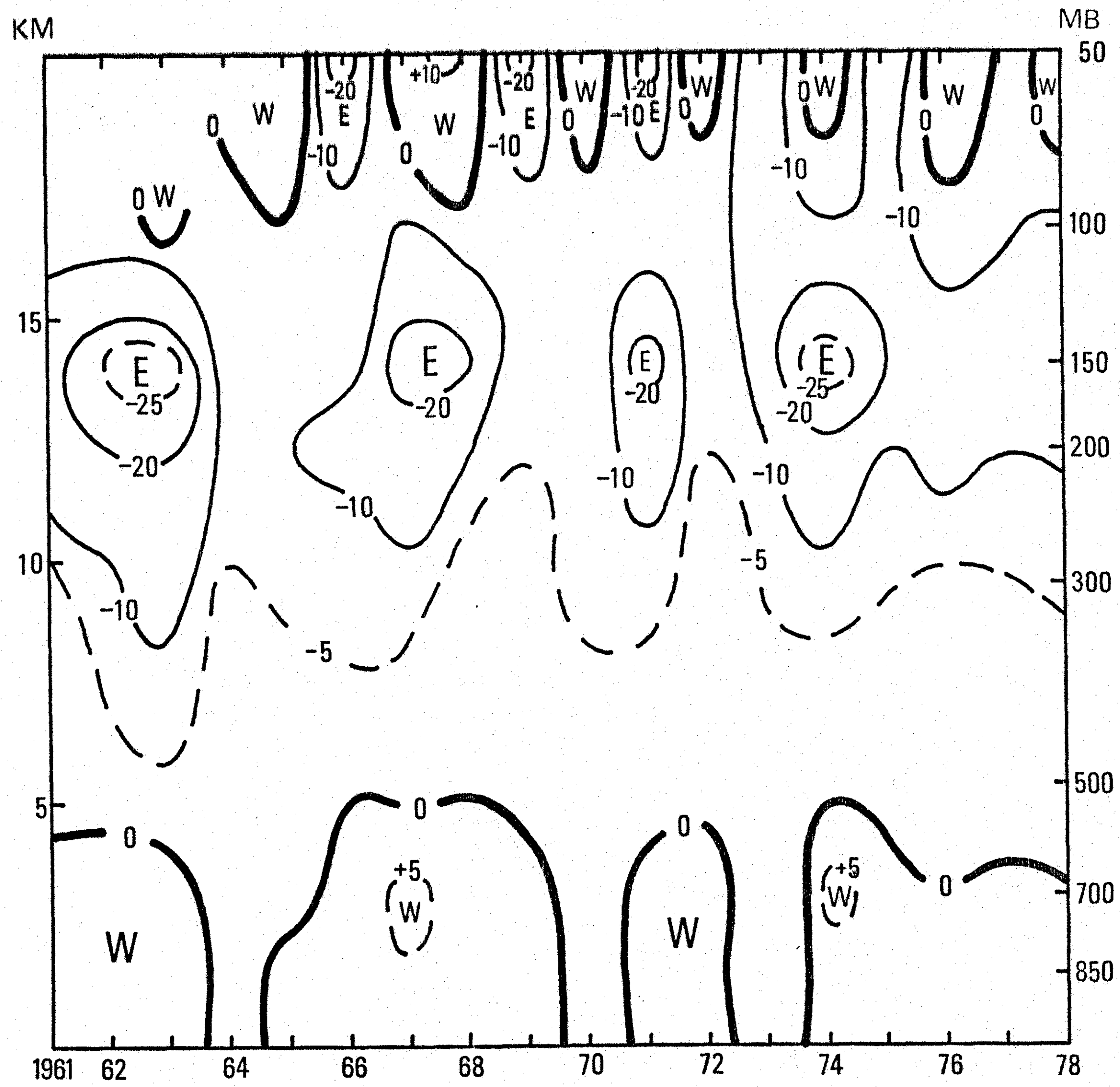

Fig. 15 Zonal wind at Singapore in January, 1961-1978

(unit: m/sec)

TABLE 1

The zonal easterly wind at the 150 mb level at Singapore in winter

(unit: m/sec)

Year (Jan.yr)	Month Dec.	Jan.	Feb.	3 month average
1961	----	15.7	19.7	(17.7)
1962	16.7	25.4	18.8	20.3
1963	9.8	28.2	21.6	19.9
1964	10.8	8.7	8.5	9.3
1965	13.8	5.6	14.1	11.2
1966	14.8	8.0	13.2	12.0
1967	18.8	23.6	16.9	19.8
1968	10.8	20.7	13.8	15.1
1969	12.2	6.6	13.2	10.7
1970	13.8	1.5	10.5	8.6
1971	20.7	20.3	26.5	22.5
1972	15.8	7.9	14.4	12.7
1973	9.8	16.0	10.8	12.2
1974	20.9	28.1	21.3	23.4
1975	17.8	20.3	17.2	18.4
1976	22.9	17.5	13.9	18.1
1977	12.0	13.5	13.5	13.0
1978	5.6	18.4	-0.3	7.9
Average	14.5	15.9	14.9	15.1
S.T.D.	4.5	7.8	5.7	4.8

TABLE 2

Monthly variation of the wind at the 150 mb level at Singapore

period of average: January 1961 to May 1978 (unit: m/sec)

		Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Zonal	U	-15.9	-14.9	-11.8	-4.4	-9.5	-17.9	-19.4	-21.4	-19.8	-14.4	-12.9	-14.5
wind	S.T.D.	7.8	5.7	5.3	3.6	3.1	3.2	3.1	5.4	3.2	4.6	4.1	4.5
Meridional	V	+4.3	+5.0	+4.0	-0.1	-3.7	-5.6	-5.7	-6.5	-6.1	-3.0	-0.8	+2.4
wind	S.T.D.	3.2	2.5	2.4	1.6	1.3	1.6	1.9	2.6	1.8	1.2	1.5	0.9

reaches its peak intensity and has large standard deviation in August and January. These peaks corresponds to the summer and winter monsoon.

The meridional component of the wind shows a pronounced annual cycle with southerly flow during the winter monsoon from December to March and northerly flow during the summer monsoon from May to October. Thus the wind at the 150 mb level constitutes a return flow of the monsoon circulation. The summer season peak of the circulation at the 150 mb level is later at Singapore than that over India, because of warm oceans near the Philippines which maintain a strong monsoon circulation through September.

3-2 Distribution of the monthly total precipitation in January

The relationship between the fluctuations of the zonal wind at the 150 mb level at Singapore and the winter monsoon at the low level was investigated by the construction of composite maps. An inspection of the monthly total precipitation has shown that every January with the easterly wind over 20 m/sec at the 150 mb level with the exception of 1967 has shown the tendency for formation of two ITCZs. Hence a composite map of the precipitation in 6 Januaries (1962,1963,1968,1971,1974,1975), when a strong easterly wind was observed at the 150 mb level at Singapore was constructed (Figure 16). The regions of heavy precipitation are North Borneo, eastern Mindanao, and Micronesia located in the northern ITCZ. The southern ITCZ is located in northern Australia.

In all of above regions, more than 400 mm of precipitation are observed. On the other hand, the regions to the east of New Guinea are dry with less than 100 mm of precipitation. In

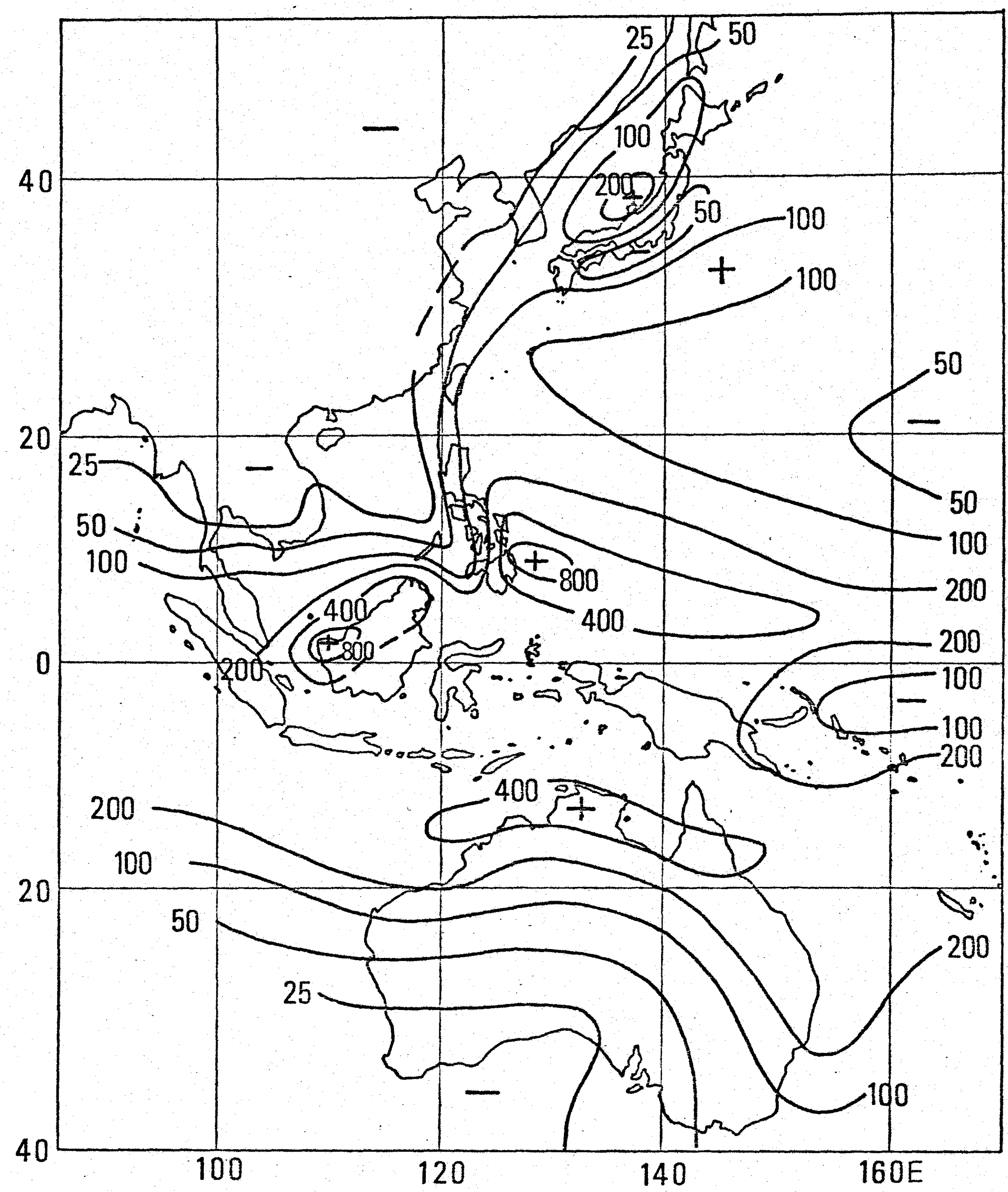

Fig. 16 Composite map of the precipitation in 6 Januaries with strong easterly wind at the 150 mb level at Singapore (unit: m/sec)

the other regions in the map the precipitation is similar to its climatological normal with a notable exception of the polar front zones to the south of Japan where the precipitation is less than normal.

Figure 17 shows the composite map of the precipitation in 6 Januaries (1964,1965,1966,1969,1970,1972), when the easterly wind observed at the 150 mb level was weak at Singapore. In general, the distributions of precipitation are opposite of the previous figure. The regions of heavy precipitation are located to the east of New Guinea and northeastern Australia.

The difference in the two composite maps was obtained by subtracting the precipitation of the map of weak easterly wind from the strong easterly wind. Figure 18 shows the difference in precipitation obtained in this manner. Two zones of increased precipitation of over 100 mm are identical with the two ITCZs observed during January with strong easterly winds. The decrease in precipitation during these years of strong monsoon circulation is conspicuous to the east of New Guinea and in the polar front zones to south of Japan.

Figure 19 shows effects of change in vertical motions associated with change in the monsoon regime. The dewpoint depression at the 500 mb level was chosen to monitor this effects. The change in this parameter clearly indicates that the precipitation patterns shown in Figure 18 are not restricted to a few island observations in Micronesia. Hence the pattern of two ITCZs and increased subsidence in the subtropical high near Taiwan is the main feature of the strong monsoon regime associated with a strong easterlies at the 150 mb level at Singapore.

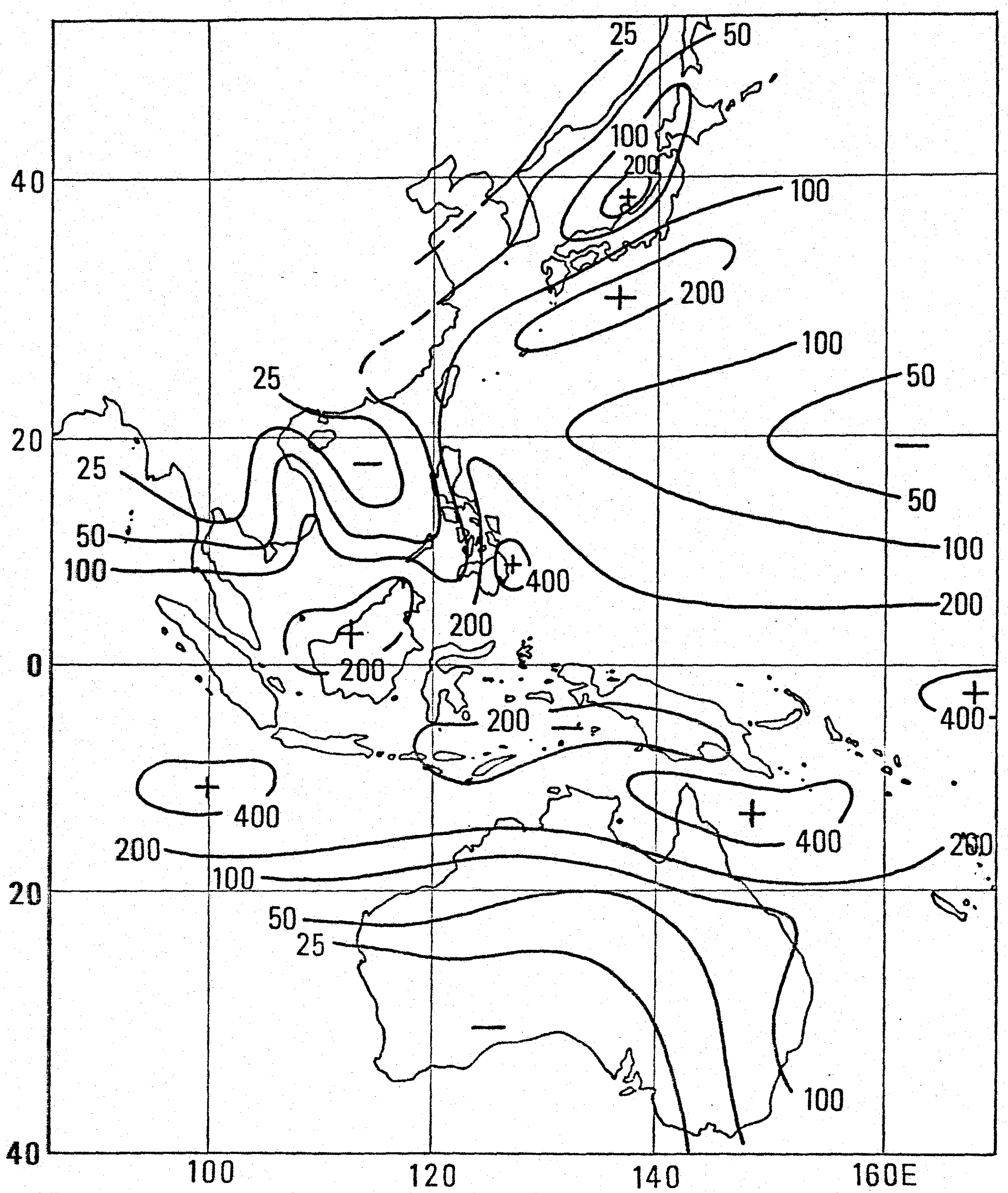

Fig. 17 Composite map of the precipitation in 6 Januaries with weak easterly wind at the 150 mb level at Singapore (unit: m/sec)

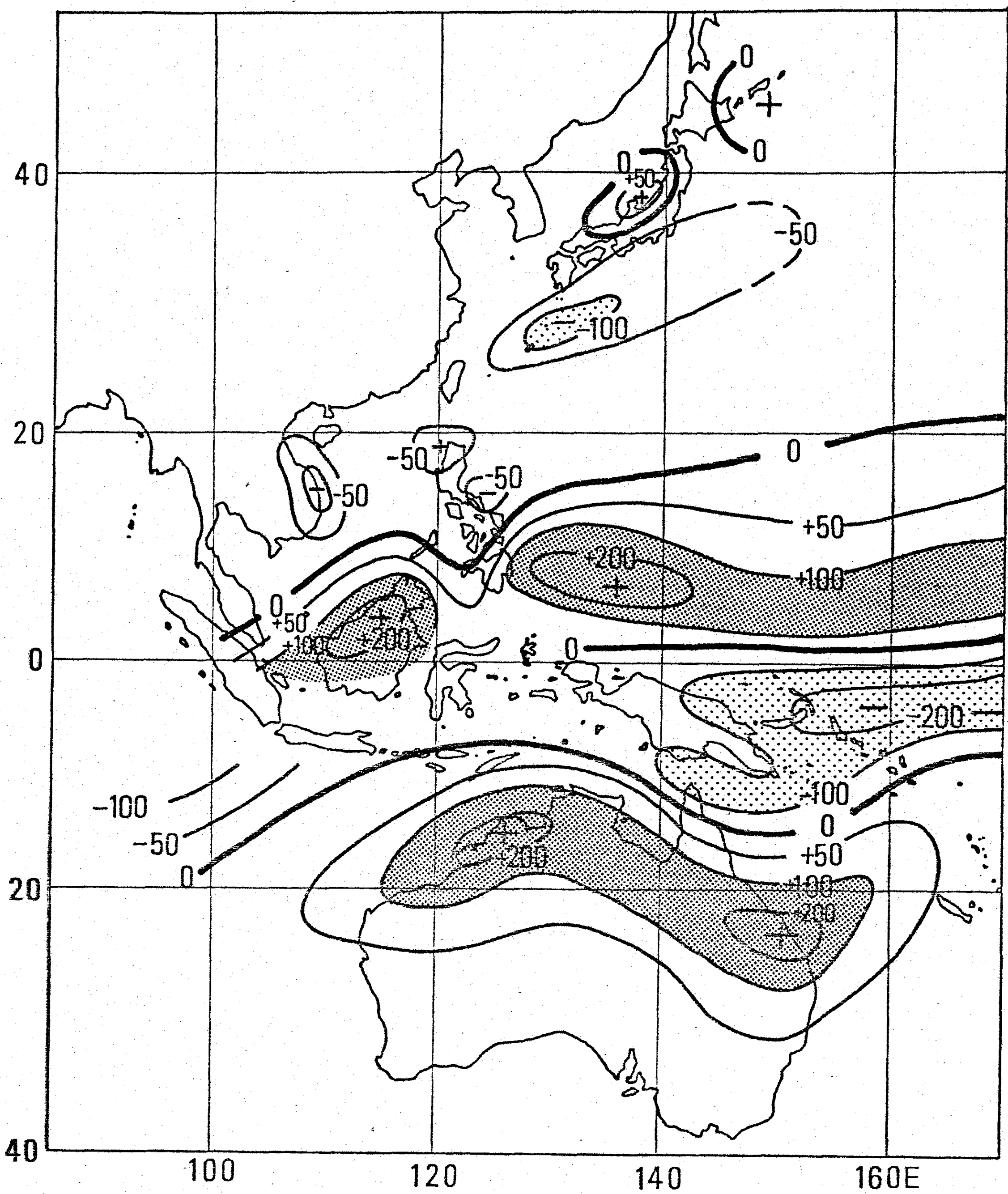

Fig. 18 . Difference in the precipitation in January.

(unit: mm)

(strong easterly case minus weak easterly case)

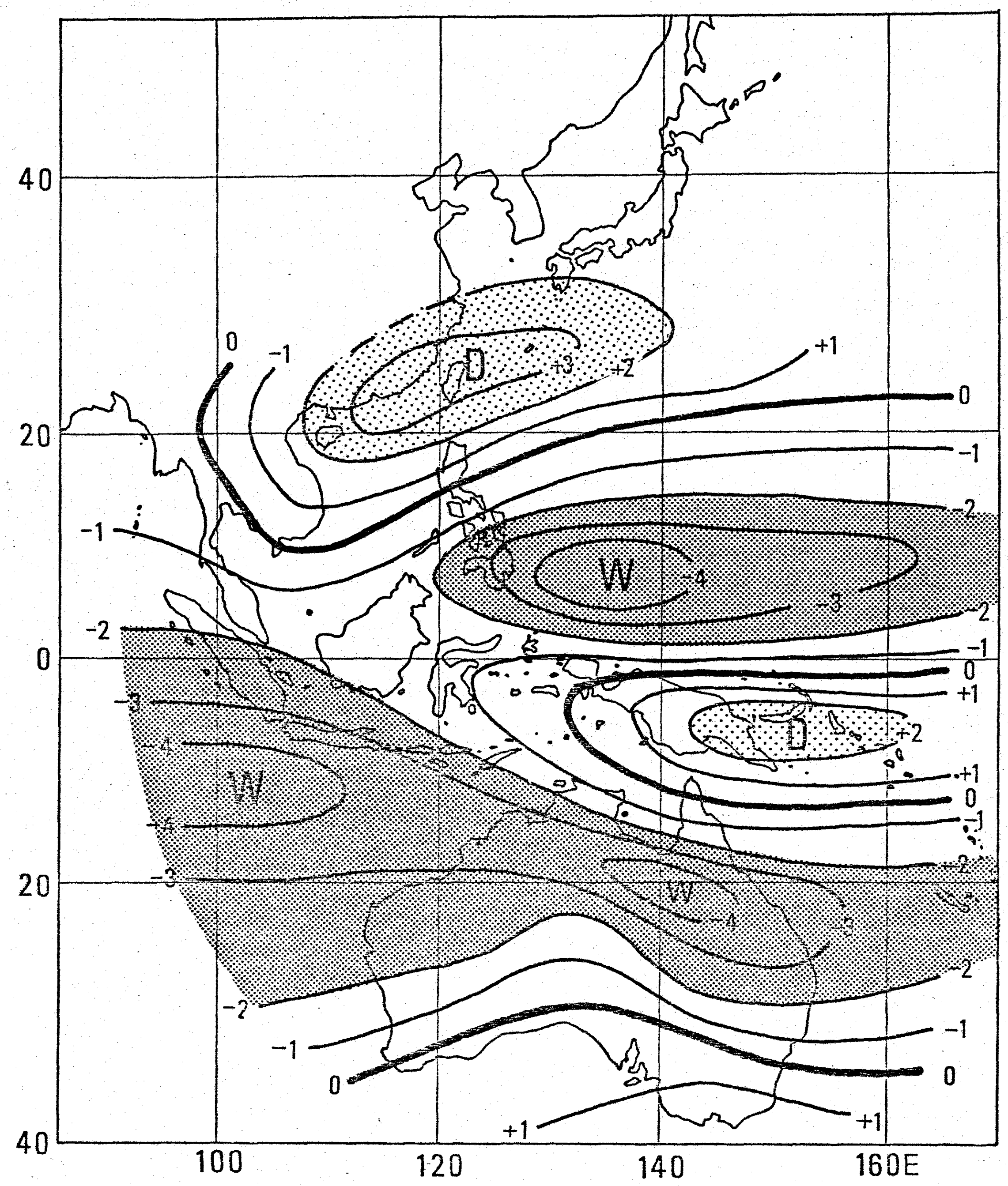

Fig. 19 Difference in the dewpoint depression at the 500 mb level in January (unit: °C)
(strong easterly case minus weak easterly case)

3-3 Circulation at the 850 mb level in January

The circulation at the 850 mb level was analyzed using the same procedure for the monthly total precipitations. The combinations of the Januaries in the composite map are also identical. Figure 20 shows the difference in the geopotential height at the 850 mb level. Comparing this map with Figure 7 (which shows the normal circulation patterns for January) the following important points can be noted: In the Indonesian region, between the two ITCZs, the equatorial westerlies are well developed during the Januaries with strong easterlies at the 150 mb level. The surge of the cold northwest monsoon from Siberia is frequent in East China Sea and southern Japan. In the Micronesian region, the northeast trade wind is weaker than the normal.

On the other hand, the northeast monsoon over the South China Sea and the southeast trades over Australia are strong and well developed. In the higher northern latitudes, the northern Japan and adjacent Siberia are under warm advection from the Pacific Ocean. The reverse conditions prevail in all the areas when the easterly wind at the 150 mb level is weak.

Thus January with a strong easterly wind at the 150 mb level corresponds to the case of a strong monsoon circulation at the low level. The combined influence of change in the cloudiness associated with fluctuations in the precipitation and the circulation change at the 850 mb level can be clearly seen in Figure 21, which illustrates the difference in the surface temperature in January. The frequent surges of the cold monsoon in the East China Sea and South China Sea are

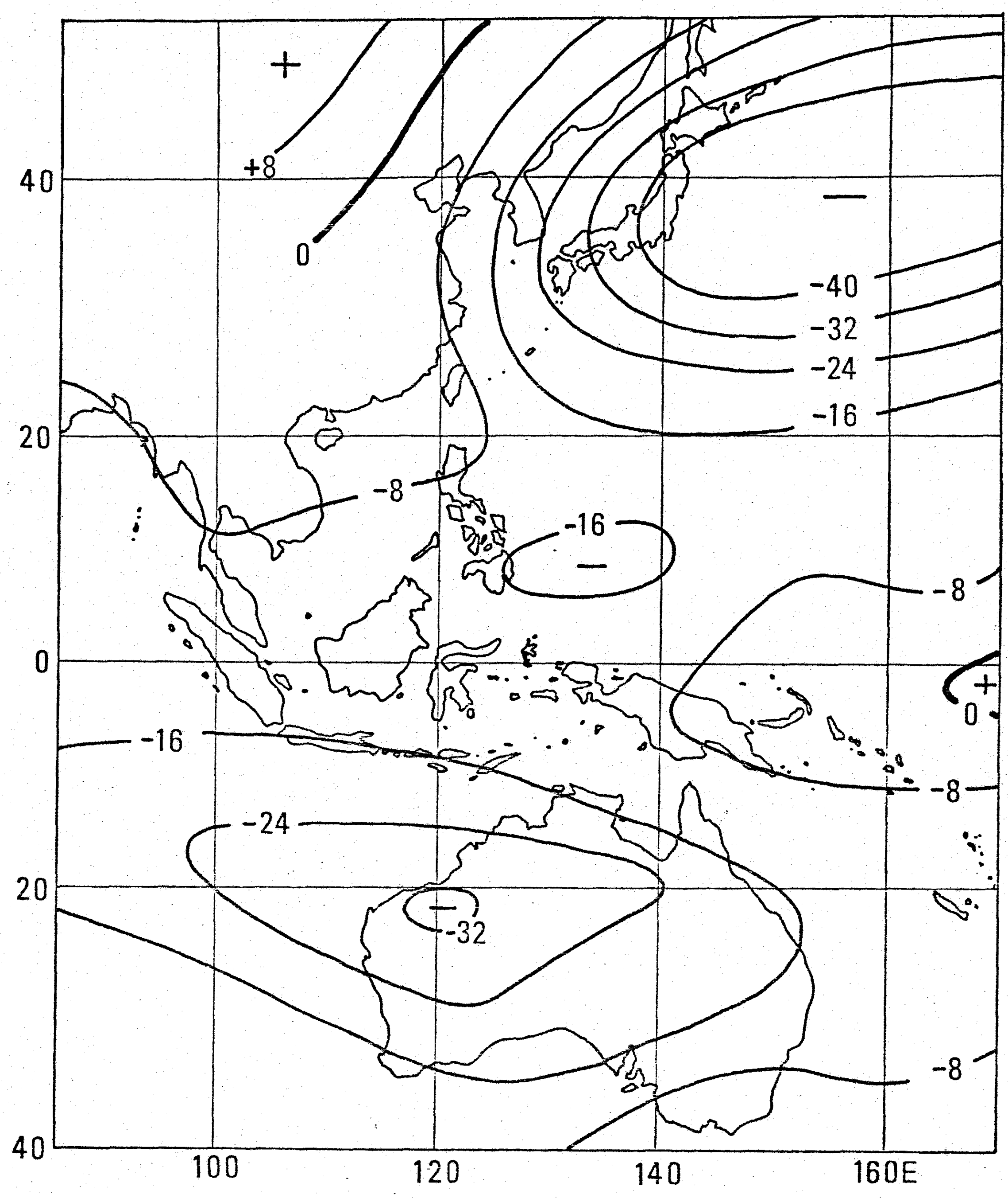

Fig. 20 Difference in the geopotential height at the 850 mb level in January
(unit: geopotential meters)
(strong easterly case minus weak easterly case)

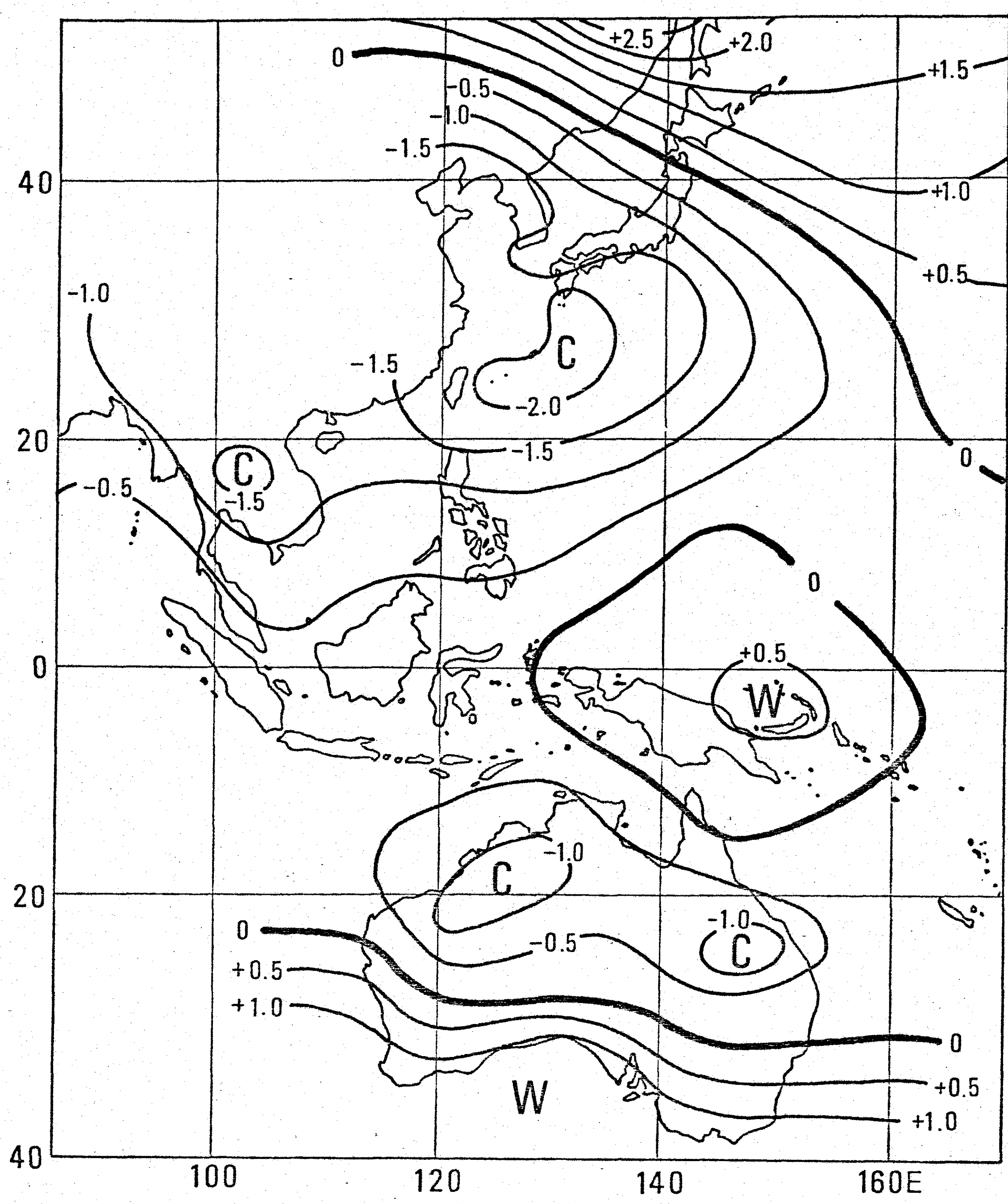

Fig. 21 Difference in the surface temperature in January (unit: ℃)
(strong easterly case minus weak easterly case)
C: cold W: warm

shown by the low temperatures during Januaries with a strong monsoon.

The increase in the cloudiness associated with heavy rain in the northern Australia is responsible for a drop in the surface temperature there. The higher temperature near New Guinea is associated with decreased cloudiness. In the high northern latitudes, warm temperature is associated with increased warm advection. Between 10°N and 25°S, the amount of the cloudiness is the dominant factor influencing the surface temperature. North of 10°N advection determines the distribution of the surface temperatures.

3-4 Circulation at the 150 mb level in January

The circulation at the 150 mb level was analyzed in the same manner as the previous section. Figure 22 shows the difference in the geopotential height at the 150 mb level. In Januaries with strong easterlies, a trough develops near Singapore.

On the other hand, in the regions of the subtropical westerly jet, general increase in the geopotential height is observed. This change is associated with poleward displacement of these jet streams near Himalaya and southern Australia. In the regions near Japan, there is a trough over the Kuroshio current and a ridge in Siberia.

