Theor. Appl. Climatol. 71, 1–16 (2002)

Theoretical and Applied Climatology

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Demarcating the worldwide monsoon

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With 9 Figures

Received June 29, 2000
Revised May 15, 2001

Summary

The monsoon is a global climate phenomenon. This paper addresses the seasonal reversal of atmospheric circulation and the transition of dry/wet spells in the monsoon regions worldwide. The NCEP/NCAR 850 hPa wind reanalysis data for 1950–1999 and the upper-tropospheric water vapour (UTWV) band brightness temperature (BT) data observed by NOAA polar orbiting satellites for 1980–1995 are used. In the tropics, there are three large wet-UTWV centres. The summer monsoon in the subtropics can be defined as the expansion of these centres associated with the influence of cross-equatorial flows. Specifically, the dry/wet spell transition is determined by the BT values that are smaller than 244 K. The regions of the South and North African monsoons, the Asian-northwest Pacific and Australian-Southwest Pacific monsoons, and the North and South American monsoons are examined with a focus on the dry/wet spell transition and stream field features.

Findings suggest that the summer monsoons over many subtropical regions can be defined by both cross-equatorial flows and dry/wet spell transitions. In the mid- and low-latitudes, there exist six major dry/wet spell transition regions with a dry or wet period lasting more than one month. The region of most significant change is located over the subtropical North Africa–Asia–northwest Pacific. The others appear over subtropical South Africa, Indonesia–Australia, southwest Pacific, the Mexico-Caribbean Sea, and subtropical South America. In addition, three regions exist where only one of the two indicators (cross-equatorial flow and dry/wet transition) is satisfied. The first is near the Equator where the directions of cross-equatorial flows alternate but a dry/wet spell transition is never experienced. The second is over North Africa where only the dry/wet spell transition can be found but not the cross-equatorial flows. The third is over the mid-latitude regions in North China, South Africa, and northern North America. These regions are influenced by cross-equatorial flows but the upper-tropospheric water vapour content is not as high as that in tropical regions.

1. Introduction

The term monsoon is most often applied to describe the seasonal reversals of wind direction and persistent convective precipitation along the shores of the Indian Ocean (see Ramage, 1971; Webster, 1987). Since changes in the monsoon exert a strong impact on the economy and human lives in many countries of the world, a better understanding of the variability and predictability of the monsoon is one of the major research activities of the CLIVAR (Climate Variability and Predictability) Initial Implementation Plan (1998). It is known that monsoon phenomena appear in many parts of the world (Tao and Chen, 1987; Menton and McBride, 1992; Lau, 1992; Webster and Yang, 1992; Adams and Comrie, 1997; Janicot et al., 1998; Zhou and Lau, 1998; Trenberth et al., 2000). Each monsoon component varies, with both its own momentum and interaction with other monsoons (Tang and Reiter, 1984; Yang and Gutowski, 1992). Obviously, a better understanding of the global monsoon requires a broad knowledge of almost all of the phenomena associated with the strong annual cycle within the tropical and subtropical continents of Asia, Australia, Africa, and the Americas.

To define the geographic extent of the global monsoon, we need to clarify first what constitutes a monsoon climate. A most common definition relies usually on the characteristics of the annual variation of both wind direction and convective rainfall. Meanwhile, the definition also requires a wet summer and a dry winter. According to this criterion, no monsoon exists in the annually-permanent wet regions near the Equator and the annually-permanent dry regions in high latitudes, as well as the steady anti-cyclonic circulation regions. The summer monsoon in the subtropics has been defined as an expansion of the atmospheric wet convective centres associated with tropical monsoon flows (Qian and Lee, 2000). The upper-tropospheric water vapour (UTWV) band brightness temperature (BT) or high-resolution infrared radiation sounders (HIRS) water vapour band BT has been used to monitor the variation of atmospheric water vapour. The seasonal march and the demarcation of major summer monsoon regions in Asia have been investigated using the BT and low tropospheric winds (Qian and Lee, 2000; Qian and Yang, 2000). It has been noted that in the subtropical anti-cyclonic regions the BT is higher than that of other regions in the tropical and high latitudes. The low BT in the Tropics represents the wet atmosphere while the low BT in the polarward part of the anti-cyclonic region shows the low tropospheric temperature. We anticipate that the results mentioned previously will provide valuable information for a more thorough investigation of the global monsoon in the present paper.

To identify the global monsoon regions, two data sets will be used in this study. The first is the NCEP/NCAR wind reanalysis data (Kalnay et al., 1996). The second is the UTWV band BT or HIRS water vapour band BT data observed by the NOAA polar orbiting satellites (Wu et al., 1993; Bates et al., 1996). There are three channels on the HIRS among which channel-12 (or HIRS12) is designed to measure the amount of water vapour at the upper troposphere. Both data sets have the same spatial resolution of 2.5 degrees. The BT is available in five-day (pentad) means from 1980 to 1995. The reanalysis wind used here is the monthly data from 1950 to 1999. In this paper, the climatological mean of BT is based on the HIRS12 data for 1980–1995 while that of the 850 hPa wind is based on the NCEP/NCAR data for 1950–1999. The interdecadal changes in the NCEP/NCAR 850 hPa reanalysis winds have been analyzed by Qian and Zhu (2001). It has been noted that the summertime 850 hPa southwesterly wind during 1950s–1960s was stronger than that during 1980s–1990s in East Asia. This indicates that the domain of summer monsoon flow influencing East Asia may vary during different decades. Therefore, it should be kept in mind that the monsoon region identified by the BT for 1980–1995 may underestimate the extension of the East Asian monsoon. To describe and demarcate the climatology of the global monsoon, longer records are desirable.

This paper first describes the seasonal variation of global BT in mid- and low-latitude regions in section 2. The regions of the African monsoon, the Asian-Australian monsoon, and the American monsoon are then identified using 850 hPa wind reversals and BT in sections 3–5, respectively. Finally, the features of the global monsoon are discussed in a more reconciled way in section 6. Only the summer monsoon is discussed in this study although the monsoon phenomena also consists of another component: the winter monsoon.

2. Global dry/wet spell transition

The HIRS12 is designed particularly to measure the amount of water vapour in the upper troposphere whose retrievals tend to peak at around 325 hPa in the wet tropics (Berg et al., 1999). High HIRS12 value represents the low water vapour content, and vice versa. Although cirrus clouds can influence the BT derived from HIRS12, with low temperature in the upper troposphere in some individual cases, we ignore this influence on the climatology. Wu et al. (1993) and Bates et al. (1996) have described that the HIRS12 can not only indicate the water vapour content in the upper troposphere but also reflect the ascending and descending motions of the atmosphere. If the ascending or convective motions occur in a wet atmosphere, the water vapour content at the upper troposphere increases. As a result, the value of BT or HIRS12 decreases. When descending, the air becomes dry and the value of BT increases. Thus, the value of water vapour BT can be considered as an indicator of vertical motion mainly at the upper troposphere. Correspondingly, it can indicate the intensities of deep convection and the subtropical high.

During the onset of the summer monsoon, deep convection develops and moisture increases at the upper troposphere over the monsoon regions. Therefore, the changes in BT can appropriately measure the onset of summer monsoon (Qian and Yang, 2000). In contrast to the outgoing longwave radiation (OLR) that depends on cloud top temperature during deep convection, the BT reflects the temperature of a cloud-free upper troposphere. The advantage of OLR lies in its indication of the centre of deep convective rainfall, while the BT shows rather smooth patterns in summer monsoon regions. Comparisons have indicated that the BT can measure the monsoon rainfall in subtropical regions while OLR is inconsistent with the rainbelts in the extratropics (Qian and Lee, 2000; Qian and Yang, 2000).

