

How to Measure the Strength of the East Asian Summer Monsoon

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ABSTRACT

Defining the intensity of the East Asian summer monsoon (EASM) has been extremely controversial. This paper elaborates on the meanings of 25 existing EASM indices in terms of two observed major modes of interannual variation in the precipitation and circulation anomalies for the 1979–2006 period. The existing indices can be classified into five categories: the east–west thermal contrast, north–south thermal contrast, shear vorticity of zonal winds, southwesterly monsoon, and South China Sea monsoon. The last four types of indices reflect various aspects of the leading mode of interannual variability of the EASM rainfall and circulations, which correspond to the decaying El Niño, while the first category reflects the second mode that corresponds to the developing El Niño.

The authors recommend that the EASM strength can be represented by the principal component of the leading mode of the interannual variability, which provides a unified index for the majority of the existing indices. This new index is extremely robust, captures a large portion (50%) of the total variance of the precipitation and three-dimensional circulation, and has unique advantages over all the existing indices. The authors also recommend a simple index, the reversed Wang and Fan index, which is nearly identical to the leading principal component of the EASM and greatly facilitates real-time monitoring.

The proposed index highlights the significance of the mei-yu/baiu/changma rainfall in gauging the strength of the EASM. The mei-yu, which is produced in the primary rain-bearing system, the East Asian (EA) subtropical front, better represents the variability of the EASM circulation system. This new index reverses the traditional Chinese meaning of a strong EASM, which corresponds to a deficient mei-yu that is associated with an abnormal northward extension of southerly over northern China. The new definition is consistent with the meaning used in other monsoon regions worldwide, where abundant rainfall within the major local rain-bearing monsoon system is considered to be a strong monsoon.

1. Introduction

The East Asian summer monsoon (EASM) is a distinctive component of the Asian climate system (Chen and Chang 1980; Tao and Chen 1987; Lau et al. 1988; Ding 1992; Wang and Li 2004) due to unique orographic forcing: huge thermal contrasts between the world's largest continent, Eurasia, and the largest ocean basin, the Pacific, and is strongly influenced by the world's highest land feature, the Tibetan Plateau. The

EASM also has complex space and time structures that encompass tropics, subtropics, and midlatitudes.

Giving this complexity, it is difficult to quantify the EASM variability with an appropriate yet simple index. For the Indian summer monsoon (ISM), the All Indian Rainfall index (AIRI) has generally been accepted as one such measure (e.g., Mooley and Parthasarathy 1984; Parthasarathy et al. 1992), partly due to the relative homogeneity of rainfall distribution. However, to quantify the EASM variability with averaged rainfall over the domain is much more difficult because the seasonal mean precipitation anomaly often exhibits large meridional variations.

For those readers who are less familiar with the

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EASM rainfall structure and circulation system, we provide in this paragraph some background information. One of the prominent features of the EASM is the rainfall concentration in a nearly east–west-elongated rainbelt. This subtropical rainbelt is most prominent during June and July, which stretches for many thousands of kilometers, affecting China, Japan, Korea, and the surrounding seas. The intense rains during that period are called mei-yu in China, baiu in Japan, and changma in Korea. The rainbelt is associated with a quasi-stationary subtropical front that is the primary rain-producing system in the EASM. The East Asian (EA) subtropical front is established when East Asian polar fronts repeatedly move southward into the subtropics where they undergo modification such that their baroclinicity weakens significantly and much of the rainfall are caused by deep cumulus convection (Chen and Chang 1980; Ding 2004; Ninomiya 2004). A significant portion of the convective rainfall is produced in organized mesoscale vortices along the narrow frontal zone (e.g., Chang et al. 1998; Ding 2004). The mei-yu/baiu/changma rainbelt has alternatively been called the EA monsoon trough since it is the main low-level trough over East Asia that produces most of the summer monsoon rainfall (Chen and Chang 1980). The impacts of floods and droughts on human lives and economics during the EASM are tremendous because the finer intraseasonal space–time structure, coupled with the orientation of the rivers that is mostly parallel to the rainbands, makes the occurrence of floods and droughts particularly sensitive to interannual variations of precipitation (Chang 2004).

In studies where the EASM variability is mainly defined by the seasonal rainfall, the complex rainfall structure encouraged some authors to use a principal component approach (e.g., Tian and Yasunari 1992; Nitta and Hu 1996; Lee et al. 2005). Since actual rainfall amount and variations differ considerably among observational stations because of terrain effects, often the variance of a few stations can dominate the total variability, but their significance may be downplayed in the EOF analysis. Some authors focused on station rainfall averaged over one or more area(s) that are determined from climatological activity zones. Most of these studies considered the mei-yu/baiu/changma rainfall as the representative feature of the EASM (e.g., Tanaka 1997; Chang et al. 2000a,b; Huang 2004). However, in most of these studies the concept of a simple EASM index was not explicitly stated.