The combination of Figures 20 and 22 yields the difference in the thickness between the 850 mb level and 150 mb level in January (Figure 23). The patterns in the higher latitudes are similar to the previous figure with exception of a trough near Japan. In the tropical region, the release of the latent heat

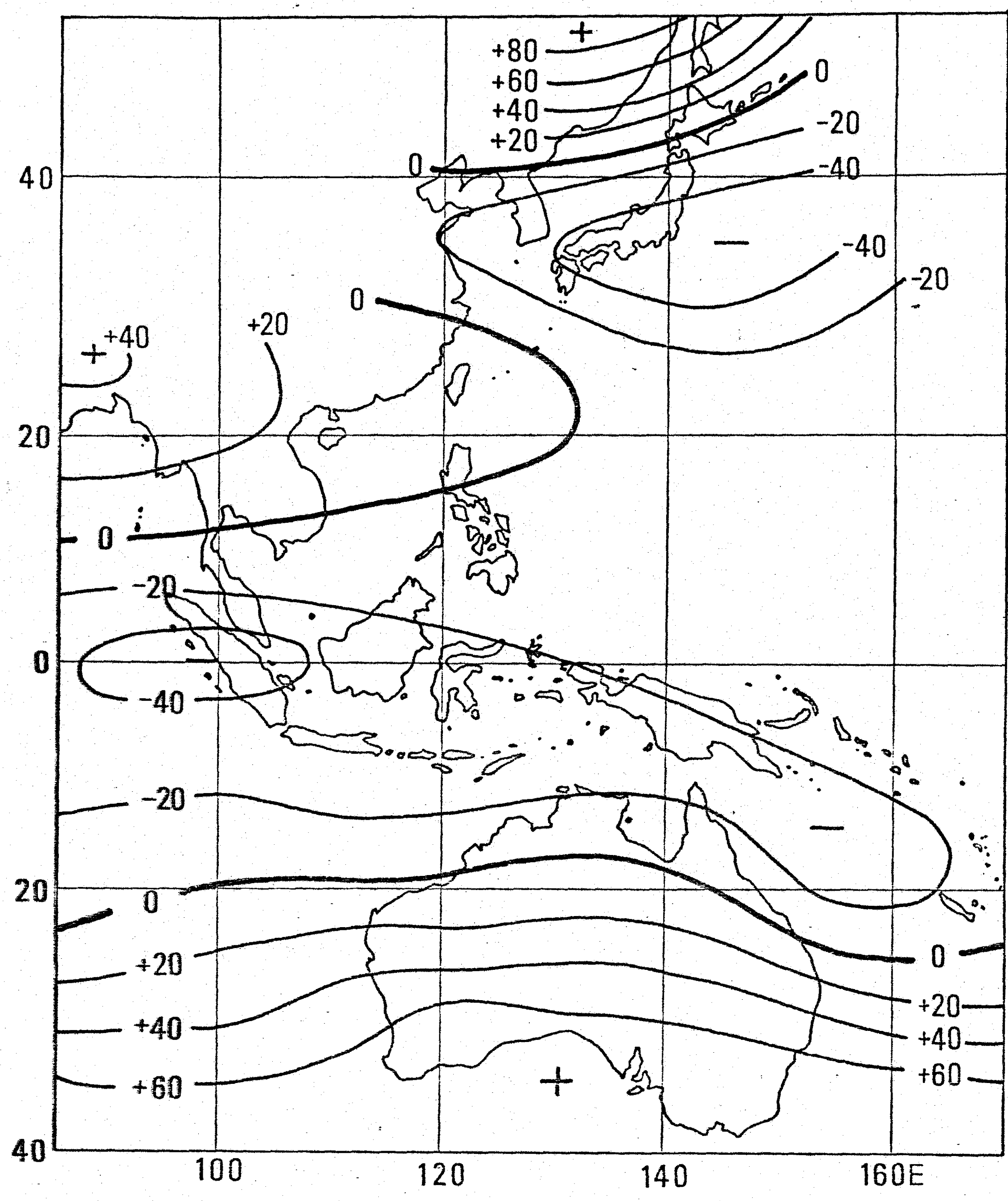

Fig. 22 Difference in the geopotential height at the 150 mb level in January

(unit: geopotential meters)

(strong easterly case minus weak easterly case)

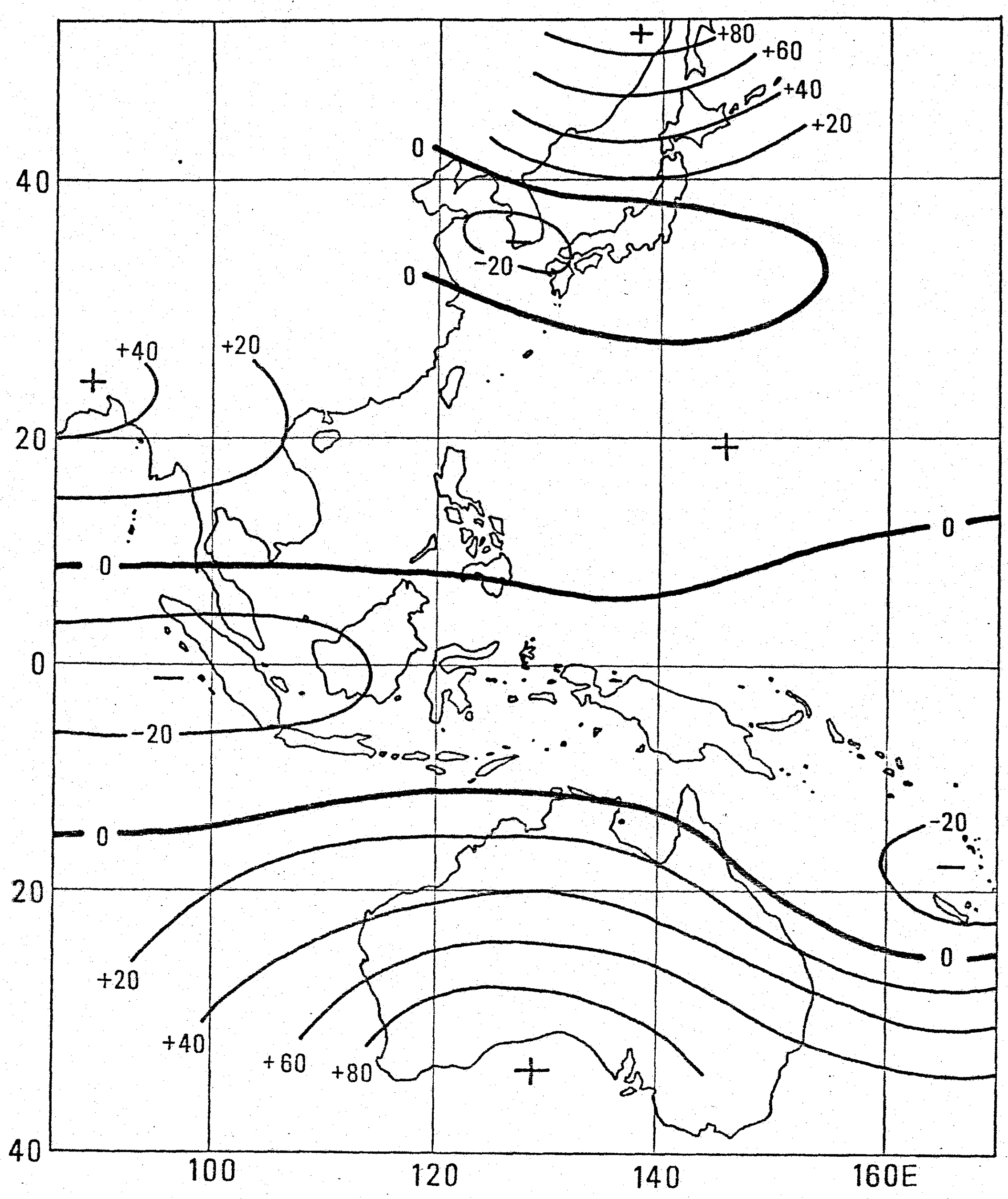

Fig. 23 Difference in the thickness between the 850 mb level and 150 mb level in January
(unit: geopotential meters)
(strong easterly case minus weak easterly case)

can be seen in the Australian region. The months with strong easterlies are characterized by a cooler troposphere near the equator and a warmer troposphere in the regions of the subtropical high.

Thus Hadley circulation is intense during the winter with strong easterly wind at the 150 mb level. This can be seen from adiabatic heating of the subsiding branch of the Hadley cell and a poleward displacement of the subtropical high.

3-5 Winter monsoon in December and February

A similar but brief survey of the winter monsoon was also conducted for December and February. Figure 24 shows the difference in the precipitation in December. The broad features of the map are similar to that of January with two ITCZs in the months with strong easterly wind at the 150 mb level at Singapore. However, the locations of the northern ITCZ are slightly north of January. In the Malay peninsula and the east coast of the central Philippines, heavy precipitation is observed in the month with a strong monsoon circulation. These regions are no longer dominated by the ITCZ in January (Figure 18).

Figure 25 shows the difference in the precipitation in February. Again, the broad features of the map are similar to that of January with development of two ITCZs. The locations of the northern ITCZ are slightly south of January and shift away from North Borneo.

Hence the fluctuations of the ITCZ in December and February are similar to that of January. In addition, there are tendencies for three winter months to have either strong easterly wind or weak easterly wind at the 150 mb level at Singapore (Table 1).

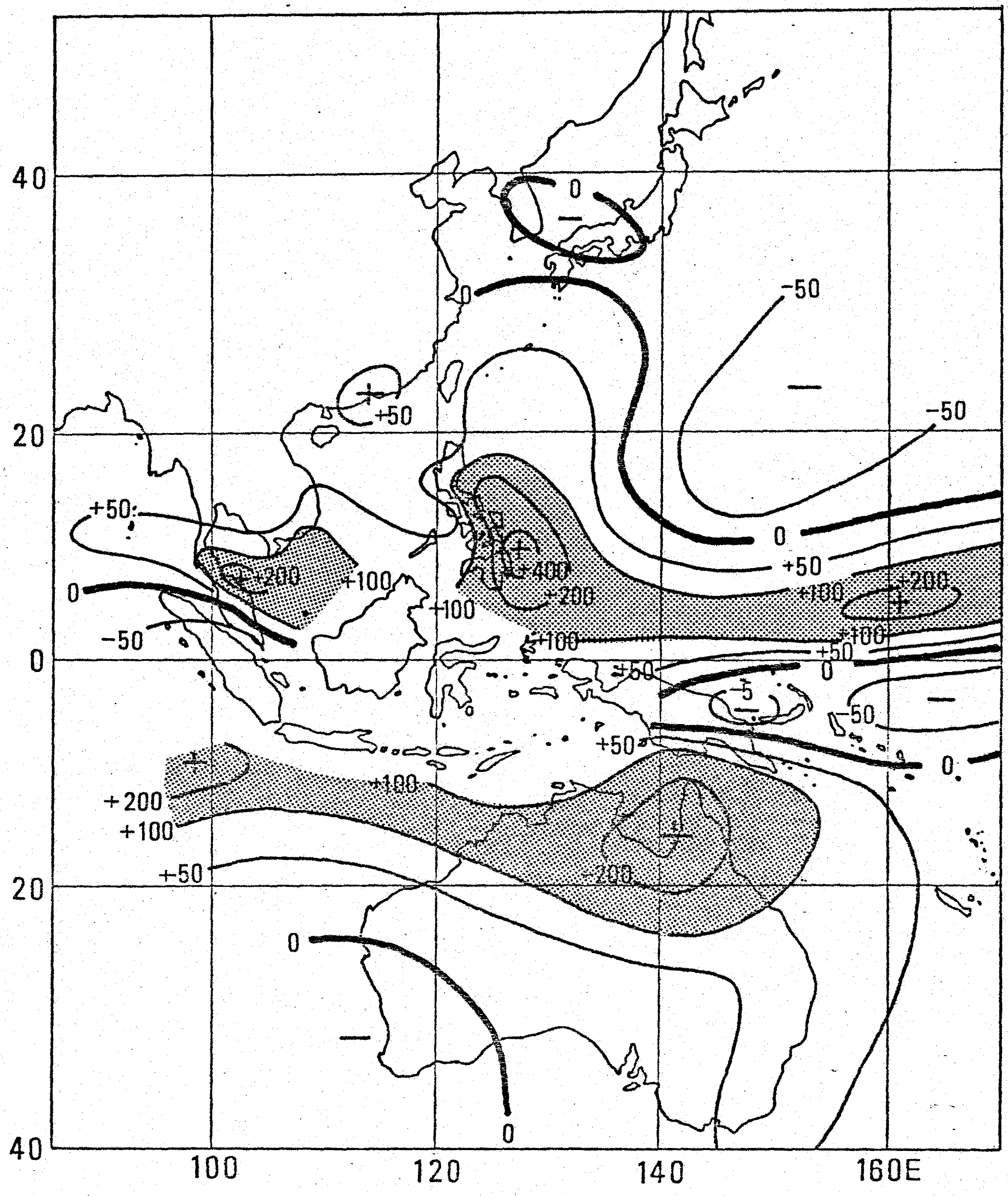

Fig. 24 Difference in the precipitation in December

(unit: mm)

(strong easterly case minus weak easterly case)

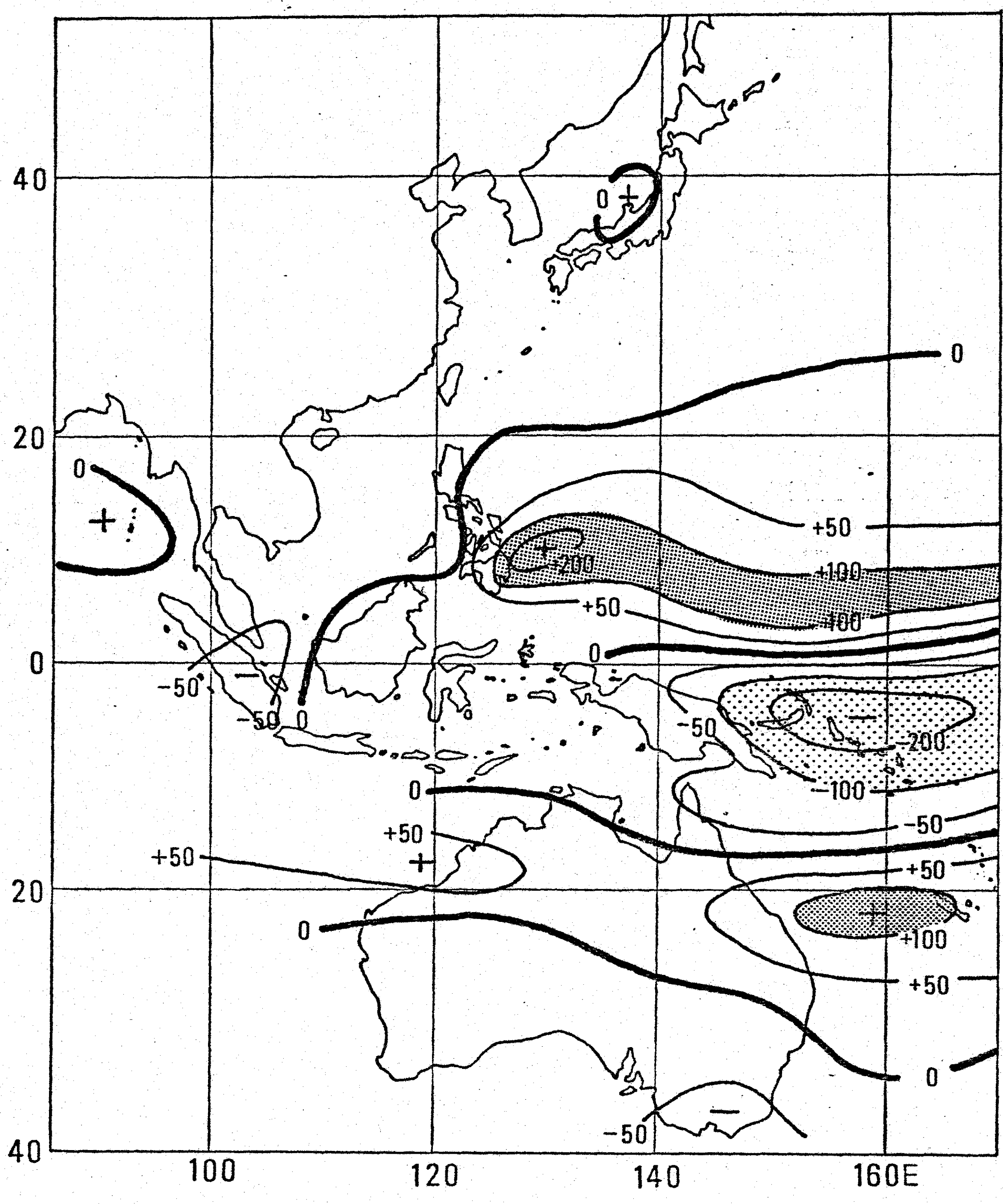

Fig. 25 Difference in the precipitation in February

(unit: mm)

(strong easterly case minus weak easterly case)

3-6 Summary

The study of the role of the circulation at the 150 mb level during the winter monsoon period has shown that fluctuations in strength of the easterly wind at the 150 mb level are closely connected to fluctuations of the winter monsoon circulation at the 850 mb level and activities of the ITCZ. When the easterly wind at the 150 mb level is strong, the winter monsoon circulation at the 850 mb level is intense and there are tendencies for formation of two ITCZs, each with heavy precipitation.

On the other hand, contrasting conditions of weak monsoon circulation and only one ITCZ prevail, when the easterly wind at the 150 mb level is weak. The summary of the monthly variation of the zonal wind at the 150 mb level in Table 2 shows that this wind is a return flow of the monsoon forming the Hadley cell.

The distributions of precipitation are closely connected to the locations of the ITCZ. In the previous chapter, it was shown that the precipitation in the east coast of Indochina, India and Philippines reaches its seasonal maximum slightly before the increase in the steadiness of the northeast monsoon.

Thus these precipitations are not orographic as traditionally recognized. Yoshimura (1973) also conducted a similar study. In general, the results of the current study are similar to his findings. However, the current study also utilizes the wind data at the 850 mb level. Hence, the relationship to the monsoon wind was found in this study.

A slow southward migration of the ITCZ continues during the period from December to February. In December, the Malay peninsula is still under the influence of the northern ITCZ (Figure 24). By January, the northeast monsoon intensifies in this region and the ITCZ retreats from the peninsula (Figure 18). In February, the ITCZ reaches its southern limit and retreats from North Borneo (Figure 25). This month coincides with coldest surface temperature of the sea in the northern hemisphere and warmest surface temperature of the sea in the southern hemisphere.

Finally, it should be noted that the month to month persistence of the upper tropospheric monsoon is high (Table 1). This fact suggests that the fluctuations of the winter monsoon are influenced by long period oscillations of the tropical general circulation.

CHAPTER 4

RELATIONSHIP BETWEEN THE WINTER MONSOON AND THE GENERAL CIRCULATION

4-1 Description of the Walker circulation

The Walker circulation (sometimes called the Southern oscillation) was discovered by Walker and Bliss (1932,1937) and subsequently became recognized as an important mode of fluctuation of the tropical atmosphere. In general, it consists of exchange of air between the eastern and western hemispheres. In the case of northern winter, the center of action lies over the South Pacific Ocean and Australian regions where climatological elements such as pressure at the sea level and temperature are negatively correlated.

Troup (1965) modified Walker's definition and retained only the station pressure to obtain the Index of circulation. He also found that the oscillation has declined in the decades from 1921 to 1950. Berlage (1966) surveyed the correlation of the pressure at the sea level in the tropical region of the world and found that the center of action of the Walker circulation lies over Indonesia and Easter Island. The correlation of the annual pressure between Djakarta and Easter Island was found to be greater than -0.8 based on a limited samples which may contain errors (Quinn and Burt, 1972). Bjerknes (1969) has noted that the Walker circulation has a close connection with the sea surface temperature in the central Pacific Ocean.

Quinn and Burt (1970,1972) and Quinn (1974) have conducted an exhaustive survey of the relationship between the Walker circulation represented by the difference in pressure at the sea level between Easter Island and Darwin and activities of the Pacific ITCZ by the rainfall in the equatorial Islands. They found that a large pressure difference between Easter Island (high pressure) and Darwin (low pressure) precedes the onset of El Niño and subsequent increase in activities of the Pacific ITCZ. El Niño is a condition of warm sea water which is observed recurrently off the west coast of northern South America. The sudden change of the oceanographic conditions frequently causes a disastrous impact on the biomass and economies of this region. Similar investigation was also conducted by Tsuchiya (1971).

Ramage (1975) has presented the first cloudiness data based on the satellite observation during the 1972-1973 El Niño. These data have shown the increase in activities of the ITCZ in the equatorial central Pacific during the period of El Niño. Trenberth(1975) has studied the fluctuations of the pressure at the sea level at Darwin, Apia, Tahiti and Easter Island. He found a phase difference of about one season between Tahiti and Easter Island (the later leading). On the other hand, Tahiti and Darwin pressures were almost exactly out of phase. Thus using a study by Kidson (1975) as a reference, an index using only Tahiti and Darwin data was suggested.

Wyrtki (1975) has scrutinized the wind observation by ships near the equatorial Pacific as well as sea level observations in the tropical Pacific area and concluded that the El Niño events are responses of the Pacific Ocean to atmospheric forcing. He postulated that the warm sea water accumulates near New Guinea when the southeast trade wind is strong. When this trade wind weakens, the accumulated warm water returns to the eastern

Pacific via Equatorial counter current and initiates an El Niño. He also noted that the wind near the coast of Peru is not necessary weak during an El Niño.

Weare et al. (1976) has conducted an empirical orthogonal analysis of the sea surface temperature in the Pacific Ocean and found that the El Niño mode of the oscillations of sea surface temperature accounts for 23 % of the total nonseasonal variance and the largest components of the nonseasonal eigenvectors. Finally, Wyrtki (1977) has updated the previous research and concluded that the strength of the trade wind and the occurrence of El Niño are best correlated if the winds in the box of 4° N to 4° S and from 180° to 140° W were used.

In the current study the role of the Walker circulation in the winter and summer monsoons are investigated. For this purpose the index of the Walker circulation incorporating the seasonal change of the center of action was made. Since the Walker circulation has a period of 3 to 5 years, the correlations of the 5 month moving averages of the sea level pressure for a period from January 1961 to June 1978 were calculated for the stations near the center of action known from the previous researchers.

Table 3 shows the monthly variation of the Walker circulation calculated in this manner. The month given in this table is the central month of the 5 month mean. For the year as a whole, the correlation between Darwin and Tahiti is the highest which is in agreement with Trenberth (1975). However during the northern winter, the Australian center of action shifts to Cocos Island. On the other hand, during the

TABLE 3

Monthly variation of the Walker circulation

Correlation of sea level pressure (5 month moving averages)
January 1961 to June 1978

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
C-T	-0.76	-0.76	-0.60	-0.57	-0.30	-0.42	-0.38	-0.63	-0.67	-0.69	-0.70	-0.69
D-T	-0.58	-0.65	-0.75	-0.73	-0.55	-0.52	-0.65	-0.84	-0.84	-0.78	-0.68	-0.64
D-E	-0.27	-0.33	-0.35	-0.47	-0.60	-0.58	-0.73	-0.61	-0.67	-0.63	-0.53	-0.22

C = Cocos Island

T = Tahiti Island

E = Easter Island

D = Darwin

northern summer, Easter Island also lies near the center of action. In all the months the sea surface temperature is cold in the equatorial Pacific Ocean when the pressure at the sea level is high in Tahiti or Easter Island and low in Darwin or Cocos Island. The reverse situation is true when the sea surface temperature is warm and frequent El Niño is observed.

A pair of stations with highest negative correlation of the pressure shown in Table 3 were used to construct the Walker circulation index defined by the fluctuations of the pressure at the sea level. Four indices were calculated to cover the four seasons of the year. These are centered on the months of January, April, July, and October.

In the case of 5 month moving average centered on January, the pressure at the sea level at Cocos Island and Tahiti was chosen. Indices are defined as positive if the pressure in Tahiti is higher than normal. In April and October, Darwin and Tahiti were chosen. Finally during the northern summer, Darwin and the normalized linear combination of Tahiti and Easter Island (R= -0.83) was chosen to represent the Walker circulation. Except for the location of the pair of stations chosen to represent the center of action, the method used to calculate the indices is similar in all seasons.

The method of calculating the Walker Circulation Index (W.C.I.) is as follows: For each of the two stations used in the index, the 5 month moving averages of the departures from the normal pressure at the sea level were obtained. These departures were then normalized by the standard deviation of the departures centered on a given month. Finally the W.C.I. is obtained by the following formula:

$$W.C.I. = \frac{P(1)-P(2)}{|P(1)+P(2)| + 1}$$

where P(1) = normalized departures of the 5 month moving averages of the pressure at the sea level in South Pacific (Tahiti or Tahiti and Easter in July index)

P(2) = normalized departures of the 5 month moving averages of the pressure at the sea level near the Australian region (Cocos in northern winter or Darwin in other season)

Figure 26 illustrates the fluctuations of W.C.I. defined by above formula for the period from January 1955 to April 1978. Solid line is W.C.I. which has a long period oscillation of three to five years. In addition, there is a short period fluctuations of about 6 months to one year. This short period fluctuation appear to be a real fluctuation because smoothing of the values will result in lower relationship to the winter monsoon.

The dashed line is the 5 month moving averages of the values of the El Niño mode of the oscillation (NS 1) of the sea surface temperature obtained by Weare et al. (1976). The scale of this mode is reduced by a factor of five from the original paper in this figure. Thus the isotherm of 0.2°C in the original figure on NS 1 will have a 1°C departure of sea surface

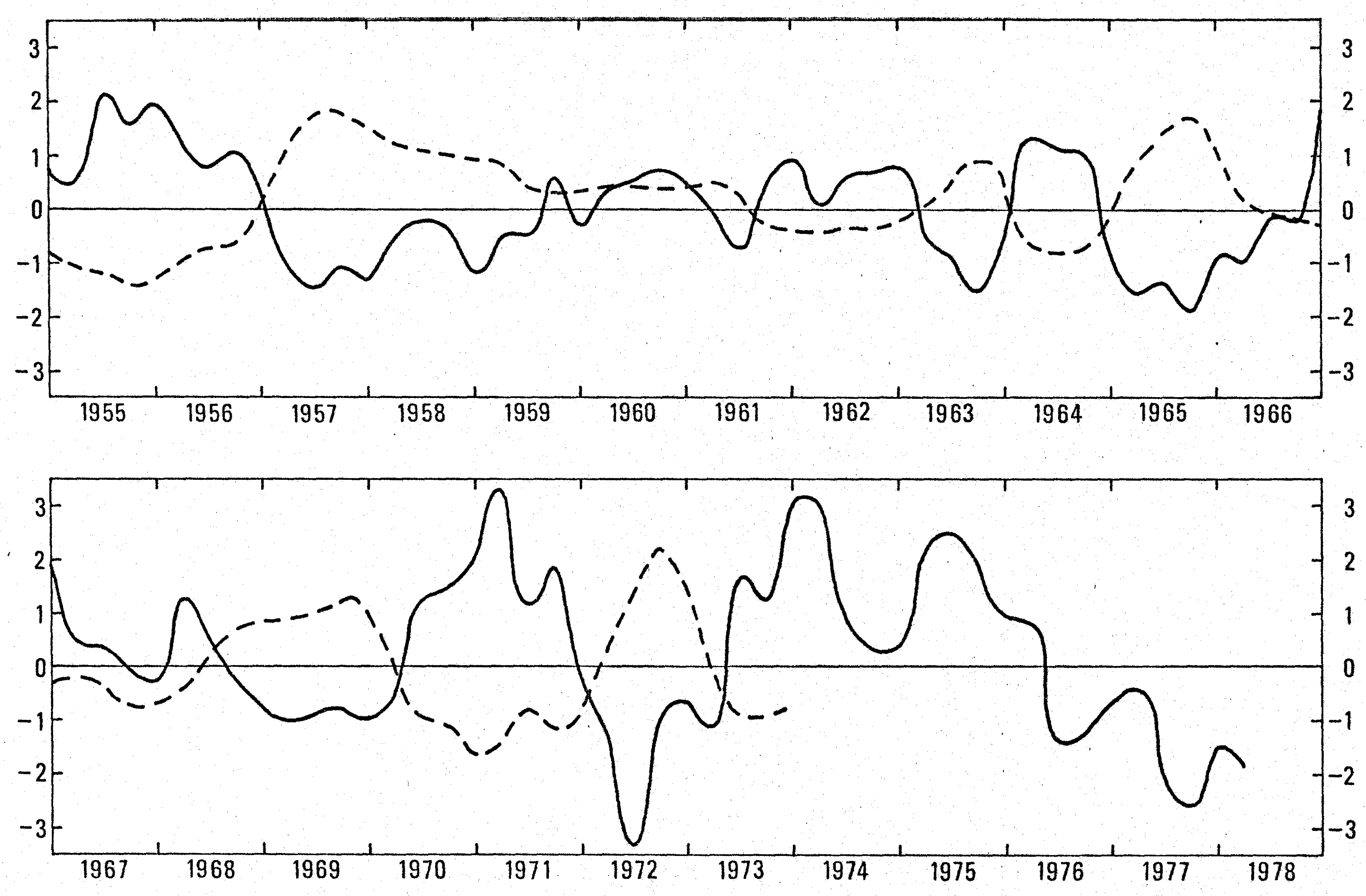

Fig. 26 Walker Circulation Index (solid line) and El Niño mode (NS 1) of the sea surface temperature in the Pacific Ocean (broken line)

temperature if the values on Figure 26 is 1°C.

An inspection of Figure 26 shows that the occurrence of El Niño is associated with the negative values of W.C.I. (higher pressure near Australia). On the other hand, the positive values of W.C.I. are associated with the cold sea surface temperature in the equatorial Pacific. Thus the Walker circulation is the coupled atmospheric and oceanic oscillation of the tropical Pacific Ocean.

4-2 Walker circulation and the winter monsoon

Since the Walker circulation has a long period of oscillations, the 3 month average of the winter monsoon was considered for a study of the interactions. Figure 27 shows the time-height cross-section of the zonal wind observed at Singapore during the winters from 1961 to 1978 (January year). The fluctuations are quite similar to that of January with a notable exception of 1978 when the seasonal average of the zonal wind at the 150 mb level is weak (Table 1).

Figure 28 shows the relationship between the winter monsoon expressed by the zonal wind at the 150 mb level at Singapore and the January W.C.I. shown previously in Figure 26. The solid line is the winter monsoon. The symbol (●) denotes the value of W.C.I.. In the 18 winters from 1961 to 1978. the correlation between these fluctuations is +0.94. This shows that the interannual fluctuations of the winter monsoon are the part of the main fluctuations of the Walker circulation.

The synoptic patterns of this oscillation were then investigated. Figure 29 shows the mean pressure at the sea level in January in the Pacific Ocean region. The source of data is

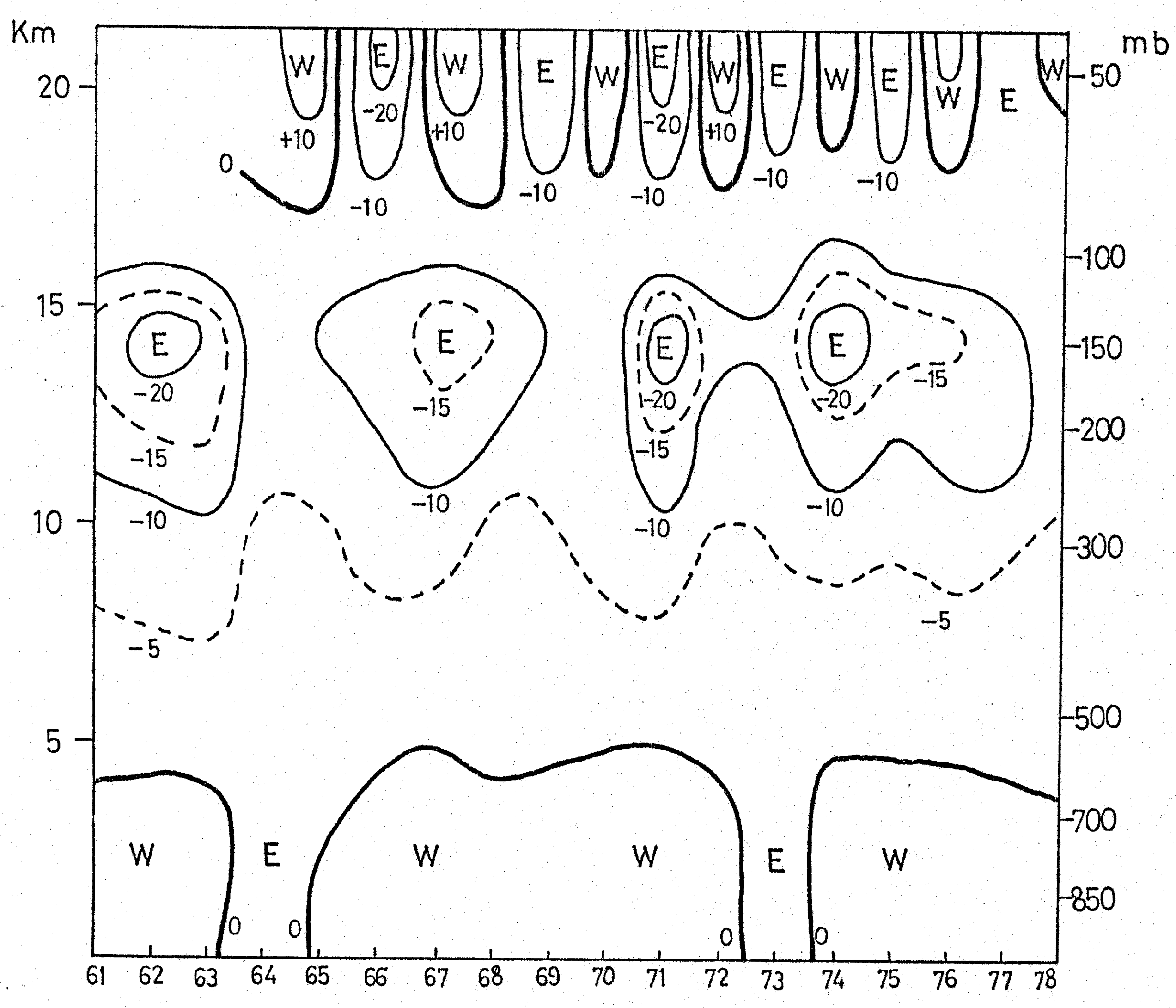

Fig. 27 Zonal wind in winter at Singapore (Dec.- Feb.)

1961-1978 (unit: m/sec)

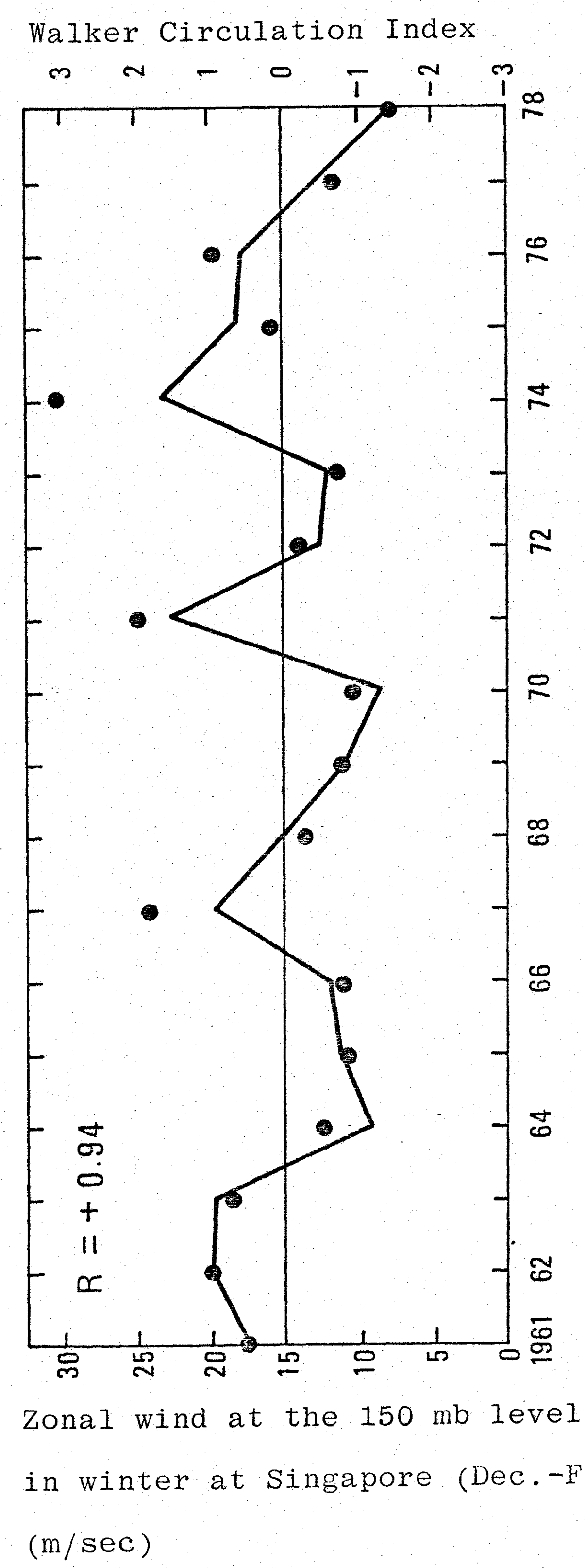

Fig. 28 Fluctuations of the winter monsoon (the solid line) and the Walker Circulation Index (large black dot)

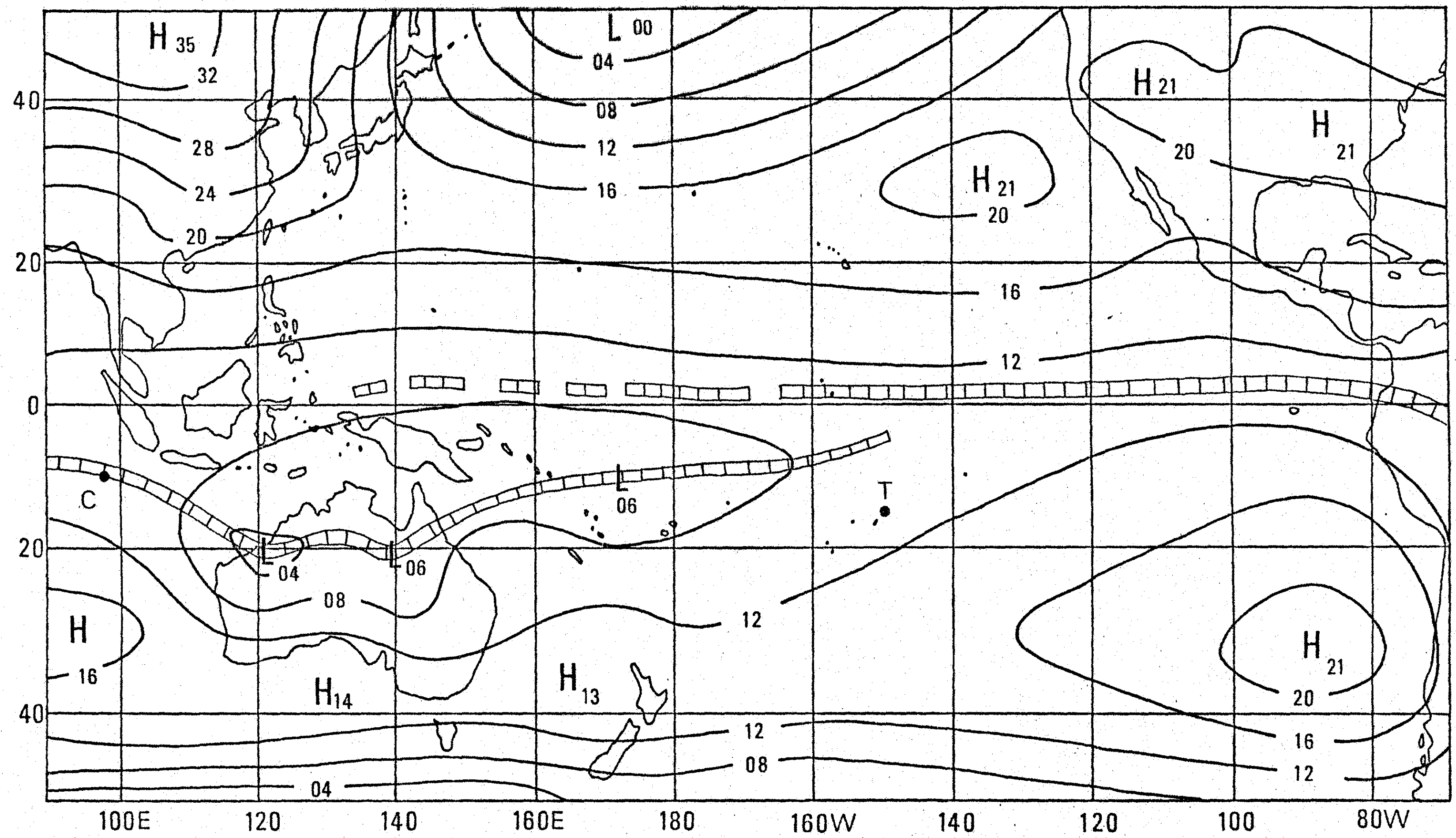

Fig. 29 Mean pressure at the sea level in January (unit: 1000 mb+)

C : Cocos Island , T : Tahiti

the climatological normal used by Japan Meteorological Agency. This data contain ship reports as well as the observations at the scattered islands. In the tropical regions, the flow patterns are quite similar to that of the 850 mb level (Figure 7). The locations of Cocos Island and Tahiti which are shown on the map suggests that the Walker circulation in winter is the longitudinal oscillation of the equatorial trough extending from Australia to about 160°W.

