ADVANCES IN ATMOSPHERIC SCIENCES, VOL. 24, NO. 5, 2007, 915–926

Interdecadal Variability of the East Asian Summer Monsoon and Associated Atmospheric Circulations

ZENG Gang*[1] (曾 刚), SUN Zhaobo[1] (孙照渤), Wei-Chyung WANG[2], and MIN Jinzhong[1] (闵锦忠)

[1] *Key Laboratory of Meteorology Disaster, Nanjing University of Information Science & Technology, Nanjing* 210044

[2] *Atmospheric Sciences Research Center* (*ASRC*)*, State University of New York at Albany, Albany* 12203*, USA*

(Received 7 June 2006; revised 9 February 2007)

ABSTRACT

Based on the National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP/NCAR) reanalysis data from 1950–1999, interdecadal variability of the East Asian Summer Monsoon (EASM) and its associated atmospheric circulations are investigated. The EASM exhibits a distinct interdecadal variation, with stronger (weaker) summer monsoon maintained from 1950–1964 (1976–1997). In the former case, there is an enhanced Walker cell in the eastern Pacific and an anti-Walker cell in the western Pacific. The associated ascending motion resides in the central Pacific, which flows eastward and westward in the upper troposphere, descending in the eastern and western ends of the Pacific basin. At the same time, an anomalous East Asian Hadley Cell (EAHC) is found to connect the low-latitude and mid-latitude systems in East Asia, which strengthens the EASM. The descending branch of the EAHC lies in the west part of the anti-Walker cell, flowing northward in the lower troposphere and then ascending at the south of Lake Baikal (40°–50°N, 95°–115°E) before returning to low latitudes in the upper troposphere, thus strengthening the EASM.

The relationship between the EASM and SST in the eastern tropical Pacific is also discussed. A possible mechanism is proposed to link interdecadal variation of the EASM with the eastern tropical Pacific SST. A warmer sea surface temperature anomaly (SSTA) therein induces anomalous ascending motion in the eastern Pacific, resulting in a weaker Walker cell, and at the same time inducing an anomalous Walker cell in the western Pacific and an enhanced EAHC, leading to a weaker EASM. Furthermore, the interdecadal variation of summer precipitation over North China is found to be strongly regulated by the velocity potential over the south of Lake Baikal through enhancing and reducing the regional vertical motions.

Key words: East Asian summer monsoon, interdecadal variability, Walker circulation, East Asian Hadley circulation

DOI: 10.1007/s00376-007-0915-y

1. Introduction

The Asian summer monsoon is a huge monsoon system consisting of two subsystems: the South Asian (or Indian) Summer Monsoon and the East Asian Summer Monsoon (EASM) system. Each has its own unique characteristics, and, at the same time, interacts with each other (Tao and Chen, 1987). The EASM, which is characterized by southerly winds at lower levels due to the land-sea thermal contrast, is distinct from the South Asian Summer Monsoon and has more complex space-time structures (Lau and Yang, 2000). Numerous studies have revealed that the EASM is one of the most important factors affecting precipitation in eastern China, Korea and Japan (Huang, 2004). The variations of the EASM, particularly associated with drought and flood events, exert strong influences upon the national economy, especially agricultural production, and the daily life of inhabitants living in these regions. In the early 1930s, Zhu (1934) first revealed that floods and droughts over East China could be attributed to the anomalous intensity of the EASM. Since then, the variations of the EASM and its relationship with floods and droughts have been important scientific issues for climate scientists.

Many studies have shown that the EASM exhibits

*Corresponding author: ZENG Gang, zenggang@nuist.edu.cn

a multi-timescale variability, from intraseasonal, interannual, and interdecadal, to quasi 80-year oscillations (Zhang, 1999; Zhu and Wang, 2001; Li et al., 2002, 2004; Huang et al., 2004; Lu et al., 2004). Interdecadal variability of the monsoon is a background for the seasonal and interannual variations. Therefore, a better understanding of the interdecdal variability may improve the predictability of seasonal-to-interannual monsoon variability. Recently, as revealed by numerous studies, the EASM and its related climate in China underwent a change in the mid-late 1970s, concurrently with the major climatic change in the tropical Pacific Ocean (Chang et al., 2000a,b; Wang, 2001; Wu and Wang, 2002; Lu, 2002; Guo et al., 2004). In Guo et al. (2004), it is revealed that the EASM experienced two interdecadal changes in the past 50 years, separately, in the mid 1960s and the mid 1970s. When the EASM is strong (weak), summer rainfalls over North China will increase (decrease), and decrease (increase) in the mid-lower Yangtze River valley. It is suggested that the summer rainfall over the Yangtze River valley has experienced a significant turning towards a wetter climate since 1979, associated with the EASM weakening and southward movement of the northwestern Pacific subtropical high (He and Gong, 2002; Zhang and Wu, 2001). Nitta and Hu (1996) investigated the spatial and temporal variations of summer rainfall in China. The primary mode shows strong signals of seesaw between the Yangtze River valley and the region to the north of it. The time coefficients of their first EOF mode also showed a jump-like change around 1979. Huang et al. (1999) analyzed the interdecadal variations of summer precipitation in China and the aridification trend in North China by using the observations of summer precipitation at 336 stations in China and SST in the Pacific from 1951–1994. Their results indicate that the climate in China underwent a turning twice (in the mid 1960s and the late 1970s) based on the variation of summer precipitation, which is attributed to the interdecadal variation in SST over the eastern tropical Pacific. Examining summer rainfall in Korea, Ho et al. (2003) revealed that the regional mean of 700-hPa geopotential heights within central-eastern Asia had a sudden increase during the Northern Hemisphere summer in the mid 1970s and an abrupt change in summer rainfall in Korea during the late 1970s. All these results indicate that the interdecadal change of the summer rainfalls over the Yangtze River valley and North China in the late 1970s is not a local phenomenon, but related to the large-scale rainfall regime change.