In a previous investigation, Qian and Lee (2000) used the threshold BT value of 244 K in mid- and low-latitude regions. The area where BT is lower than 244 K marks the wet and deep convective atmosphere, as opposed to the dry and descending one. Figure 1 displays the monthly climatological-mean BT for January, April, July, and October. The lightly dashed line covers BT lower than 244 K. Note the $BT < 244\,K$ in the tropical regions and the $BT > 244\,K$ in the subtropical regions. In January (Fig. 1a), three wet centres appear to the south of the Equator over western Africa, the marine continent in the equatorial western Pacific, and South America. The wet belt indicated by the heavily dashed line from Africa through the Indian Ocean to the southwestern Pacific coincides with the Inter-Tropical Convergence Zone (ITCZ) in this season. Another wet belt appears from subtropical South America to the equatorial Atlantic. Along 20° N, there is a dry zone with three descending centres over the Arabian Sea, the North Pacific, and the North Atlantic. A dry centre also appears over the Southeast Pacific, linking to the descending zone of the Northern Hemisphere. These dry or descending areas are helpful for outlining the anti-cyclonic circulation in the troposphere.

In April (Fig. 1b), three wet centres are located separately in the equatorial region. The values of the dry and wet centres in this transient month from the boreal winter to summer decrease relative to those in January. Two wet areas can be identified in equatorial western Africa and the Amazonian Basin in South America. The largest wet area appears over Indonesia. At the same time, the wet belts (see the dashed lines) move northward but become weaker relative to January. Two locations indicating the northward advance of wet areas, or the weakened descending motion, can be found near the Indochina Peninsula in Southeast Asia and Panama in Central America.

In July (Fig. 1c), the three wet centres move to the north of the Equator and become significantly more extensive. Compared to the previous months, both the wet and dry centers intensify. In this month, north subtropical Africa, South Asia, Southeast Asia, the Northwest Pacific, and the Mexico-Caribbean Sea are all confined within the wet areas. The deep convective line similar to the ITCZ is basically along 10° N while a strong descending zone appears in the southern subtropical latitudes. Strong descending areas in Southwest Asia, the southern Indian Ocean, and the Southeast Pacific are consistent with the anti-cyclonic circulation. In October (Fig. 1d), three wet centres appear again over the Equator but their orientation is somewhat meridional and isolated in this transient month from the boreal summer to winter.

From Fig. 1, it can be noted that the three wet centres shift in location and change in intensity from month to month. The centre over western Africa extends northward to 19° N in the boreal summer and shifts southward to 18° S in the boreal winter. The wet domain over the South–North Americas extends southward to 19° S along the South American continent in January and reaches 40° N over the Northwest Atlantic in the boreal summer. For the largest wet domain over Asia–Australia and the western Pacific, the southern boundary extends to 10° S near Australia and 22° S over the central South Pacific in January. In the boreal summer, this domain covers India, Southeast Asia, and the Northwest Pacific. Geographically, three continental belts, i.e., the African continent, the Asian-Australian continental bridge, and the South-North

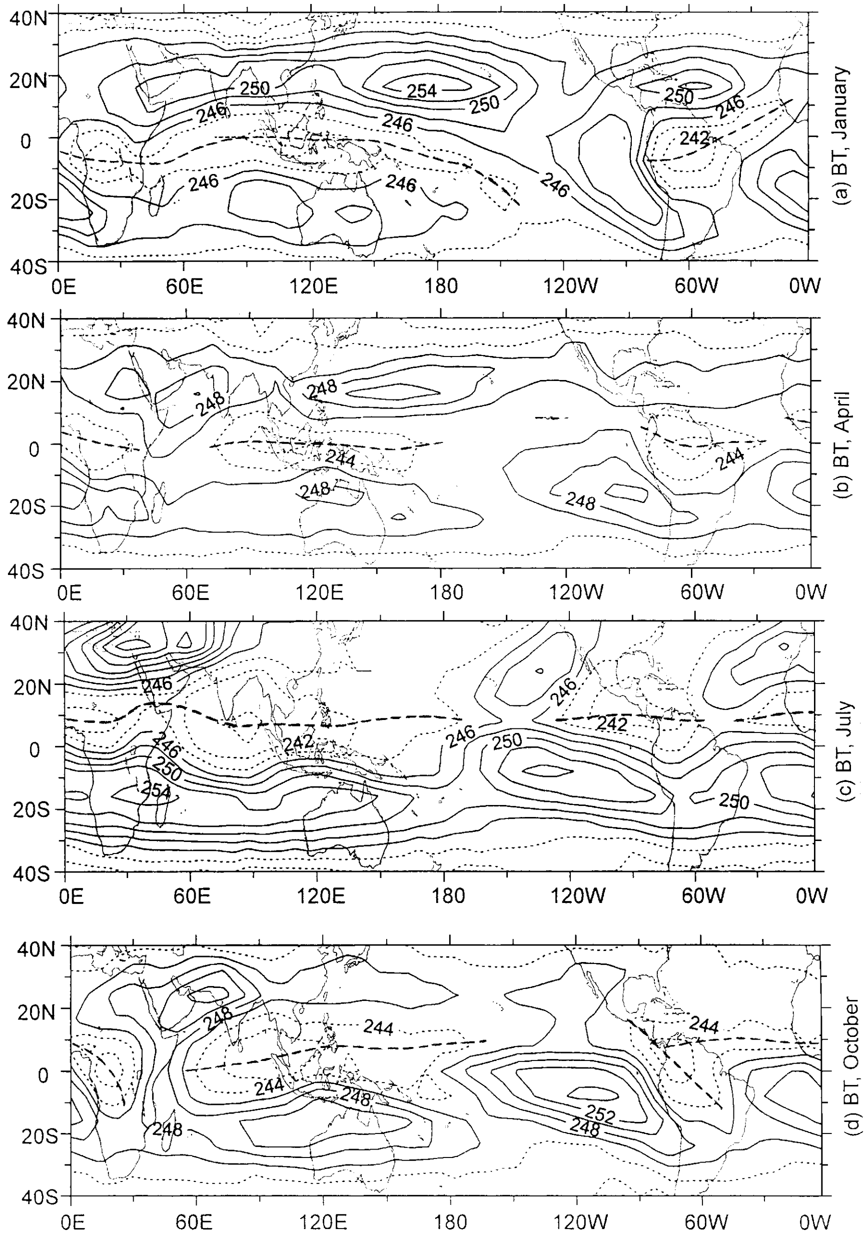

Fig. 1. Climatology of global BT (K) for (**a**) January, (**b**) April, (**c**) July, and (**d**) October based on the HIRS12 data from 1980 to 1995. Lightly dashed areas outline BT < 244 K. Heavily dashed lines indicate the wet maximum or deep convective zone

American continent separate global ocean in the mid- and low-latitudes into the three basins. The orientation of the three continents is from north-northwest to south-southeast so that the three wet centres shift along this direction from the south of the Equator in the boreal winter to the north of it in the boreal summer. The variations of these wet areas indicate the global seasonal march of the monsoons. It is clear that the region of the Asian-Australian monsoon should include both the Indian Ocean and the central-western Pacific while the region of the American monsoon should include part of the North Atlantic. The African monsoon mostly prevails over the continent only.