Figure 30 shows the difference in the pressure at the sea level between the winters with strong easterlies at the 150 mb level (1962,1963,1971,1974) and the winters with weak easterlies (1964,1969,1970,1978). In winters with strong monsoon circulation, the equatorial trough is active to the south of Java and northern Australia. On the other hand, the trade wind to the north of Tahiti is strong and the sea surface temperature is cold. Hence the active zone of ITCZ shifts toward Indian Ocean. The northeast trade winds near the 180° date line are weak. Strong monsoon circulation in the winter monsoon regions are compensated by a weak Hadley cell activity in central Pacific.

In the higher northern latitudes, the Aleutian low is displaced southwest toward Japan and slightly weaker than the normal. The correlation of the central pressure of this low to the wind at the 150 mb level at Singapore is only +0.39 in 18 winters from 1961 to 1978. The intensity of the Siberian high also has a similar low correlation (+0.29).

In winters with weak easterlies at the 150 mb level, weak monsoon circulation is compensated by a strong Hadley cell activities in the central Pacific and the warm sea water. The interesting aspect of the winter Walker circulation is the

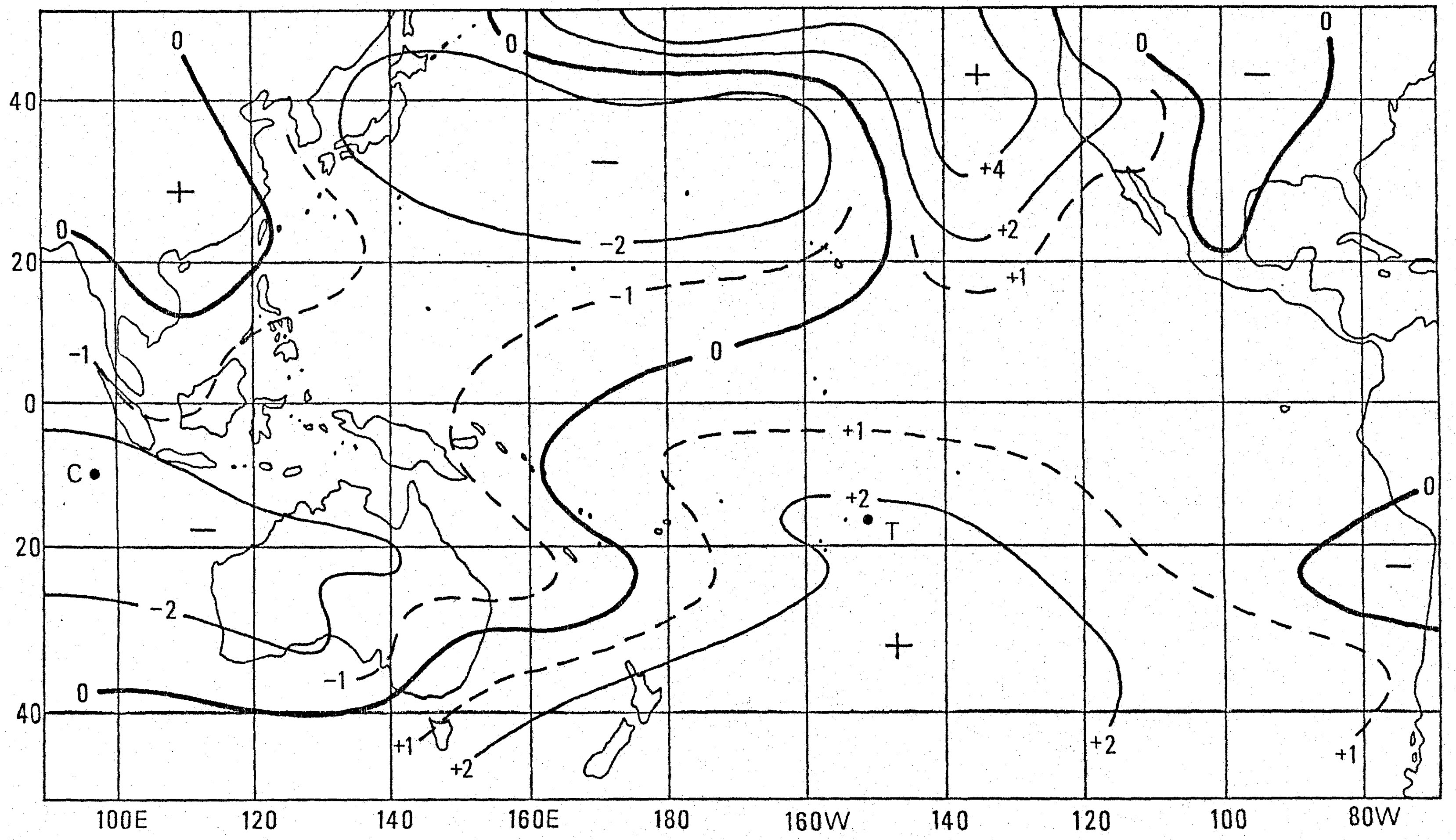

Fig. 30 Difference in the pressure at the sea level in winter (unit: mb)

(strong easterly case minus weak easterly case)

C : Cocos Island , T : Tahiti

relationship to the El Niño. None of the major El Niño years 1957,1965, and 1972 is included in the winters with weakest easterly wind. This shows that the interaction between the oscillations of the equatorial trough (northern winter only) and the subtropical high in the South Pacific may be the key to the initiations of the El Niño. In addition, a seasonal change of the center of action of the Walker circulation (Table 3) may be a vital factor in initiating the El Nino.

4-3 Circumpolar westerlies in the northern hemisphere at the 500 mb level

The relationship between the middle latitude circulation and the Walker circulation was first suggested by Bjerknes (1969). He postulated that the strength of the Aleutian low is linked to the Walker circulation. The current study suggests that only the location of the Aleutian low has a high relationship to the fluctuations of the Walker circulation which includes the tropical winter monsoon.

Figure 31 shows the climatology of the geopotential height at the 500 mb level in winter. The trough near 140°E is associated with the cold monsoon surges to Japan and the developments of the Aleutian low to the east. Figure 32 shows the departure from the zonal mean of the climatology of the geopotential height at the 500 mb level. The standing eddy component is near Hokkaido and is advecting cold Siberian air to Japan and warm Pacific air to the northeast Siberia. The ridge of the positive anomaly coincides with the Siberian high. Figure 33 shows the difference in the geopotential height at the 500 mb level between the winters with strong and weak tropical monsoon.

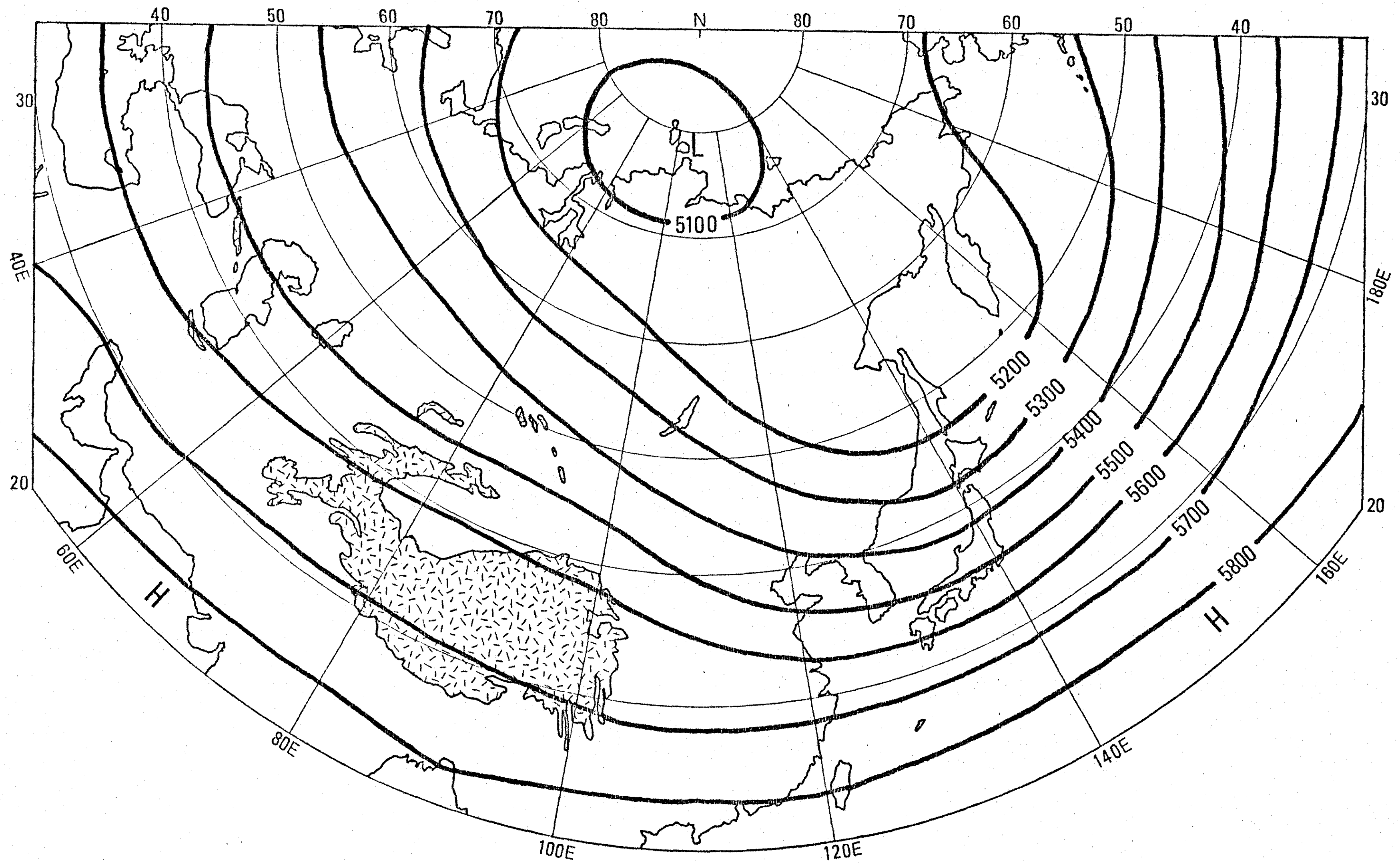

Fig. 31 Geopotential height at the 500 mb level in winter (1946-1975 average)
(unit: geopotential meters) (data adopted from Kisetsu Yoho Shiryo)

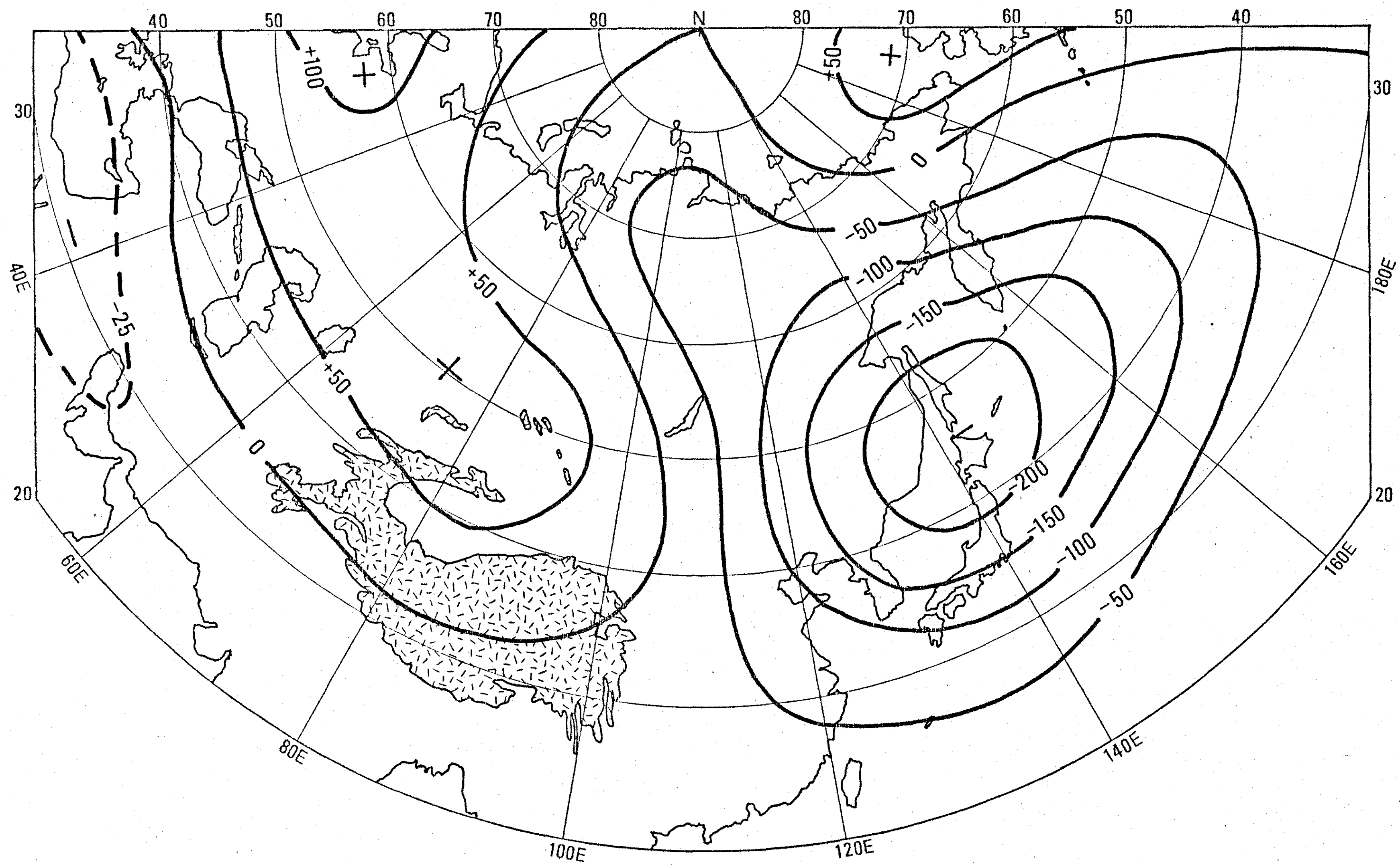

Fig. 32 Departure of the geopotential height from the zonal mean at the 500 mb level in January (1946-1975 average) (unit: geopotential meters)

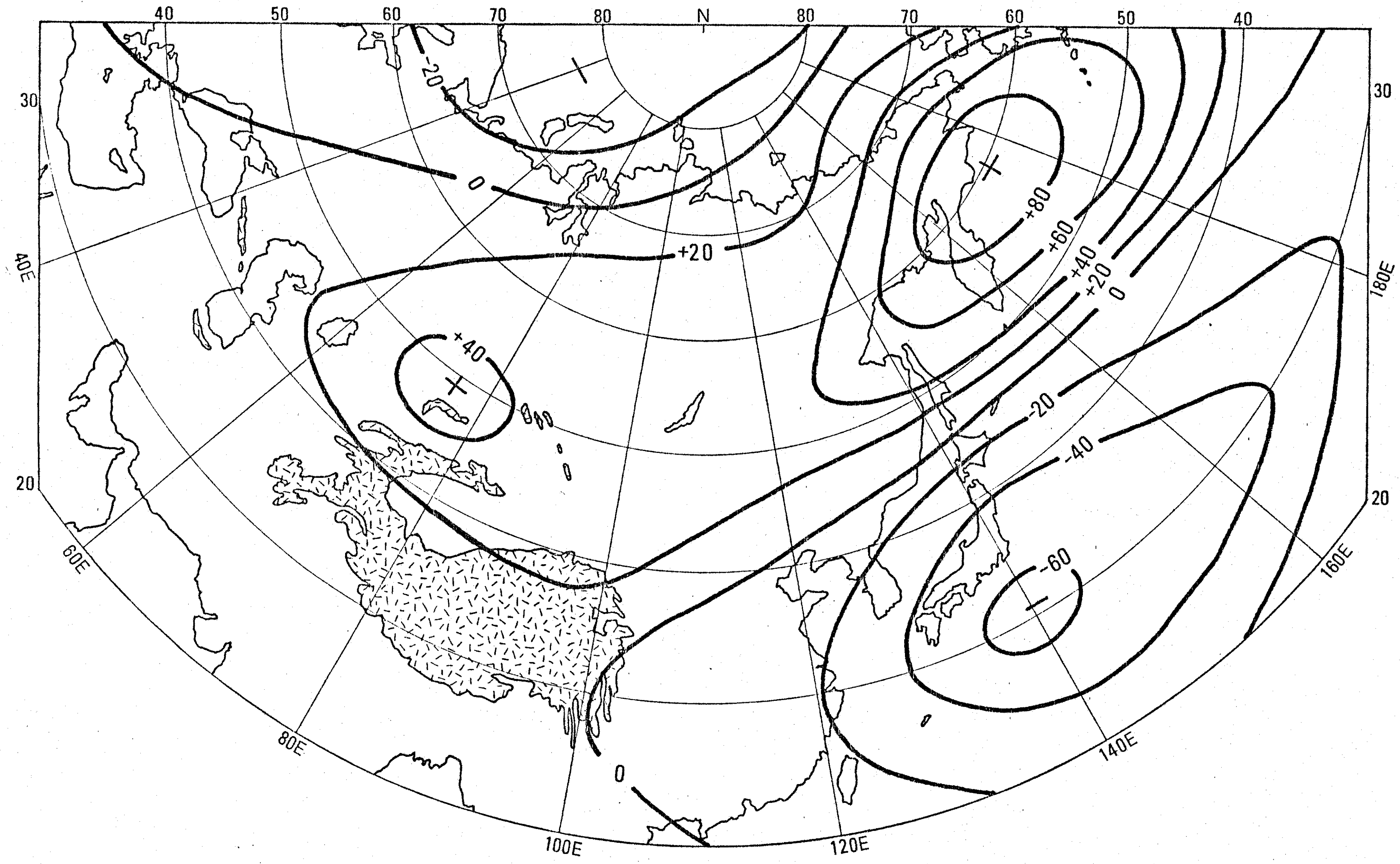

Fig. 33 Difference in the geopotential height at the 500 mb level in winter
(unit: geopotential meters) (strong easterly case minus weak easterly case)

The southward displacement of the trough near Japan and a blocking to the east of Kamchatka with weaker zonal westerlies between these centers characterize the extratropical circulation associated with the strong tropical monsoon. The two grid points with highest correlation to the wind at the 150 mb level at Singapore are 30°N 140°E (R=-0.56) and 60°N 170°E (R=+0.48) which corresponds to the two large centers of anomalies.

Since the center of blocking near Kamchatka does not coincide with the Aleutian low, the displacement of the location of this low toward Japan occurs more frequently during the winter with strong tropical monsoon. The reverse condition is common when the tropical monsoon is weak.

These displacement of the locations of the Aleutian low is probably influenced by the strength of the subtropical high in the Pacific. If the winter monsoon is strong, the subtropical high is weak (weak Hadley cell) and the tracks of the migratory cyclones shift southward. This results in a southward displaced Aleutian low. On the other hand, the strong subtropical high in the Pacific will shift the tracks of cyclones to the north. Consequently, the Aleutian low which is a statistical average of activities of the cyclones also shift northward.

CHAPTER 5

GENERAL DESCRIPTION OF THE SUMMER MONSOON

5-1 Climatology

In this section the mean climatological conditions during the peak of the summer monsoon are discussed. The seasonal change toward the summer monsoon regime is discussed in the next section. Figure 34 shows the mean precipitation in June. In most of the regions between India and southern Japan, there is at least 100 mm of precipitation with the few exceptions of the southeast coast of the Indian peninsula and east coast of Indochina. These dry regions are located in the rainshadow of the southwest monsoon. On the other hand, the west coast of the Indian peninsula and coastal Burma have heavy precipitation of over 400 mm. Unlike the case of the winter monsoon, this rainfall appears to be orographic in nature and the heaviest precipitation coincides with the month of strongest southwest monsoon which is July in India and July or August in Burma.

The role of the ITCZ is also important. The rainfall in most of India near the Ganges river, the southern coast of China, and Micronesia is due to activities of the ITCZ. The heavy precipitation in southern Japan is associated with the Baiu front. The large regions with scanty precipitation in the Middle East, central Asia, western China and Australia coincide with the location of the subtropical high.

Figure 35 shows the geopotential height at the 850 mb level in June. There is a low in the northern India and the

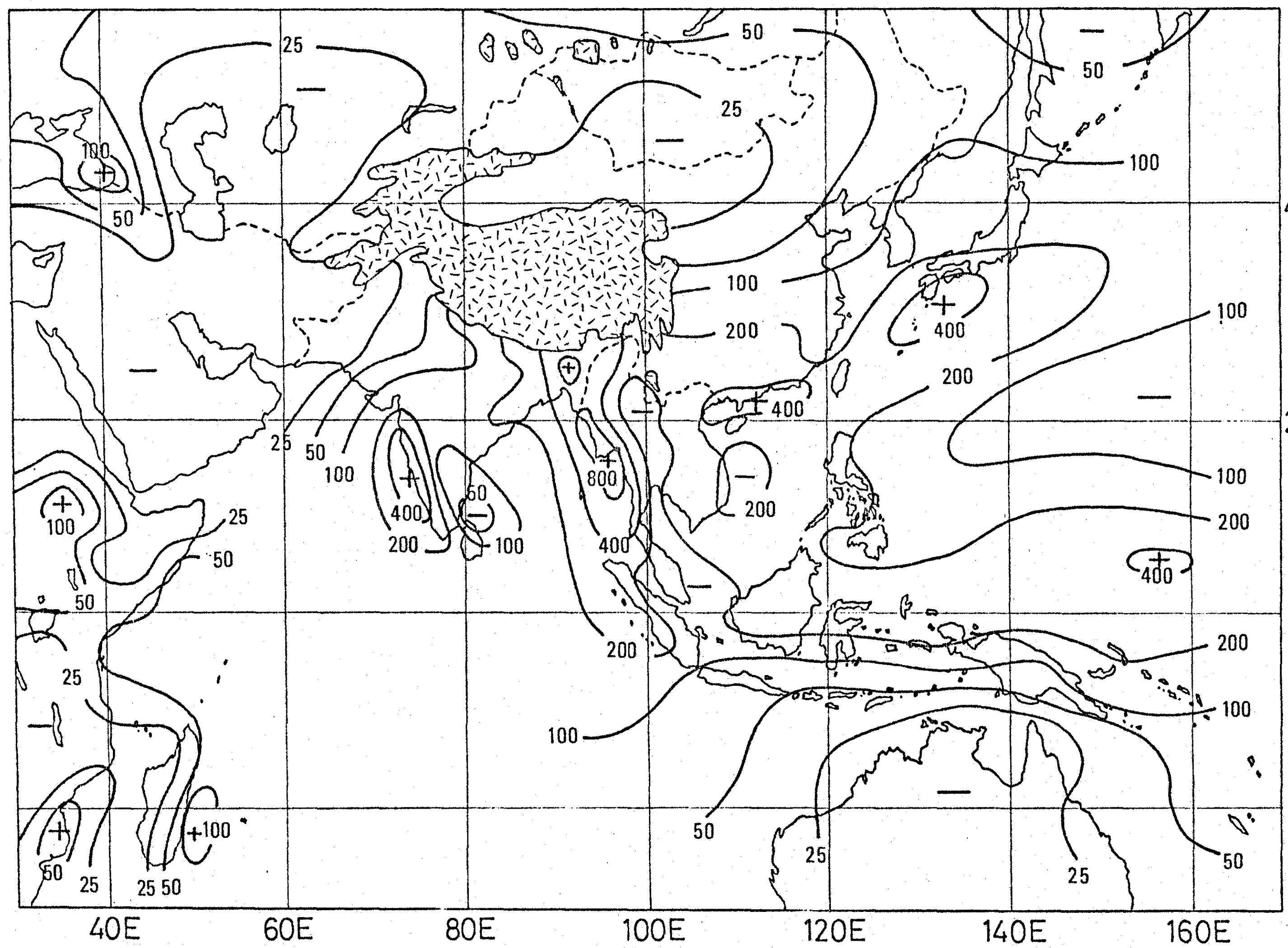

Fig. 34 Precipitation in June (1951-1960 average) (unit: mm)

(data adopted from World Weather Records,1967)

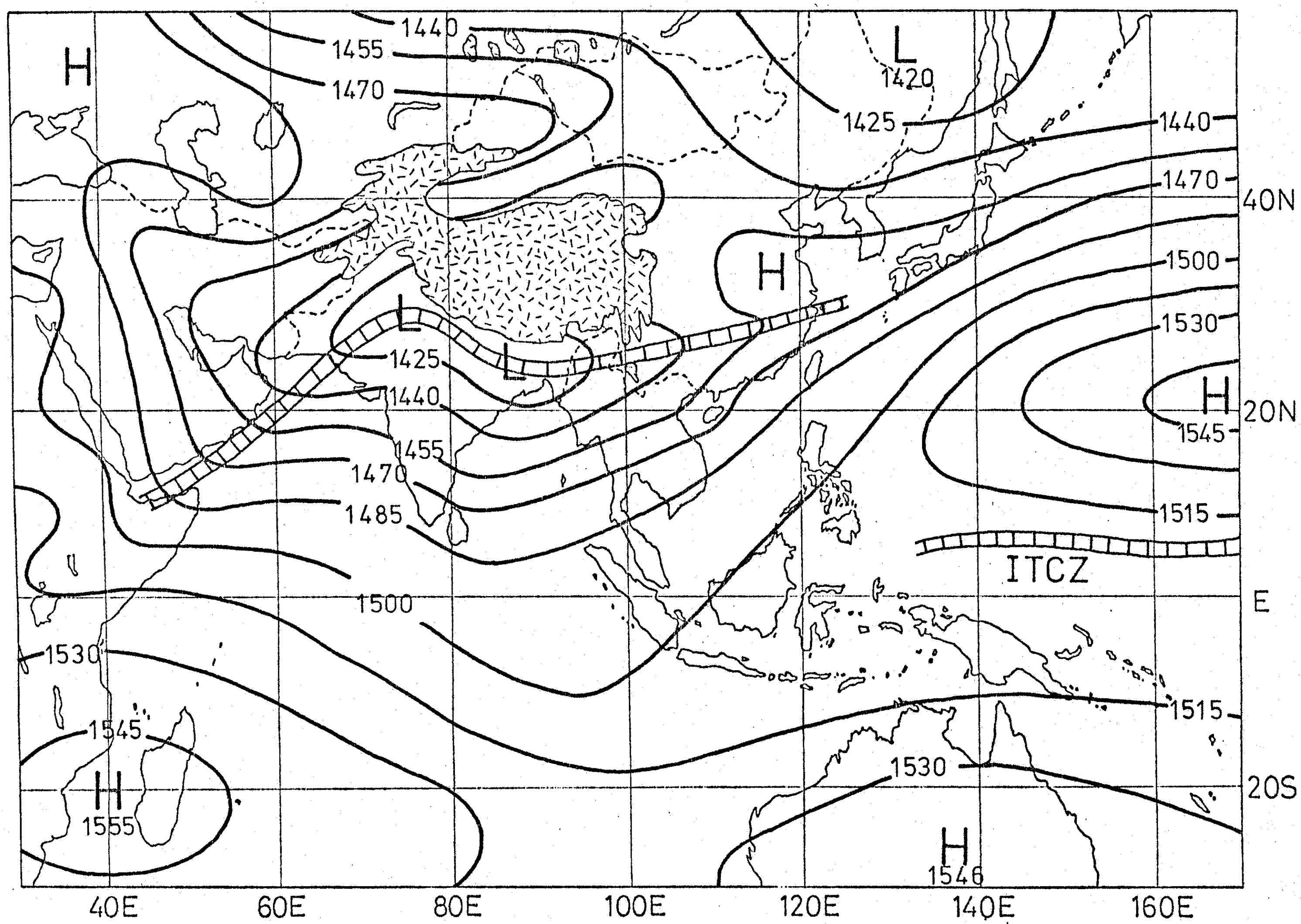

Fig. 35 Geopotential height at the 850 mb level in June (1964-1975 average)
(unit: geopotential meters)

subtropical high is located near Madagascar and Australia. In Indian Ocean, the southeast trade wind prevails to the south of the equator and the southwest monsoon to the north of the equator. The cross equatorial flow at the low level is most intense near the east coast of Africa where the Somali jet is located.

Findlater (1969a,1969b) has analyzed pilot balloon data in the western Indian Ocean and discovered the existence of the Somali jet. The jet has a core of maximum wind near the 850 mb level and accounts for about 50 % of the cross equatorial transport in July. Saha (1970) has compared the flux of air and water vapor transport at the west coast of India and equatorial transport by the Somali jet and concluded that 63 to 84 % of air and 65 to 73 % of water vapor flux at the west coast of India are transported from the southern hemisphere by the Somali jet. Further investigation by Findlater (1974) has shown that this jet is a part of the Asian monsoon circulation. The jet splits into 2 branches off the coast of Somalia. The developments of the northern branch may coincide with a burst of monsoon rain in the west coast of India.

The ITCZ located over the Ganges river region of India is called the monsoon trough and plays a vital role in the distribution of the monsoon rainfall. The locations of the monsoon trough are under the strong influence of the development and tracks of the monsoon depression. Raghavan (1967) has conducted a statistical survey of the role of the monsoon depression on activating and surpressing monsoon rainfall in India. Anjaneyulu (1969) has conducted an estimate of sensible and latent heat over the monsoon trough region. The computation shows that this region is importing latent heat and exporting sensible heat.

Raghavan (1973) has shown that the interaction between the monsoon depression and monsoon trough plays an important role in the distributions of the monsoon rainfall in India. If the monsoon depression moves north to the foothill of the Himalaya mountains, the monsoon trough also moves north and dry rainless period called the break monsoon begins in the Ganges river region. The monsoon rain resumes only when the monsoon trough retreats southward to its normal location.

Finally Krishnamurti et al. (1975) have conducted a detailed analysis of the structure of the monsoon depression. It was found that this depression is an intense closed vortex and extends up to about 9 km. The vortex is cold core in the lower troposphere and a warm core above the 500 mb level. To the west of the cold core, there exists a warm core in the lower troposphere created by the advection of desert air by the storm circulation.

The existence of the different types of the disturbances is the primary cause of the inhomogeneous distribution of the Indian monsoon rainfall. Subbaramayya (1968) has shown that the monsoon rainfall regime in India can be divided into eastern and western parts with opposite tendencies. In some years a break monsoon period may result in a severe drought (Ayyar et al., 1973) . Fortunately in most regions of India, the monsoon rainfall is steady or increasing in recent years (Parthasarathy and Dhar, 1974).

Returning to Figure 35, the ITCZ over the northern India extends eastward toward southern Japan. As it approaches Japan, the convergence zone gradually loses its tropical characteristics and becomes a polar front near Japan (Kurashima and Hiranuma, 1970). Yabana (1973) has conducted a survey of the large scale aspects of the Baiu front and reached a similar conclusion.

Finally, the subtropical high in the northern hemisphere is separated by orography into three cells. In the regions to the west of 100°E, the axis of the high is located near 42°N. There is a small high in the North China plain and the western extension of the Pacific high is located near 20°N to the east of 130°E. The difference in the latitudes of the subtropical high in Eurasia and Pacific Ocean reflects the thermal contrast between land and ocean.

Figure 36 shows the geopotential height at the 150 mb level in June. There is an immense South Asian high extending along 25°N from the Red Sea to Taiwan. To the north of this high, the subtropical jet stream flow eastward from Turkey through the northern fringe of Tibetan plateau to central Japan. South of the high a strong tropical easterly jet (here after called TEJ) is observed with maximum wind speed near the southern Indian peninsula where the geostrophic wind is largest. The trough located in the equatorial Indian Ocean plays an important role in the upper level monsoon. South of 20°S there is another subtropical jet stream flowing eastward from Africa to Australia. With the exception of the Baiu front zone near Japan, the location of the subtropical jet stream generally coincides with the position of the subtropical high of the respective hemisphere.

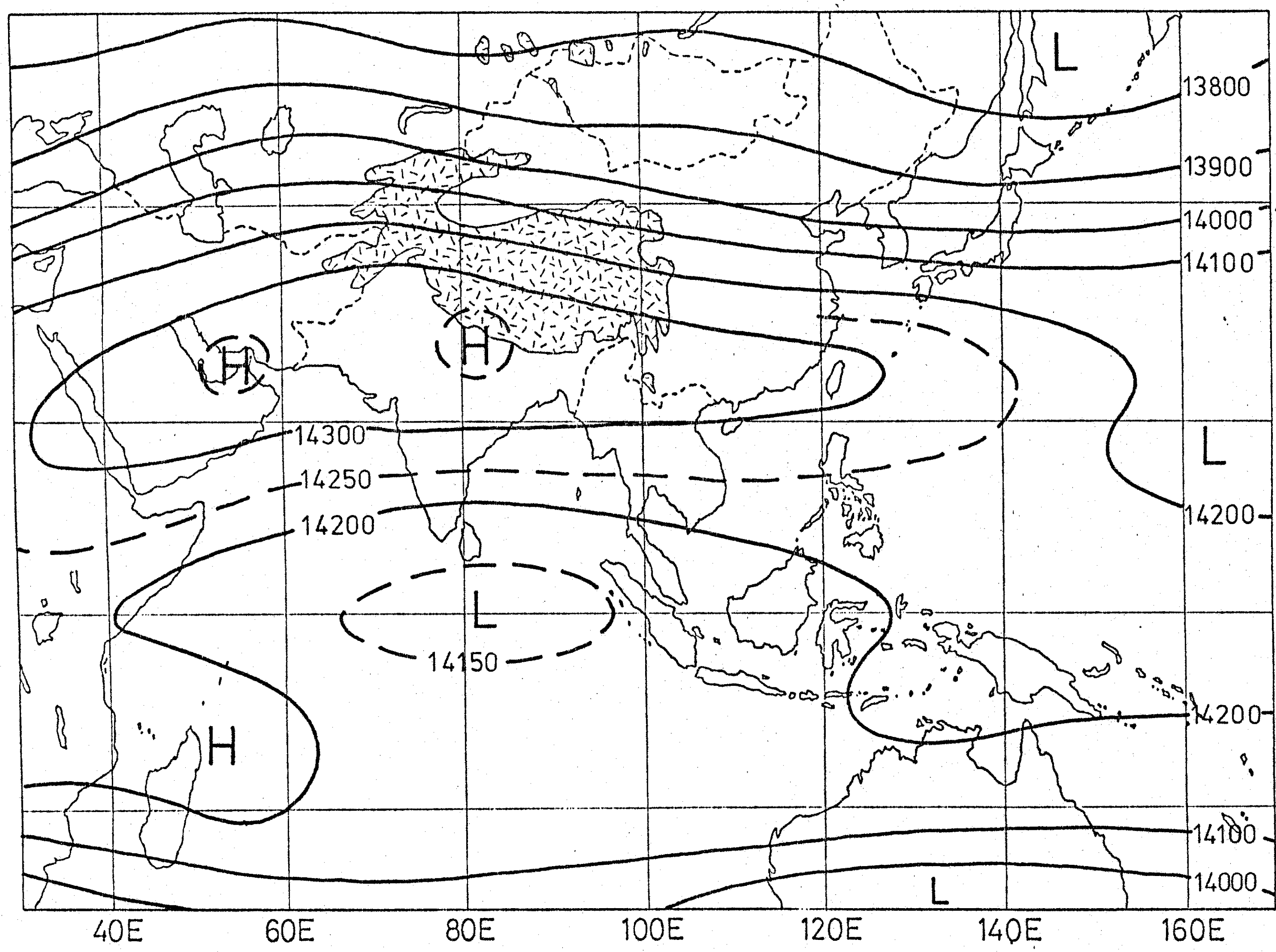

Fig. 36 Geopotential height at the 150 mb level in June (1970-1975 average)
(unit: geopotential meters)

Figure 37 shows the mean precipitation in July. The basic distribution patterns are similar to that of June. However, important seasonal changes have occurred from June. In India, the North China plain, Korea and central Japan, a marked increase in the precipitations have occurred. This increase in release of latent heat intensifies the monsoon circulation to their seasonal maximum. The orographic precipitation in the west coast of India reaches its maximum as the southwest monsoon reaches its peak development.

Figure 38 shows the geopotential height at the 850 mb level in July. In the Indian region, the southwest monsoon reaches its seasonal maximum. The Pacific subtropical high migrates north to 25°N. The ITCZ near 120°E is in the transition season. In the first half of the month, the ITCZ from India gradually changes its characteristics in the Baiu front zone and extends northeastward from Assam to the southern tip of the Korean peninsula.

In the middle of July, the Baiu season ends in southern Japan which coincides with northward expansion of the South Asian high and a rapid northward shift of the subtropical jet to Hokkaido. Simultaneously, the typhoon activities begin to intensify in the tropical Pacific and a new zone of the ITCZ begins to develop near the Philippines.