Most of the investigators searching for a simple index for the EASM strength use circulation parameters instead of rainfall partly due to the complex rainfall struc-

ture and partly due to a preference of using large-scale winds to define the broad-scale monsoon. Unfortunately, representation of the EASM circulation's strength remains highly controversial. To our knowledge, at least 25 circulation indices have been proposed to measure the EASM intensity (Table 1). This raises several questions: Why have there been such a large number of indices proposed? Is it possible to construct an appropriate “index” that is more broadly applicable? What is the physical basis for such an index? How should we measure the strength of the EASM?

To address these questions, it is important to first elaborate on the climatic meanings of these existing indices and examine their relationships to the large-scale precipitation and circulation anomalies associated with the EASM (see section 2). In this regard, an objective and efficient way is perhaps to try to establish the fundamental characteristics of the major modes of EASM interannual variability (see section 3). These major modes can then be used to set up objective metrics for gauging an index's performance in measuring EASM and helping to illuminate the meanings of each index (see section 3). Based on these analyses, a unified EASM index and a simple EASM index are recommended (see section 4). The robustness, advantages, and significance of the unified EASM index are also demonstrated in section 4. In the last section, the complexity, the definition of a strong Chinese summer monsoon, and the potential limitations of the new unified index are discussed.

2. Analysis of the existing indices

The existing 25 circulation indices listed in Table 1 may be classified into five categories. The first category can be labeled an “east–west thermal contrast” index, which is constructed by the sea level pressure (SLP) difference between a land longitude over the EA and an oceanic longitude over the western North Pacific (WNP). The original idea was proposed by Guo (1983), and her index was subsequently modified (e.g., Shi and Zhu 1996; Peng et al. 2000; Zhao and Zhou 2005). The notion behind this early definition was that the east–west land–ocean thermal contrast may determine the southerly monsoon strength over the EA.

The second category reflects the “north–south thermal contrast” by using vertical shear of zonal winds, as in Webster and Yang (1992). Most of the indices in this category represent the zonal thermal winds between 850 and 200 hPa that result from north–south thermal contrast between the EA land and the South China Sea (SCS; e.g., Wang et al. 1998; Zhu et al. 2000; He et al. 2001). The idea behind these indices emphasizes the

TABLE 1. Description of the 25 EASM circulation indices. Here, u = zonal winds, v = meridional winds, Φ = potential height, D = divergence, and PV = potential vorticity. Their correlation coefficients with the first two PCs of the MV-EOF of the EASM are shown. The definition of the combined skill is given in the text. The bold italic correlation coefficients exceed 99% confidence level based on the Student's t test and the bold italic values in the combined skill column represent the high performers.

Index	Reference	Defining variable(s), level (hPa), and regions	Correlation with PC1 and PC2		Combined skill (%)
			PC1	PC2	
I_{GOY}	Guo (1983)	SLP gradient, (10°–50°N, 110°–160°E)	–0.34	0.49	12.6
I_{SZ}	Shi and Zhu (1996)	SLP gradient, (20°–50°N, 110°–160°E)	0.02	0.70	8.3
I_{PSN}	Peng et al. (2000)	Φ gradient, 500, (10°–50°N, 110°–150°E)	0.25	0.63	12.2
I_{ZZ}	Zhao and Zhou (2005)	SLP gradient, (30°–50°N, 110°–160°E)	–0.01	0.78	8.9
I_{WY}	Webster and Yang (1992)	u , 200–850, (10°–40°N, 110°–140°E)	–0.57	–0.40	16.6
I_{WDJ}	Wang et al. (1998)	u , 850–200, (5°–15°N, 90°–130°E)	–0.86	–0.15	20.0
I_{ZHW}	Zhu et al. (2000)	u , 850–200, (0°–10°N, 100°–130°E); SLP	–0.55	–0.51	17.4
I_{HSX}	He et al. (2001)	u , 850–200, (0°–10°N, 100°–130°E)	–0.89	0.04	19.3
I_{WF}	Wang and Fan (1999)	vorticity, 850, (5°–32.5°N, 90°–140°E)	–0.97	0.06	21.1
I_{ZTC}	Zhang et al. (2003)	vorticity, 850, (10°–35°N, 100°–150°E)	–0.93	–0.03	20.1
I_{LKY}	Lau and Yang (2000)	vorticity, 200, (20°–50°N, 110°–150°E)	–0.38	–0.39	12.3
I_{HY}	Huang and Yan (1999)	vorticity, 500, (20°–60°N, 125°E)	–0.38	–0.08	8.9
I_{LZ}	Li and Zeng (2002)	u , v , 850, (10°–40°N, 110°–140°E)	0.93	0.03	20.0
I_{WHJ}	Wang (2002)	u , v , 850, (20°–40°N, 110°–125°E)	0.70	–0.14	16.3
I_{OCZ}	Qiao et al. (2002)	u , v , 850, (20°–40°N, 110°–140°E)	0.81	–0.20	14.0
I_{JOC}	Ju et al. (2005)	u , v , 850, (22°–32°N, 112°–135°E); OLR	0.59	0.14	14.0
I_{WN}	Wu and Ni (1997)	v , 850, (20°–30°N, 110°–130°E)	0.56	–0.02	12.2
I_{WVO}	Y. F. Wang et al. (2001)	v , 850, (20°–40°N, 110°–140°E)	0.69	0.28	17.8
I_{LZh}	Li and Zhang (1999)	D200–850, (7.5°–17.5°N, 105°–125°E)	–0.44	0.05	9.8
I_{LWY}	Liang et al. (1999)	u , v , 850, (5°–20°N, 105°–120°E); OLR	–0.89	0.10	19.8
I_{WL}	Wu and Liang (2001)	u , v , 850, (5°–20°N, 105°–120°E); OLR	–0.35	0.23	10.0
I_{ZLY}	Zhang et al. (2002)	u , v , 850, (5°–20°N, 105°–120°E); OLR	–0.57	0.21	14.5
I_{DXZ}	Dai et al. (2000)	u , v , 850, (5°–20°N, 105°–120°E)	–0.93	0.07	20.7
I_{LC}	Lu and Chan (1999)	v , 1000, (7.5°–20°N, 107.5°–120°E)	–0.51	0.12	12.1
I_{YO}	Yao and Qian (2001)	Moisture PV, 850, (10°–20°N, 105°–120°E)	0.06	–0.42	12.7