So what might be the causes of this interdecadal shift of the EASM? This issue has received increasing attention in the last two decades. The causes should include, in the main: (1) decadal change of the spring snow depth over the Tibetan Plateau in the mid 1970s, which may significantly impact the EASM (Zhang et al., 2004; Huang et al., 2006); (2) decadal variations of the Arctic Oscillation (AO) and land-sea thermal contrast, which play an important role in decadal variations of the EASM (Ju et al., 2006); (3) the blocking high in the vicinity of Lake Baikal and China-Mongolia, which is also an important factor for variation of the EASM (Li et al., 2002; Liu and Guo, 2005); (4) precipitation trends in China over recent decades, with sufficiency (deficit) rainfall in the south (north) of China, which might be related to the increased content of black carbon aerosols in the past 20 years (Menon et al., 2002); (5) monsoon variability, in particular that may be bound up with the interdecadal variation of El Niño events in the eastern tropical Pacific after the mid 1970s (Huang, 2001); and (6) the 500-hPa western Pacific subtropical high, which is influenced by the equatorial Pacific SST, and is a crucial factor to this interdecadal shift of the EASM and summer climate over China (Chang et al., 2000a,b). Most of these results have revealed that sea surface temperature anomaly (SSTA) in the eastern tropical Pacific is a very important factor for variation of the EASM. However, exactly how the eastern tropical Pacific SSTA exerts impacts upon the EASM remains poorly investigated.

The purpose of this study is to explore a possible mechanism for interdecadal variation of the EASM and to investigate, separately, how the Walker, anti-Walker and Hadley cells vary during the strong and weak EASM periods. The paper is organized as follows:

In section 2, a description of the datasets is presented. In section 3, the definition of the EASM index and its relationship with velocity potential are described. Section 4 discusses the atmospheric circulations of the Walker and East Asian Hadley cells during strong and weak EASM. Interdecadal variability of summer precipitation over North China and its possible causes are investigated in section 5. Section 6 briefly explores the SSTA distribution during strong and weak EASM. Finally, a discussion and conclusions are given in section 7.

2. Data and methodology

Several datasets are used in this study. The National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis data (Kalnay et al., 1996) from January 1950 to December 1999 are used, in which fields are extracted for 50 summer periods (June to August) from

1950–1999. Variables used are monthly horizontal wind and isobaric vertical velocity at 12 levels: 1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, and 100 hPa at a $2.5^\circ \times 2.5^\circ$ resolution. The sign of the isobaric vertical velocity is changed to positive to give values representing the upward motion of air parcels, and for discussing variation of the EASM, sea level pressure fields are also extracted from the NCEP/NCAR reanalysis data to compute the EASM index.

Monthly mean rainfall data of 26 out of 160 stations in China during June, July, and August (JJA) from 1951–1999 are used—compiled by the China Meteorological Administration—for indicating the variations over North China. Twenty-six stations have been chosen from Huang et al. (1999), including Chengde, Beijing, Tianjin, Shijiazhuang, Dezhou, Xingtai, Anyang, Yantai, Qingdao, Weifang, Jinan, Linyi, Heze, Zhengzhou, Changzhi, Taiyuan, Linfen, Chaoyang, Chifeng, Zhangjiakou, Hohhot City, Yulin, Yan'an, Xuzhou, Xinpu, and Xi'an.

Monthly SST data are also used, taken from the NOAA extended reconstructed sea surface temperature (ERSST) dataset on a $2^\circ \times 2^\circ$ grid from January 1950 to December 1999 (see Smith and Reynolds, 2004).

According to the Helmholtz theorem (Holton, 1992), horizontal wind velocity can be divided into two components, one being nondivergent (or rational) wind, and the other being divergent (or nonrational) wind (Krishnamurti, 1971; Krishnamurti et al., 1973).

$$V = V_\Psi + V_\Phi = \boldsymbol{k} \times \nabla\Psi + \nabla\Phi\ , \tag{1}$$

where Ψ is a streamfunction, and Φ is a velocity potential.

In Cartesian coordinates,

$$u_\psi = -\frac{\partial\Psi}{\partial y}, v_\Psi = \frac{\partial\Psi}{\partial x}\ , \tag{2}$$

$$u_\Phi = \frac{\partial\Phi}{\partial x}, v_\Phi = \frac{\partial\Phi}{\partial y}\ , \tag{3}$$

and thus:

$$\xi = \Delta\psi\ , \tag{4}$$

$$D = \Delta\Phi\ , \tag{5}$$

where ξ is vorticity and D is divergence.

The first part of Eq. (1) does not contribute to atmospheric divergence associated with atmospheric vertical motion, however the second part of Eq. (1) is of particular importance in the tropics, where it involves the famous Walker and Hadley circulations, which are thermal driven circulations in relation to atmospheric convergence/divergence. For this reason, we address the atmospheric cell(s) by means of isobaric vertical velocity and the divergent component of horizontal winds.

3. EASM variability and its relationship with velocity potential

To discuss the variation of the EASM, it is necessary to define a monsoon index that measures its intensity. Many researchers have attempted to describe the strength of the EASM using various variables and criteria (Guo, 1983; Webster and Yang, 1992; Shi et al., 1996; Qiao et al., 2002; Huang, 2004), however there are both advantages and disadvantages in their definitions (Huang, 2004). In this study, Shi et al. (1996) definition of the EASM index is adopted, which measures the intensity of the EASM with the difference of sea level pressure between the Eurasian continent and the North Pacific Ocean. The EASM index (hereafter SMI) is defined as:

$$\mathrm{SMI} = N_{\mathrm{or}}\left\{\sum_{\varphi=20}^{50}\left[\overline{P}(110,\varphi) - \overline{P}(160,\varphi)\right]\right\}\ , \tag{6}$$

where $\overline{P}(110,\varphi)$ and $\overline{P}(160,\varphi)$ are the standardized seasonal means of sea level pressure at a gridpoint with latitude φ and longitude 110°E and 160°E, respectively. $N_{\mathrm{or}}(X)$ means "normalized X". Here, φ is set from 20°N to 50°N with an interval of 2.5°. According to this definition, a positive (negative) SMI represents a weak (strong) EASM.