Figure 39 shows the mean precipitation in August. Compared to July, the precipitations have decreased in the west coast of India and increased near the ITCZ and typhoon zone near the Philippines. The regions near Hong Kong and Okinawa experience their 2nd peak in precipitation. In southern Japan, precipitation

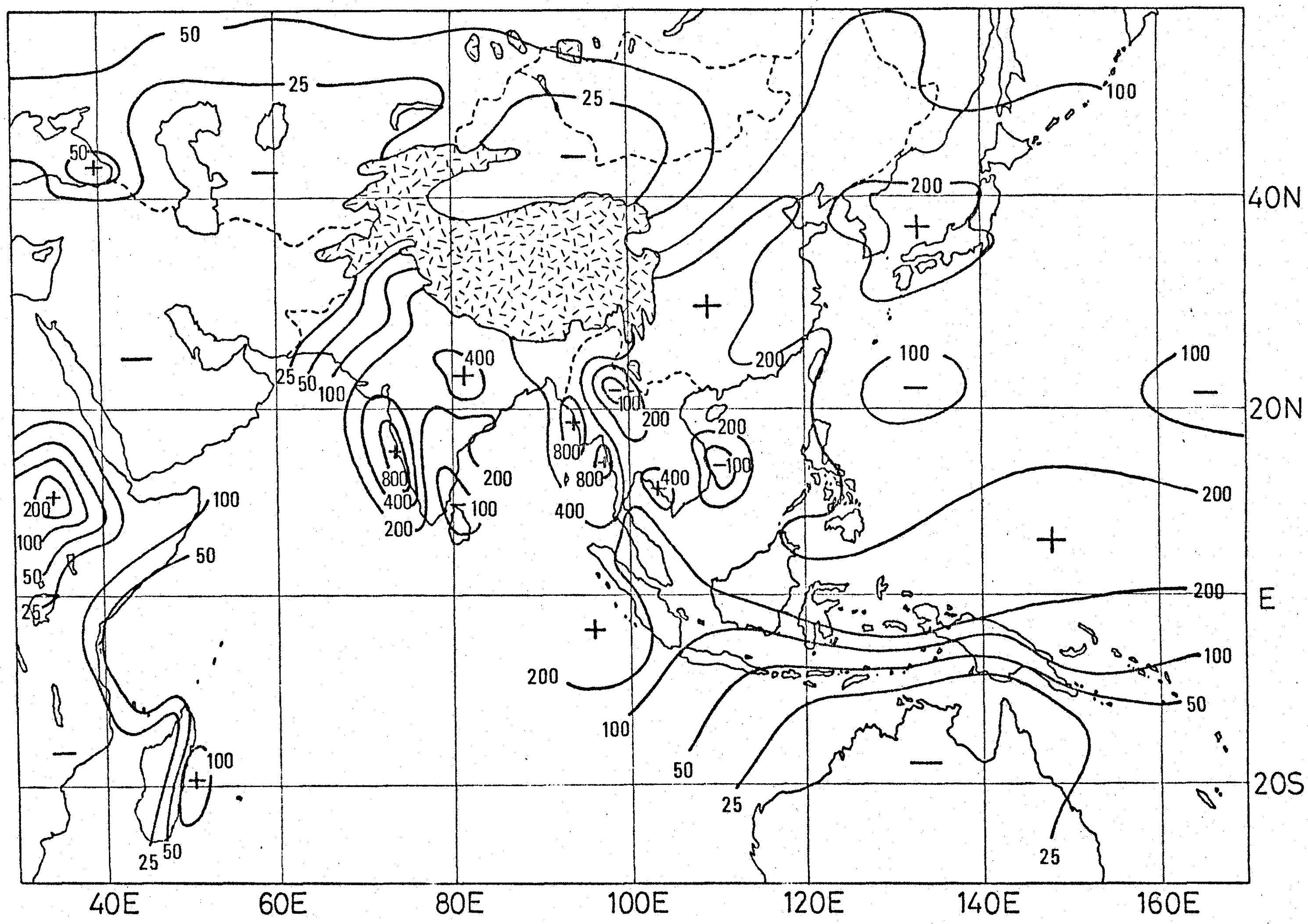

Fig. 37 Precipitation in July (1951-1960 average) (unit: mm)

(data adopted from World Weather Records,1967)

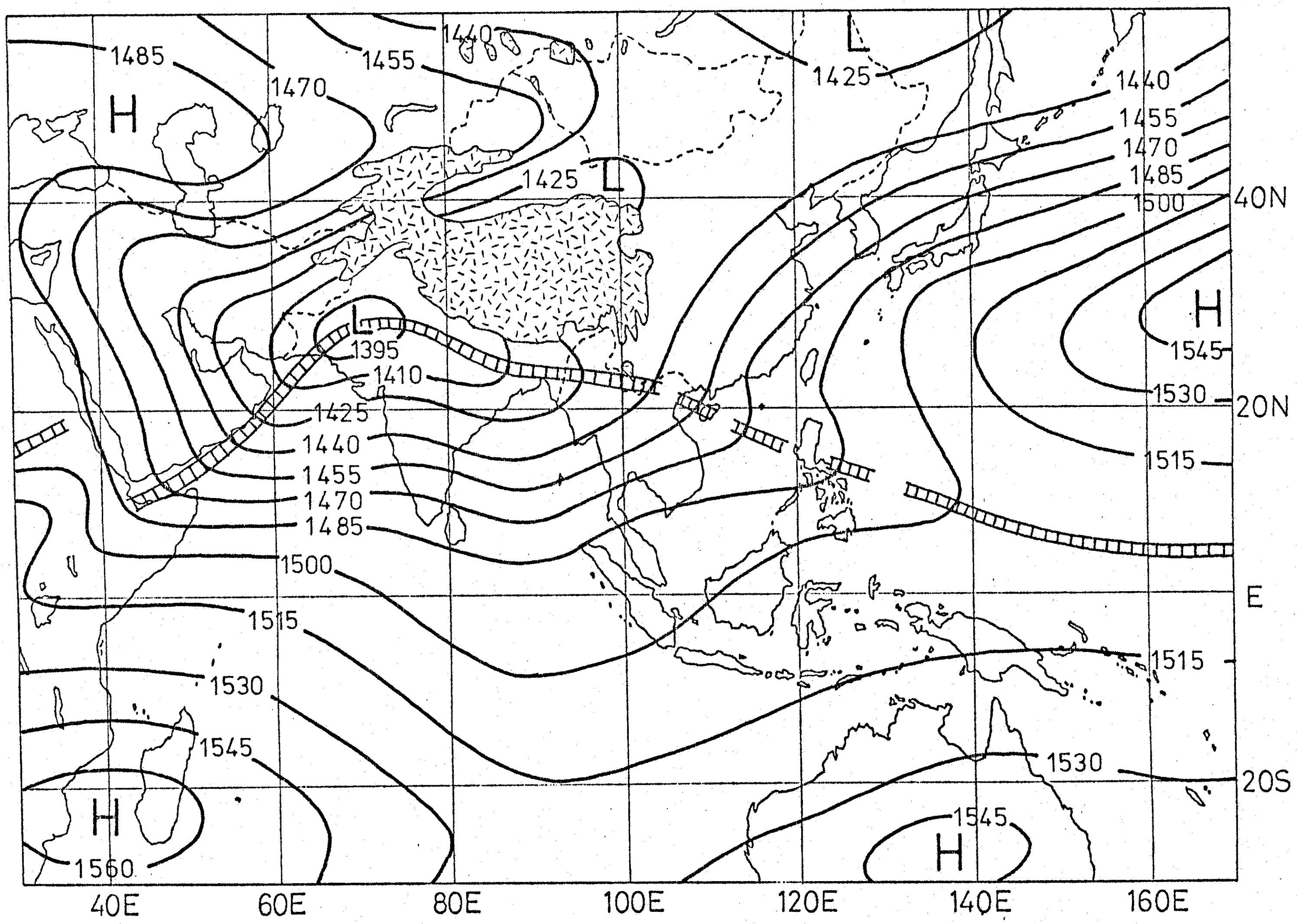

Fig. 38 Geopotential height at the 850 mb level in July (1964-1975 average)

(unit: geopotential meters)

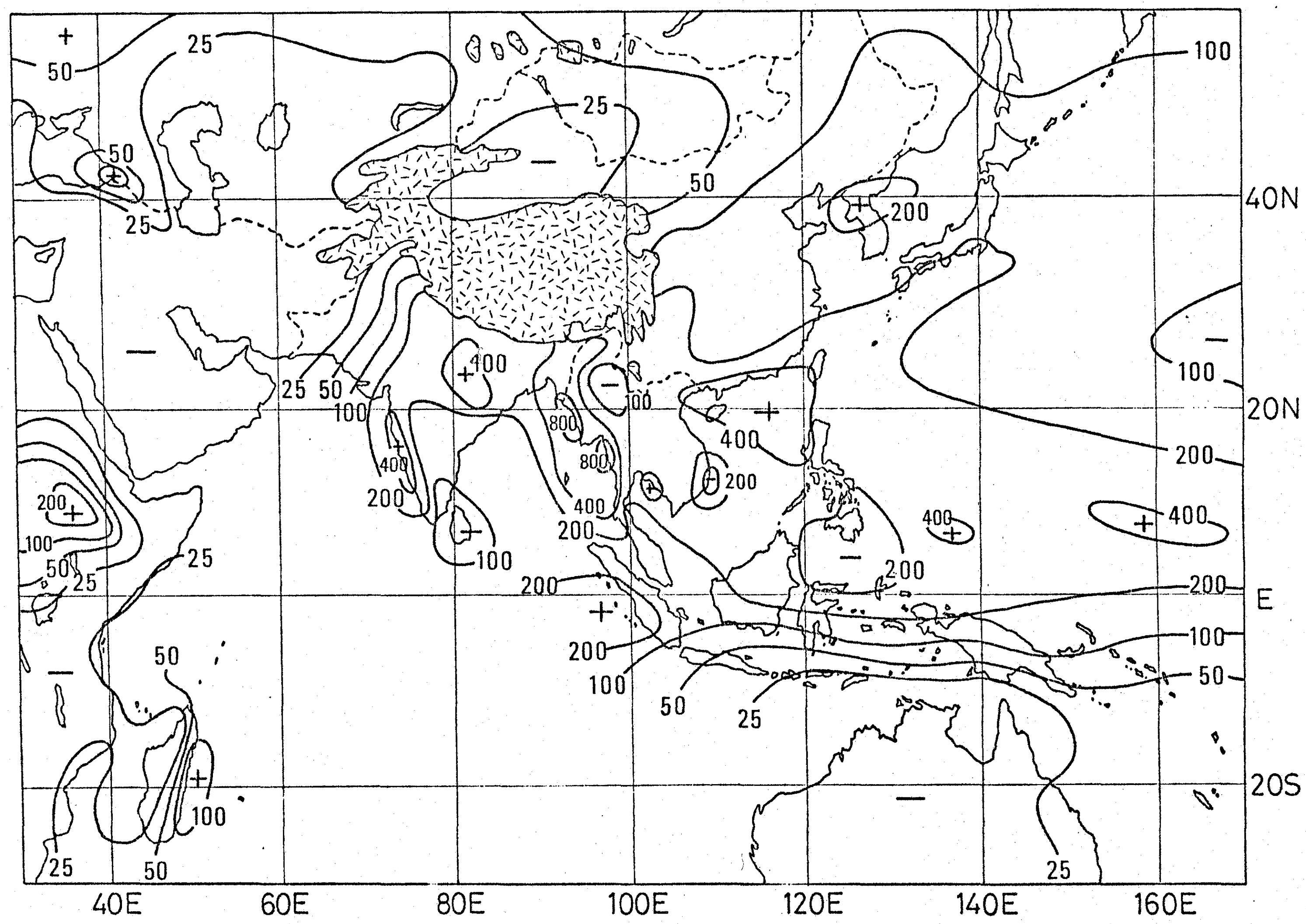

Fig. 39 Precipitation in August (1951-1960 average) (unit: mm)

(data adopted from World Weather Records,1967)

generally decreases in most areas except near the Pacific coast where typhoon activities maintain heavy precipitation. If entire Asian region is considered as a single source of latent heat, the center of the heat source shifts eastward toward Pacific Ocean in August. The warming of the Pacific and the increased activities of the typhoons are the main reasons for this change.

Figure 40 shows the geopotential height at the 850 mb level August. In India, the southwest monsoon is weaker than in July. The Pacific subtropical high reaches its highest latitude of the year and its western extension is located near 33°N. The ITCZ near 120°E is under the strong influence of the typhoon activities. In a given year there is only one ITCZ. If the activities of the typhoons are weak, the ITCZ is located near the Philippines and the axis of the Pacific subtropical high is located near 30°N and extends westward to the East China Sea. On the other hand, if the activities of the typhoons are intense, the ITCZ is located near Okinawa and the western extension of the Pacific subtropical high is located near 36 N. Many typhoons frequently take stationary positions for several days near Okinawa.

Figure 41 shows the geopotential at the 150 mb level in August. The South Asian high located near 25°N in June shifts to near 30°N in August. The subtropical jet in the northern hemisphere is now located near 45°N and the influence of the Himalaya mountains is much weaker than in June. The TEJ between the South Asian high and the trough in the equatorial Indian Ocean is most intense during July and August. The isotachs of the TEJ are shown by the thin lines. Compared

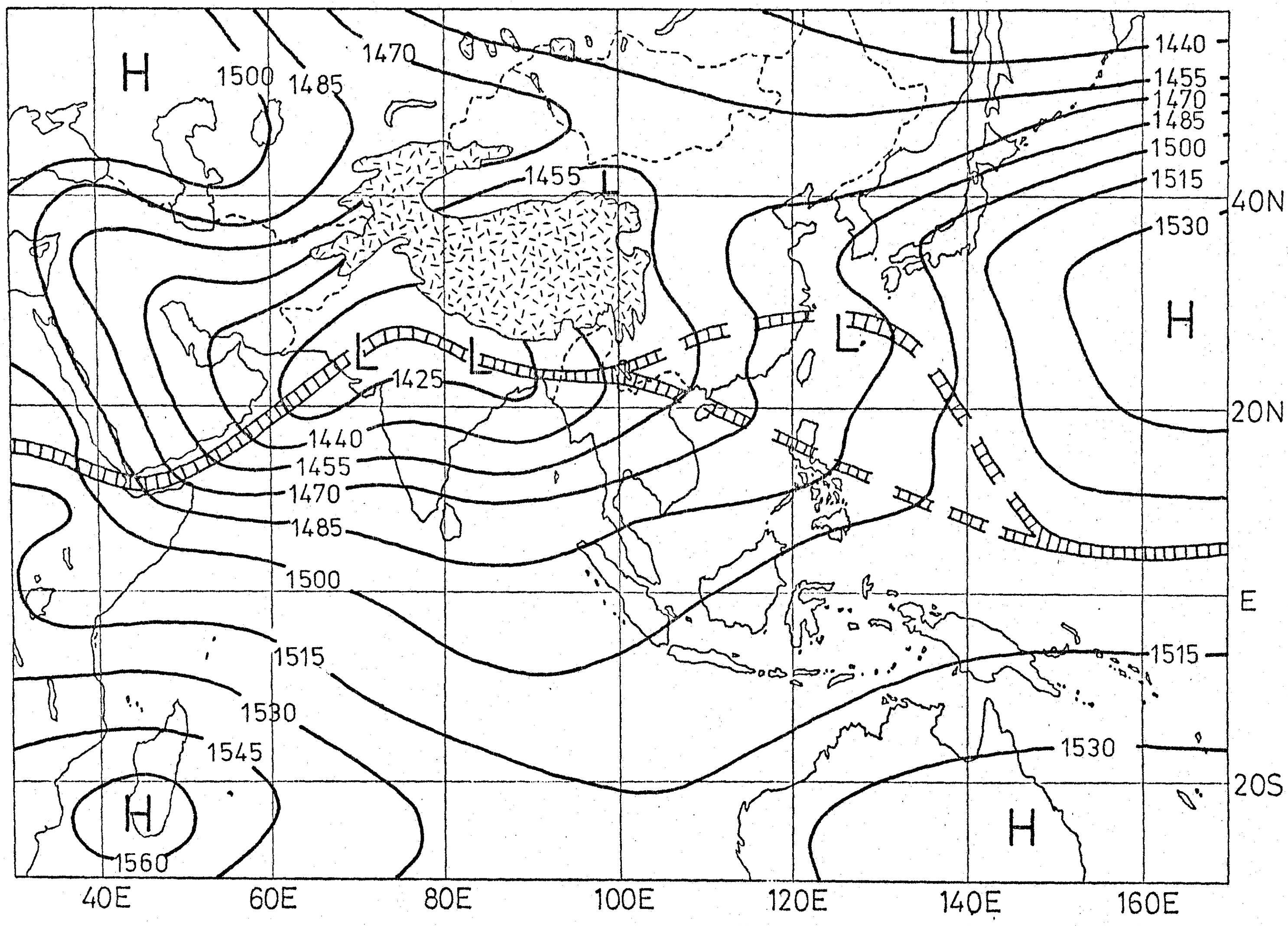

Fig. 40 Geopotential height at the 850 mb level in August (1964-1975 average)
(unit: geopotential meters)

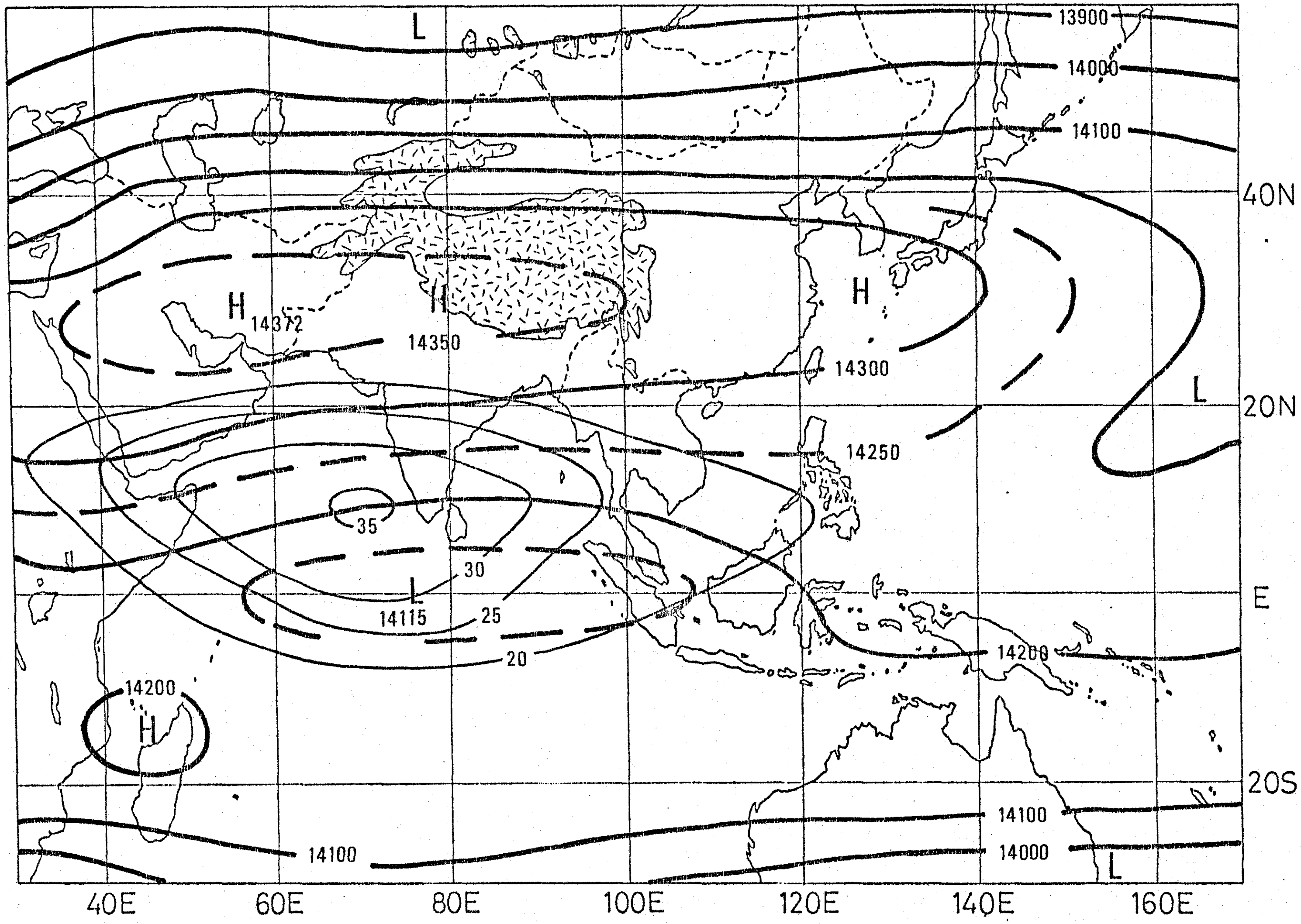

Fig. 41 Geopotential height at the 150 mb level in August (1964-1977 average)
(unit: geopotential meters)
(the isotachs of the TEJ are shown by the thin lines)

to July the shift of the latent heat source towards Pacific Ocean in August is also reflected in the speed of the TEJ at 10°N at the 150 mb level. The average wind speed is slightly stronger in August between the longitudes of 80°E to 110°E and weaker than in July between the longitudes of 40°E to 70°E. In the southern hemisphere the circulation is similar to June.

Figure 42 shows the seasonal migration of the circulation at the 150 mb level in June, July and August. The TEJ which was located near 6°N in June, is located near 10°N in both months of July and August. The isotach of 30 m/sec shown by a thin ellipse shows that the TEJ is best developed in July and August. The broken line located near 25°N in June is the location of the South Asian high. The eastern extension of this high has the largest seasonal shift and reaches 33°N over Kyushu in August. The contour of the 14,350 geopotential meter for August is shown by a thin ellipse which is centered near Iran. However, inspections of Figure 41 shows that this high extends from Sahara to southern Japan.

Hence the heating of the Tibetan plateau is insufficient to maintain this high pressure system. The massive release of latent heat by the entire Asian-African monsoon system is the main source of the heat required to warm the deep layer of the troposphere and maintains this high.

Returning to Figure 42, the subtropical westerly jet in the northern hemisphere also shows a marked northward shift with the advance of summer. The orographic effect of the Himalaya mountains and the effects of cold sea temperature near Japan in the month of June is reflected in the southward

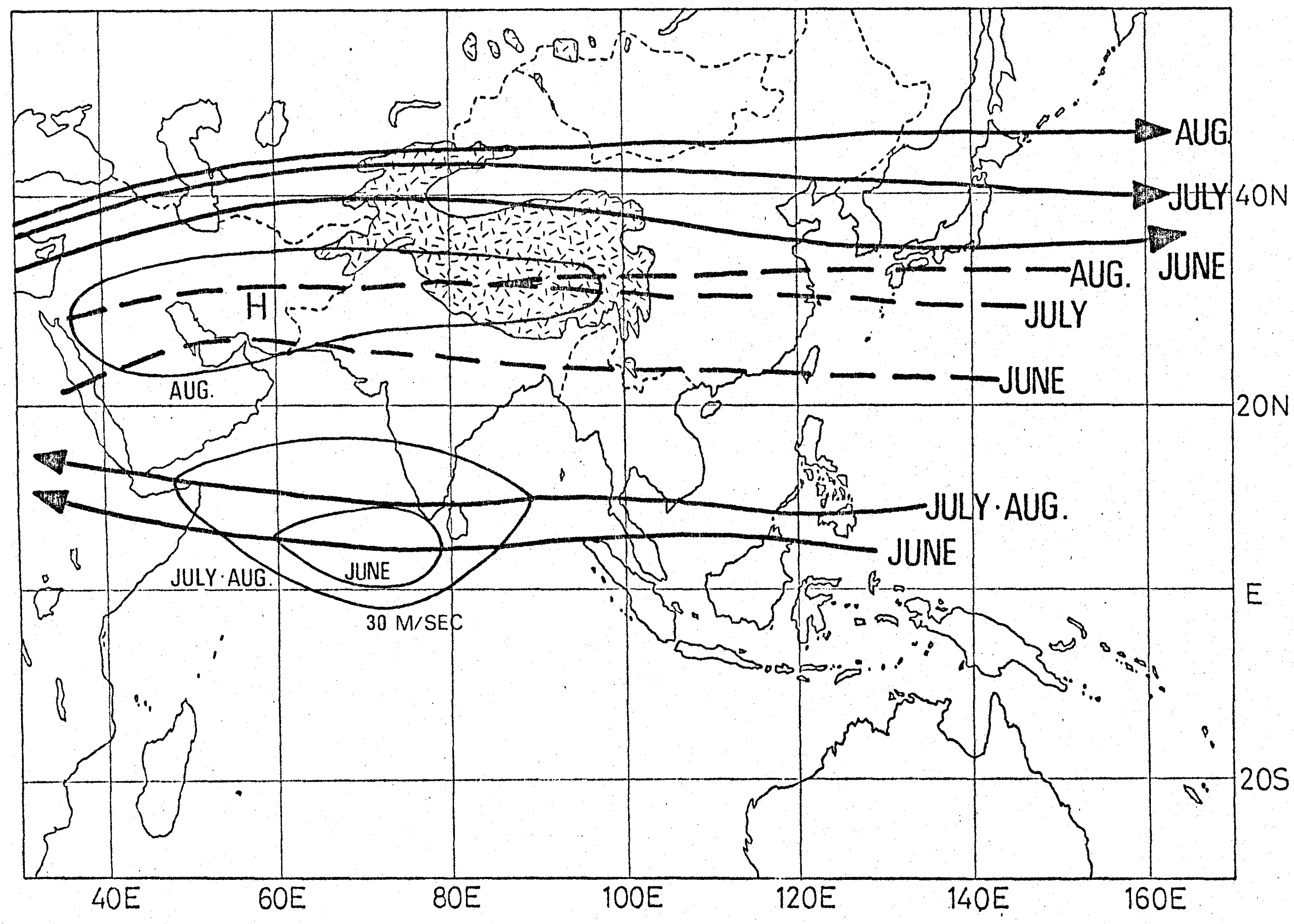

Fig. 42 Seasonal migration of the circulation at the 150 mb level in summer

meander of the jet stream near Japan. However by August, the warming of the sea water and the northward migration of the TEJ near 80°E both act to reduce the effects of Himalaya and the southward meander of the jet stream near Japan. It is a well known fact that this northward migration of the subtropical jet near Japan coincides with the end of Baiu season in the middle of July.

5-2 Regional characteristics of the monsoon regime

In this section the seasonal change to the summer monsoon regime is discussed. The sources of the data and the period of analysis are same as that of section 2-2. Figure 43 shows the northward migration of the ITCZ defined by the minimum in the steadiness factor and the shift of the prevailing winds to the southwest quadrant associated with equatorial westerlies. In the case of the winter monsoon, it took 4 months from the month of August before the northern ITCZ retreated southward to its winter position (Figure 10).

In the case of the summer monsoon, a rapid northward advance is observed. In April, the southern ITCZ to the south of the equator dissipates over the Indonesian region. On the other hand, the northern ITCZ is still located near its winter position near the equator. Two months later by the end of June, the ITCZ is located near the summer locations extending from northern India to southern China. The dashed line indicates the regions where the positions of the ITCZ are unstable or areas in the Baiu front region where the ITCZ slowly changes its characteristics to the polar front. The major break in the ITCZ near the longitude of 130°E is due to the land-sea temperature difference.

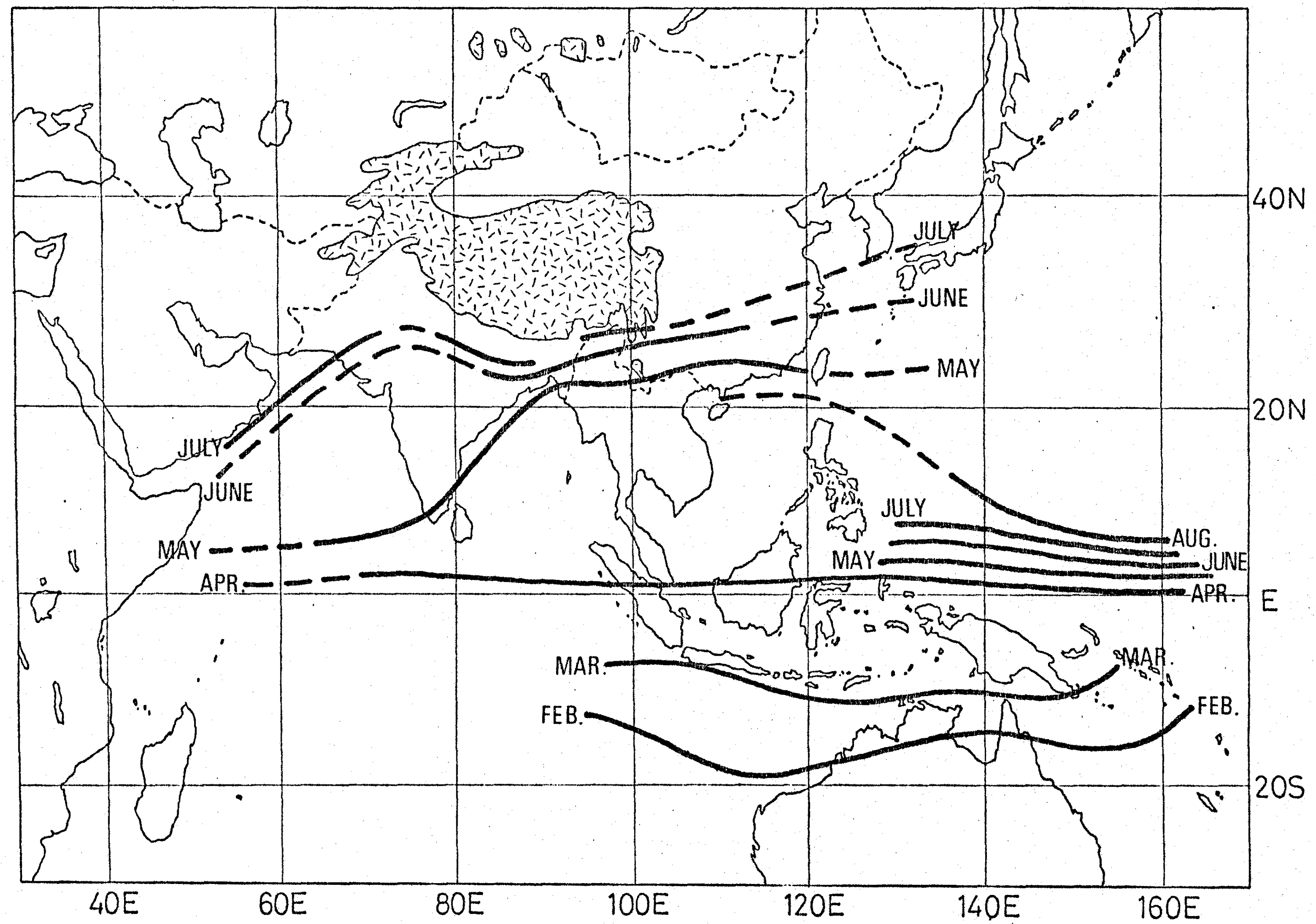

Fig. 43 Northward migration of the ITCZ at the 850 mb level from February to August

Because of the rapid northward advance of the ITCZ, the rainfall maximum does not follow the progress of the ITCZ. However the equatorial westerlies to the south of the ITCZ which are called the southwest monsoon are extremely moist and a favorable environment for the initiation of rainfall. Thus the approach and the subsequent passage of the ITCZ coincides with the beginning of a summer rainy season in the tropical monsoon regions.

Figure 44 shows a time height cross-section of the tropospheric temperature at Lucknow, India. The vertical decrease in temperature was removed by expressing the temperature at a given level as a departure from the annual mean temperature. At the surface level, the highest temperature coincides with the onset of monsoon rainfall.

Once the rainy season starts, the warm troposphere between the 500 mb level (5.8 km) and 200 mb level (12.5 km) is maintained by latent heat. This latent heating of the atmosphere maintains the South Asian high by expansion of the atmosphere. The heating of the ground by the solar radiation is insufficient to initiate a deep monsoon circulation because the troposphere remains cold (eg. in May).

In the upper troposphere above the 200 mb level, the evaporation of the cirrus cloud and the adiabatic cooling of the ascending air cools the atmosphere and maintains the cold tropopause. Thus like the case of the winter monsoon, the latent heating near the ITCZ plays an important role in the maintenance of the monsoon circulation.

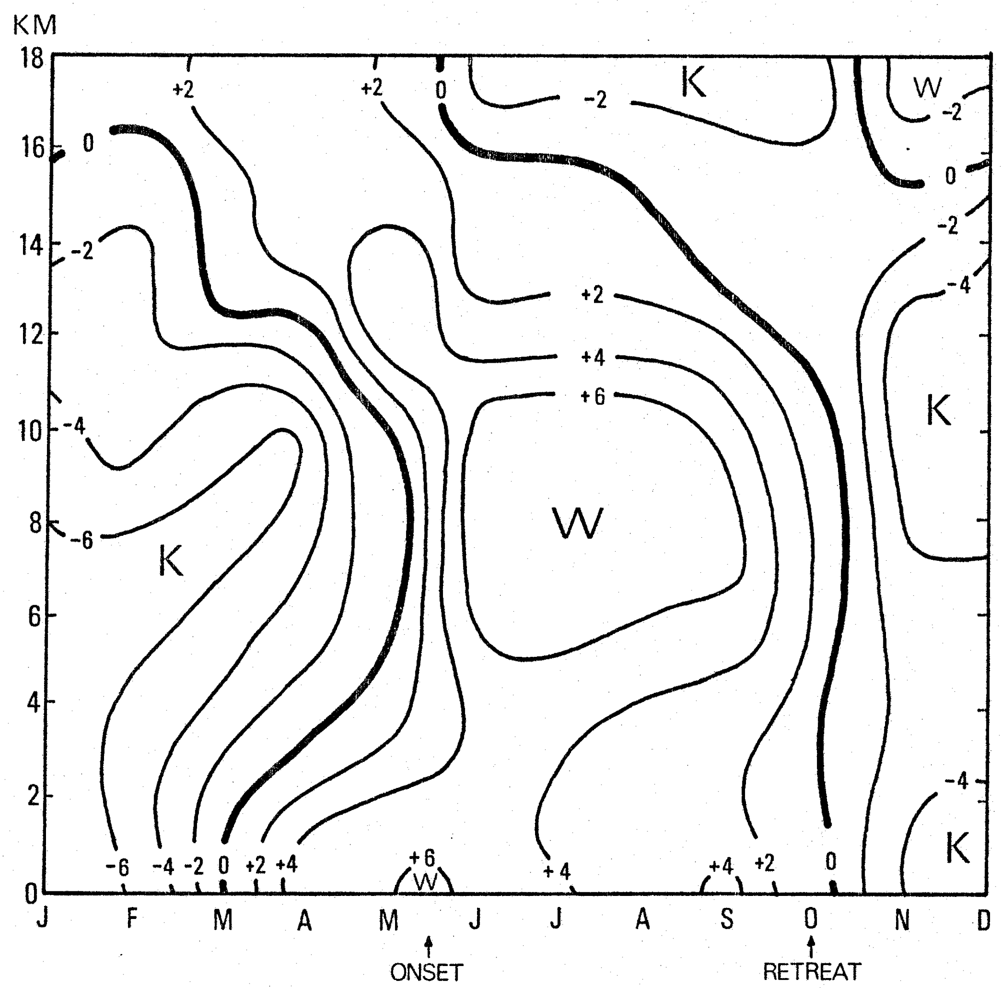

Fig. 44 Vertical distribution of temperature at Lucknow, India (unit: °C)

Figure 45 shows the northward migration of the beginning of the wet season defined by the 100 mm precipitation. In the regions to the west of 100°E or south of 20°N in south east Asia, the presence of the ITCZ or the equatorial westerlies is the necessary condition for the precipitation rate of at least 100 mm/month. In the southern parts of China and Japan, the heavy precipitation in April is associated with the activities of polar fronts. In the subsequent months the frontal zone migrates northward. In southern China and Japan, the Baiu front (which has subtropical characteristics) is associated with heavy rain in early summer.

Figure 46 shows the numbers of the months with more than 100 mm precipitation. The regions near the equator where the NITCZ or SITCZ exists throughout the year have heavy precipitation in all months of the year. The regions to the south of Japan also have heavy precipitation associated with the polar front, Baiu front and the typhoons. On the other hand, inner Asia and the subtropical desert are dry throught the years.

In most of the regions where the monsoon wind prevails, three to six wet months are observed. The regions of India and Indochina are prime examples of such regions. It is important to note that the regions with the highest seasonal concentration of precipitation lie near the fringe of the monsoon wind system. These regions generally coincide with the summer locations of the ITCZ.