3. The African monsoon

The southernmost position of the ITCZ usually appears in January and the northernmost position

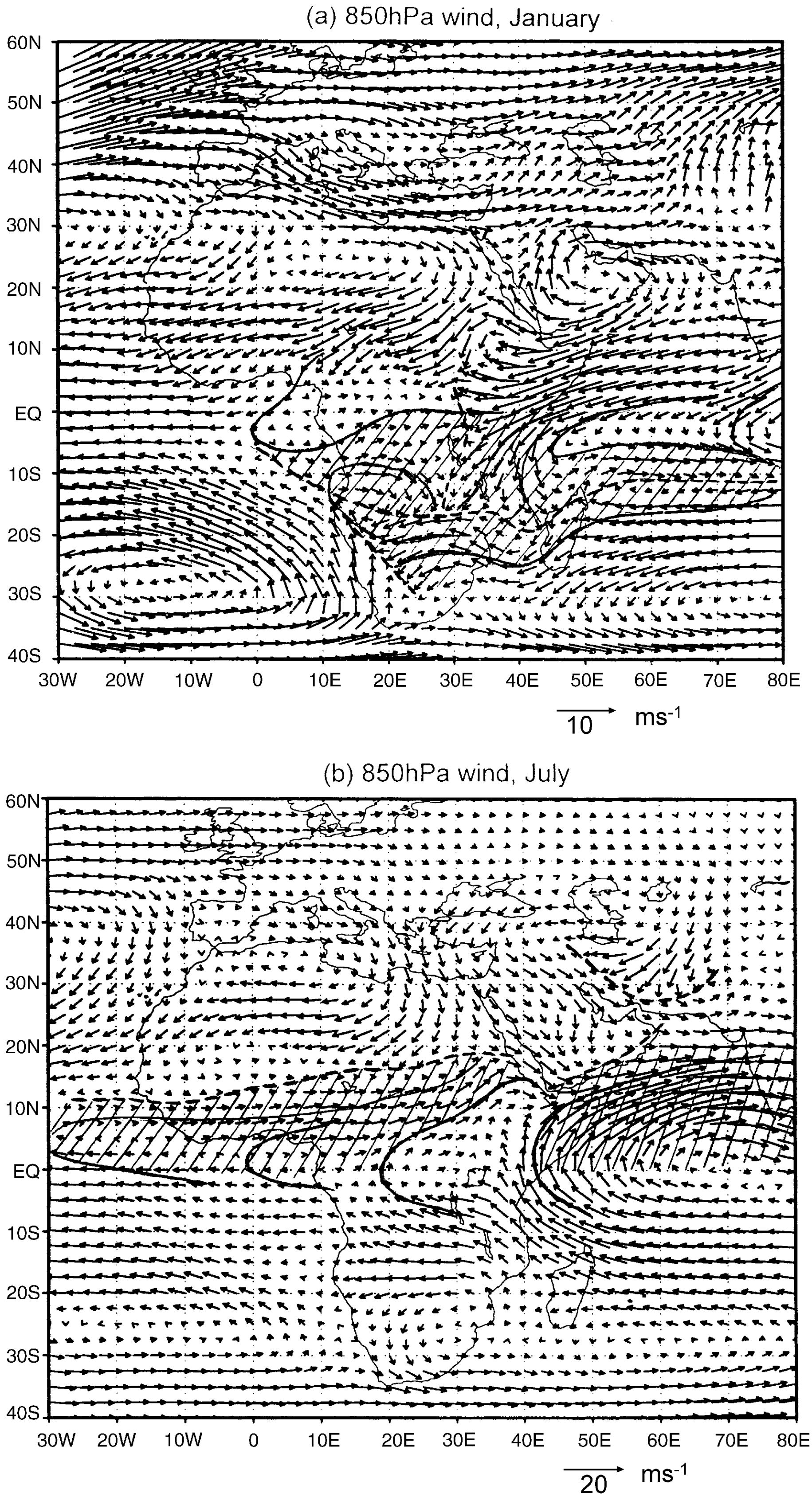

Fig. 2. Climatologically mean 850 hPa winds (ms^{-1}) for the African continent in (**a**) January and (**b**) July based on the NCEP/NCAR reanalysis data from 1950 to 1999. Shaded areas denote the domain influenced by the cross-equatorial flows. The solid lines are streamlines and the dashed lines indicate the ITCZ

in July. Figure 2 shows the climatological mean 850 hPa winds over Africa and the adjacent regions in January and July. From Fig. 2a, a convergence line with several convergence centres stretches between 19° S and 10° S from South Africa through Madagascar to the south of the equatorial Indian Ocean in January. The strong northeast trade winds appear over the Arabian Sea and the southward cross-equatorial flows prevail over the equatorial eastern Africa and the Indian Ocean (see the shadings in Fig. 2a). Only a weak southward cross-equatorial flow can be noted over equatorial western Africa. In January, there are four different flows that

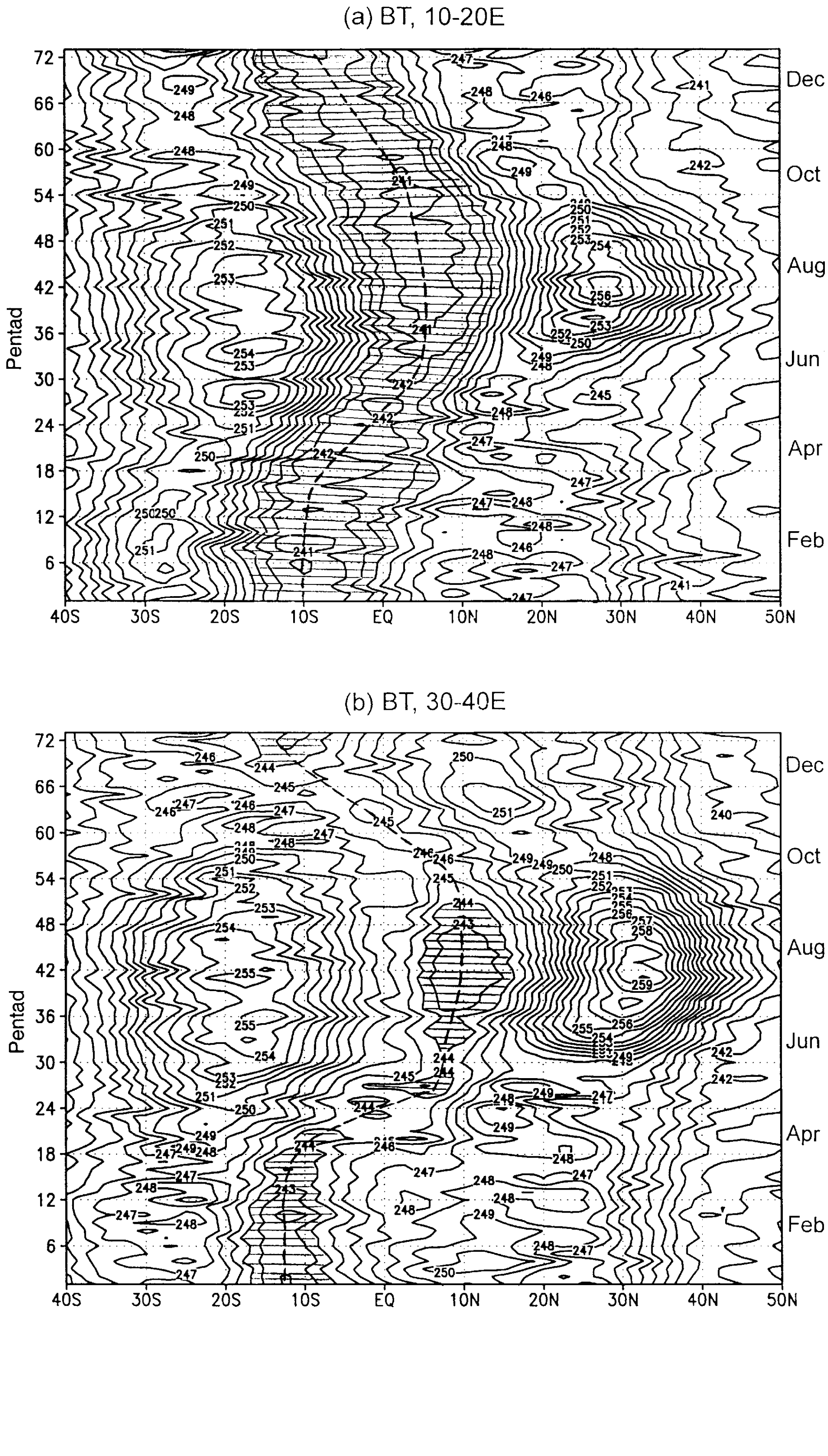

Fig. 3. Time-latitude section of climatological pentad-mean BT (K) along (**a**) 10–20° E and (**b**) 30–40° E based on the HIRS12 data from 1980 to 1995. Shaded areas denote the BT lower than 244 K and dashed lines the maximum of BT