Figure 1 depicts a distinct interdecadal variation of the EASM. Based on the SMI, a strong EASM is maintained from the 1950s to the mid 1960s and a weak EASM from the mid 1970s to the 1990s. Therefore, the 1950–1964 (1976–1997) periods were chosen as a strong (weak) EASM stage for the composite analysis below.

It is well known that Walker and Hadley cells are thermal driven circulations associated with atmospheric convergence/divergence. For documenting the relationship between the EASM and Walker and Hadley circulations, it is necessary to investigate the variability of velocity potential that indicates atmospheric convergence/divergence. EOF analysis is first used to explore the variation of velocity potential at the 850-hPa level in boreal summer (see Figs. 2a and 2b). The first mode explains 41.6% of the total variance and exhibits a wave-2 structure in the tropics. From a global view, the distributions over the eastern and western hemispheres are almost opposite. In the western hemisphere, the maximum center is located around the eastern tropical Pacific and the Amazon Basin. For the eastern hemisphere, the distribution

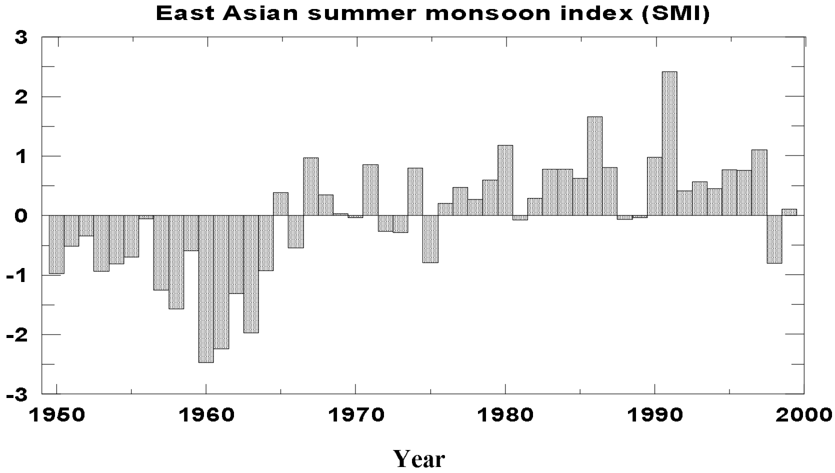

Fig. 1. Time series of the SMI from 1950–1999.

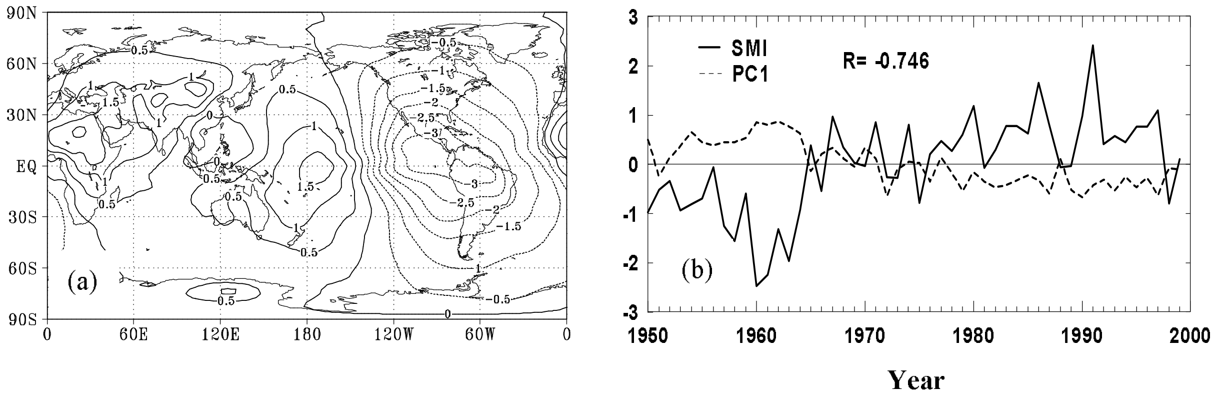

Fig. 2. (a) The first mode of summer velocity potential at 850 hPa, and (b) its principle component (PC1) and SMI (solid line=SMI; dashed line=PC1).

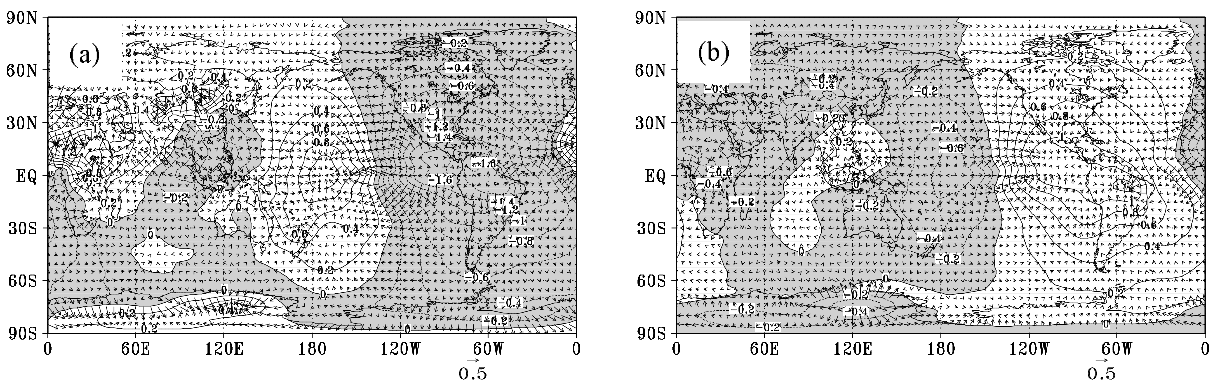

Fig. 3. Composites of anomaly velocity potential (10^6 m^2 s^{-1}) and divergent wind (m s^{-1}) at 850 hPa during (a) the strong EASM (1950–1964) and (b) the weak EASM (1976–1997). The contour is 0.2×10^6 m^2 s^{-1}. The shaded area is negative velocity potential.