Finally a brief survey of the precipitation and wind at the 850 mb level for the entire year was conducted. Figure 47 shows the synoptic situations associated with the wettest months over

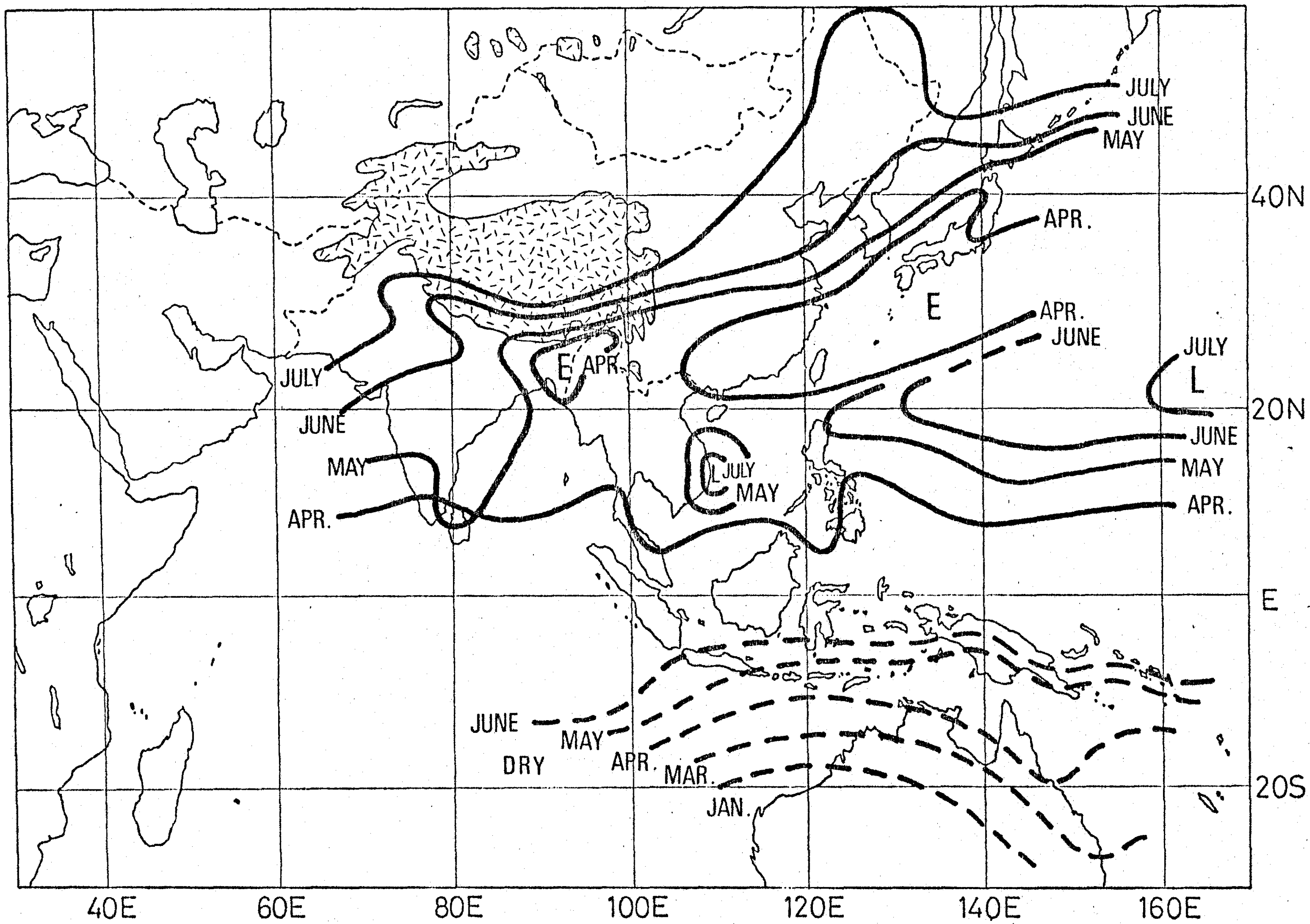

Fig. 45 Northward migration of beginning of wet season defined by 100 mm precipitation from January to July

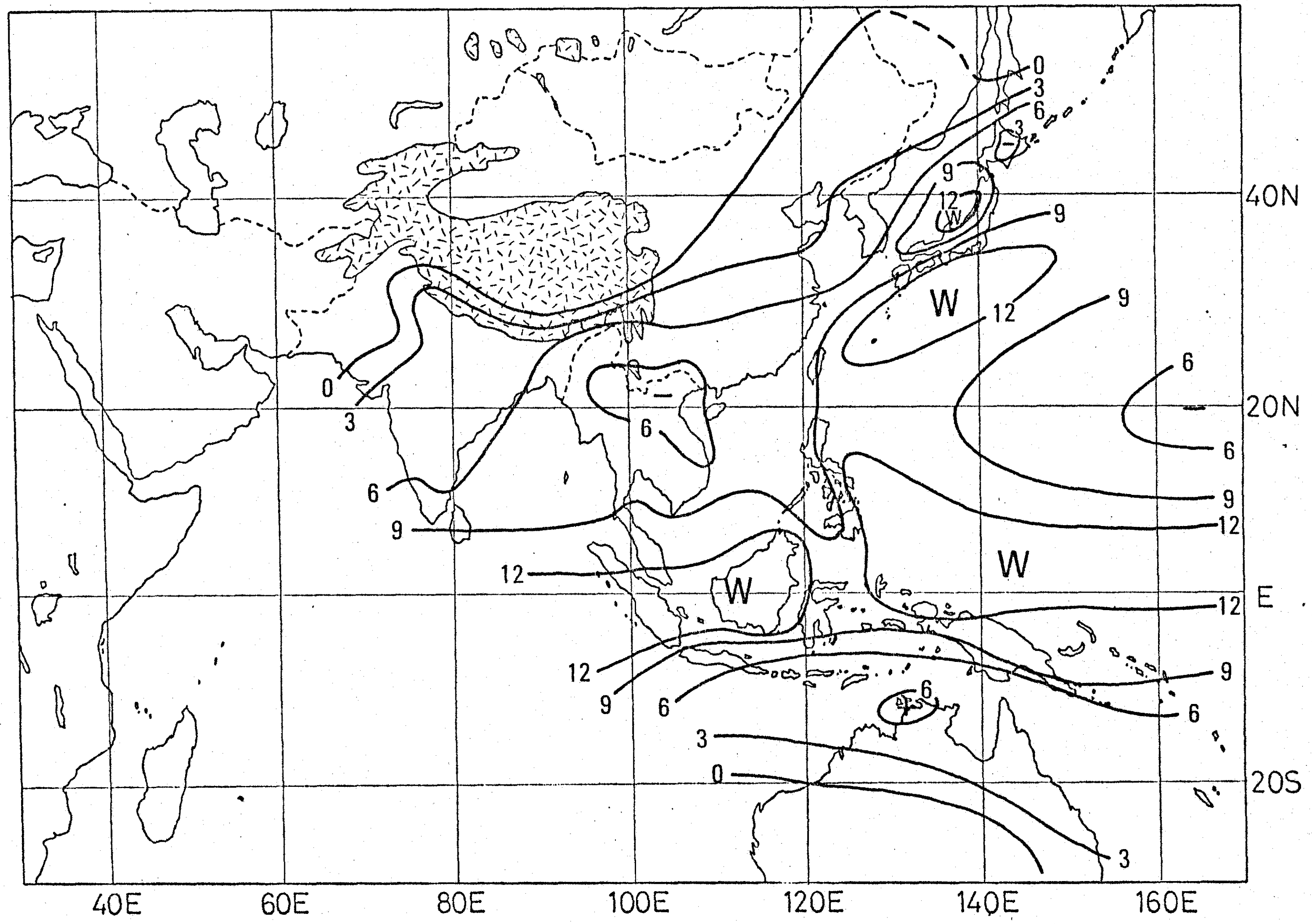

Fig. 46 Number of the months with more than 100 mm precipitation

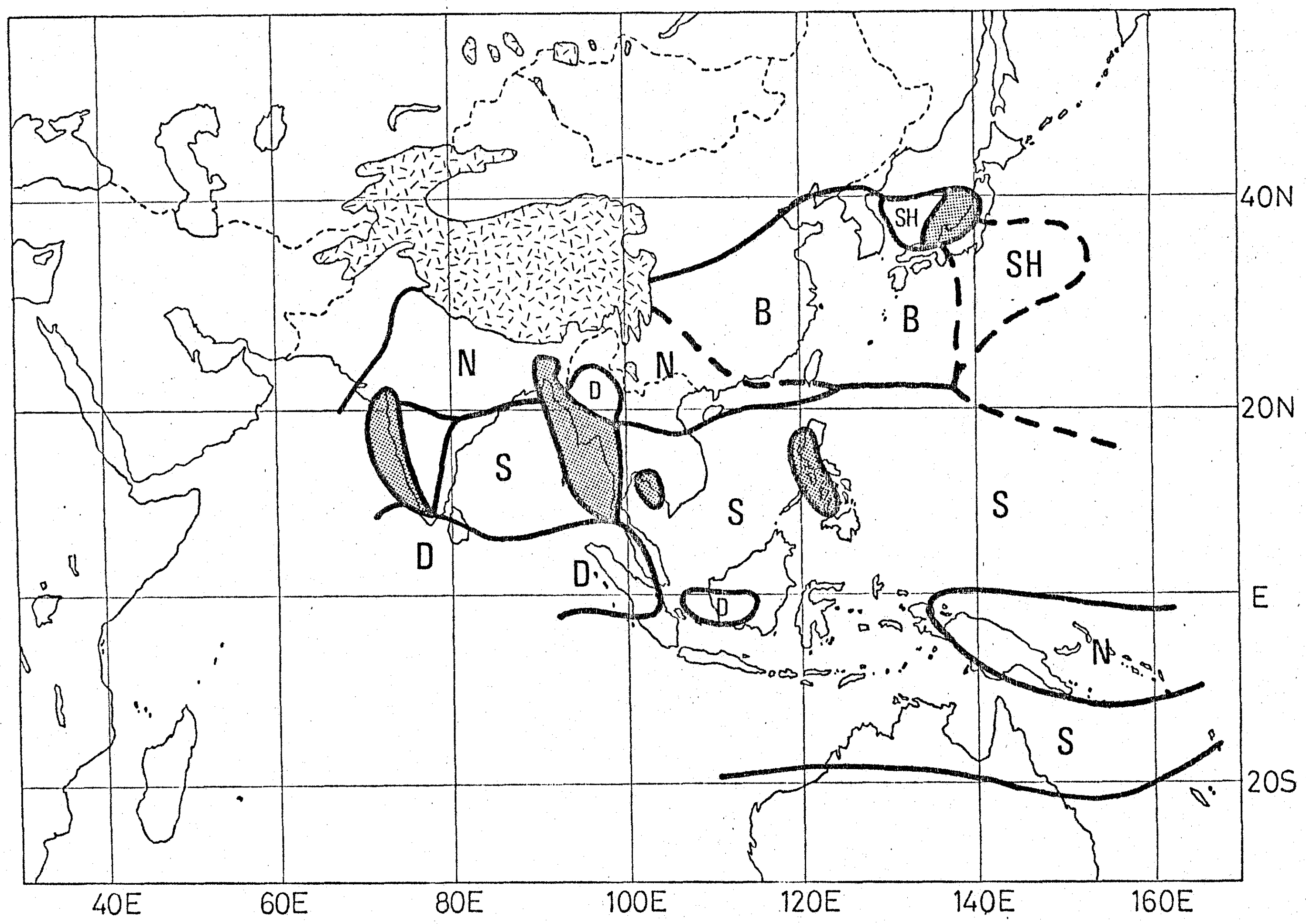

Fig. 47 Synoptic situation associated with wettest months (> 200 mm)

S = wettest month following the southward migration of the ITCZ.
N = wettest month followinf the northward migration of the ITCZ.
D = regions with two maximum of precipitation with passage of the ITCZ.
B = wettest month following the northward migration of the Baiu front.
SH= wettest month during Shurin-typhoon season.

200 mm/month. The regions marked by the letter "S" have their wettest month immediately following the southward migration of the ITCZ in autumn and early winter months from September to January. The regions marked by the letter "N" have their wettest month following the northward migration of the ITCZ from February to August. The regions which experience two maximum of precipitation associated with the passage of the ITCZ are marked by the letter "D". The regions with the orographic precipitation during strong monsoon wind are shaded. Finally in the subtropical regions,the regions marked by the letter "B" have their wettest month following the northward migration of the Baiu front. On the other hand, the regions with maximum precipitation during the Shurin-typhoon season are marked by the letter "SH".

In most of the tropical monsoon regions, the presence of the slowly moving ITCZ coincides with the wettest month. In the winter monsoon months the high level subsidence above the northeast monsoon effectively rules out the orographic maximum of precipitation which coincides with the months of strongest northeast monsoon. The importance of the convective stability of the atmosphere is the determining factor governing the maintenance of heavy orographic precipitation. In all the regions with the orographic precipitation (defined as the near coincidence of the maximam monsoon wind and the wettest month), the atmosphere is convectively unstable.

Figure 48 shows the monsoon regions at the 850 mb level defined by a seasonal shift of at least 120° in wind direction and a steadiness factor of at least 60 % for both summer and

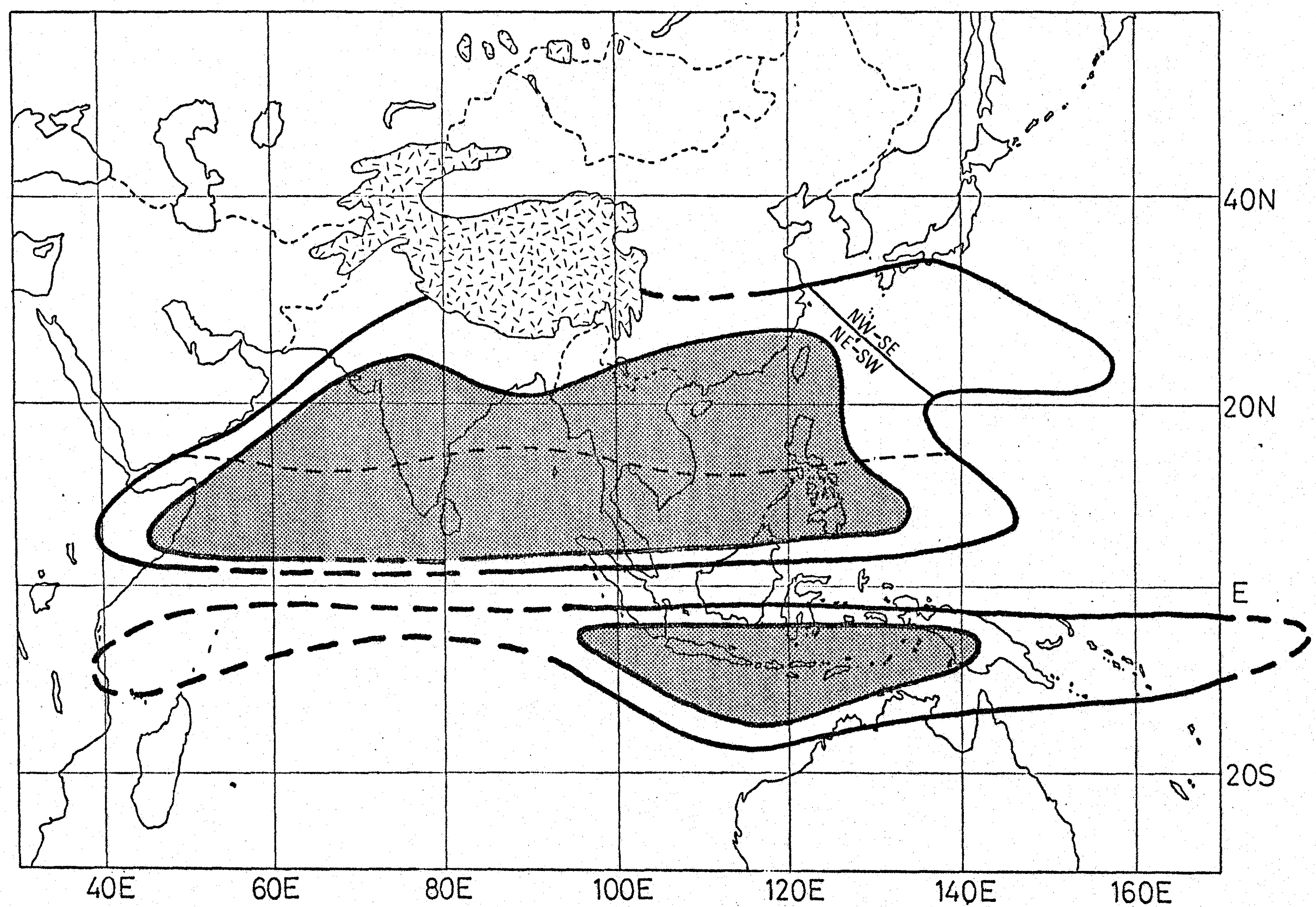

Fig. 48 Monsoon region at the 850 mb level, thin broken line indicates the boundary between the two types of the tropical monsoon.

and winter monsoon. The shaded regions are the core regions which satisfy both of the above definitions. The unshaded regions are the regions which satisfy the wind shift definition but with a low steadiness factor. These regions can be considered as regions with monsoon tendencies. The thin line near Okinawa separates two different types of the monsoon regimes: The tropical northeast-southwest monsoon and the subtropical northwest-southeast monsoon regime.

The thin dashed line near 10°N separates two different character of the tropical monsoon regime. North of this line, the northeast monsoon is primarily an autumn phenomenon. During the middle of the winter, the subtropical high dominates this region and consequently, the winds at the 850 mb level are variable. Thus this region has a subtropical character. South of 10°N the behavior of the tropical monsoon is similar to a textbook example with northeast monsoon during the middle of winter months. The regions extremely near the equator lie between two ITCZs throughout the year. They are dominated by equatorial westerlies, hence no monsoon is observed in this narrow belt.

Finally the regions of the Korean peninsula and most of Japan (traditionally considered monsoonal based on the surface wind) do not experience the pronounced reversal of the wind at the 850 mb level. In this region, the winds at the 850 mb level are from northwest in winter and from southwest in summer. The so-called southeast monsoon beyond the subtropical margin of Kyushu and Shikoku is a shallow phenomenon and not a major part of the general circulation of the atmosphere.

CHAPTER 6

ROLE OF THE CIRCULATION AT THE 150 MB LEVEL DURING THE SUMMER MONSOON

6-1 Interannual fluctuations of the tropical easterly jet stream

Unlike the case of the winter monsoon, the TEJ extends over a large area of southern Asia. Since the axis of the TEJ lies near 10°N in July and August, the wind speed at the 150 mb level at 10°N between the longitude from 40°E to 110°E were averaged to obtain the monthly wind speed of the TEJ.

Figure 49 shows the wind speed at the 150 mb level at 10°N in June in a 14 year period from 1964 to 1977. The wind speed reaches its peak near 75°E at the southern end of the Indian subcontinent. The wind was strong in June of 1964,1969,1970 and 1971 with peak speeds of more than 32 m/sec. On the other hand, in 1968,1972,1973 and 1977 the TEJ was notably weak.

Figures 50 and 51 show similar cross sections at 10°N for July and August. The longitude of the peak wind speed is similar to that in June, but the wind speed reaches 35 to 40 m/sec. These wind speed data covering 80° of longitudes were then averaged to obtain the monthly wind speed of the entire TEJ over southern Asia (average from 40°E to 110°E). Figure 52 and Table 4 show the interannual fluctuations of the TEJ obtained in this manner.

An inspection of Figure 52 shows that in June, the fluctuations of the TEJ are not in phase with those of July and August. The year 1972 is an outstanding example of the consistantly weak TEJ (Kanamitsu and Krishnamurti, 1978).

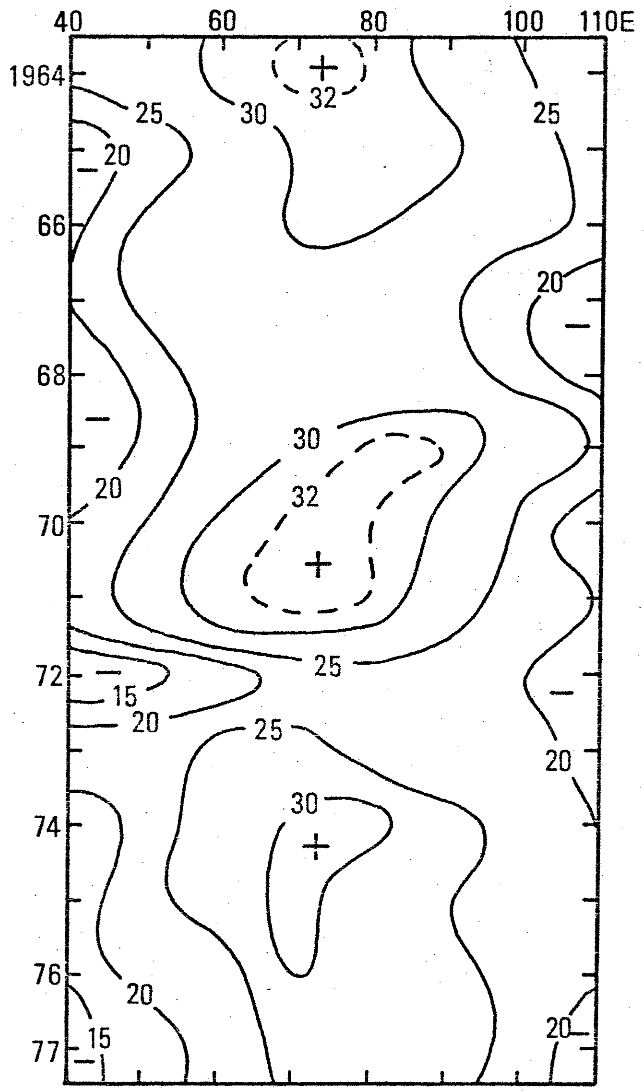

Fig. 49 Wind speed at the 150 mb level at 10°N in June (unit: m/sec)

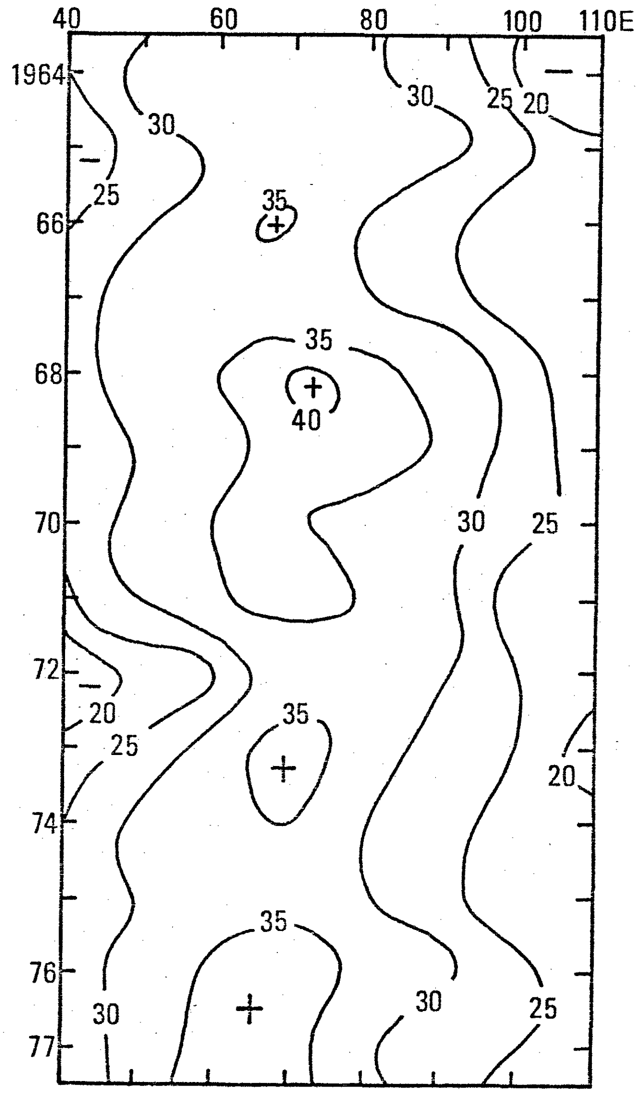

Fig. 50 Wind speed at the 150 mb level at 10°N in July (unit: m/sec)

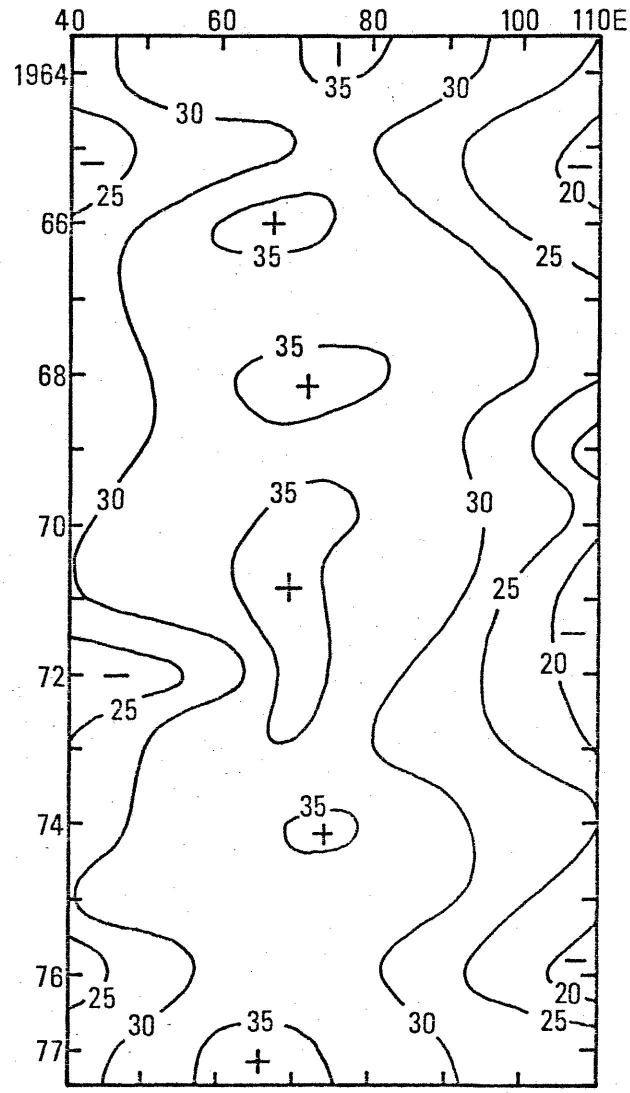

Fig. 51 Wind speed at the 150 mb level at 10°N in August (unit: m/sec)

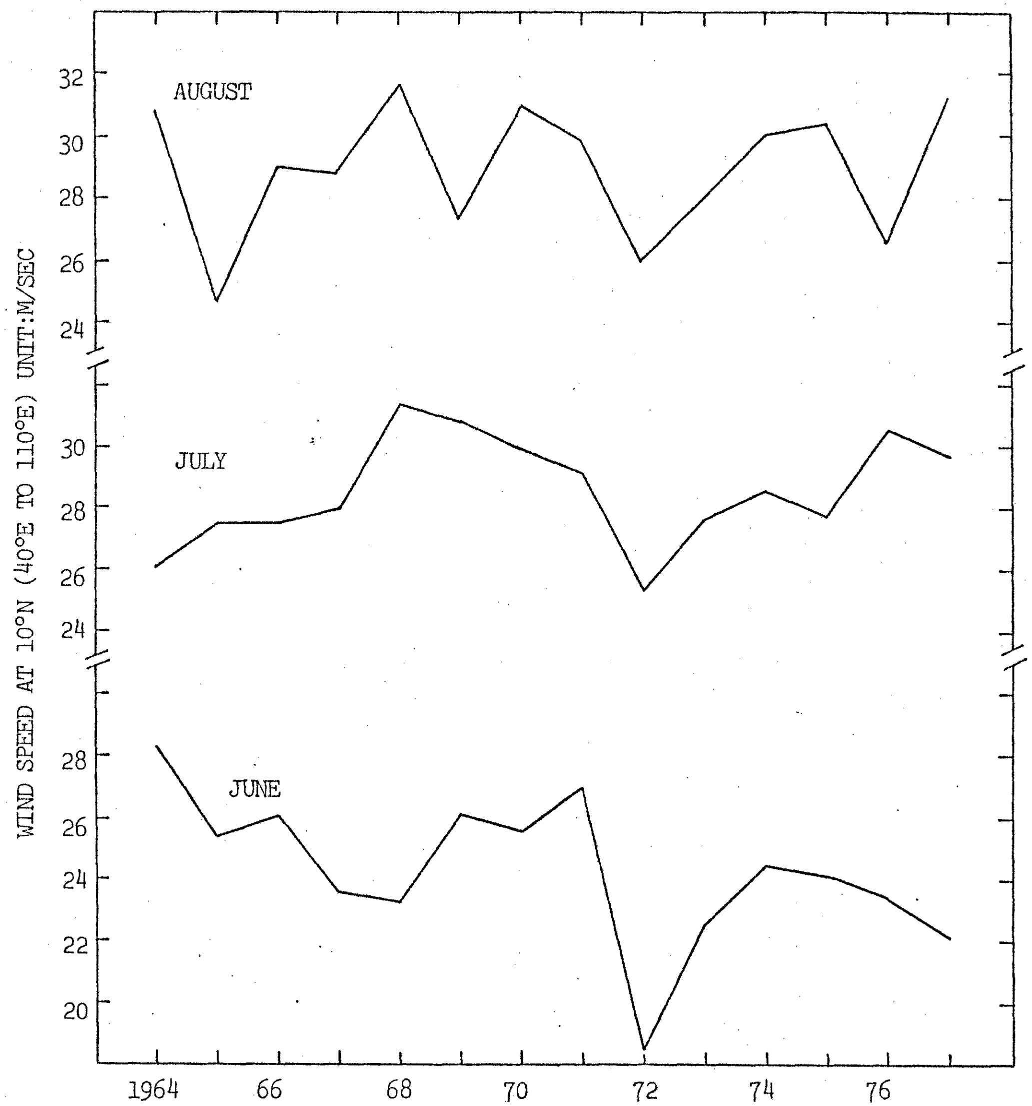

Fig. 52 Wind speed at the 150 mb level at 10°N in summer
(average of the longitudes from 40°E to 110°E)

TABLE 4

Wind speed at the 150 mb level at 10°N in summer (average from 40°E to 110°E unit: m/sec)

Year	Month			3 month
	June	July	August	average
1964	28.3	26.1	30.8	28.4
1965	25.4	27.5	24.6	25.8
1966	26.1*	27.5	29.0	27.5
1967	23.6	28.0	28.8	26.8
1968	23.3	31.4	31.6	28.8
1969	26.1	30.8	27.4	28.1
1970	25.6	29.9	31.0	28.8
1971	27.0	29.1	29.8	28.6
1972	18.5	25.4	26.0	23.3
1973	22.5	27.6	28.0	26.0
1974	24.5	28.5	30.0	27.7
1975	24.1	27.5	30.4	27.3
1976	23.4	30.5	26.6	26.8
1977	22.1	29.6	31.3	27.7
Average	24.3	28.5	29.0	27.3
S.T.D.	2.3	1.7	2.1	1.4

* estimate (poor data)

On the other hand, in 1970 the TEJ is strong in all three months of the summer monsoon. In most of the other years, the strength of the TEJ varies during the summer monsoon months.

Sets of composite maps of precipitation and other parameters were made in each of the three summer months of June, July and August based on the data shown in Table 4. In all of these months four years with strong TEJ were compared with four years with weak TEJ. The only exception is June of 1966 which was omitted because the rawinsonde data was unreliable. The standard deviations of the wind speed of the TEJ shows that the inter-annual fluctuation is much smaller than the case of the winter monsoon (Table 2).

6-2 Monsoon circulation and the precipitation in June

The relationship between the fluctuations of the TEJ at the 150 mb level and the monsoon circulation and precipitation was obtained by the same procedure as the winter monsoon. Since topographical influences are larger than the interannual fluctuations of the precipitation in many localities, only the maps of the difference in the two composite maps of the years with strong TEJ minus weak TEJ will be shown.

Figure 53 shows the difference in the precipitation in June. The dark shaded regions show the regions with heavy precipitation when the TEJ is strong. In these years with strong TEJ, the onset of monsoon rain is early in India while orographic precipitation in Burma and southern Thailand is also heavy. The ITCZ is active near Micronesia. In the higher northern latitudes, the Baiu rain is heavy in southern Japan. If the Asian monsoon region

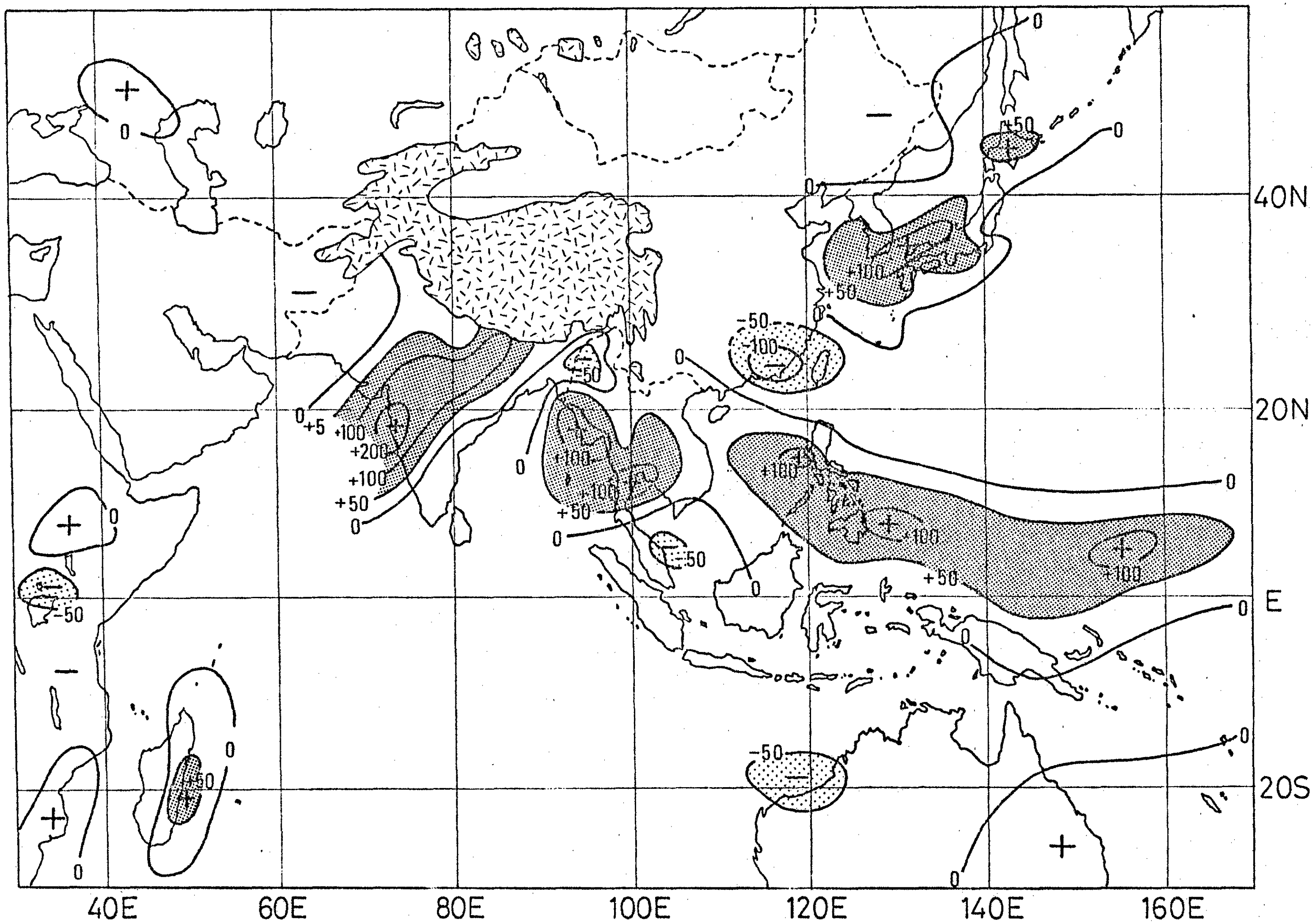

Fig. 53 Difference in the precipitation in June (unit: mm)
(strong TEJ minus weak TEJ)

is considered for analysis, the release of latent heat increases when the TEJ is strong.

Figure 54 shows the difference in the geopotential height at the 850 mb level in June. The positive sign shows an increase in the geopotential when the TEJ is strong. Figure 54 clearly indicates that the monsoon circulation is intense in most of southern Asia when the TEJ is strong. Thus the Hadley cell nature of the monsoon circulation has an excellent coupling between the converging branch at the 850 mb level and the diverging branch at the 150 mb level. In the higher northern latitudes, there is a blocking high to the north of Caspian Sea and the Okhotsk high develops when the TEJ is weak. The heavy Baiu rain in southern Japan is a result of increased transport of moisture by the strong western extension of the Pacific high.

Figure 55 shows the difference in the geopotential height at the 150 mb level in June. Since the strength of the TEJ was used to obtain the change in circulations, the South Asian high is strong near the Tibetan plateau. However there is another important feature which influences the intensity of the TEJ. This is the broad trough near the equator. In some years, the deep trough near the equator is the primary factor maintaining the intense TEJ. For this reason, the fluctuations of the geopotential height near Himalaya do not show a high correlation to the fluctuations of the TEJ.