influence South Africa. The wettest flow comes from the equatorial western Indian Ocean. The other three relatively drier flows originate from North Africa, the South Atlantic, and the southern Indian Ocean. It can be noted by comparing Figs. 1a and 2a that the wet belt is to the north of the convergence line. In July, the strong northward cross-equatorial flows are found over the East African coast and they change directions over the Arabian Sea before reaching India. In equatorial central-eastern Africa, an anti-cyclonic circulation can be seen at the 850 hPa level. To the west of the anti-cyclonic circulation, relatively weak northward cross-equatorial flows

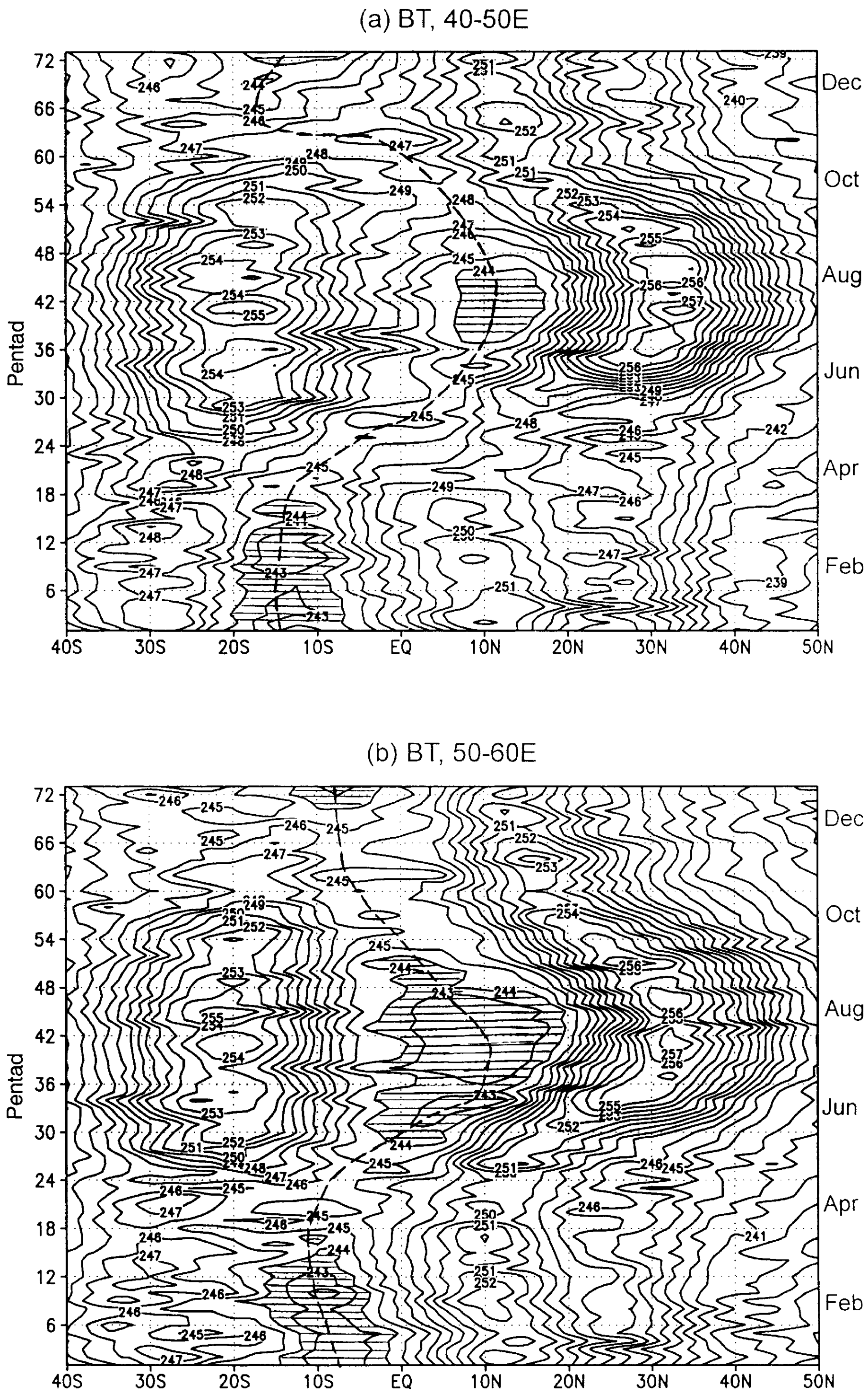

Fig. 4. Same as in Fig. 3 but for (**a**) 40–50° E and (**b**) 50–60° E

are noted along 20° E, 0° E, and 30° W. To the north of the Equator over the Africa continent, the convergence line stretches along around 15° N. To the south of the line, northward cross-equatorial flows prevail but an anti-cyclonic circulation dominates to its north. By comparing Figs. 2b and 1c, it can be seen that the wet centre covers an area between the ITCZ and the southwesterly monsoon flow that influences central Africa. The centre of convergence in Fig. 2 shifts with season along the central-western coast of Africa.

In the boreal winter, southward cross-equatorial flows near the African continent mainly appear along the East Coast. The streamline (heavy solid line) outlines the domain influenced by the cross-equatorial flows. This domain in January coincides basically with that of wet BT shown in Fig. 1a, indicating a linkage between the lower-tropospheric convergence of cross-equatorial flows and the upper-tropospheric water vapour content. According to Fig. 2, the summer monsoon in central-western Africa should be continuously shifting but there is a break in eastern Africa due to an anti-cyclonic circulation in the transition seasons. The shaded domains in the figure indicate not only the seasonal reversal of wind direction but also the influence of the cross-equatorial flow. Some other regions such as Southwest Asia experience reversal of wind direction but the dominant winds do not originate from the cross-equatorial flows.

To depict the expansion of the wet centre over the African continent, we show in Fig. 3 the time-latitude sections of climatological pentad-mean BT along 10–20° E and 30–40° E from January to December. In Fig. 3a, the dry/wet spell transition takes place in the latitudes between 19° S and 14° N along the longitude band 10–20° E. The wet maximum in central-western Africa continuously shifts across the Equator from 10° S to 10° N. In mid-February the wet boundary extends to the southernmost position and, in the boreal summer, it reaches the northernmost position. Different from the BT distribution in central western Africa, the wet maximum along the longitudes of 30–40° E does not always reach the threshold BT value of 244 K (Fig. 3b). It is noted that the seasonal migration of deep convection along eastern Africa is terminated in the two transient seasons of spring and autumn. In Fig. 3b, wet areas cover 8° S–18° S from December to mid-April and 4° N–16° N from June to mid-September. As the deep convection appears to the north of the Equator in July and August, strong descending motion is found over the subtropical northern and southern Africa.