importance of the north–south land–sea thermal contrast.

In the third category, the shear vorticity (often expressed by a north–south gradient of the zonal winds) is used. Wang and Fan (1999) first proposed such a shear vorticity index to quantify the variability of the WNP summer monsoon. This index was defined by the U_{850} in (5°–15°N, 90°–130°E) minus U_{850} in (22.5°–32.5°N, 110°–140°E), where U_{850} denotes the zonal wind at 850 hPa. Zhang et al. (2003) used a similar vorticity index but defined it in a slightly modified domain, that is, U_{850} (10°–20°N, 100°–150°E) minus U_{850} (25°–35°N, 100°–150°E). Lau and Yang (2000) applied a shear vorticity index to 200-hPa zonal winds to measure changes in the upper-tropospheric westerly jet stream, which affect the EASM. Huang and Yan (1999) introduced an atmospheric teleconnection index that reflects 500-hPa vorticity at three grids in the EA–WNP region.

The fourth category may be called “southwest monsoon” indices, which directly gauge the strength of the low-level EA monsoon winds using the 850-hPa southwesterly winds. The area where the winds are averaged

primarily covers the subtropical EASM region with various latitudinal extents (Li and Zeng 2002; Wang 2002; Qiao et al. 2002; Ju et al. 2005). Some indices used southerly wind component (Wu and Ni 1997) or meridional variation of the southerly component (Y. F. Wang et al. 2001).

The fifth category may be classified as “South China Sea monsoon” indices, because in this category, the SCS monsoon is thought to be a critical tropical portion of the EASM and its variations are often indicative of the changes in the EASM. Chang and Chen (1995) was an earlier adopter using a low-level southwesterly wind index, but they used it only to indicate the monsoon onset rather than the monsoon strength because they defined EASM principally on the pre–mei-yu and mei-yu rainfall system. The SCS monsoon indices have been denoted by several variables such as vertical differential divergence (Li and Zhang 1999), a combination of 850-hPa southwest winds and outgoing longwave radiation (OLR; Liang et al. 1999; Wu and Liang 2001; Zhang et al. 2002), 850- and 1000-hPa southwesterly winds only (Dai et al. 2000; Lu and Chan 1999), and moist potential vorticity (Yao and Qian 2001).

3. How well do the circulation indices represent the major modes of variability?

a. Major modes of variability of the EASM

One way to evaluate the suitability of the many circulation indices is to see how each of them reflects the leading modes of EASM variability. The first step is to determine the leading modes of EASM variability. To capture the EASM circulation system discussed by Tao and Chen (1987), we choose an analysis domain (0° – 50° N, 100° – 140° E) that includes both tropical WNP and subtropical EA as they are closely coupled. Further, since the EASM has unique characteristics in both rainfall distribution and associated large-scale circulation systems, we decide to use multivariate EOF analysis (MV-EOF) on a set of six meteorological fields in June–August (JJA), including precipitation and five atmospheric circulation fields (the zonal and meridional winds at 850 and 200 hPa, and SLP). The MV-EOF analysis method was described in detail in Wang (1992); it has the advantage of capturing spatial phase relationships among the various circulation and precipitation fields. In this paper, an area-weighted correlation coefficient matrix is constructed for the combined six meteorological fields to carry out the MV-EOF. As such, the eigenvectors (spatial patterns) are nondimensional.