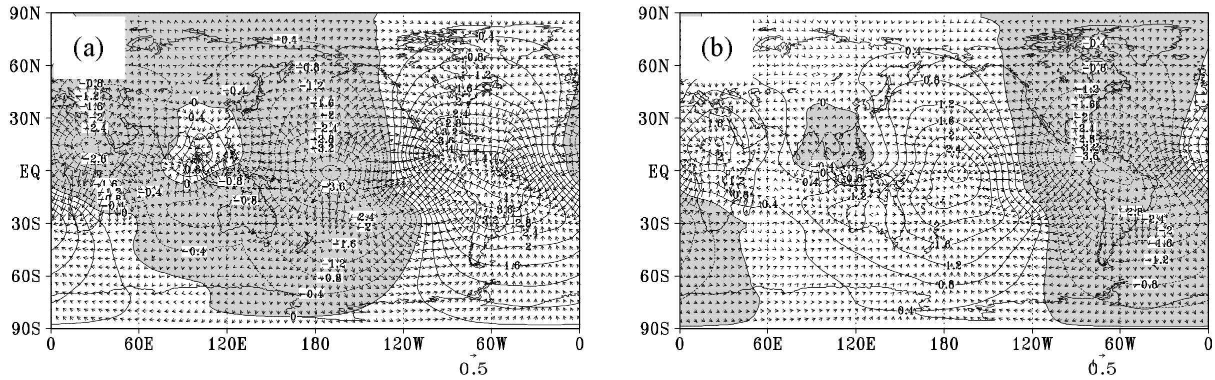

Fig. 4. Same as Fig. 3 but for 200 hPa. The contour is 0.4×10^6 m^2 s^{-1}.

of velocity potential anomalies over the South China Sea is opposite to that in Africa, Europe, Asia, Australia, and the central Pacific. Three centers are located in western Africa, the south of Lake Baikal, and in the central tropical Pacific, separately. In the Asian-Pacific region, a wavetrain may propagate from the eastern tropical Pacific to East Asia according to the anomalous pattern (Fig. 2a). The temporal evolution of the first mode of summer velocity potential is shown in Fig. 2b. There is a distinct declining trend of the first principal component (PC1), which shows a shift in the 1970s. To discuss the relationship between summer velocity potential and the EASM, their correlation coefficient has been calculated, which reaches -0.746 and exceeds the 99.9% confidence level, thus indicating a very close relationship between them.

In order to discuss the convergence/divergence distribution in the strong and weak EASM periods, 1950–1964 summers (strong) and 1976–1997 summers (weak EASM) have been chosen for the composite analysis based on the variation of the EASM, respectively. Thus, corresponding anomalous velocity potentials and divergent winds in the upper and lower troposphere are shown in Figs. 3 and 4, respectively. At 850 hPa, the anomaly pattern of velocity potential in 1950–1964 exhibits a wave-2 structure in the tropics, i.e. two convergent (divergent) centers are located separately over the central tropical Pacific and western Africa (the eastern Pacific to the Amazon Basin and the South China Sea), and the divergent center over the South China Sea is weak. Over East Asia, the anomalous convergent center is located in the south of Lake Baikal and an anomalous divergent center in the South China Sea, which, in combination, produces an anomalous strong monsoon circulation. At 200 hPa, the divergent/convergent centers correspond fairly well to their counterparts at 850 hPa, except that the center over the south of Lake Baikal disappears in the upper troposphere. The opposite behavior exists during the weak EASM (1976–1997). This configuration of convergence/divergence in the lower and upper troposphere suggests that several anomalous vertical circulations are present over the tropical Pacific Ocean and East Asia, which will be discussed in the following section.

4. Walker and East Asian Hadley cells

4.1 *Walker cell*

Walker circulation refers to the zonally-oriented overturning circulation cell in the equatorial Pacific with westerlies at upper levels and easterlies at the surface (Bjerknes, 1969). Many studies have revealed that the South Asian summer monsoon-ENSO relationship through Walker circulation changed obviously around the late 1970s (e.g., Kinter et al., 2002; Krishnamurthy and Goswami, 2000). However, the relationship between Walker circulation and the EASM has been poorly investigated. Hence, the variation of Walker circulation during strong and weak EASM periods will be discussed in this section.

To better understand anomalous atmospheric circulation associated with Walker circulation, the climatic mean summer atmospheric cell (averaged from 1950–1999) is given in Fig. 5a. The air ascends in the western and central tropical Pacific, flows eastward in the upper troposphere, sinks in the eastern tropical Pacific, and then returns to the western Pacific in the lower troposphere.

Anomalous Walker cells of the tropical Pacific during the strong EASM and weak EASM are presented in Figs. 5b–c. In this study, the averaged zonal component of divergent wind and isobaric vertical velocity between 5°S and 5°N are used to describe the Walker cell. During the strong EASM period, anomalous ascending motion of air parcels occur in the central tropical Pacific, flowing eastward and westward in the upper troposphere, sinking in the eastern tropical Pacific