The two maps of the geopotential height at the 850 mb level and 150 mb level can be used to derive a map of change in the tropospheric thickness (Figure 56). Since atmospheric conditions are hydrostatic, the change in thickness of the atmosphere

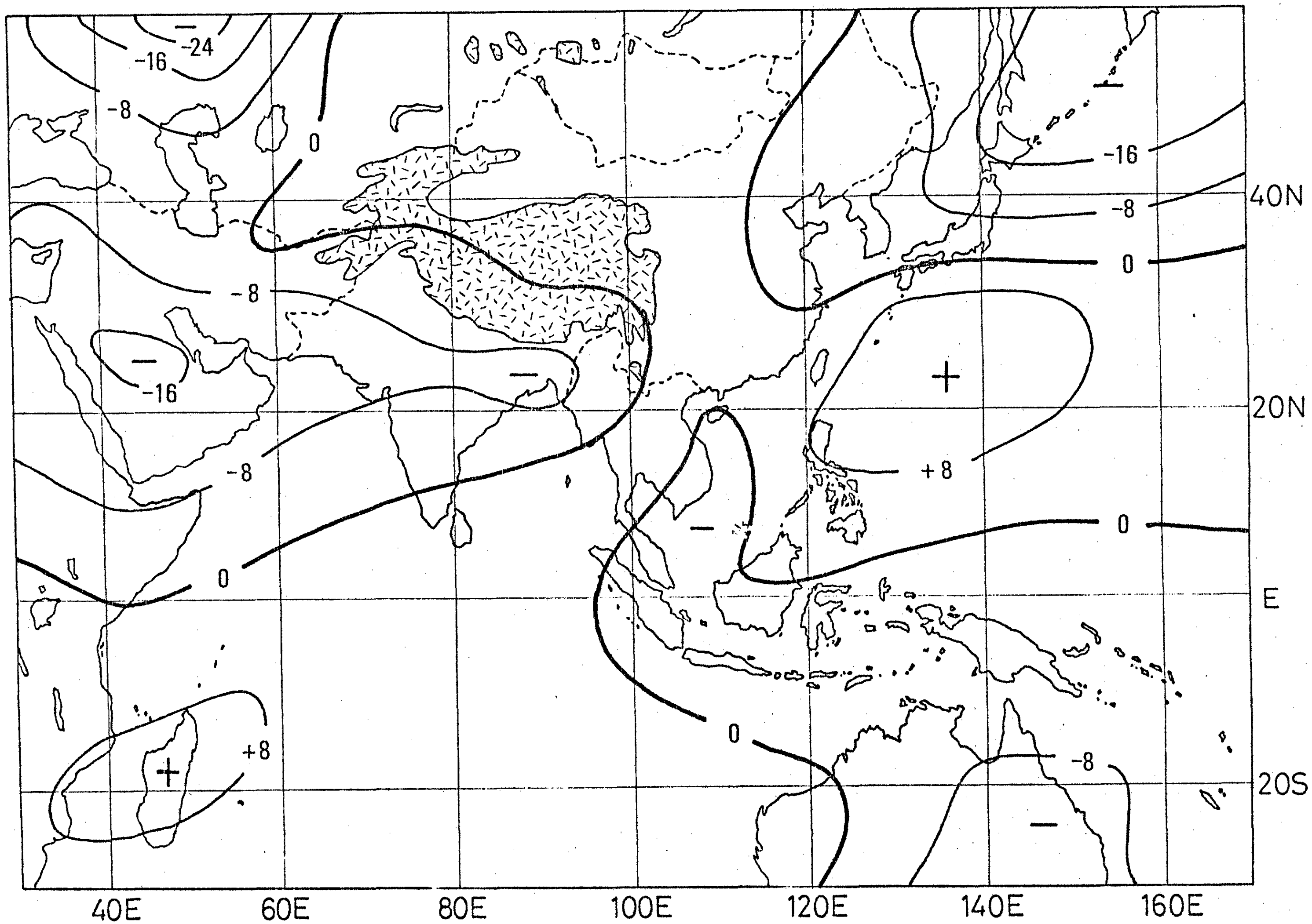

Fig. 54 Difference in the geopotential height at the 850 mb level in June

(unit: geopotential meters) (strong TEJ minus weak TEJ)

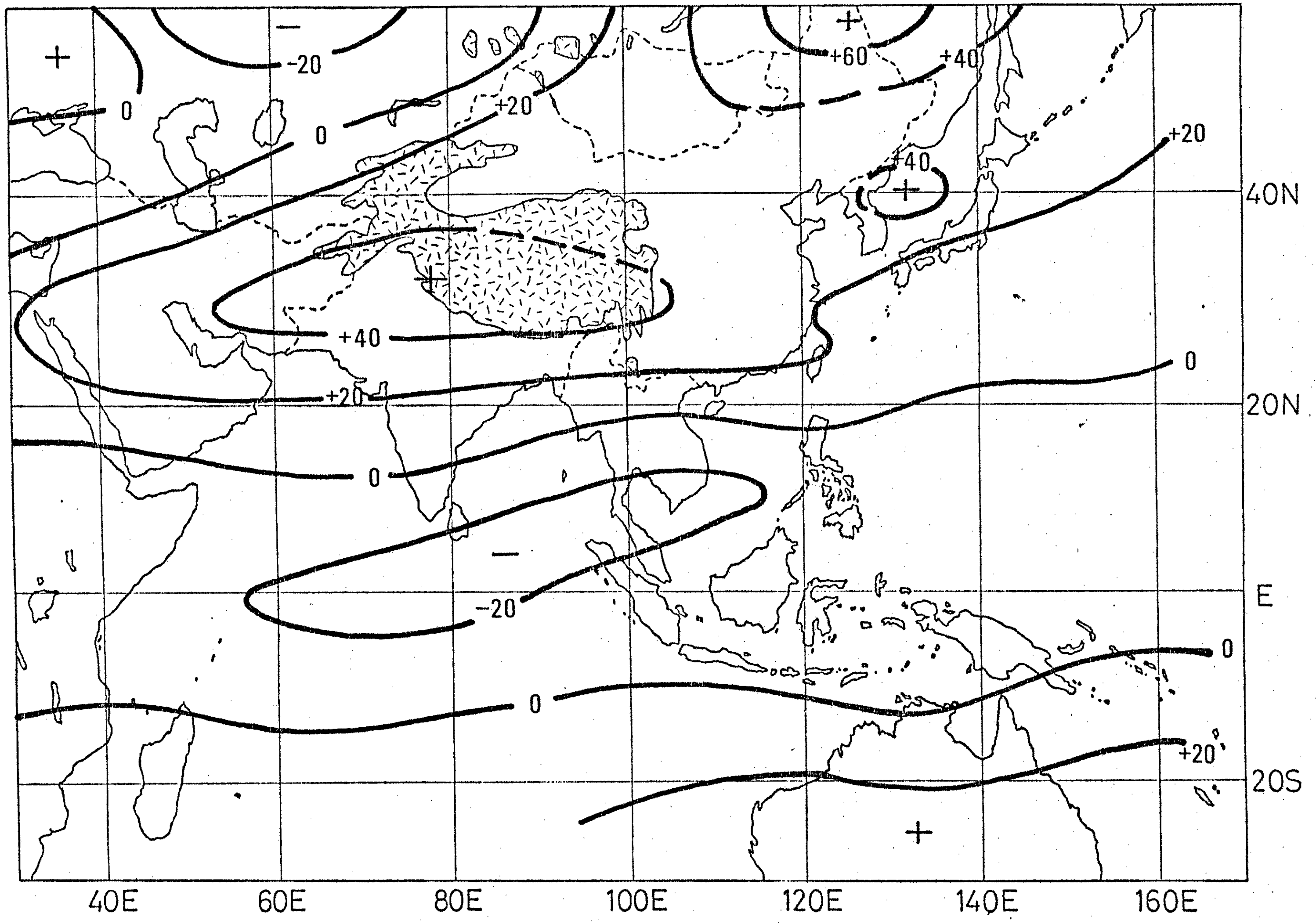

Fig. 55 Difference in the geopotential height at the 150 mb level in June (unit: geopotential meters) (strong TEJ minus weak TEJ)

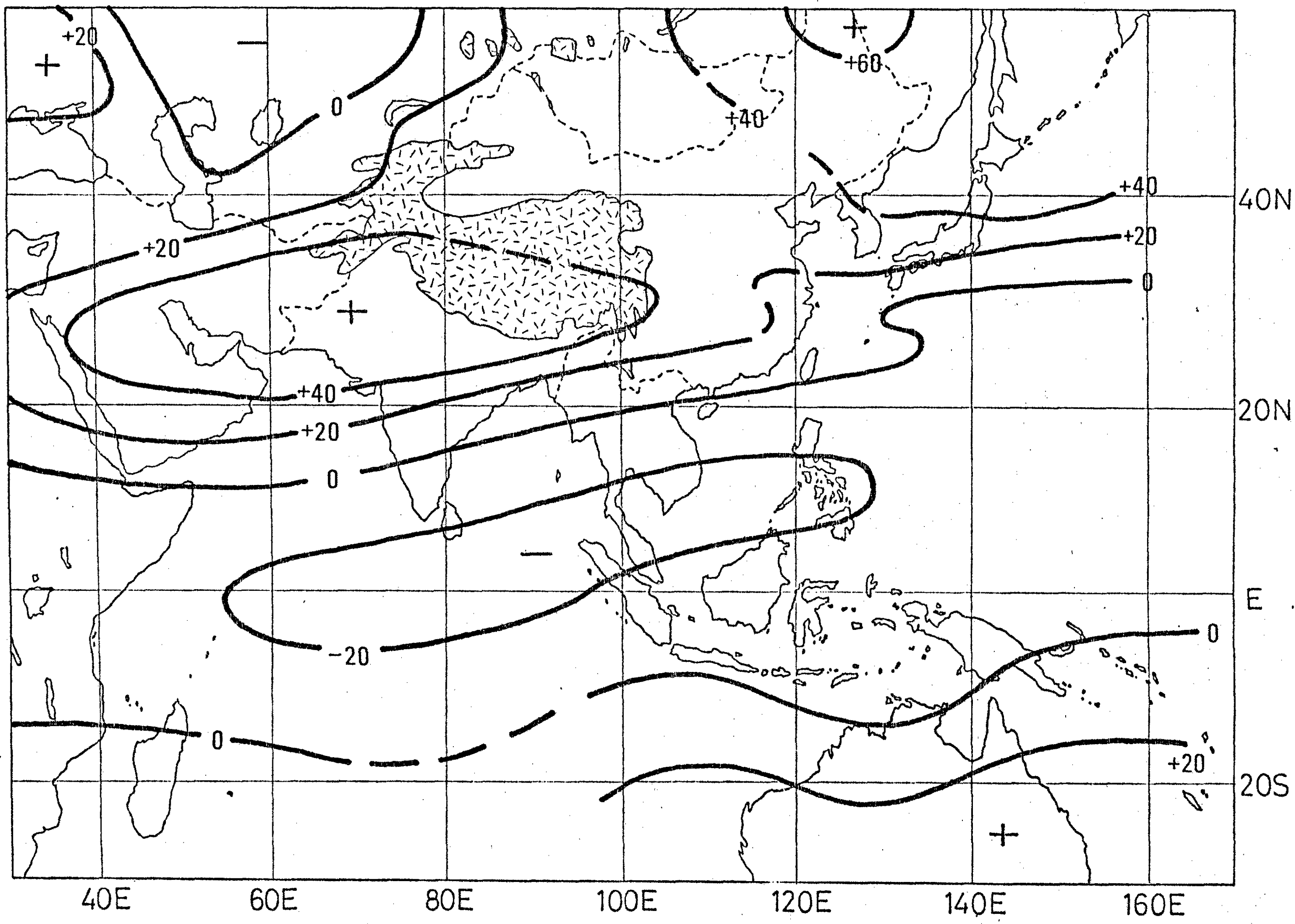

Fig. 56 Difference in the thickness between the 850 mb level and 150 mb level in June (unit: geopotential meters) (strong TEJ minus weak TEJ)

is equivalent to the change in temperature between two layers of the atmosphere. The thickness map indicates that when the TEJ is strong in June, the Asian continent between 20°N and 35°N is warm, while the equatorial atmosphere is cold. As shown in Figure 44, much of the heating between the 850 mb level and 150 mb level is due to increased release of latent heat associated with the monsoon rainfall. The slight warming of the subtropics in the southern hemisphere is associated with adiabatic warming caused by increased subsidence near the subtropical high. In June with a weak TEJ, the conditions are opposite to the discussion in this section.

6-3 Monsoon rainfall and the precipitation in July

In July the summer monsoon reaches its peak intensity in the Indian subcontinent. The intensity of the TEJ also reaches its maximum development in July and August. However the interannual fluctuations of the TEJ reaches its seasonal minimum in the month of July (Table 4). For this reason, July was found to be most difficult with regards to analyzing the fluctuations of the monsoon systems.

Figure 57 shows the difference in the precipitation in July. The dark shaded regions with heavy precipitation in the case of strong TEJ are rather patchy. In spite of this irregularity, some order can be found. On the west coast of India and Indochina regions, there are regions of heavy orographic precipitation when the TEJ is strong. The northward migration of the Baiu rain is delayed near Japan. The anomalous decrease in precipitation in the Philippines is a result of the flood-producing severe rainstorm associated with the typhoons in 1972. This

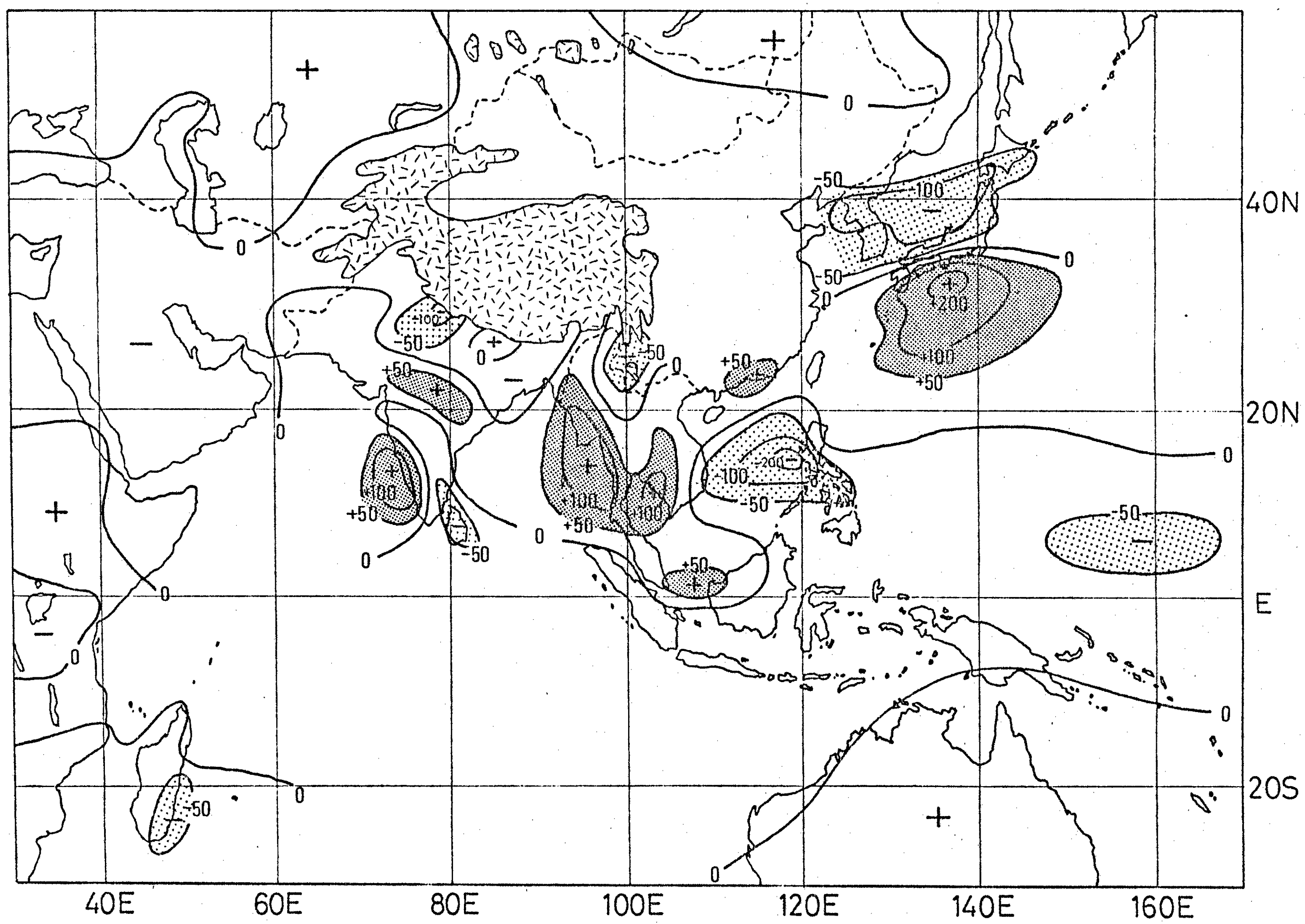

Fig. 57 Difference in the precipitation in July (unit: mm)

(strong TEJ minus weak TEJ)

storm has occured in a year with an unusually weak TEJ. In the same year, India and the Indochina regions experienced a severe drought. Consequently, the release of latent heat over the Asian continent was less than the normal years.

Figure 58 shows the difference in the geopotential height at the 850 mb level in July. When the TEJ is strong, the Indian low is intense and the southwest monsoon is strong near the west coasts of India and Indochina. The Baiu rain in southern Japan is associated with a localized area of cyclonic circulation. The subtropical high in the southern hemisphere is strong. In the higher northern latitudes, there is a ridge to the east of Lake Baikal when the TEJ is strong. The negative departure near Caspian Sea is a result of a blocking high which develops when the TEJ is weak.

Figure 59 shows the difference in the thickness between the 850 mb level and 150 mb level in July. The warm troposphere in southern Asia to the west of 100°E is maintained by the release of latent heat in India and Indochina(Figure 57). There is intense subsidence in the Arabian peninsula where the descending branch of the indirect cell which consumes the energy of the TEJ is located (Tanaka,1976). The equatorial atmosphere is cold when the TEJ is strong and this maintains the trough in the equatorial Indian Ocean.

The subtropical jet is strong to the west of Himalaya and is displaced south of its normal position near Japan. In the southern hemisphere, the strong subtropical high at the 850 mb level is supported by subsidence which heat the atmosphere. When the TEJ is weak, the reverse condition is true and the blocking high develops to the north of the Caspian Sea.

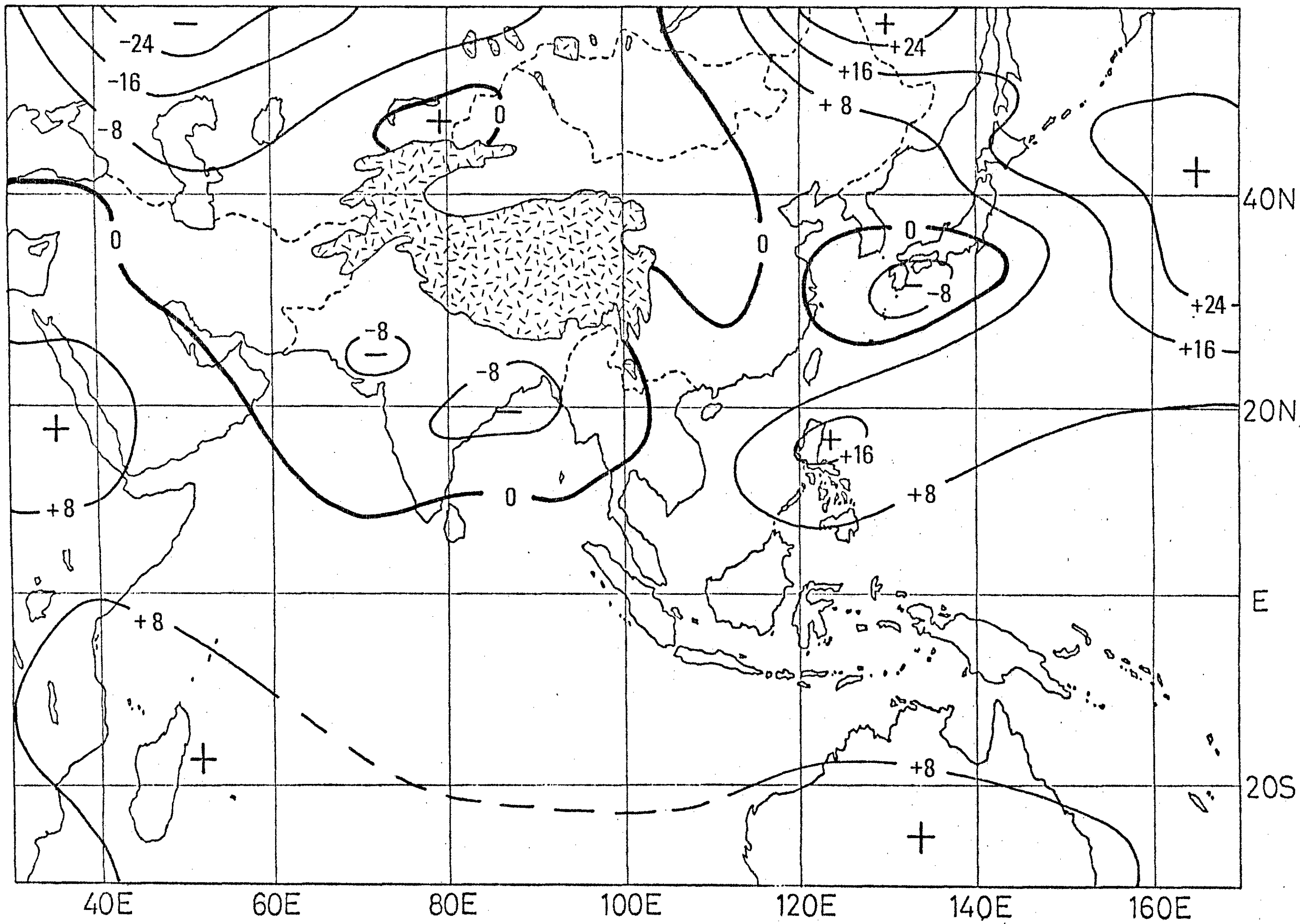

Fig. 58 Difference in the geopotential height at the 850 mb level in July (unit: geopotential meters) (strong TEJ minus weak TEJ)

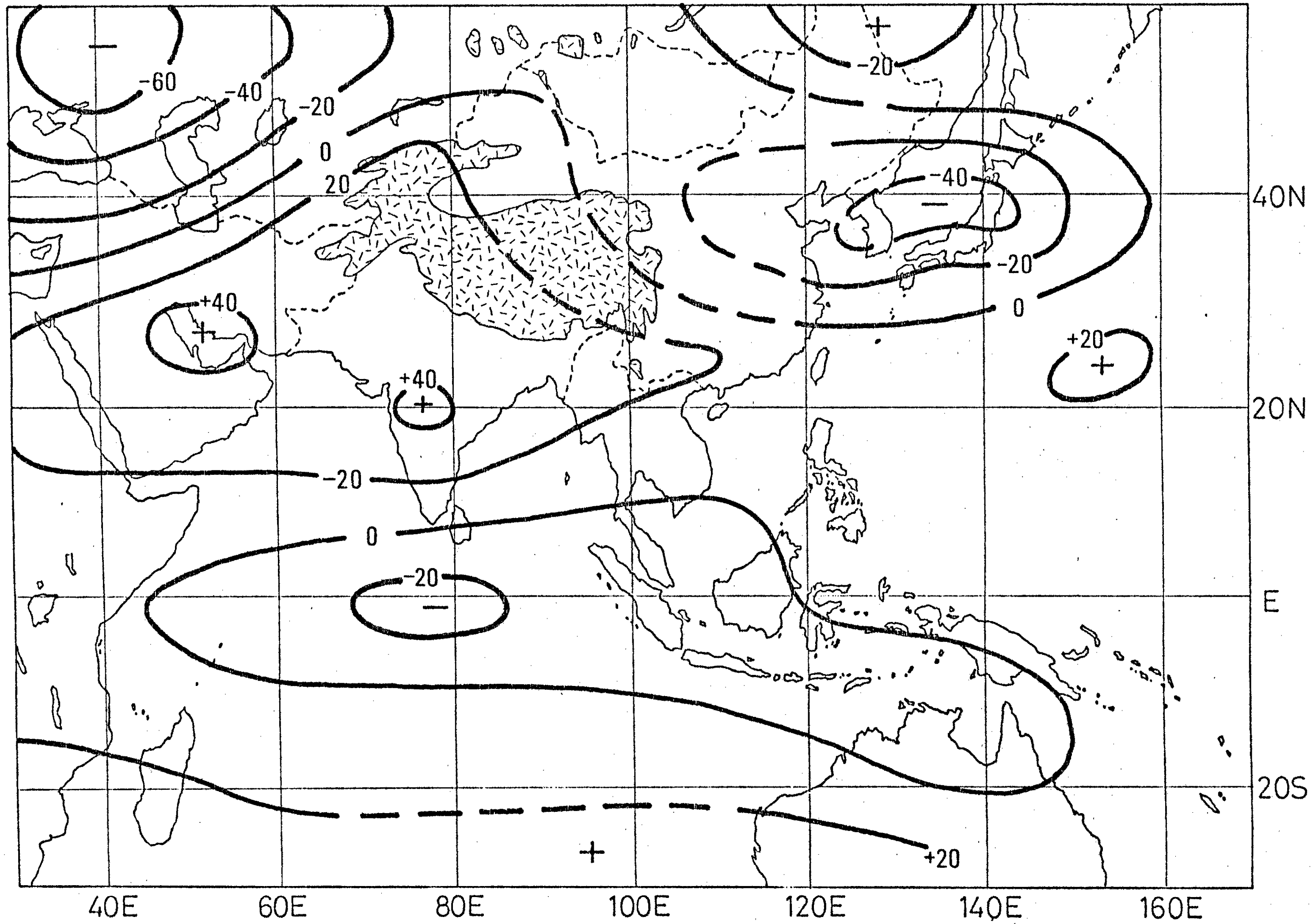

Fig. 59 Difference in the thickness between the 850 mb level and 150 mb level in July (unit: geopotential meters) (strong TEJ minus weak TEJ)

6-4 Monsoon circulation and the precipitation in August

In August, the summer monsoon reaches its peak intensity near the Philippines. The ITCZ penetrates deep in to Micronesia and the typhoon activities reach their seasonal maximum. The release of latent heat increases in the west Pacific and the adjacent regions. Hence the main source of latent heat moves east toward the Pacific Ocean.

Figure 60 shows the difference in the precipitation in August between strong TEJ and weak TEJ. In the most part of India and the west coast of Indochina, heavy precipitation is associated with a strong southwest monsoon. The heavy precipitation along the southern coast of Japan is a result of frequent typhoon activities. The decrease in the precipitation in the Korean peninsula is associated with a weak southwest monsoon when the typhoon is active near Okinawa.

In August, the change in the circulation regime at the 850 mb level was found to be extremely large to the east of 100°E. For this reason, two composite map of the circulation at this level will also be shown. Figure 61 shows the composite map of the geopotential height at the 850 mb level with strong TEJ in 1964,1968, 1970, and 1977. Compared to Figure 40 which illustrated the normal circulation, the ITCZ is located near the Okinawa region reflecting the strong southwest monsoon near the Philippines. Inspection of daily weather maps near Japan shows that on many days in the month, the typhoon stagnates for a long period near Okinawa.

The southeast monsoon in the Kyushu and Shikoku regions is well developed. Heavy precipitation on the southeast coast of these island is associated with converging currents of moist

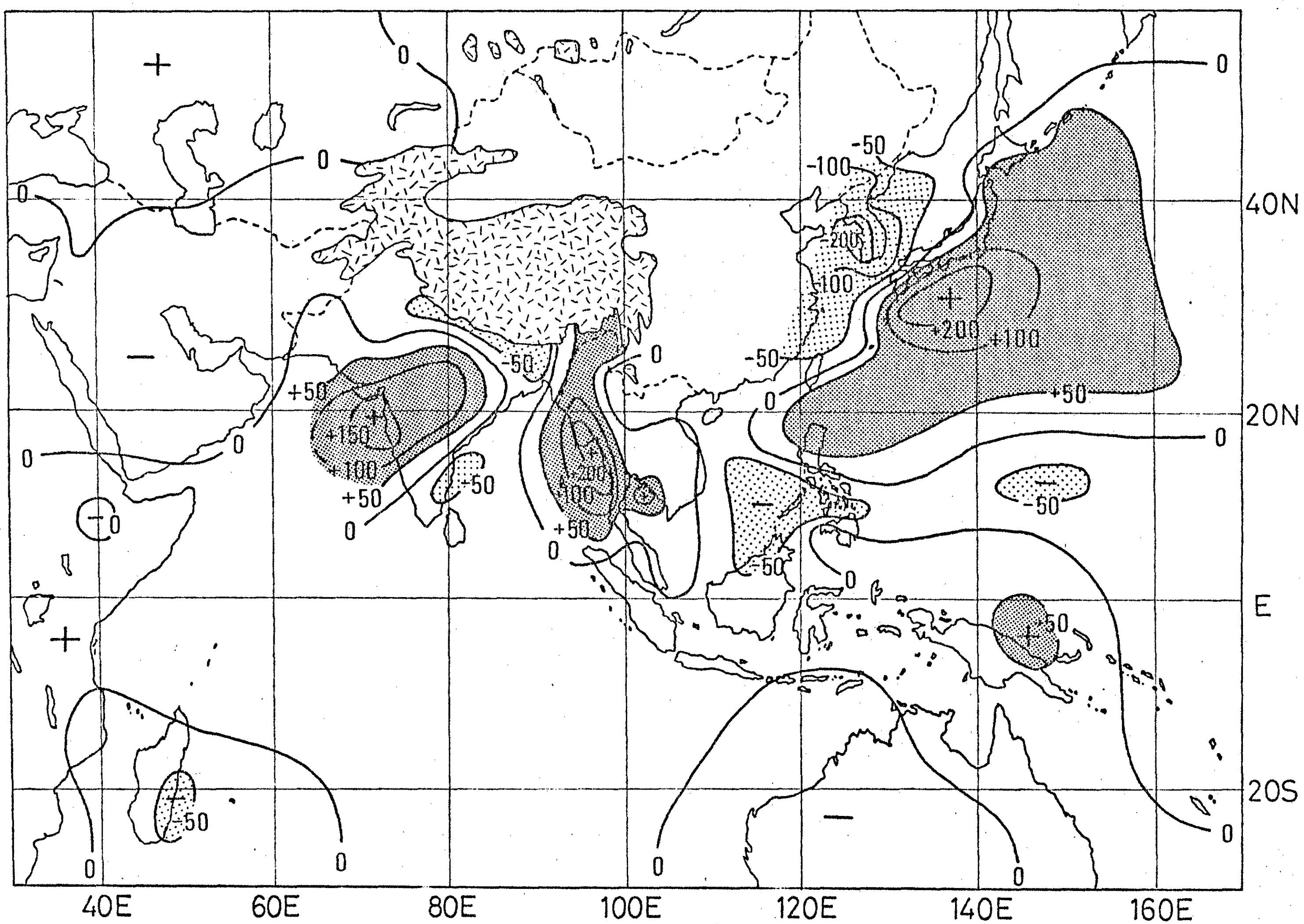

Fig. 60 Difference in the precipitation in August (unit: mm)
(strong TEJ minus weak TEJ)

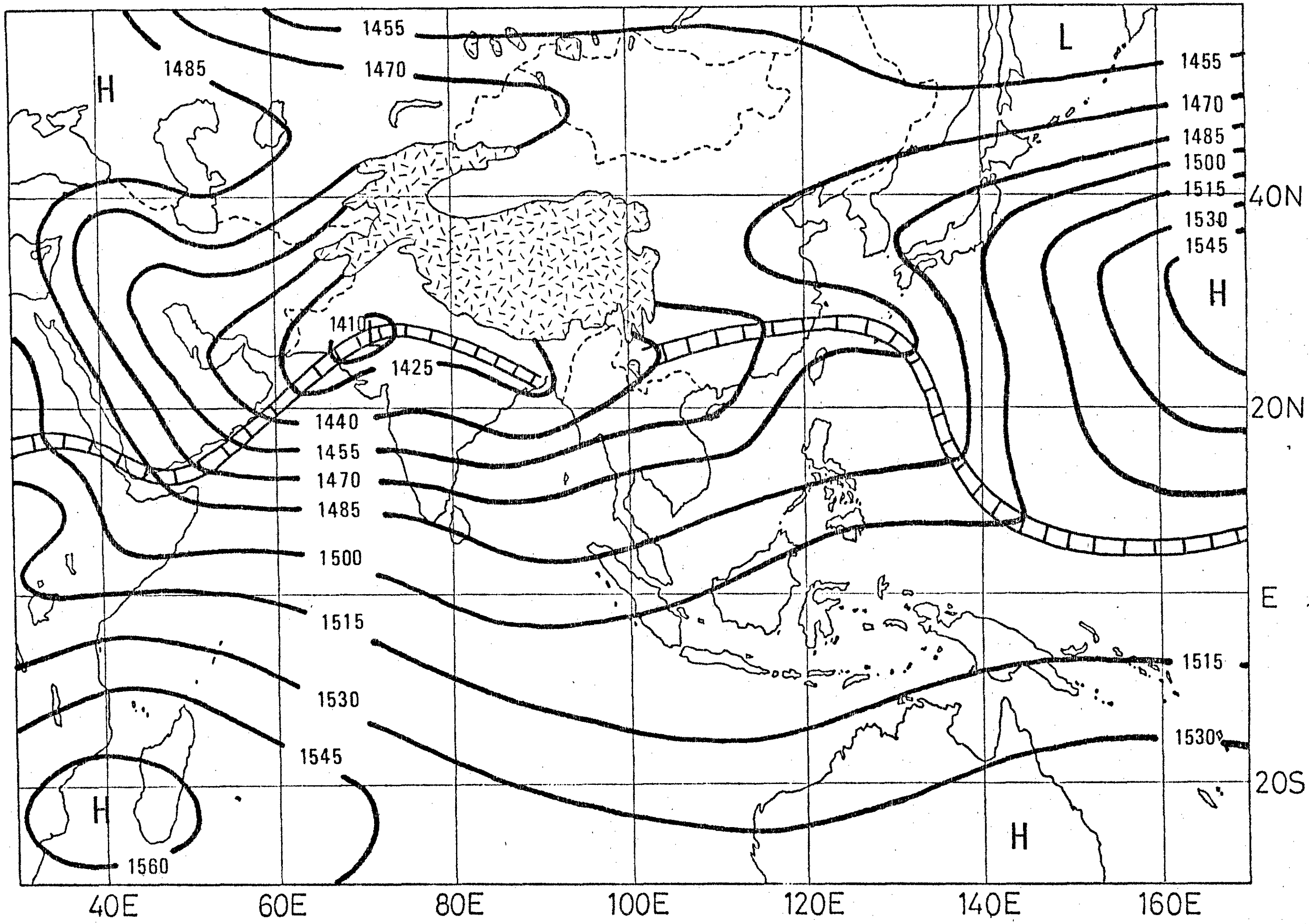

Fig. 61 Composite map of the geopotential height at the 850 mb level with strong TEJ in August (unit: geopotential meters)

southeast flow around the large typhoon stagnating near Okinawa. The western extension of the Pacific high extends across the Korean peninsula. Consequently, the summer rain in this peninsula is reduced by the subsidence. Finally, the middle latitude westerlies are displaced north of their normal position in northern Japan.

Figure 62 shows the composite map of the geopotential height at the 850 mb level with weak TEJ in 1965,1969,1972,and 1976. The map shows a considerable change in the regions to the east of 100°E. The ITCZ is displaced south of their normal position near the longitude of 130°E.

The western extension of the Pacific high reaches northern Ryukyu island along 30°N. Consequently, the southeast monsoon is restricted to Ryukyu islands. The Korean peninsula receives heavy precipitation associated with converging southwest flow. This summer rainy season in Korea is called Changma.

The westerly flow near northern Japan is much stronger than normal. In the other areas, the Indian low and Madagascar high are both weaker than normal. In addition, a blocking high develops near the Caspian Sea and a trough exists near 50°N 130°E which is associated with frequent cyclogenesis along the Eurasian polar fronts.