Figure 4 shows the time-latitude sections of climatological pentad-mean BT distributions along the East Africa coast (40–50° E and 50–60° E). Wet convections appear within 20° S–5° S from January to March and 6° N–18° N from early July to mid-August (Fig. 4a). Between 5° S and 6° N, there is no dry/wet spell transition. Along this longitude band, the spell of wet convection is short, compared with other longitudes over Africa, and it is accompanied by the weak cross-equatorial flow over northeast Africa. Over the western Indian Ocean (50–60° E), dry/wet spell transition occurs between 19° S and 20° N (Fig. 4b). Two terminations take place in April–May and in October–December. The strengthened wet centre can be noted to the north of the Equator during the boreal summer (Fig. 4b), which reflects strong northward cross-equatorial flows over the western Indian Ocean.

A feature associated with Africa monsoon's seasonal expansion is that the area of dry/wet spell transition has extended northward and moved away from the area affected by the northward cross-equatorial flows by the boreal summer. However, in winter, the southward cross-equatorial flows can affect the more southern region than the wet upper-tropospheric water vapour. The changes in location and intensity of the wet centre are determined by the cross-equatorial flows.

4. The American monsoon

Figure 5 displays the climatological means of 850 hPa winds over North and South America in January and July. In January, the southward cross-equatorial flows mainly prevail over the mid-low latitudes of South America. In southwestern South America, a convergence line separates the anti-cyclonic circulation over the Southeast Pacific from the southward cross-equatorial flow. Another anti-cyclonic circulation over the South Atlantic influences Southeast Brazil. The southward cross-equatorial monsoon flow can be clearly found in the deltatic area in South America (Fig. 5a).

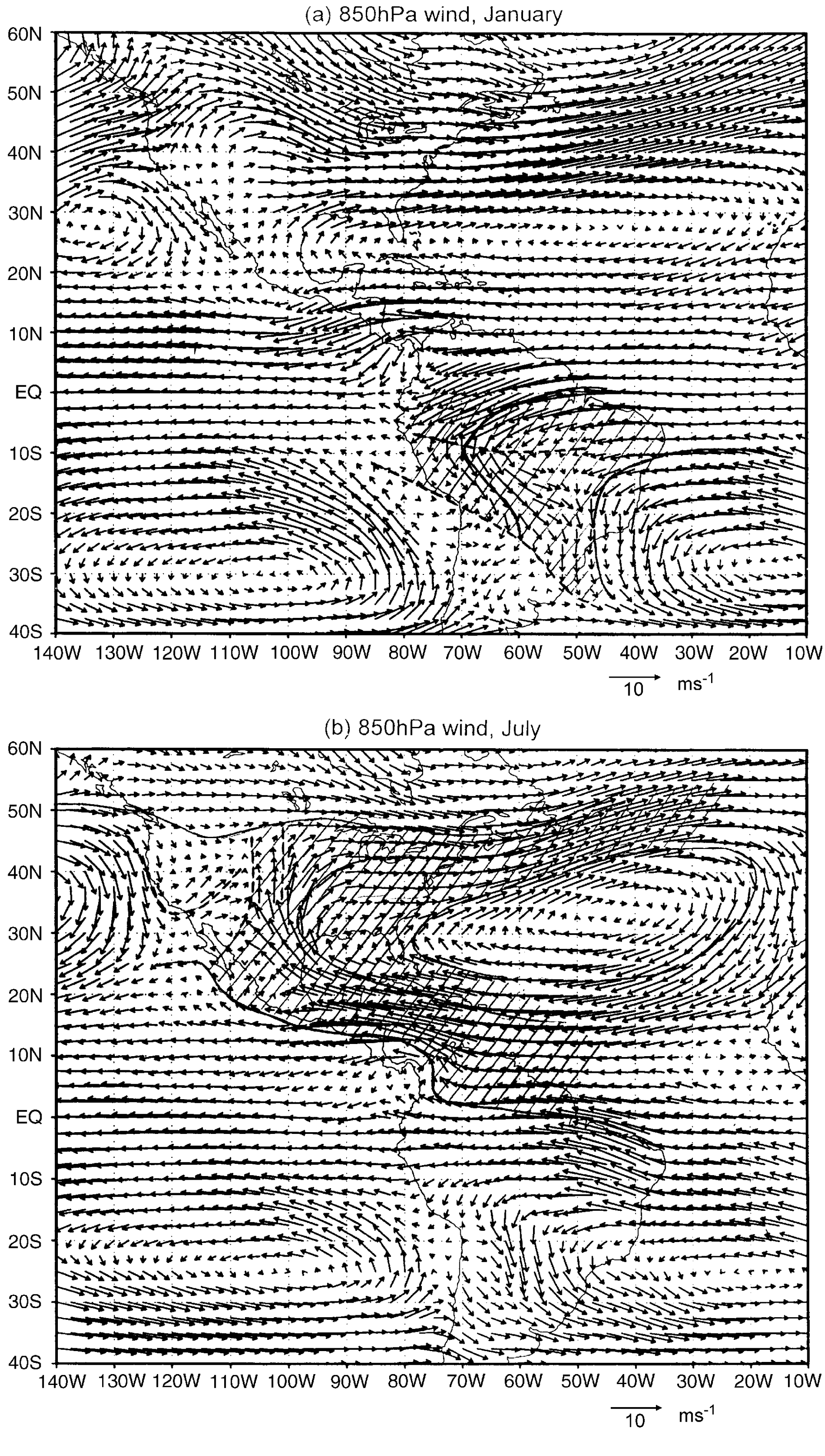

Fig. 5. Same as in Fig. 2 but for the South and North American continents. Dashed lines in the central North America indicate the convergence and divergence axes

In the boreal summer, the northward cross-equatorial flow over equatorial South America is not as strong as that in other seasons and for other continents. However, the warm-wet air is mainly concentrated over the warm pool of the tropical western Atlantic and the Caribbean Sea. Thus, the strong southeasterly winds over the Caribbean Sea are the warm-wet flows. The flow turns its direction southwesterly over North America and over the mid-latitude Atlantic. The westerly flow dominates north of 47° N and the western part of the North American continent. The air mass with warm-wet flows is different from that over the subtropical Atlantic (with dry-warm air) and that over northern North America and western North America (with dry-cool air). A convergence line between the warm-wet and dry-cool flows can be noted in western North America near 105° W. A divergence line could be noted along 100° W in North America. To the east of the convergence line, the southwesterly warm-wet flow and the westerly flow may form frontal processes near 45–50° N. From Fig. 5a–b, the flows over central-southern