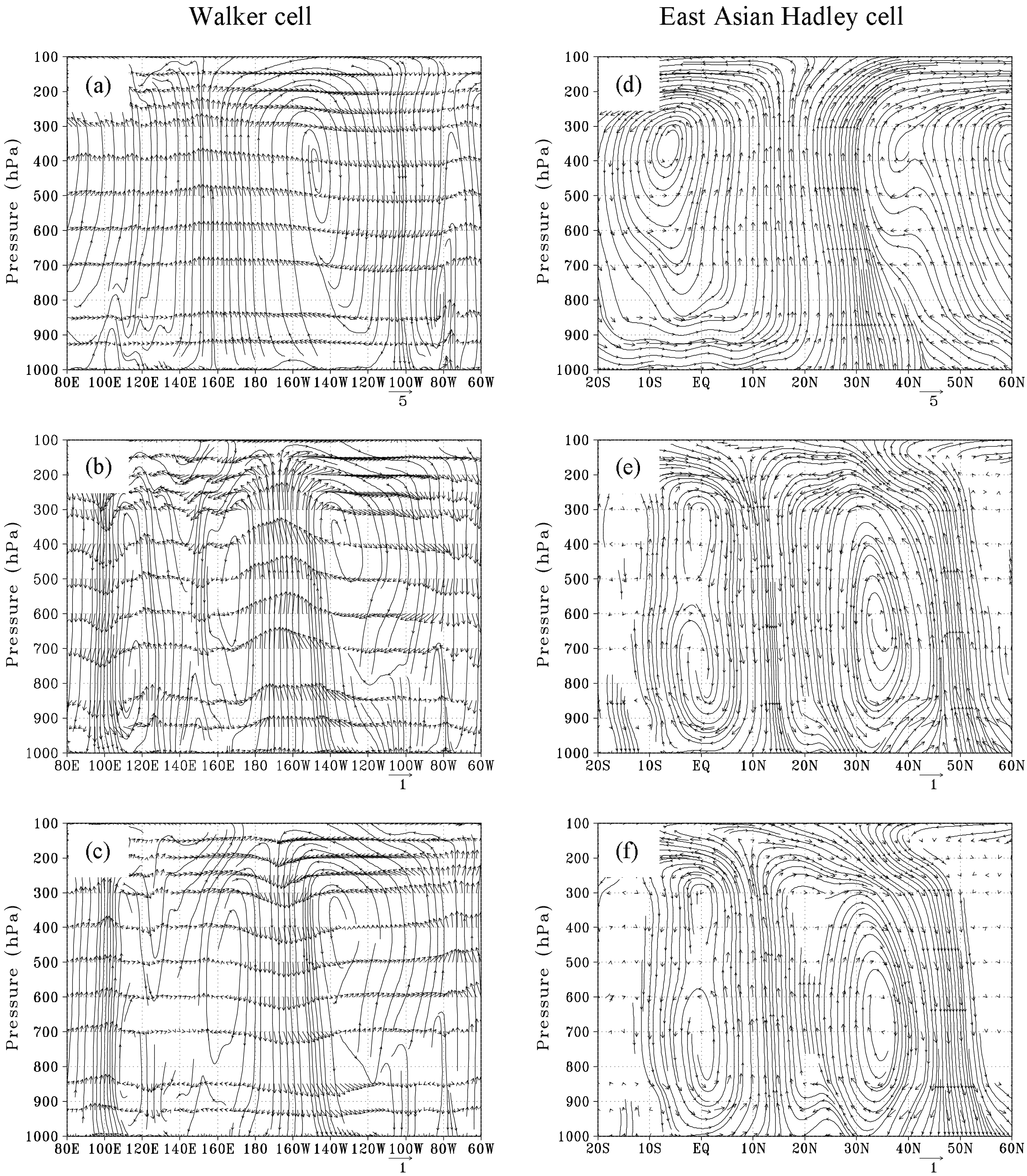

Fig. 5. Summer tropically-zonal (meridional) vertical atmospheric circulation averaged over 5°S–5°N (100°–130°E) in East Asia on the left (right) side, with climatological mean (top), anomaly in the strong EASM (1950–1964; middle), and anomaly in the weak EASM (1976–1997; bottom). (Vertical velocity scales 50; units: hPa s^{-1}; divergent wind: m s^{-1}).