Figure 63 shows the difference in the geopotential height at the 850 mb level between strong TEJ and weak TEJ. The frequent activities of the typhoons in the years with strong TEJ are reflected in a fall of geopotential centered south of Kyushu. In a broad region from India to western Micronesia, the southwest monsoon is well developed when the TEJ is strong. Thus August also shows a strong monsoon circulation at the

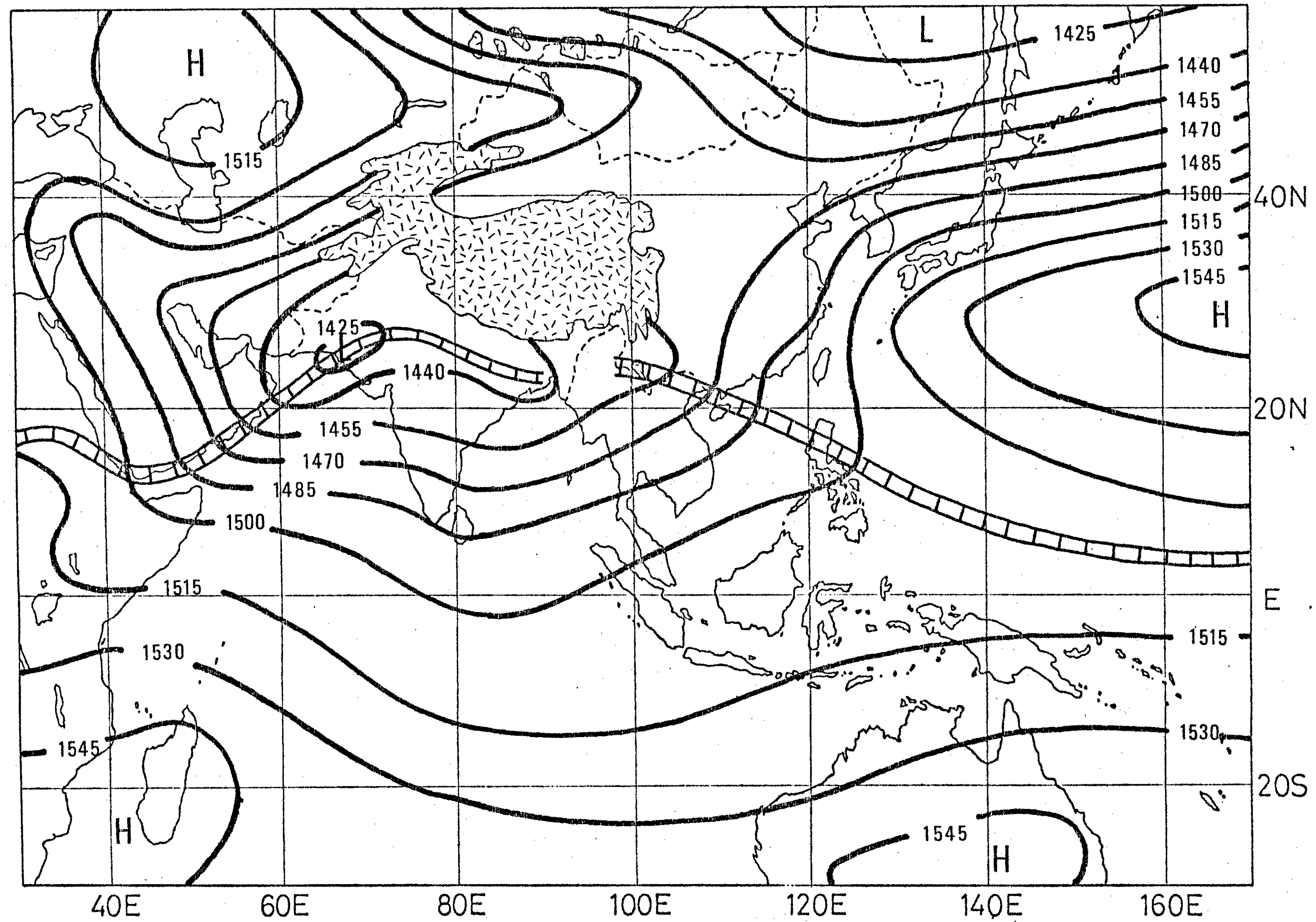

Fig. 62 Composite map of the geopotential height at the 850 mb level with weak TEJ in August (unit: geopotential meters)

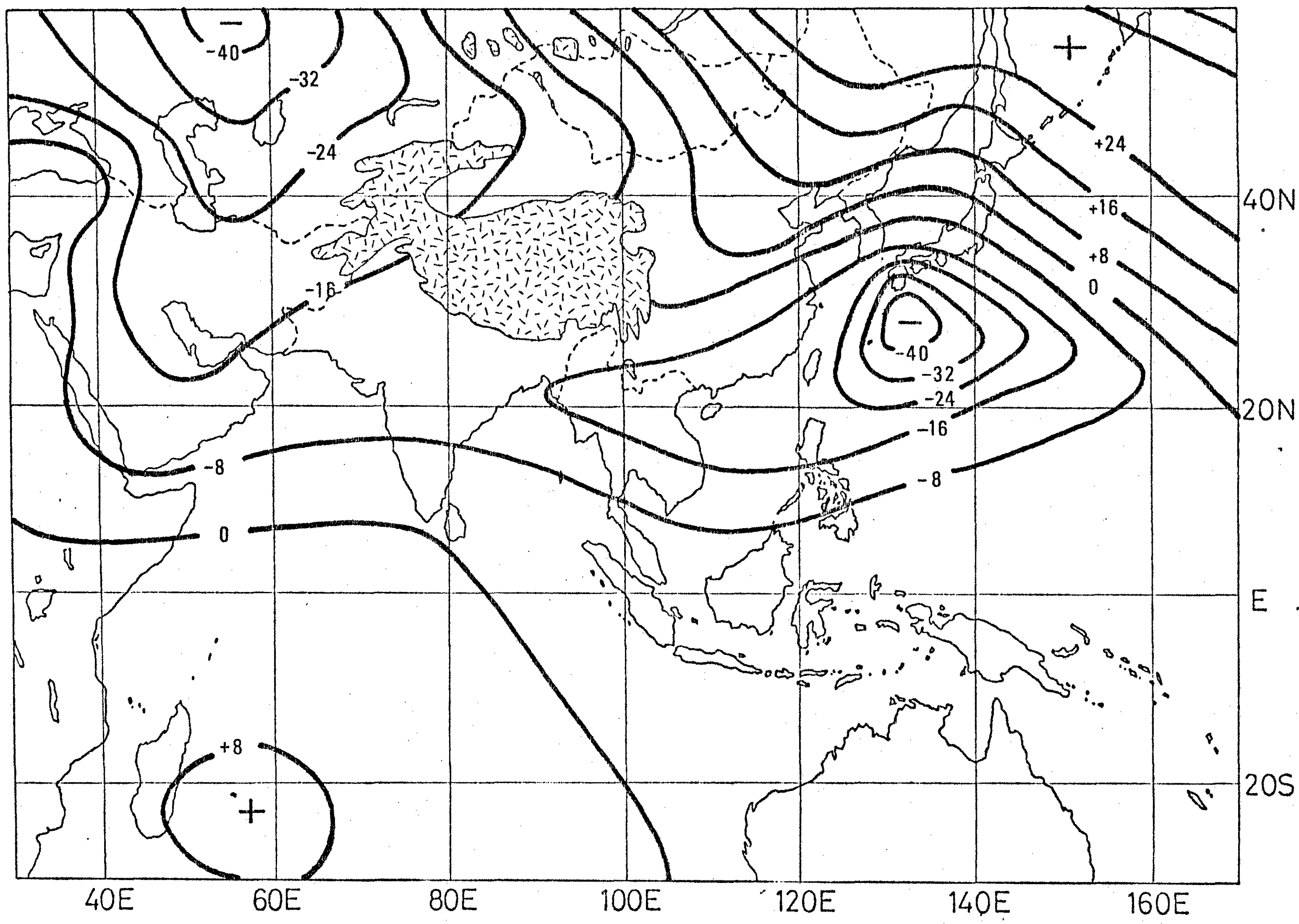

Fig. 63 Difference in the geopotential height at the 850 mb level in August (unit: geopotential meters) (strong TEJ minus weak TEJ)

850 mb level when the TEJ is strong.

Figure 64 shows the difference in the thickness between the 850 mb level and 150 mb level in August. The most conspicuous feature is the large increase of thickness associated with the intense activities of the typhoons near Okinawa when the TEJ is strong. The rest of the patterns are similar to that of July. These are strong subtropical jet associated with subsidence in the Middle East and a cold trough near the equatorial Indian Ocean. These maps clearly indicate the strong coupling between the activities of typhoons and the fluctuations of the monsoon in August.

6-5 Summary

The study of the role of the circulation at the 150 mb level during the summer monsoon has shown that the fluctuations in the strength of the TEJ at the 150 mb level are closely connected to the fluctuations of the summer monsoon at the 850 mb level and the distributions of precipitation. Unlike the case of the winter monsoon, the response of the monsoon circulation at the 850 mb level to the fluctuations of the TEJ at the 150 mb level is different in each of the three summer months considered in the current study.

In all three summer months the Indian monsoon circulation is strong when the TEJ is strong. In June the Baiu front and ITCZ near Micronesia are active when the TEJ is strong. In July the end of Baiu is delayed near Japan when the TEJ is strong. Finally in August, many typhoons stagnate near Okinawa when the TEJ is strong. The strong outflow from these typhoon helps to maintain the strong TEJ.

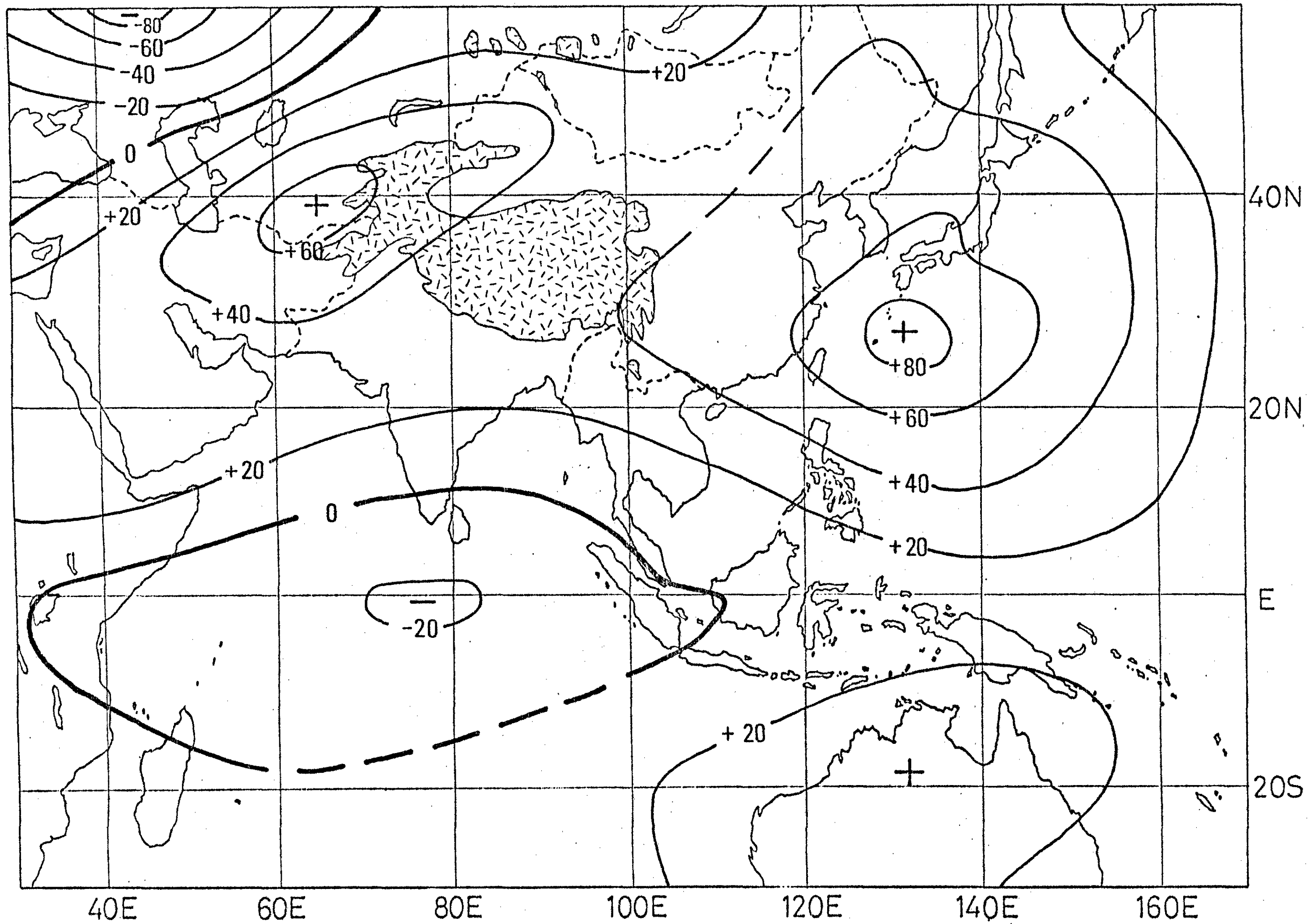

Fig. 64 Difference in the thickness between the 850 mb level and 150 mb level in August (unit: geopotential meters) (strong TEJ minus weak TEJ)

In the higher northern latitudes, there is a blocking high to the north of the Caspian sea and a trough of cold air to the east of Lake Baikal when the TEJ is weak in all three months of the summer. These results show that the release of latent heat is a critical factor in maintaining the strong monsoon circulations. The developments of the blocking high to the north of the Caspian Sea shows that the fluctuations of the summer monsoon are strongly coupled with the middle latitude circulations.

In this study the tropical and subtropical monsoon was discussed. At this point, it is interesting to compare the results of this paper and the research works on the East Asian monsoon. The role of Himalaya and adjacent Tibetan plateau is one of the important issues.

Staff members of Academica Sinica in Peking (1958) have estimated that the Tibetan plateau acts as a heat source in the summer months. However the extent of the South Asian high, which is far larger than the Tibetan plateau as shown in Figure 41, suggests that release of latent heat is probably more important mechanism for the maintenance of this system.

Another interesting topic is the relationship of the Baiu rain in Japan to the tropical monsoon. Figure 53 shows the difference in precipitation (strong minus weak TEJ) in June. In both India and southern Japan, there are regions with heavy precipitation when the TEJ is strong. Because the rainy season in these regions initiates during June, the date of onset of the Baiu rain and the monsoon rain has large influence on the precipitation in June.

Thus the case of strong TEJ is characterized by the early onset of the rainy season in both of these regions. On the other hand, when the TEJ is weak, the onset of the rainy season is late in both of these regions. Suda and Asakura (1955) found the tendencies for the simultaneous initiations of both the Indian monsoon and the Baiu rainy season. The results of the current study supports their discoverly.

The structure of the Baiu front and associated circulation regime were investigated by many researchers. The Chinese works such as Hsu (1965) treated the Mei-yü fronts as a part of the ITCZ. On the other hand, Japanese researchers such as Saito (1966), Kurashima and Hiranuma (1970) and Yabana (1973) considered the Baiu front as a polar front. This study suggests that the transition from the ITCZ to that of the polar front is more gradual processes. Between the longitudes from 110°E to 130°E, the characteristics of the Baiu front is subtropical with properties intermediate between the ITCZ and the polar front.

The locations of the ITCZ is also an important features of the monsoon circulation. Thompson (1951) has analyzed the locations of the ITCZ using the primitive radiosonde and pilot balloon data. More recently, Yoshino (1969,1971) has analyzed the distributions of the steadiness factor of the wind at the 2,000 feet level for a period from 1956 to 1960. He has identified the location of the ITCZ as the regions with low steadiness factor. In both of above studies, only the wind data was used to define the location of the ITCZ.

In the current study, the ITCZ was defined to exist at the ascending branch of the Hadley cell. For this reason, the wind at the 850 mb level, the geopotential height at the 850 mb level and the precipitation data were used as a additional reference for locating the ITCZ. The wind at the 850 mb level was the primary source of the data for the current study.

Since the regions of low steadiness factor can be associated with both the ITCZ and the subtropical high (Figure 14), some of the regions of the low steadiness in Yoshino (1969) were found to be a part of the subtropical high. These are located near Taiwan and Calcutta in January. In July they are located to southeast of Kyushu.

The relationship between the TEJ and precipitation distribution is another interesting problem. Kobayashi (1974) has conducted a survey of the relationship between the TEJ measured at India (150 mb-100 mb level), and the rainfall anomalies at tropical southern Asia. He found that there is a general increase in precipitation at India and Burma when the TEJ over India is strong. However domain of the rainfall and the wind data at the 150 mb level and 100 mb level considered in his study might be too small compared to the size of the regions which has a relation to the TEJ. His results are confirmed by the present study which covers entire Asian region. The general increase in precipitation was found to be distributed over much broader regions (such as southern Japan) when the TEJ is strong.

In the closely related problem of the precipitation distribution and role of the monsoon circulation at the low level, Yoshino and Aihara (1971) have classified four types of the distribution patterns of precipitation in monsoon Asia. The current study has shown that Type IIb is observed in August when the TEJ is strong and many typhoons stagnate near Okinawa. The other three types have less clearly defined relationship to the fluctuations of the TEJ.

On the other hand, the relationship to the monsoon circulation at the low level is better defined. Figure 61 which shows the northward displaced ITCZ near Japan is similar to their Type Ib and Type IIb. The contrasting situation of the southward displaced ITCZ near the longitude of Japan shown in Figure 62 is similar to their Type Ia and Type IIa.

Finally the definition of the monsoon region is an important problem. Ramage (1971) recognizes only the tropical monsoon located away from the regions of travelling cyclones and anticyclones which characterizes the Rossby regime of the middle latitudes. On the other hand, Kurashima (1968) has defined four monsoon regions. These are the tropical, subtropical, middle latitude and the polar monsoon regions.

In the current study as discussed earlier, the wind at the 850 mb level were used to define the monsoon regions (see Figure 48). This level was chosen so that the orographic distortion can be ignored. In addition, the current at the low level has a sufficient depth to be considered as a major current on the scale of the general circulation. In the middle and high latitudes, the seasonal shift of the wind direction at the 850 mb level usually does not reach 120°. This value was used by Ramage (1971) to define his monsoon regions. Using this windshift criterion, only the tropical and subtropical regions can be included in the Asian monsoon regions.

Hence northern Japan whose prevailing wind at the 850 mb level is from northwest in January and southwest in July and August is considered to lie wholly within the middle latitude westerlies and consequently not a monsoon region. Thus the monsoon regions defined by using three dimensional data in this study are not as limited as Ramage (1971) but not as broadly defined as the study of Kurashima (1968).

CHAPTER 7

RELATIONSHIP BETWEEN THE SUMMER MONSOON AND THE GENERAL CIRCULATION

7-1 Walker circulation and the summer monsoon

The Walker circulation discribed in the beginning of Chapter 4 was investigated for a possible relationship to the summer monsoon. Since the Walker circulation is a long period phenomenon, the average of three summer months was compared. Figure 65 shows the mean pressure at the sea level in July in the Pacific Ocean region. The data sources are same as Figure 29. The location of Easter Island and Darwin which was found to lie near the center of action suggests that the summer Walker circulation is a longitudinal oscillation of the subtropical high.

Figure 66 shows the difference in the pressure at the sea level between 1964,1970 and 1965,1972, the years when major El Niño have occurred. The subtropical high in the South Pacific is strong while the subtropical high in Australia is weak in the year before El Niño when the sea surface temperature in the Pacific Ocean is cold. The El Niño years are associated with a weak subtropical high in the South Pacific near Easter Island.

The relationship between the Walker Circulation Index (defined in Chapter 4) and the fluctuations of the TEJ from June to August (shown on the right hand column of Table 4) shows a correlation coeficient of +0.56 for a 14 year period from 1964 to 1977. When compared with the relationship to the circumpolar westerlies at the 500 mb level, this correlation turned out to be slightly lower. However the years with major El Niño

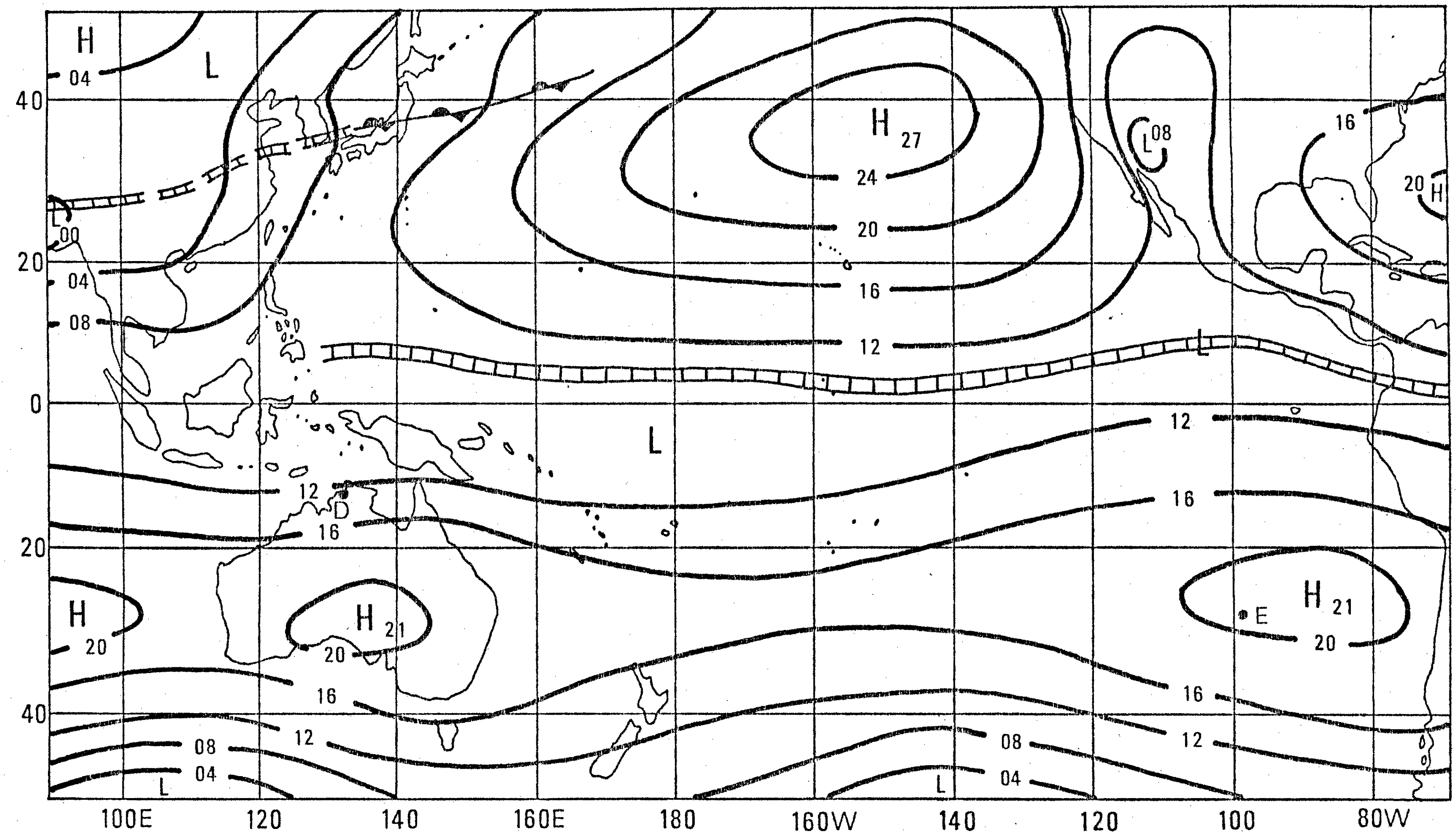

Fig. 65 Mean pressure at the sea level in July (unit: 1000 mb+)

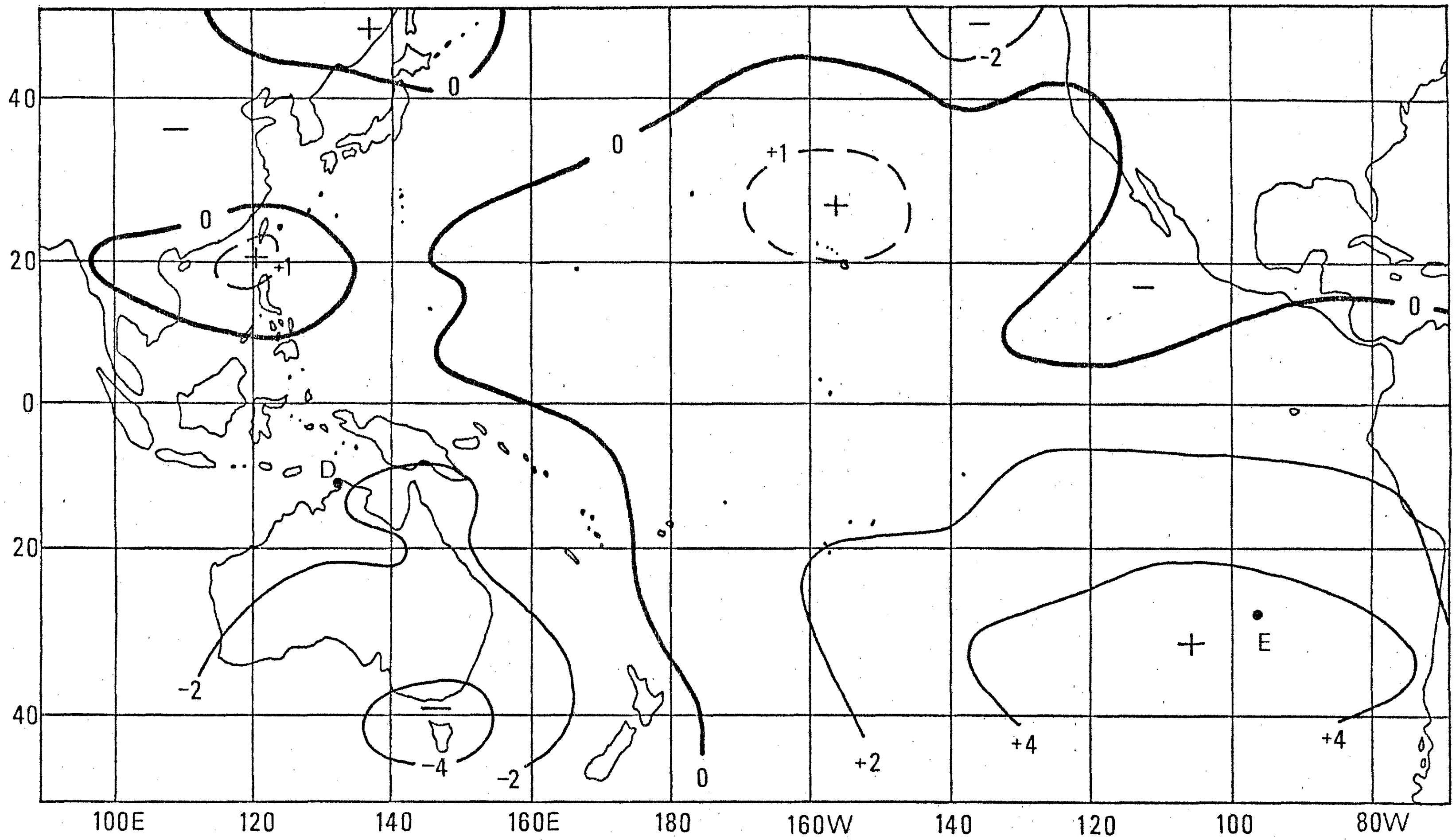

Fig. 66 Difference in the pressure at the sea level in summer:
1964.1971 minus 1965.1972 (unit: mb)

are also the years with unusually weak TEJ(1965,1972).

The relationship to the El Niño component of the sea surface temperature in the Pacific Ocean (Weare et al.,1976) shows a correlation of -0.55 with contemporary 5 month average centered on July. A slightly higher value of -0.60 was obtained for a 5 month average centered on October. However both the TEJ and sea surface temperature are available for only a 10 year period from 1964 to 1973. These results show that the fluctuations of the Asian summer monsoon are influenced by many other factors of the general circulation of the atmosphere.

7-2 Circumpolar westerlies in the northern hemisphere at the 500 mb level

The response of the atmospheric circulation to the change in the intensity of TEJ showed persistent blocking near the Caspian Sea when the TEJ was weak. This suggest a strong coupling between the fluctuations of the TEJ and the middle latitude circulations. For this reason, the circulation at the 500 mb level in the middle and high latitudes was investigated in June, July and August.

Figure 67 shows the mean geopotential height at the 500 mb level in July. The intensity of the circumpolar westerlies is much weaker than in winter (Figure 31). This is a direct result of the smaller temperature difference between the pole and the equator in summer. In addition, the center of the polar vortex is near the pole in summer. The southward displacement of the polar vortex toward Siberia in winter is a result of the extremely cold temperature in northeast Siberia. Returning to Figure 67, the month of July was shown because it represents average conditions between June and August in the oceanic regions

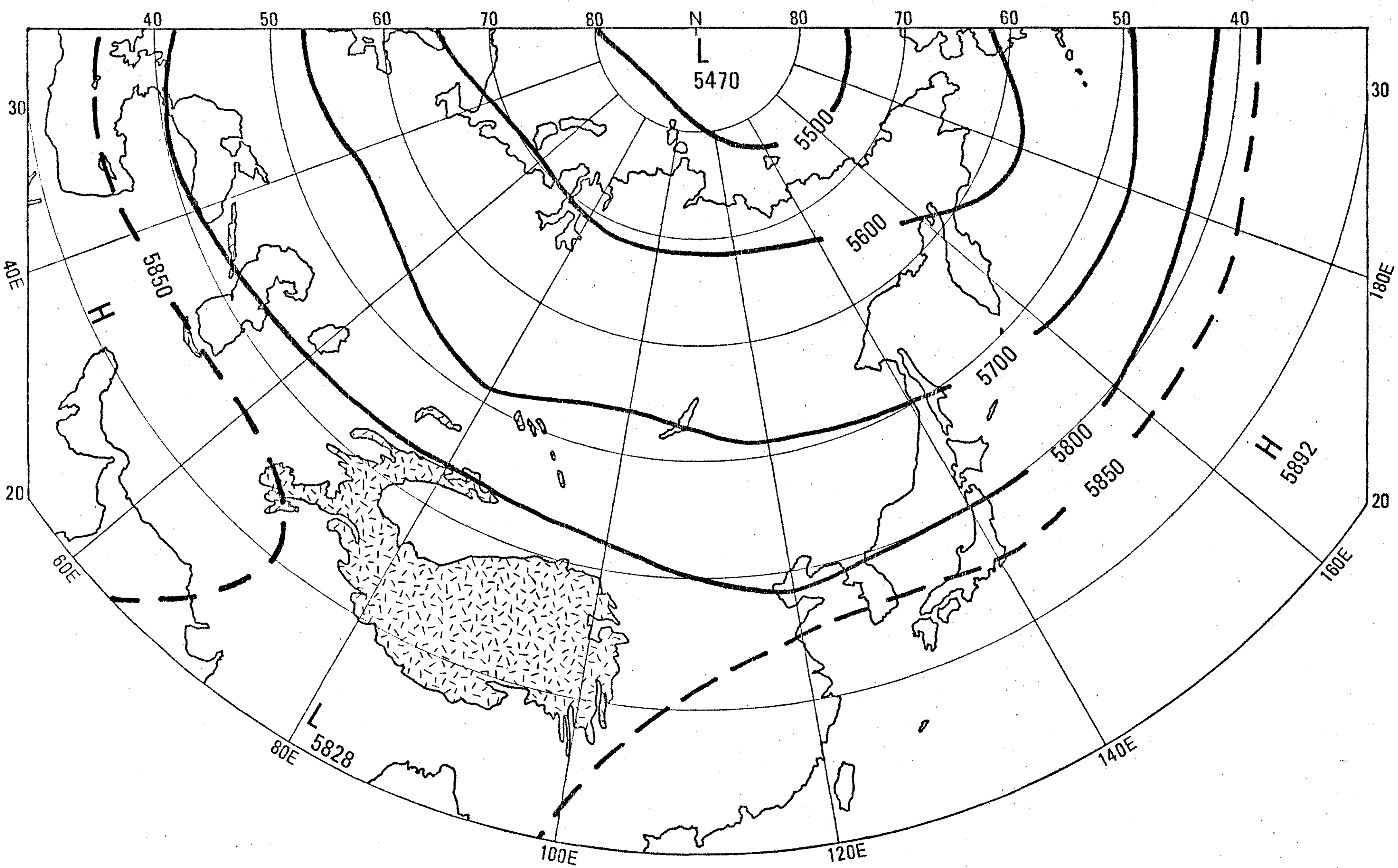

Fig. 67 Geopotential height at the 500 mb level in July (1946-1975 average) (unit: geopotential meters) (data adopted from Kisetsu Yoho Shiryo)

and the peak of summer in the continental regions. Along 50°N the height is still low over the Pacific Ocean reflecting the cold ocean in the northern Pacific. South of 35°N the circulation is weak and disorganized. This level is near the transition level between the southwest monsoon at the low level and the TEJ at the high level.

Compared with the winter circulation, the southern limit of the strong westerly wind has shifted north from 20°N to 35°N. This seasonal contraction of the circumpolar westerlies is less than the shift of the northern ITCZ (2°N in January and 25°N in July-August). Consequently, the interaction between the tropical Hadley circulation and the circumpolar westerlies is expected to be more significant.

Figure 68 shows the change of the geopotential height at the 500 mb level between June and August. The large area in the Pacific near 45°N shows an increase of more than 120 geopotential meters. This increase of height is a reflection of the seasonal warming of the Pacific Ocean which is warmest in late August.

The other area of notable increase in height is located near 40°N to the west of Himalaya. In this region, the subtropical jet migrates northward between June and July. The northward migration of the subtropical jet is also reflected in an increase of the westerlies in the latitudes from 45°N to 55°N and a decrease of the westerlies in the latitudes from 30°N to 40°N between June and August.

The relationship to the fluctuations of the TEJ was then investigated by a comparision of the two composite maps based on the strength of the TEJ. Figure 69 shows the difference

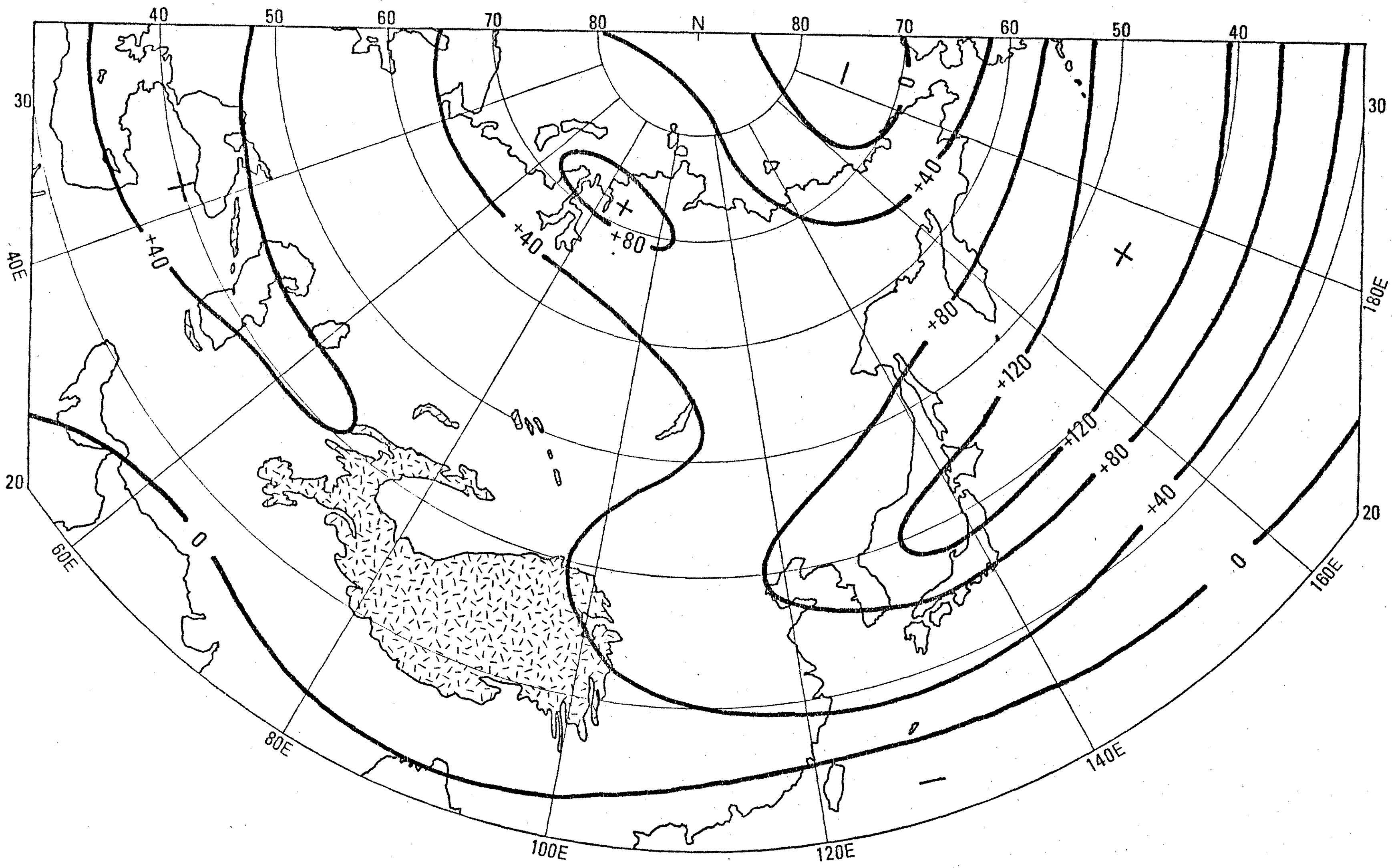

Fig. 68 Change in the geopotential height at the 500 mb level: August minus June (unit: geopotential meters)

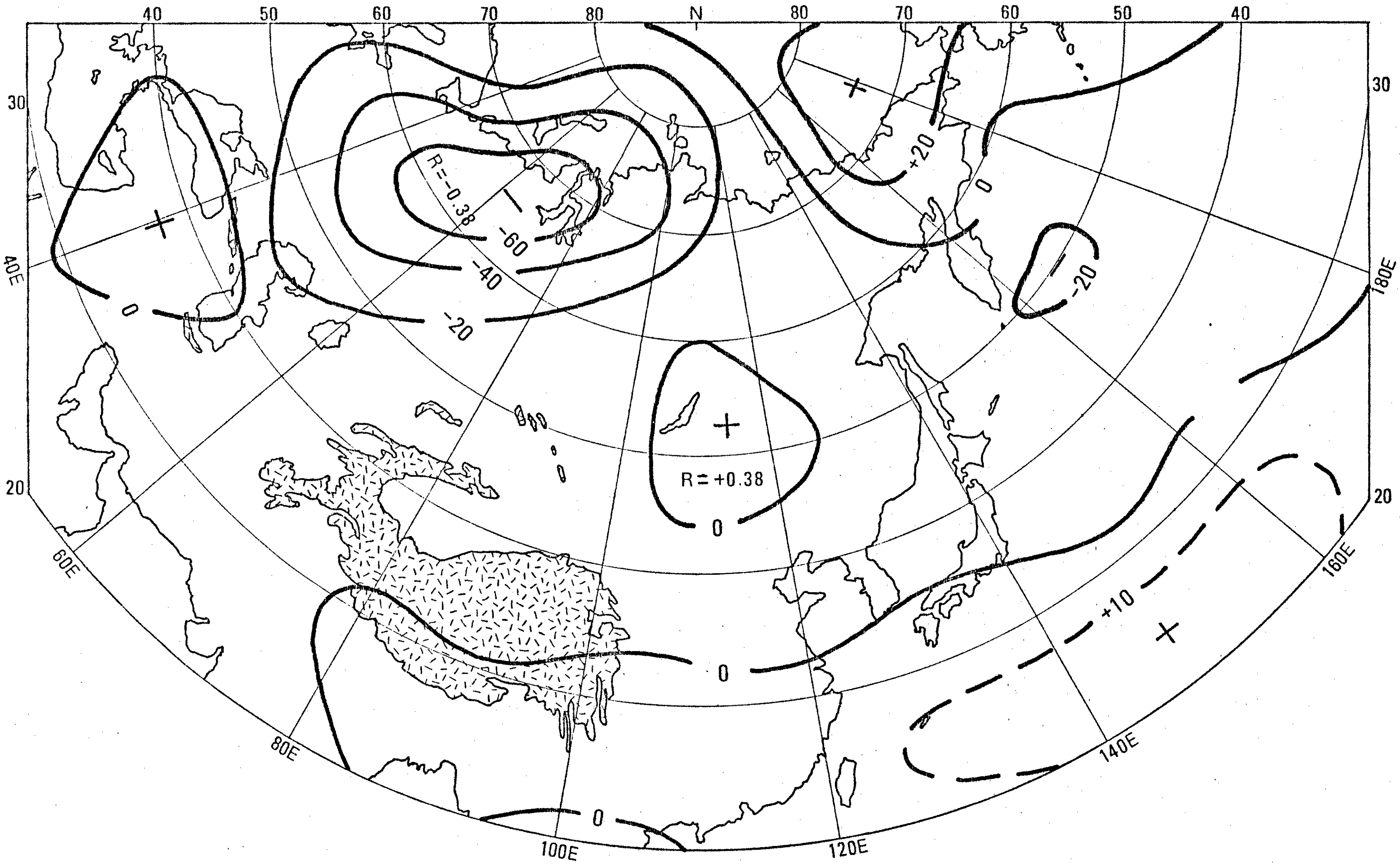

Fig. 69 Difference in the geopotential height at the 500 mb level in June (unit: geopotential meters) (strong TEJ minus weak TEJ)

in the geopotential height at the 500 mb level between the strong and weak TEJ in June. The developments of the blocking high to the north of the Caspian Sea and the trough to the east of Lake Baikal when the TEJ is weak are similar to that of July and August. However the monsoon circulation is much closer to the equator in June (Figure 42). Consequently the coupling with the middle latitude circulation is weak with correlation coeficient of -0.38 at 60°N 50°E and +0.38 at 50°N 120°E.