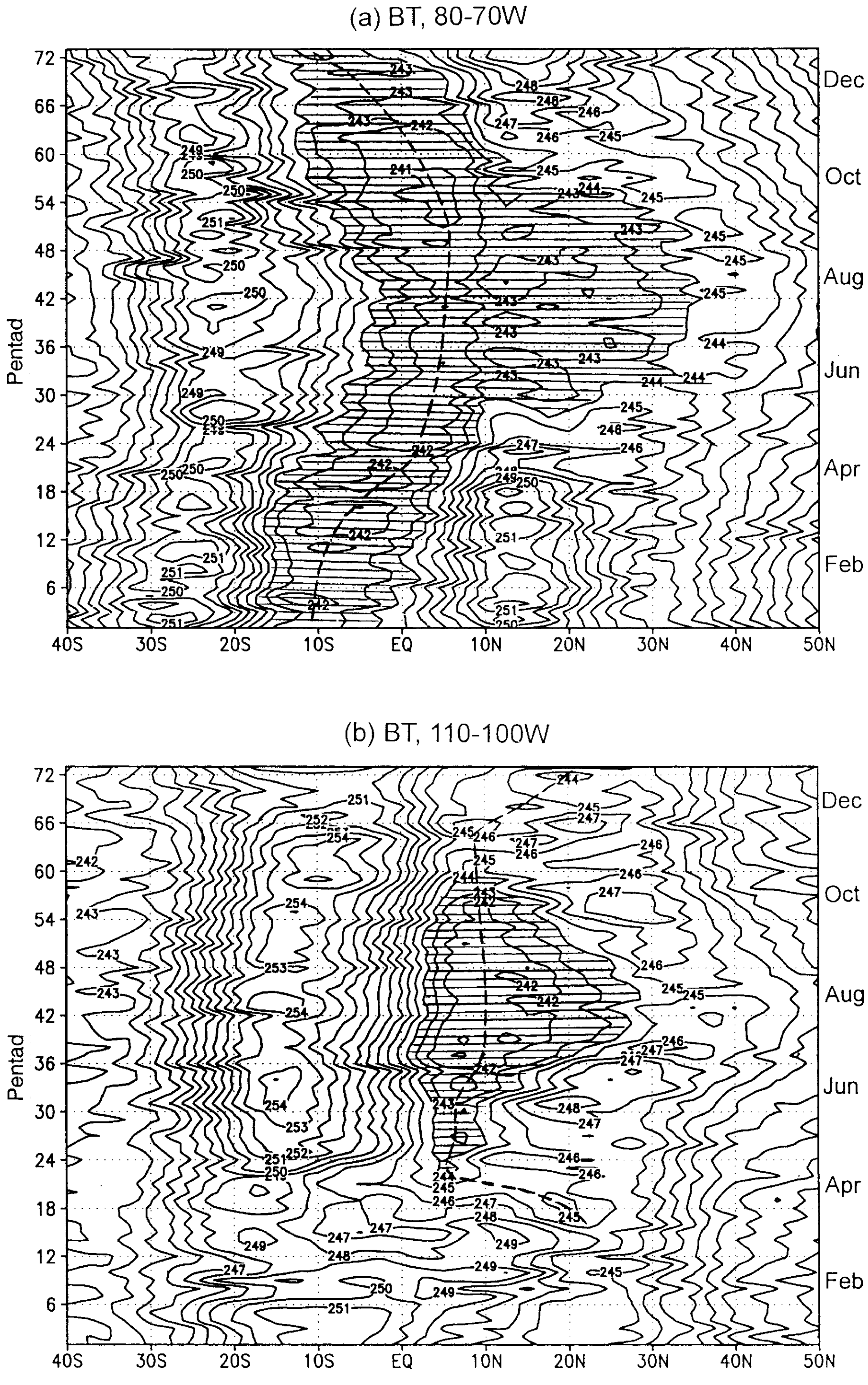

Fig. 6. Same as in Fig. 3 but for (**a**) 80–70° W and (**b**) 110–100° W

North America and central-southern South America do not reverse their direction with season.

To demonstrate the seasonal expansion of water vapour, Fig. 6 shows the time-latitude sections of climatological pentad-mean BT along the longitudes averaged for 80–70° W and 110–100° W. Along 80–70° W (Fig. 6a), the dry/wet spell transition appears from January at 18° S to late July at 35° N. The dry/wet spell transitions over North and South America are not symmetrical relative to the Equator. The dry/wet spell transition in South America gradually advances toward the south and north but to the north of equatorial America the expansion of water vapour with BT lower than 244 K takes place abruptly in two stages in late May and early June. Its retreat in early October is also rather rapid. It can been seen clearly from Fig. 5 that the cross-equatorial flow influences the region of 20° S–45° N along the longitudes of 80–70° W. For the longitudes 110–100° W, the dry/wet spell transition appears only over North America from late April to late October, and at 1° N in early July and 29° N in mid-August.

5. The Asian-Australian monsoon

The strongest monsoon worldwide appears in the Asian-Australian (A-A) region. Figure 7 shows the climatological 850 hPa winds in January and July. In Fig. 7a, the southward cross-equatorial flows can be found between the western Indian Ocean and western Pacific. The solid and dashed lines as well as the shaded areas denote the regions influenced by the southward cross-equatorial flow and the ITCZ. The region affected by the southward cross-equatorial flow appears only near the Equator and to the south. The northern part of Australia is affected by the monsoon flow. In East Asia, strong northwesterly with dry-cool air prevails. In Southeast and South Asia, the northeasterly trade is dominant.

The features shown in Fig. 7b are more complex because of the multiple troughs, wind shears, and convergence/divergence lines. Northward cross-equatorial flows can be noted between the western Indian Ocean and the western Pacific. The three major components over the western Indian Ocean, 105° E, and the western Pacific influence the regions of the Asian continent, the coastal areas, and the western Pacific. The northward cross-equatorial flow over the western Indian Ocean turns its direction over the Arabian Sea into the southwesterly flow that influences South Asia, Southeast Asia, and the Southeast Tibetan Plateau. In the northern South China Sea, the southwesterly flow changes direction again into southerly, influencing central-eastern China, Northeast China, and the Korean Peninsular. The northward cross-equatorial flows over Indonesia directly affect the Philippines and Japan, while the northward cross-equatorial flow over the western Pacific travels to the northwest Pacific.

More complex monsoon flows can be noted over China. In summer, southerly flows are dominant over eastern China. According to Li and Qu (1999), the monsoon circulation comes from the cross-equatorial flows. Here, we adopt this view to determine the locations and seasonal transitions of local monsoons. Along 110° E, southerly flows are strong, but there exists a meridional divergence line (heavy dashed line) over central China. This divergence line is perfectly consistent with the dry zone and divides eastern China into two parts. The south-southwesterly winds prevail in the eastern part. In the west (100° E), a meridional convergence line (dashed line) appears. This convergence line is just along the eastern edge of the Plateau and coincides with the meridional rainfall maximum. Southeasterly and southwesterly winds prevail to the east and west of the convergence line respectively. Some southwesterlies, which come from the northern Bay of Bengal (BOB), may be tracked back to the northward cross-equatorial flows. Over the central Plateau, there is a zonal shear line with northeasterlies to its north and southwesterlies to its south. Although the elevation of the central Tibet Plateau exceeds 1500 meters, the flow pattern is consistent with results from previous studies using observed surface winds (Tang et al., 1979; Bao, 1987). The westerly system is found to the north of the Tibetan Plateau. In Northeast China, the monsoon southwesterlies converge with the westerly flows.

It can been noted from Fig. 7a–b that the flow direction is different between the boreal winter and summer in all regions (see the shadings in the figure). Completely opposite directions can be found in South Asia, Southeast Asia, and