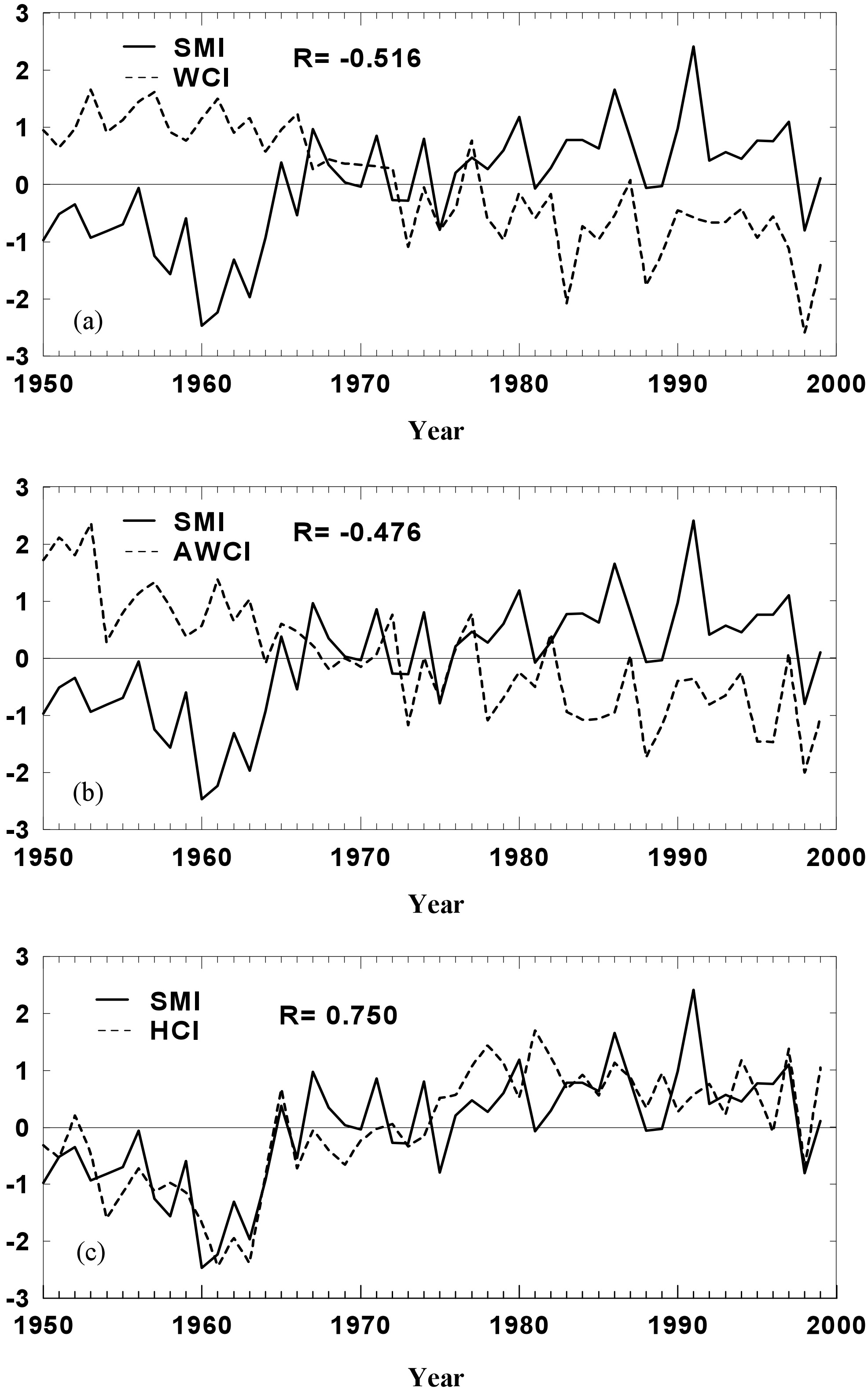

Fig. 6. Time series of (a) Walker circulation index (WCI), (b) anti-Walker circulation index (AWCI), (c) East Asian Hadley circulation index (HCI) and SMI (solid curves).

and tropical eastern Indian Ocean, before returning to the central Pacific. Therefore, anomalous westerlies and easterlies in the lower troposphere occur in the western and eastern tropical Pacific, respectively. An almost completely opposite anomalous cell appears during the weak EASM, except for a more complicated structure in the western Pacific during the strong EASM period. With regards to the climatic mean Walker Cell, it is enhanced (weakened) in the strong (weak) EASM over the central and eastern tropical Pacific. In the western Pacific and eastern Indian Ocean, rising is dominated in the summer climate mean cell. However, in the anomalous Walker cell, there is an anomalous anti-Walker cell between the eastern tropical Indian Ocean and the central Pacific, indicating stronger upward motion in the eastern tropical Indian Ocean during the weak EASM. These anomalous cells arranged over the tropical region greatly modulate the Walker circulation during the strong and weak EASM periods. Their effect is to enhance the mean Walker circulation and produce an anomalous strong anti-Walker circulation during the strong EASM but weaken them during the weak EASM.

To further discuss the relationship between the EASM and the Walker cell, a Walker circulation index (WCI) and an anti-Walker circulation index (AWCI) are defined based on 500-hPa isobaric vertical velocity. The former (latter) is defined as the normalized differences in average vertical velocity between 180°–160°W and 100°–80°W (180°–160°W and 90°–110°E) within the 5°S–5°N band. Interannual variations of the Walker and anti-Walker circulations are shown in Figs. 6a and 6b, respectively. Figure 6a shows that the Walker circulation has a distinct descending trend and interdecadal variation. During the period from the early 1950s to the early 1970s (late 1970s to the 1990s) the Walker circulation is stronger (weaker) than average. Compared with the EASM index, the Walker index has a good out-of-phase relationship. This indicates that a strong (weak) EASM is related to a strong (weak) Walker circulation. The correlation coefficient (CC) between indices of the Walker circulation and the EASM is −0.516, which exceeds the 99% confidence level. Similar to the Walker circulation index, the anti-Walker circulation has a good out-of-phase with the EASM, with a CC of −0.476, exceeding the 99% confidence level.

4.2 *East Asian Hadley cell*

It is well known that East Asia is a typical monsoon region. The variation of the EASM often influences summer precipitation and temperature over this region. From the first mode (EOF1) of summer velocity potential at 850 hPa (Fig. 2a), it is easy to see that the velocity potential varies in an opposite manner at the lower and mid latitudes over East Asia. Hence, it is necessary to discuss the meridional cells over East Asia during the strong and weak EASM, separately. Here, the meridional cell, the East Asian Hadley cell (EAHC), is described by the mean over 100°–130°E. The climatological-mean meridional-vertical circulation over East Asia is shown in Fig. 5d, where ascending (descending) motion dominates to the north (south) of the equator, indicating thermal contrast between land and ocean in boreal summer.

The anomalous EAHCs during the strong and weak EASM periods are shown in Figs. 5e and 5f. A dominant feature in these two figures is that there exists two strong anomalous closed cells, one located in the tropics with a twofold structure and the other in the mid latitudes. During the strong EASM (1950–1964) period, anomalous strong sinking of air parcels occurs over 0°–30°N, with convergence in the upper troposphere at around 10°N, flows moving northward and southward in the lower troposphere, ascending over 35°–55°N and 10°–5°S, before returning back to $\sim$ 10°N. Compared with 850-hPa anomaly velocity potential during the strong EASM (Figs. 3a and 3b), this sinking is attributed to convergence (divergence) in the upper (lower) troposphere over the South China Sea. An inverted anomalous cell emerges during the weak EASM period (Fig. 5f). All these results show that an anomalous anti-Hadley (typical Hadley) circulation cell exists during the strong (weak) EASM phase. The anomalous northward (southward) blowing wind at the lower level of the anomalous anti-Hadley (typical Hadley) circulation results in anomalous strong (weak) EASM.

To further address the relationship between the EASM and EAHCs, an East Asia Hadley circulation index (HCI) is defined based on 500-hPa isobaric vertical velocity. The EAHC index is defined as the normalized differences in average vertical velocity between 40°–50°N, 100°–130°E and 10°–20°N, 100°–130°E. Here, the positive (negative) index indicates strong (weak) Hadley circulation over East Asia. The variation of the EAHC index is shown in Fig. 6c. Over the past 50 years, there was an increasing trend and a distinct interdecadal variation in the East Asia Hadley circulation, which was weak (strong) before the mid 1960s and after the late 1970s. Also presented is the variation of the EASM and its correlation coefficient with the EAHC index is computed. Their very close relationship can be seen in Fig. 6c, with the correlation coefficient 0.750 exceeding the 99.9% confidence level.