The unique feature in June is strong westward extension of the Pacific high, when the TEJ is strong. This is the only month with this relationship. This probably reflects a smaller monsoon circulation and a concentration of the latent and sensible heat source in the continental area.

Figure 70 shows the difference in the geopotential height at the 500 mb level between the strong and weak TEJ in July. The patterns of the blocking high to the north of Caspian Sea and the trough near Lake Baikal when the TEJ is weak are similar to June. However the expansion and northward displacement of the monsoon circulation result in a close coupling between the circumpolar westerlies and the fluctuations of the TEJ. This is shown by the high correlation of -0.56 at 50°N 40°E and +0.84 at 50°N 110°E in 14 Julies from 1964 to 1977.

The subtropical jet stream is strong to the west of Himalaya and meanders south near 120°E when the TEJ is strong. The Pacific high is displaced toward the east and the Baiu rain near Japan is heavy while drought prevails in the Korean peninsula. Another notable development in July is a ridge extending from Lake Baikal to the Pacific Ocean near 30°N 170°W.

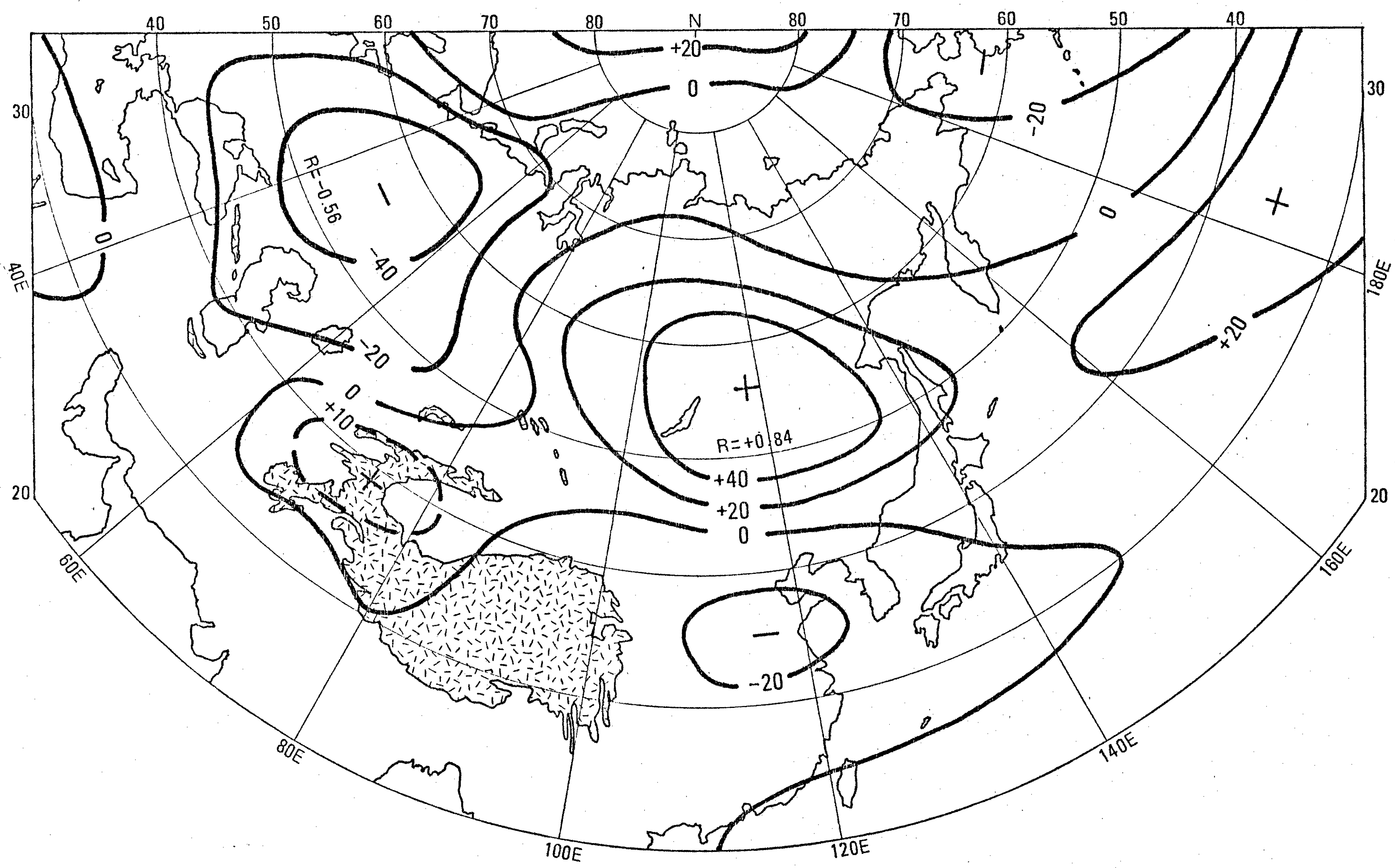

Fig. 70 Difference in the geopotential height at the 500 mb level in July

(unit: geopotential meters) (strong TEJ minus weak TEJ)

Finally, Figure 71 shows the difference in the geopotential height at the 500 mb level between strong and weak TEJ in August. The patterns near 50°N are similar to July. The close coupling between the circumpolar westerlies and the fluctuations of the TEJ continues in August. The regions of high correlation lie at 50°N 50°E (R= -0.65) and at 50°N 120°E with a correlation of +0.66. The response of the subtropical jet stream is different from July. To the west of Himalaya, this jet stream is strong when the TEJ is strong. However to the east of the Himalaya, this jet is displaced north of its normal position and flows near 50°N.

The regions of decreased height to the south of Kyushu are associated with the frequent activities of typhoons and a northward displacement of the Pacific high. It is important that this decrease in the geopotential height at the 500 mb level coincides with an increase in the thickness of the atmosphere (Figure 64). A comparision of Figures 63 and 71 also shows that the increase of thickness extends to the lower half of the troposphere with an increase of about 20 geopotential meters between the 500 mb level and 850 mb level.

Figure 71 also shows an important features of the circumpolar westerlies which favors the existence of slow moving typhoons near Okinawa when the TEJ is strong. This is a ridge extending from Lake Baikal to the Pacific Ocean near 40°N 160°E. The development of this ridge is associated with weak westerlies near Japan. Consequently, the typhoons are unable to recurve and therefore remain stationary near Okinawa. The strong outflow at the high level from these typhoons helps maintains a strong TEJ in August.

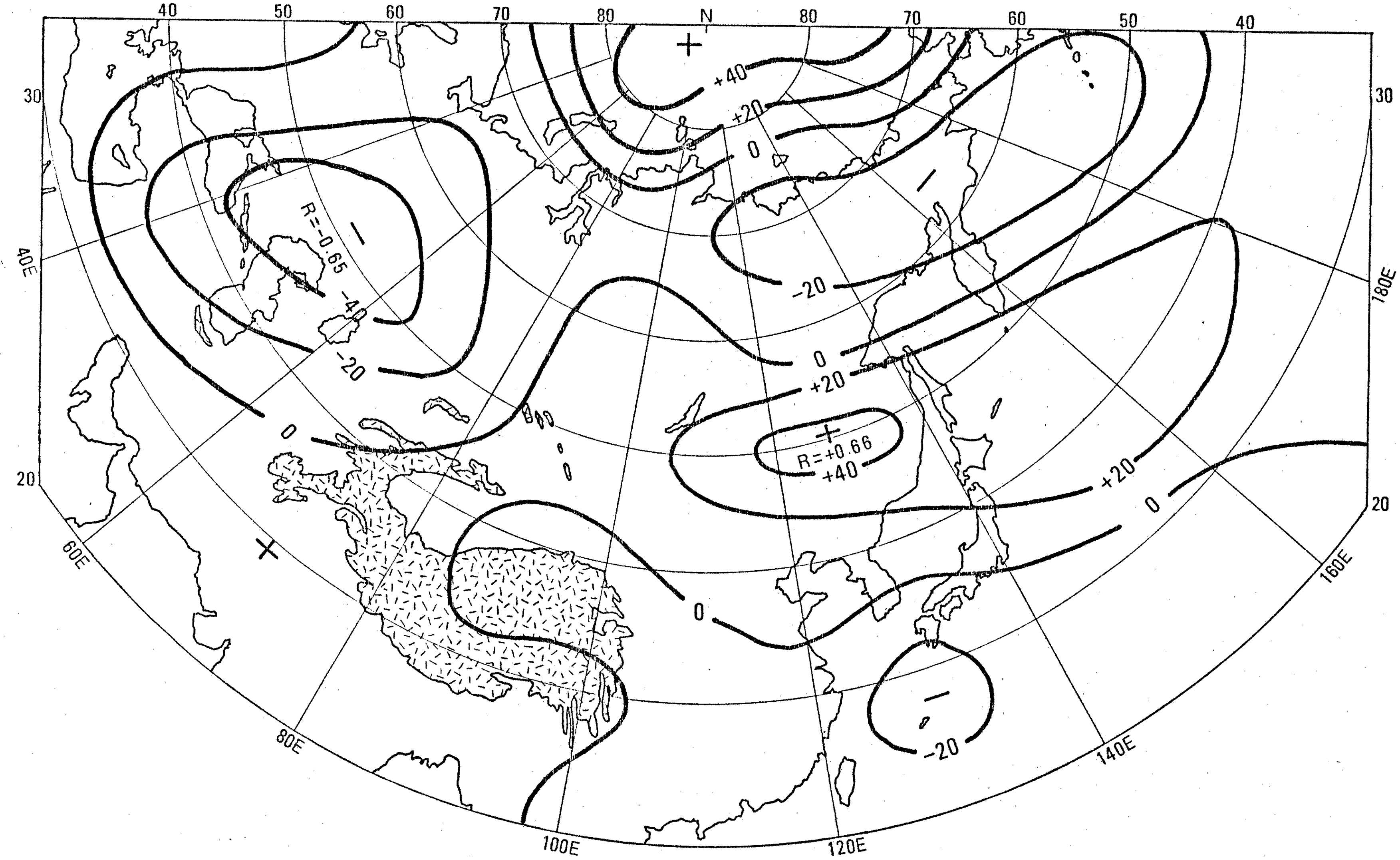

Fig. 71 Difference in the geopotential height at the 500 mb level in August (unit: geopotential meters) (strong TEJ minus weak TEJ)

The coupling between the fluctuations of the TEJ and the circumpolar westerlies is summarized in Table 5. In all three summer months, there is a tendency for the development of a blocking high to the north of the Caspian Sea and a trough to the east of Lake Baikal when the TEJ is weak. The coupling is weak in June. On the other hand, the tropical and middle latitude circulations show a strong coupling in July and August.

The right hand column in Table 5 shows the correlation for the average values of the three summer months. When compared with the correlation to the Walker circulation (which was +0.56 for the same period from 1964 to 1977) the coupling with the circumpolar westerlies is stronger. However compared with the case of the winter monsoon when the correlation to the Walker circulation was +0.94 for a period from 1961 to 1978, the summer monsoon system can be considered to be coupled with many systems of the general circulation of the atmosphere.

Finally a brief survey was conducted on the role of the standing wave in the circumpolar westerlies on the maintenance of the monsoon system in south Asia. It is well known fact that these standing wave are associated with the poleward transport of heat. The standard deviations of the monthly geopotential height at every 10° of longitude along the 50°N latitude circle between the longitudes from 30°E to 140°E were used as a measure of the activity of the standing waves.

The correlations to the fluctuations of the TEJ were found to be -0.36 in June, -0.65 in July, and -0.59 in August. This shows that the standing waves efficiently remove heat from the regions of the tropical monsoon when the monsoon circulation

TABLE 5

Relationship to the circumpolar westerlies at the 500 mb level at 50° N in summer

Correlation with TEJ (10°N) 1964-1977 (n=14)

Lat. °N	Long. °E	Month: June	July	Aug.	3 month period
50	40	-0.24	-0.56	-0.43	-0.58
50	50	-0.36	-0.48	-0.65	-0.60
50	60	-0.20	-0.33	-0.55	-0.33
50	100	-0.28	+0.54	+0.36	+0.20
50	110	+0.37	+0.84	+0.46	+0.75
50	120	+0.38	+0.63	+0.66	+0.68
50	130	+0.21	+0.49	+0.63	+0.57

represented by the TEJ is weak. The blocking high which frequently forms to the north of the Caspian Sea when the TEJ is weak is a manifestation of this tendencies.

7-3 East-west circulation

The east-west circulation, which is a part of the zonal asymmetry of the tropical general circulation was discovered by Krishnamurti (1971). As discussed in Chapter 1, this circulation is thermally direct with ascending warm air over Burma and descending cold air over the oceanic tropics. The intensity of this circulation was found to be comparable to that of Hadley circulation. The trough in the Pacific Ocean located near 30°N to the northwest of Hawaii was considered to be the descending branch of this circulation.

The few island rawinsonde data on Lihue and Hilo, Hawaii Island, Johnston Island and Midway Island were inspected for the possible connection to the fluctuations of the TEJ. The result was negative with little relationship to the trough at the 150 mb level in the Pacific Ocean. The intensity of the center of the Pacific subtropical high near Midway at the 850 mb level also showed little relationship to the fluctuations of the TEJ. Thus the east-west circulation in its original form suggested by Krishnamurti (1971) has little or no relationship to the fluctuations of the TEJ based on the monthly data.

On the other hand, the western extension of the Pacific high is intimately related to the fluctuations of the TEJ. In June the western extension of this high is strong when the TEJ is strong. This suggests a sort of east-west circulation on a smaller scale. In July and August, the western extension of the Pacific high is strong when the TEJ is weak. This is a complete reversal of the relationship. The activity of the Baiu front in July and the typhoons in August act as a source region of the TEJ. The response of the regional circulation to the fluctuations of the TEJ in June and July-August is different. This is probably related to the different locations and size of the monsoon circulation in the respective months.

CHAPTER 8

CONCLUSIONS

The study of the fluctuations of the monsoon circulation and its relationship to the general circulation of the atmosphere has shown many interesting characteristics of the monsoon fluctuations during the winter and summer monsoons.

In the winter monsoon, the fluctuations represented by the zonal wind at the 150 mb level at Singapore were found to be a part of the Walker circulation. The correlation between the seasonal values of this easterly zonal wind and the Walker Circulation Index was +0.94 in the 18 winters from 1961 to 1978.

The relationship between the zonal wind at the 150 mb level at Singapore and the monsoon circulation at the low level was well defined in winter. When this easterly wind is strong, the monsoon circulation at the low level is also found to be strong between the longitudes from 100°E to 140°E and two active ITCZs are observed near 140°E. When this easterly wind is weak, only one ITCZ is observed near 140°E and the monsoon circulation at the low level is weak.

The inspection of the entire Walker circulation has shown that, when the zonal wind at the 150 mb level at Singapore is strong, the Pacific Ocean near the equator to the west of Peru is cold and the activity of the ITCZs is weak in the central Pacific near 170°E. When this easterly zonal wind (150 mb level) is weak, the circulation in the central Pacific is strong and there is a development of two ITCZs near 170°E.

These facts clearly show that the fluctuations of the Walker circulation in winter are longitudinal oscillations of the equatorial trough and associated ITCZ. The ITCZ is active in the winter monsoon regions when the pressure is low near Cocos Island and high near Tahiti Island. When the active regions of the ITCZ shift eastward, the reverse conditions of the weak monsoon are observed. These fluctuations of the Walker circulation are coupled to the sea surface temperature in the equatorial Pacific, suggesting a complex atmospheric and oceanographic feedback mechanism.

In the case of the summer monsoon, the fluctuations represented by the wind speed of the tropical easterly jet (TEJ) measured along 10°N between the longitudes from 40°E to 110°E are found to be influenced by the circumpolar westerlies near 50°N and less often by the Walker circulation.

In June, the monsoon circulation is still in its development stage and because of its smaller size and relatively lower latitudes, the interactions with the other circulations of the atmosphere were found to be small. The relationship between the fluctuations of the TEJ to the monsoon circulation at the low level and precipitation in June is as follows: The increase in the precipitation associated with the early onset of the rainy season is observed in India and southern Japan, while the monsoon circulation at the low level is strong when the TEJ is strong. When the TEJ is weak, the monsoon circulation at the low level is weak and the onset of the rainy season is delayed.

In July and August with strong TEJ, the monsoon circulation at the low level is strong, and heavy orographic precipitation is observed in western India and the west coast of Indochina. In southern Japan, heavy precipitation during the Baiu season in July and the typhoons in August are observed. The outflow at the high level from these systems helps maintain the strong TEJ.

When the TEJ is weak, the monsoon circulation at the low level is weak and there is a decrease of the precipitation over the large areas of Asia. The western extension of the Pacific high is strong and the southwest flow around this high is transporting the moisture toward the Korean peninsula where heavy precipitation is observed.

The relationship of the fluctuations of the TEJ to the circumpolar westerlies was found to be the most important interactions of the summer monsoon. In all three summer months, there is a tendency for the development of a blocking high to the north of the Caspian Sea and a trough to the east of Lake Baikal when the TEJ is weak. Hence the poleward transport of heat by the standing waves in the circumpolar westerlies increases when the TEJ is weak.

The coupling between the TEJ and the circumpolar westerlies was found to be much stronger in June than in July and August. This is a direct consequence of the expansion and the northward displacement of the monsoon circulation during July and August.

The relationship of the TEJ to the Walker circulation was found to be slightly lower than the circumpolar westerlies. However the years with major El Niños are also the years with unusually weak TEJ. The monsoon circulations in 1965 and 1972 are prime examples of such years. Unlike the case of the winter Walker circulation, this circulation in the northern summer is the longitudinal oscillation of the subtropical high located in southern Australia and near Easter Island.

In both the winter and summer monsoons, the release of latent heat and consequent warming of the middle troposphere is the critical factor in the developments of the monsoon circulation.

Intense sensible heating such as that observed in Sahara is insufficient to maintain a deep monsoon circulation. The area of the intense release of latent heat estimated by heavy precipitation also coincides with a mass source at the high level confirming the Hadley cell nature of the monsoon circulation.

Thus this study has shown that the upper branch of the monsoon circulation represented by the wind at the 150 mb level can serve as an excellent method for monitoring the fluctuations of the monsoon circulation. This circulation at the high level has the additional advantege that it is not deformed by small scale topography. The schematic maps of the strong and weak monsoon in winter and summer are shown in Figures 72,73,74 and 75.

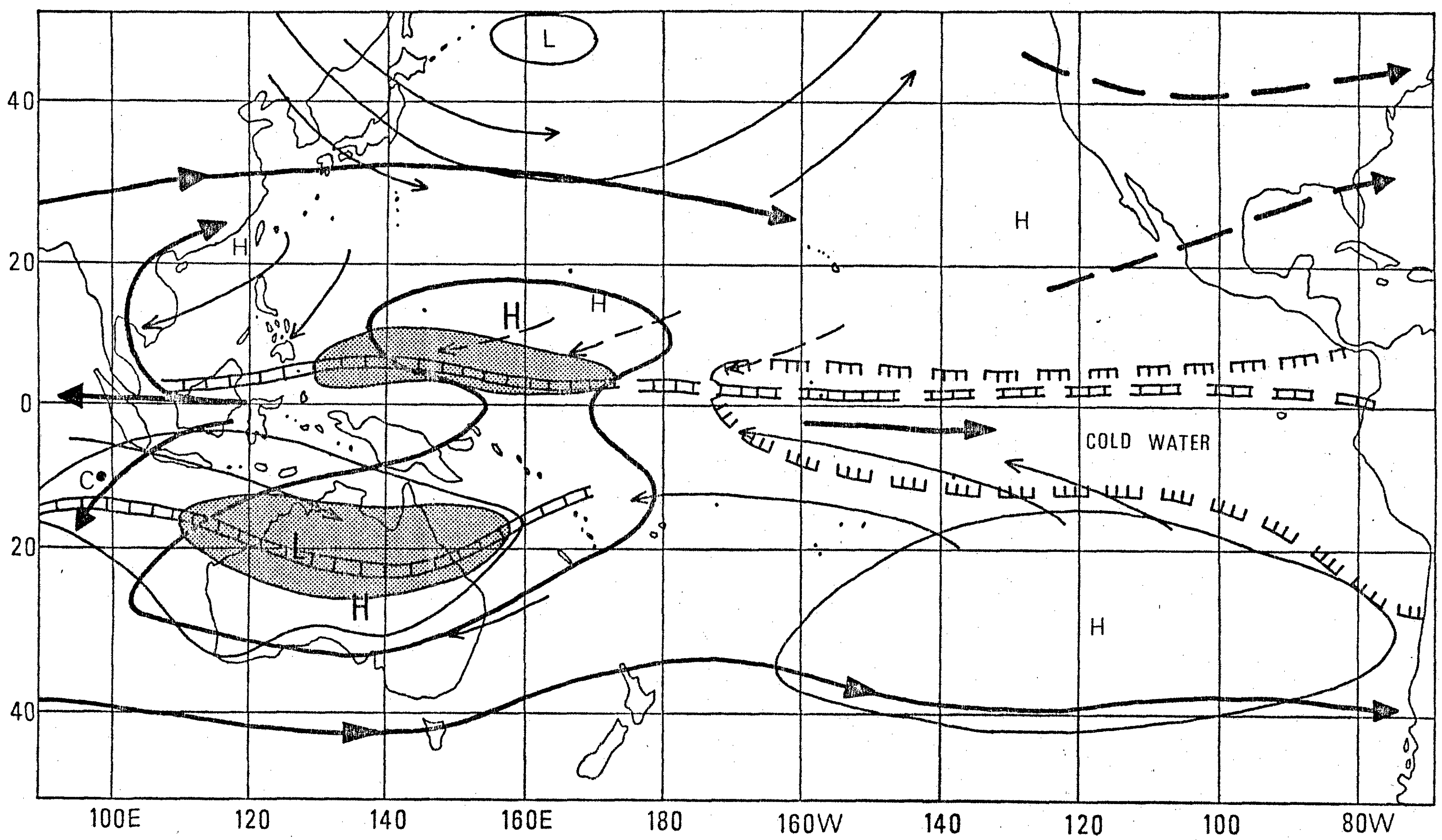

Fig. 72 Schematic map of strong winter monsoon

Thick solid (dashed) lines are strong (weak) circulation at the 150 mb level.
Thin solid (dashed) lines are strong (weak) circulation at the 850 mb level.
Shaded areas are the regions of heavy precipitation.
ㄇㄇ = the region of cold water.

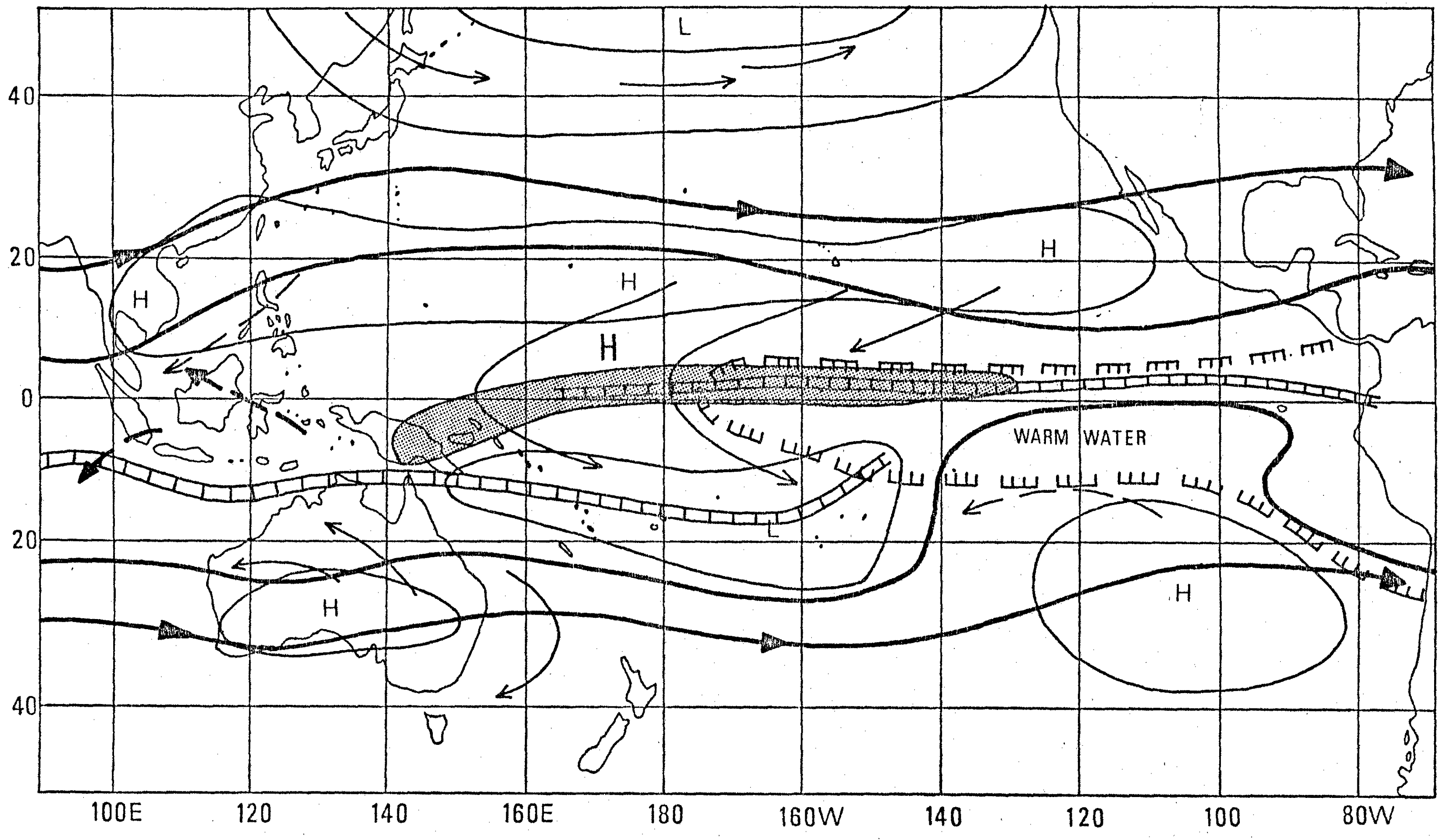

Fig. 73 Schematic map of weak winter monsoon

Thick solid (dashed) lines are strong (weak) circulation at the 150 mb level.
Thin solid (dashed) lines are strong (weak) circulation at the 850 mb level.
Shaded area is the region of heavy precipitation.
ՊՊ ՊՊ = the region of warm water.

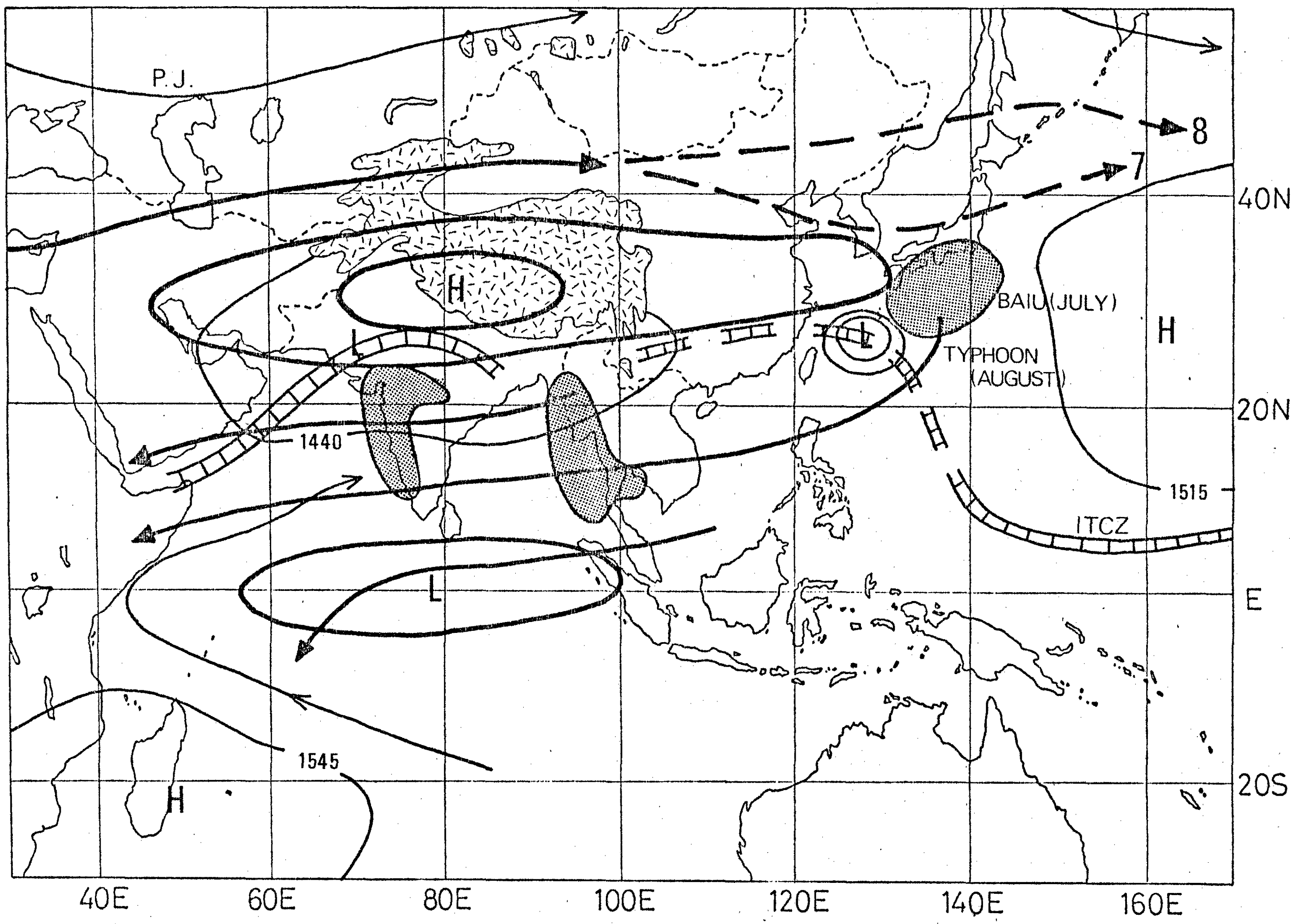

Fig. 74 Schematic map of strong summer monsoon (July and August)

Thick solid (dashed) lines are strong (weak) circulation at the 150 mb level.
Thin solid (dashed) lines are strong (weak) circulation at the 850 mb level.
Shaded areas are the regions of heavy precipitation.

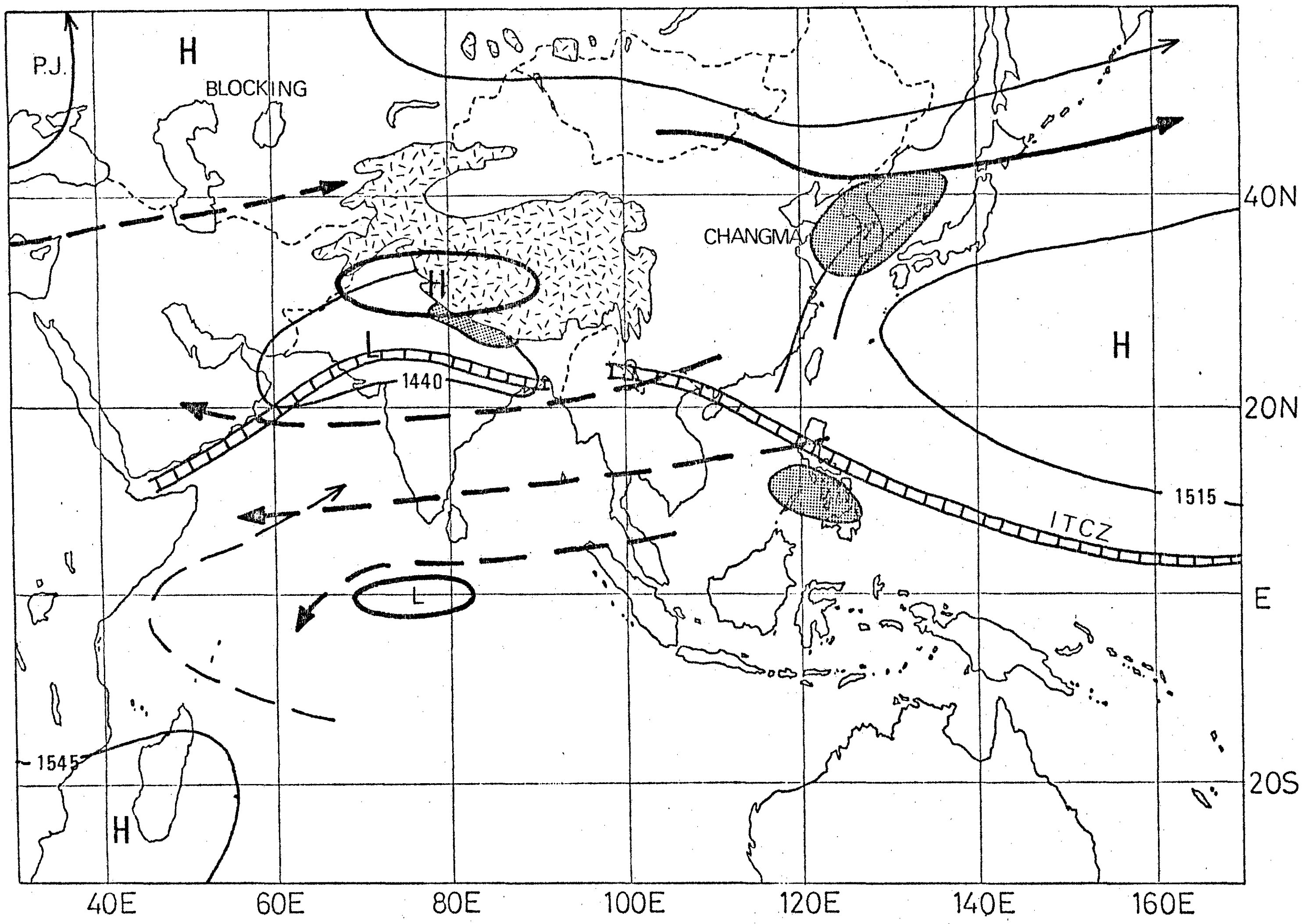

Fig. 75 Schematic map of weak summer monsoon (July and August)

Thick solid (dashed) lines are strong (weak) circulation at the 150 mb level.
Thin solid (dashed) lines are strong (weak) circulation at the 850 mb level.
Shaded areas are the regions of heavy precipitation.

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