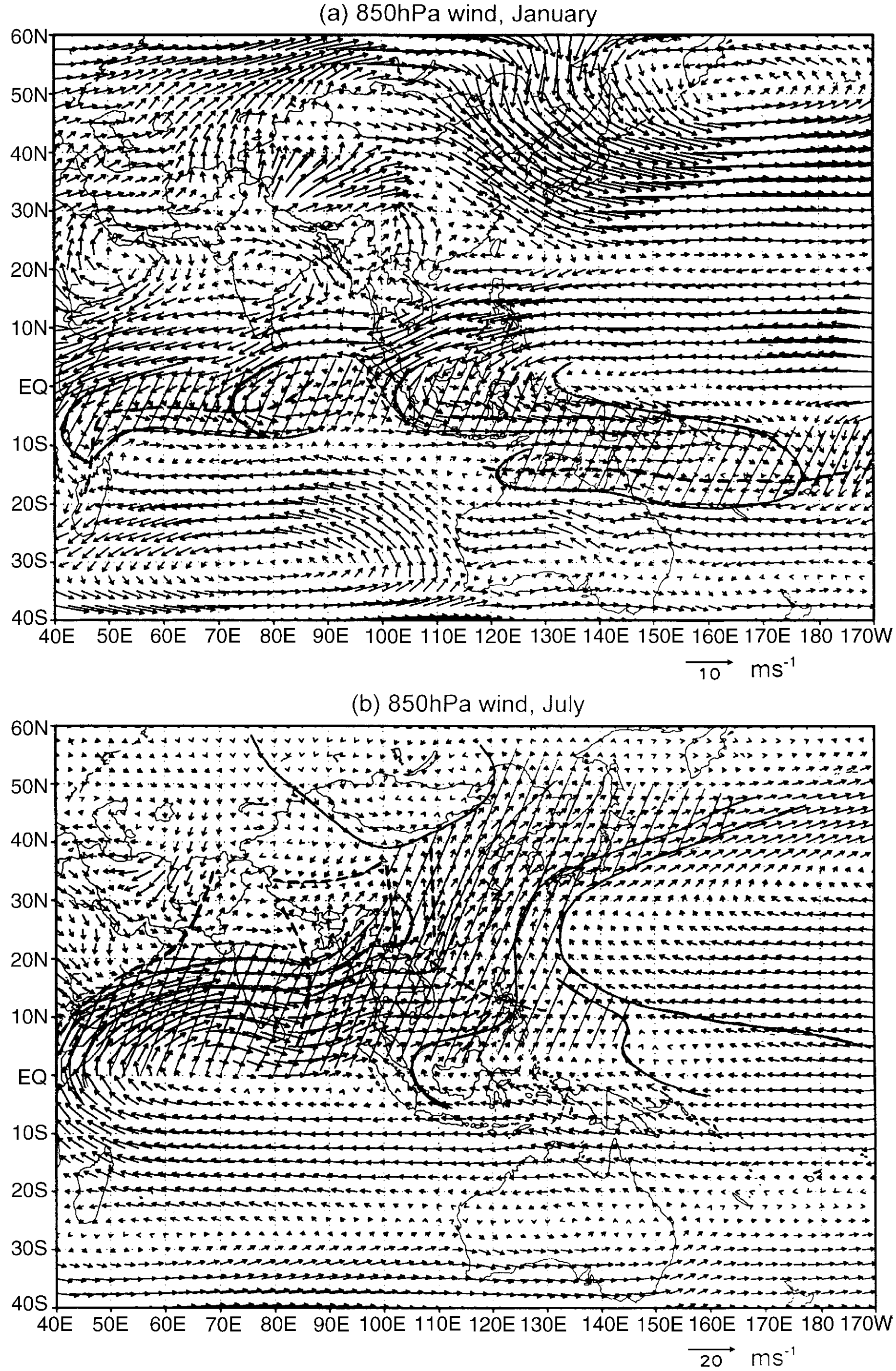

Fig. 7. Same as in Fig. 2 but for the Asian–Australian region. Dashed lines in central China indicate the convergence and divergence axes

Indonesia. In East Asia, the wind direction changes about 90° from northwesterly in winter to southwesterly in summer. In Southeast and East Asia, the southwesterly is merged with the northward cross-equatorial flow. These regions are also influenced by the subtropical anti-cyclone over the northwestern Pacific and the westerly flow at mid-high latitudes. Obviously, the convergence line, the trough line, the shear line, and the divergence line distinguish four monsoon sub-divisions. The South Asian monsoon system exists over the Arabian Sea and India. The Southeast Asian monsoon system ranges from the Bay of Bengal to Indochina and the southern South China Sea (SCS). The East Asian monsoon system includes the monsoon phenomena over the northern SCS, eastern China, the Korean Peninsula, and Japan. The Tibetan Plateau monsoon system occurs to the west of the divergence line along 110° E. It is interesting to note that in the boreal summer, similar circulation features exist between the East Asia-Tibetan Plateau and North American monsoons. For example, a pair of divergence and convergence lines can be found over central-western China and central-western North America.

To identify the dry/wet spell transition over the A-A region, we show in Fig. 8 the time-latitude