Table 1. Correlations between summer precipitation over North China (NCR) and VPA, SLPA, SMI, PC1 of VP-850 hPa, HCI, WCI, AWCI, and Niño-3 SST indices in 49 summers.

	VPA	SLPA	EAHCI	SMI	PC1	WCI	AWCI	Niño-3
Correlation	0.588	−0.477	−0.498	−0.380	0.483	0.332	0.363	−0.425

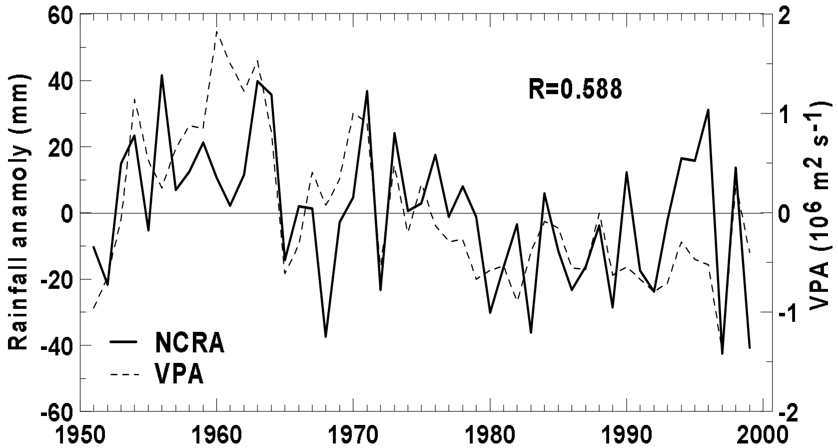

Fig. 7. Summer rainfall anomaly in North China (NCRA) and 850-hPa velocity potential anomaly over the area (40°–50°N, 95°–115°E).

5. Interdecadal variability of summer precipitation over North China and its possible cause

Summer rainfall over North China, which exhibits the most significant interdecadal feature (Zhang, 1999), is sensitive to the intensity of the EASM. The subtropical high and ITCZ usually move northward when a strong EASM occurs, bringing plenty of vapor into North China, resulting in anomalous precipitation over this region.

In this section, the interdecadal variability of summer precipitation in North China and its possible mechanism are documented. Figure 7 shows the variability of summer precipitation over North China based on regional (covering 26 stations) means. In Fig. 7, two shifts occur in the past 50 years, one in 1965 and the other in the late 1970s. Before 1965, positive anomaly precipitation occurs over North China. During the mid 1960s to the late 1970s, a transition period emerges when drought and flood come about alternately. The second shift occurs in the late 1970s. During the 1980s and 1990s, the major feature is that negative anomaly precipitation appears over the region. In the 1980s, severe drought came about frequently. These results are consistent with those of Yang et al. (2005). Compared with the variation of the EASM, the variation of summer rainfall over North China is in accord with its variability on an interdecadal scale. Their correlation coefficient reaches −0.38 and exceeds the 99% confidence level.

Many studies have dealt with the possible causes of interdecadal variability of summer precipitation in North China (Zhang, 1999; Li et al., 2002; Lu et al., 2004; Yang et al., 2005; Liu and Guo, 2005). Most of these studies revealed that the cause is attributable to the interdecadal variation of the EASM and the western Pacific subtropical high (e.g., Li et al., 2002; Lu et al., 2004). The blocking high in the vicinity of Lake Baikal and China-Mongolia is also an important factor, which results in the variation of the EASM (Li et al., 2002; Liu and Guo, 2005). However, the cause of the variation in the blocking high is not yet clear. In this study, therefore, the possible cause is investigated. The velocity potential anomalies (VPAs) averaged over 40°–50°N, 95°–115°E at 850 hPa are shown in Fig. 7, indicating that the variation of velocity potential around North China is consistent with its summer precipitation anomaly at a correlation coefficient of 0.588, exceeding the 99% confidence level. Table 1 presents the relationship between North China rainfall (NCR) and some indices, including velocity potential anomaly (VPA) at 850 hPa, sea level pressure anomaly (SLPA) averaged over 40°–50°N, 95°–115°E, the East Asian Hadley circulation index (EAHCI), SMI, the PC1 of velocity potential at 850 hPa, the WCI, AWCI and the Niño-3 SST index. All these correlation coefficients exceed the 98% confidence level. NRC has the closest relationship with VPA (correlation coefficient of 0.588), indicating that VPA may be the di-

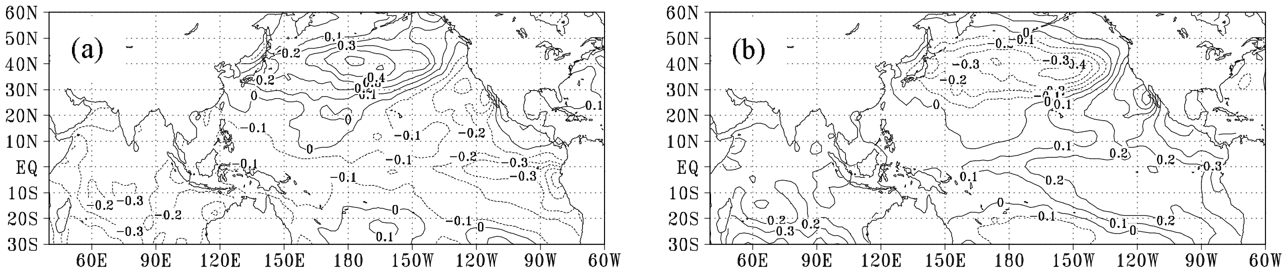

Fig. 8. Composite SSTA in boreal summer during (a) the strong EASM period (1950–1964) and (b) the weak EASM period (1976–1997). (Units: °C).

rect factor. The variation of positive/negative VPA in the south of Lake Baikal (40°–50°N, 95°–115°E) could induce anomaly convergence/divergence, thereby producing anomaly ascending/descending motion, corresponding sea level pressure over the region becoming weaker/stronger, and anomaly southerly/northerly flows dominating East Asia, accompanied by a stronger/weaker EASM, and in that case North China rainfall in summer will increase/decrease. On the other hand, the variation of VPA to the south of Lake Baikal is strongly connected to the variation of the tropical Walker circulation, which is related to the central eastern tropical Pacific SSTA.

6. Relationship of the EASM with SSTs

The SSTA in summer during the strong and weak EASM are presented in Figs. 8a and 8b, respectively. During the strong EASM period, negative (positive) SSTA dominates in the tropical central and eastern tropical Pacific and the Indian Ocean (North Pacific), and vice versa for the weak EASM. This SSTA configuration during the strong/weak EASM is very similar to the Pacific Decadal Oscillation (PDO) structure in cool/warm phases (Mantua et al., 1997), indicating that the EASM is related in structure to PDO. Huang (2001) demonstrated that the cause of interdecadal variation of the EASM is attributed to the interdecadal El Niño phenomenon in the eastern tropical Pacific after the mid 1970s. From previous analysis in section 4, it is suggested that the EASM is closely related to the intensity of the Walker circulation. The correlation coefficient between PC1 of 850-hPa velocity potential and the Niño-3 SST index in summer reaches -0.464, which exceeds the 99% confidence level, indicating that there is a close relationship between the eastern tropical Pacific SSTA and the atmospheric convergence/divergence over the tropical Pacific Ocean and East Asia.

From the above analysis, it is proposed that there is a possible mechanism between the interdecadal variation of the EASM and the eastern tropical Pacific SST. Warm SSTA in the eastern tropical Pacific induces anomalous ascending motion, resulting in an anomalous weak Walker and anti-Walker circulation, enhancing the EAHC, and resulting in a weak EASM.

7. Discussion and conclusions

By using 1950–1999 NCEP/NCAR reanalysis data, the atmospheric circulation associated with anomalous interdecadal variation of the EASM and its possible mechanism have been investigated. The major findings of this study can be summarized as follows:

(1) The EASM exhibits a distinct interdecadal variation, with a strong (weak) summer monsoon maintained between 1950–1964 (1976–1997).

(2) There are close connections between the EASM, Walker and East Asian Hadley cells (see Figs. 9a–9b). During a strong EASM (Fig. 9a), there is a strong anomalous zonal Walker cell and an anti-Walker circulation in the tropical Pacific. This zonal atmospheric cell is characterized by air rising in the central tropical Pacific, flowing eastward and westward in the upper troposphere, descending in the eastern and western tropical Pacific, before returning to central tropical Pacific. It is also found that the anomalous strong EAHC connects the low- and mid-latitude circulations in East Asia and weakens the EASM. Its ascending branch lies in the western part of the anti-Walker cell, flowing northward in the upper troposphere and descending to the south of Lake Baikal, before returning to low latitudes in the lower troposphere, which thus weakens the EASM. All anomalous atmospheric cells are almost completely reversed during a weak EASM (Fig. 9b) compared to that in a strong EASM (Fig. 9a).

(3) The relationship between the EASM and SST in the eastern tropical Pacific has been discussed. It is proposed that there is a possible mechanism on the interdecadal variation of the EASM associated with eastern tropical Pacific SST. Warm SSTA in the east-

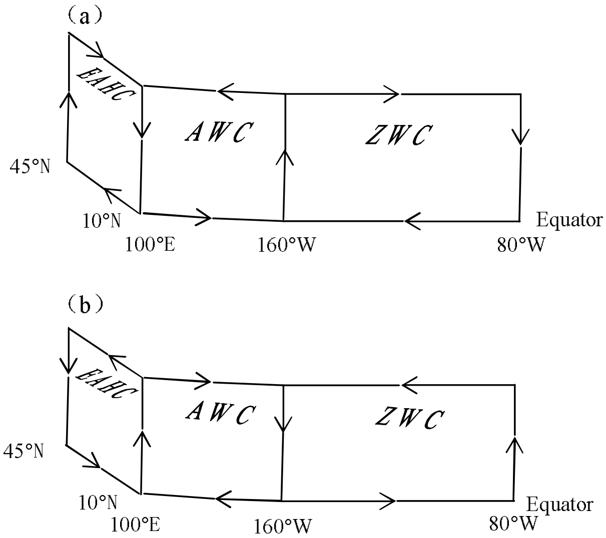

Fig. 9. Schematic diagrams of anomalous atmospheric cells during (a) strong EASM (1950–1964) and (b) weak EASM (1976–1997). Shown are the zonal Walker cell (ZWC), anti-Walker cell (AWC), and EAHC.

ern Pacific induces anomalous ascending motion, thus resulting in an anomalous weak Walker and an anti-Walker circulation, enhancing the EAHC, and resulting in a weak EASM.

(4) The interdecadal variation of summer precipitation over North China is closely related to the velocity potential of the south of Lake Baikal, which produces strong ascending and descending motion.

The mechanism of anomalous upward/downward motion over the tropical central Pacific occurring in the strong/weak EASM in relation to cool/warm SSTA in the central Pacific is obscure. This feature does not conform with the fact that warm SSTA over the central and eastern equatorial Pacific often causes ascending motion (Oort and Yienger, 1996). Is this anomalous motion over the central tropical Pacific related to the SSTA difference between the eastern tropical Pacific and the Indian Ocean? The mechanism needs to be further explored through numerical experiments.

Acknowledgements. The authors wish to thank the two anonymous reviewers for their insightful comments and suggestions. This work was completed during Zeng Gang's visit to the Atmospheric Science Research Center of State University of New York at Albany (ASRC/SUNYA) and was jointly supported by (a) the joint agreement between the U. S. Department of Energy and the China Ministry of Sciences and Technology Project of Climate Sciences, (b) the National Natural Science Foundation of China (Grant No. 40331010), (c) the project for basic Natural Science Research of Jiangsu High Institutions (Grant No. 07KJB170067), and (d) the foundation of Jiangsu Province-level key laboratory of Meteorological Disasters (KLME060301).

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