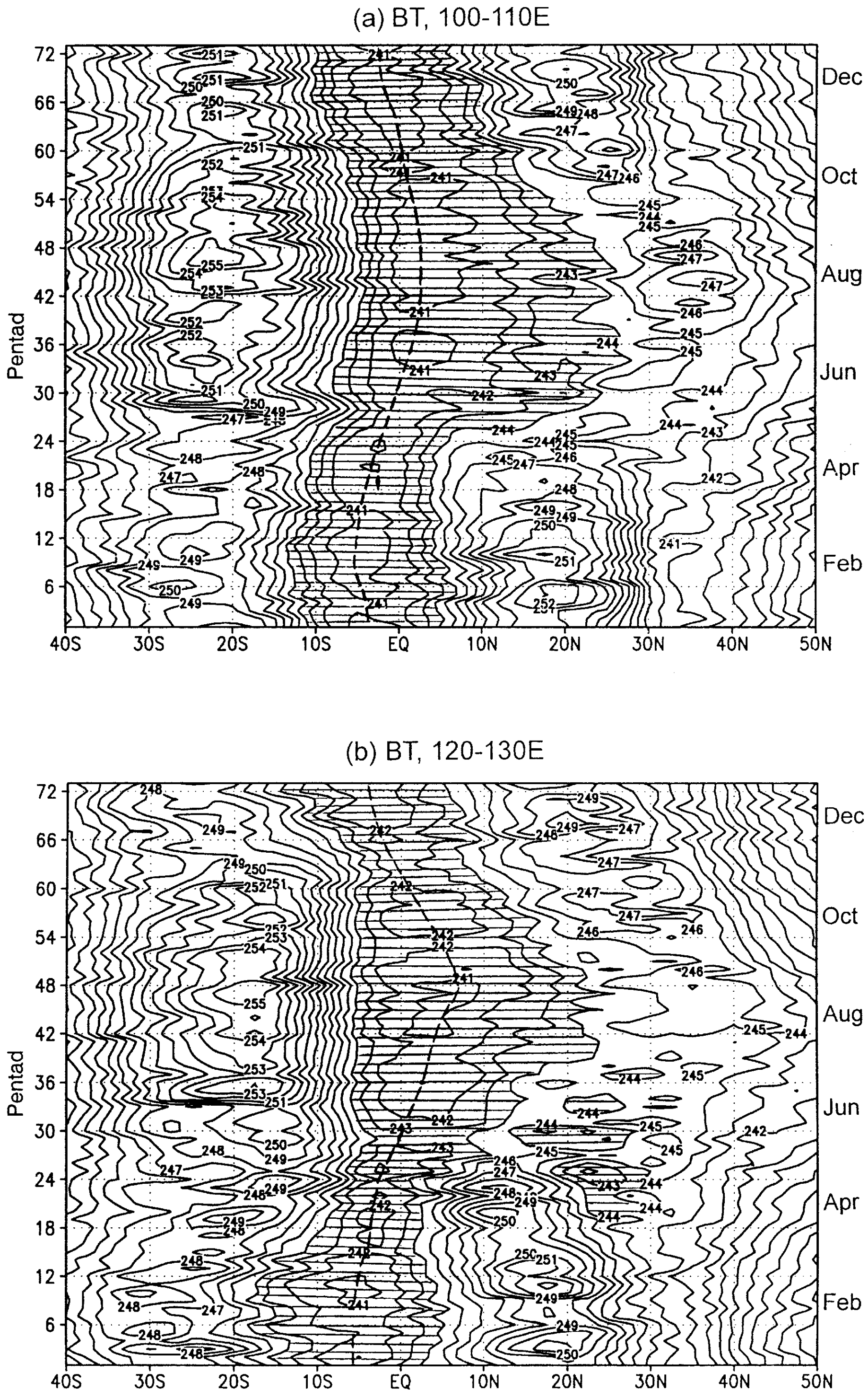

Fig. 8. Same as in Fig. 3 but for (**a**) 100–110° E and (**b**) 120–130° E

sections of climatological BT averaged within 100–110° E and 120–130° E. It is noted that there is no dry/wet spell transition between 5° S and 5° N near the Equator. Recently, several studies have showed that the earliest summer monsoon onset in Asia occurs not over the SCS region but over the land of the Indochina Peninsula and the eastern coast of the Bay of Bengal (Matsumoto, 1997; Wu and Zhang, 1998). Over the SCS, sudden onset of the summer monsoon has been found in mid-May or the fourth pentad of the month by previous studies (Lau and Yang, 1996, 1997; Xie, et al., 1998) based on different data sets. In Fig. 8a, the sudden onset of the wet area appears in early May over the Indochina Peninsula and it retreats gradually from mid-August. A similar expansion can be found over South Asia (India) but its onset date is after that of the SCS summer monsoon.

The expansion of the tropical wet area also gradually starts from the Philippines in mid-May (Fig. 8b). The wet BT (the shaded area between 15° N and 30° N) in April and May represents the subtropical westerly frontal convection during the period of pre-rainy season from South China to the East China Sea (Bao, 1987). It can been shown, by comparing Fig. 8b with Fig. 7b, that the southwesterly monsoon flow affects North-east China, but the BT fails to indicate the monsoon characteristics at the relatively high latitudes. In Northwest Australia, the south-north width of summer monsoon is about 12 degrees of latitude between December and April.

6. Discussion of the global monsoon

In this study, we have described some basic characteristics of the global summer monsoon using 850 hPa wind and upper-tropospheric water vapour band BT. Areas of BT lower than 244 K are consistent with those influenced by the cross-equatorial flows over many regions in the world. To improve understanding of the distributions of both the cross-equatorial flows and BT, we show in Fig. 9 some synthesized characteristics. Three deep convective or relatively wet areas at the tropical upper troposphere can be noted over Africa, the A-A region, and the South American continent, separately. The largest wet area is located over the equatorial eastern Indian Ocean and the equatorial western Pacific and its centre lies over the equatorial eastern Indian Ocean. As mentioned in the introduction, a monsoon should exhibit the dry/wet spell alternation. However, some areas within the wet centres should not be classified a monsoon region because they remain wet all the year round. On the other hand, the cross-shaded areas may shift in location and change in extension with season. As described from sections 3–5, the areas of wet centres or low values of BT shift from the northernmost position in July to the southernmost location in January. According to the criterion for the wet convection measured by BT, the wet areas with monthly mean BT lower than 244 K in January and July are marked with "$-$" and the other areas with "$+$". The areas covered by the heavy solid line display the difference between July and January, indicating the monsoon dry/wet spell transition. Obviously, along the equatorial zone many areas do not experience the wet/dry spell transition. At the mid-low latitudes, there exist six major dry/wet spell transition regions. The largest one appears over the subtropical North Africa–Asia–northwest Pacific. The others are located over subtropical South Africa, Indonesia–Australia, the Southwest Pacific, the Mexico-Caribbean Sea, and subtropical South America. In particular, Asia, India, the Indochina Peninsula, South China, and southern Japan all fall into the dry/wet spell transition regions. These regions experience a dry or wet period lasting more than one month. At the mid-low latitudes, many regions (except the eastern part of the Pacific) are influenced by the cross-equatorial flow in summer. It is found that the domain of dry/wet spell transition is affected by the cross-equatorial flow except for North Africa.

Figure 9 implies that both the cross-equatorial flow and the dry/wet spell transition can be applied to define the monsoon in many regions in the subtropics. However, caution should be taken for the following three regions. The first is located near the Equator where the wind directions alternate because of the hemispheric thermal contrast but there is no dry/wet spell transition. The second includes subtropical North Africa where the dry/wet spell transition can be found but the cross-equatorial flow never reaches. The third is over the middle latitude regions such as North China, South Africa, and the northern North America where cross-equatorial flows

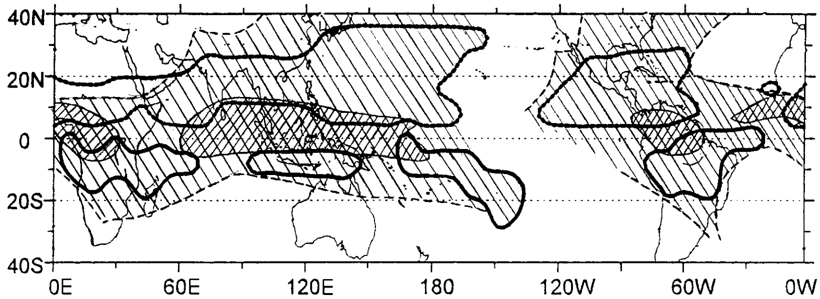

Fig. 9. Global monsoon distributions summarized based on the analysis of cross-equatorial flow reversals and dry/wet spell transition. Three cross-shaded areas near the Equator denote the annual-mean BT < 244 K. Six regions in the subtropics are affected by the dry/wet spell transition. The single-shaded areas represent the regions influenced by cross-equatorial flows

exert an influence but the upper-tropospheric water vapour is not as high as in tropical regions.

In fact, the areas of dry/wet spell transition at the mid-latitudes become larger if pentad instead of monthly means of BT are used. As revealed by pentad data (Qian and Lee, 2000), the northern boundary of the Asian summer monsoon can reach 30° N in South Asia, 35° N in the East Tibetan Plateau, and 37° N in the Korea Peninsula. The isolated areas of dry/wet spell transition in Fig. 9 indicate that the monsoon can be influenced by not only the wet cross equatorial flow but also the local topography or local thermal contrast.

It should also be pointed out that interdecadal variability exists for the features discussed in this analysis. For example, more summer monsoon rainfall occured in northern China during 1950s–1960s while since the mid-1980s more rainfall is concentrated to the south of the Yangtze River (Qian and Zhu, 2001). Furthermore, during 1950s–1960s the southwesterly summer monsoon flow at the 850 hPa reached North China and Northeast China, but since the late 20th century it mainly prevails over the south of the Yangtze River (Qian and Zhu, 2001). Therefore, the global monsoon regions demarcated by BT only represent the later period while the 850 hPa flow shows longer-term monsoon features.

In this paper, the global monsoon regions have been identified using BT and the reversal of cross-equatorial flow. For the definition of summer monsoon, the lower BT and the cross-equatorial flow are considered as two representatives that extend from the wet tropics to subtropical regions. In winter, the dry-cold continental flow extends to the middle and lower latitudes in Asia, mainly in the lower troposphere. Thus, the lower-tropospheric wind originated from the continent can be used to indicate the winter dry-cold monsoon at lower latitudes. On the other hand, the HIRS12 reveals the dry-warm upper-tropospheric atmosphere in winter season, which can be clearly seen from Fig. 1a. Therefore, BT may not be suitable for defining the winter monsoon component.

Acknowledgement

We wish to thank Dr. Song Yang and Prof. S.W. Wang for their help with the revision of the text. We also thank the anonymous reviewers and the editor whose comments and suggestions are helpful for improving the quality of this paper. This research was supported by the Key Innovation Project of Chinese Academy of Sciences (KZCX1-10-07), the NNFC (No. 49825504 & No. 49975023), and the National Key Program for Developing Basic Sciences in China (No. G1999043405).

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