The data used include 1) monthly precipitation data from the Global Precipitation Climatology Project (GPCP) for the 1979–2006 period (Adler et al. 2003); 2) the wind and SLP data, gridded at $2.5^{\circ} \times 2.5^{\circ}$ resolution, taken from the National Centers for Environmental Prediction–Department of Energy reanalysis (NCEP-2; Kanamitsu et al. 2002); and 3) the Niño-3.4 (5° N– 5° S, 170° – 120° W) sea surface temperature (SST) index calculated from the improved extended reconstructed SST, version 2 (ERSST V2; Smith and Reynolds 2004). In this study, summer (JJA) mean anomalies are defined by the deviation of JJA mean from the long-term (1979–2006) mean climatology. Thus, the anomalies contain both interannual and decadal variations.

The leading mode accounts for 21.1% of the total variance for all six fields together (Fig. 1a). The fractional (percentage) variances of the first four MV-EOF eigenvectors and the associated unit standard deviation of the sampling errors are all shown in Fig. 1a. According to the rule given by North et al. (1982), the leading mode is statistically distinguished from the rest of the eigenvectors in terms of the sampling error bars. The second mode, which accounts for 11.1% of the total variance, is not separable from the higher mode. Nevertheless, the meteorological meaning of the first two

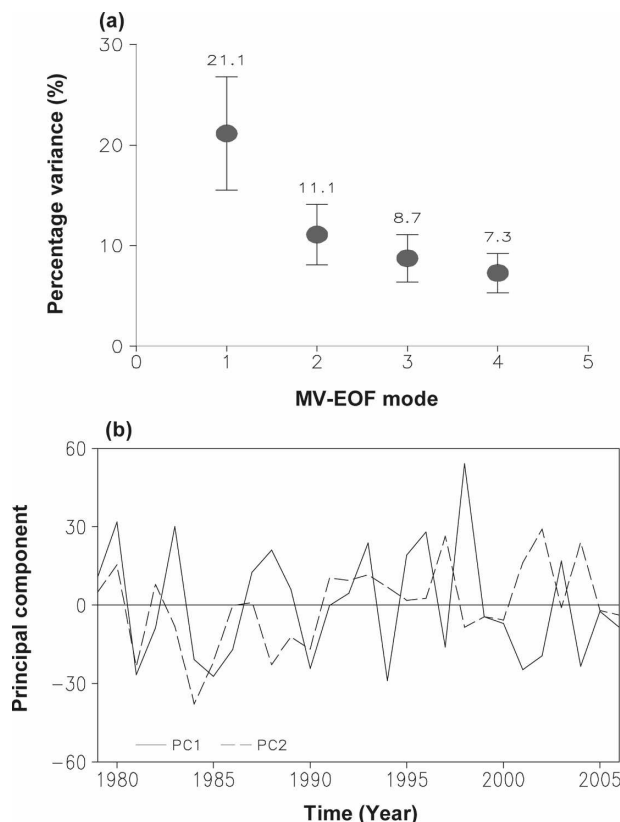


FIG. 1. (a) Fractional variance (%) explained by the first four MV-EOF modes and the associated unit standard deviation of the sampling errors. (b) PCs of the first and second MV-EOF modes.

modes is examined here. The time series of the principal components (PCs) of the first and second modes exhibit considerable interannual variations (Fig. 1b).

The spatial patterns of the first MV-EOF mode (Fig. 2a) show a north–south dipole pattern with dry anomalies over the northern SCS and Philippine Sea (PS) and enhanced precipitation along the Yangtze River valley to southern Japan, which covers the prevailing mei-yu/baiu/changma frontal area. At 850 hPa, a prominent feature is the anomalous subtropical high with enhanced southwesterly winds prevailing on its northwest flank from South China to the middle and lower reaches of the Yangtze River and strengthened easterly anomalies between 5° and 20° N (Fig. 2a, upper panel). The lower panel of Fig. 2a shows rising SLP extending from the eastern Philippine Sea to the northern South China Sea, which is consistent with a large-scale 850-hPa anticyclonic anomaly and suppressed rainfall anomaly. At 200 hPa, a cyclonic anomaly is over the Philippines and a large-scale anticyclone anomaly covers South China and expands eastward (Fig. 2a, lower panel).

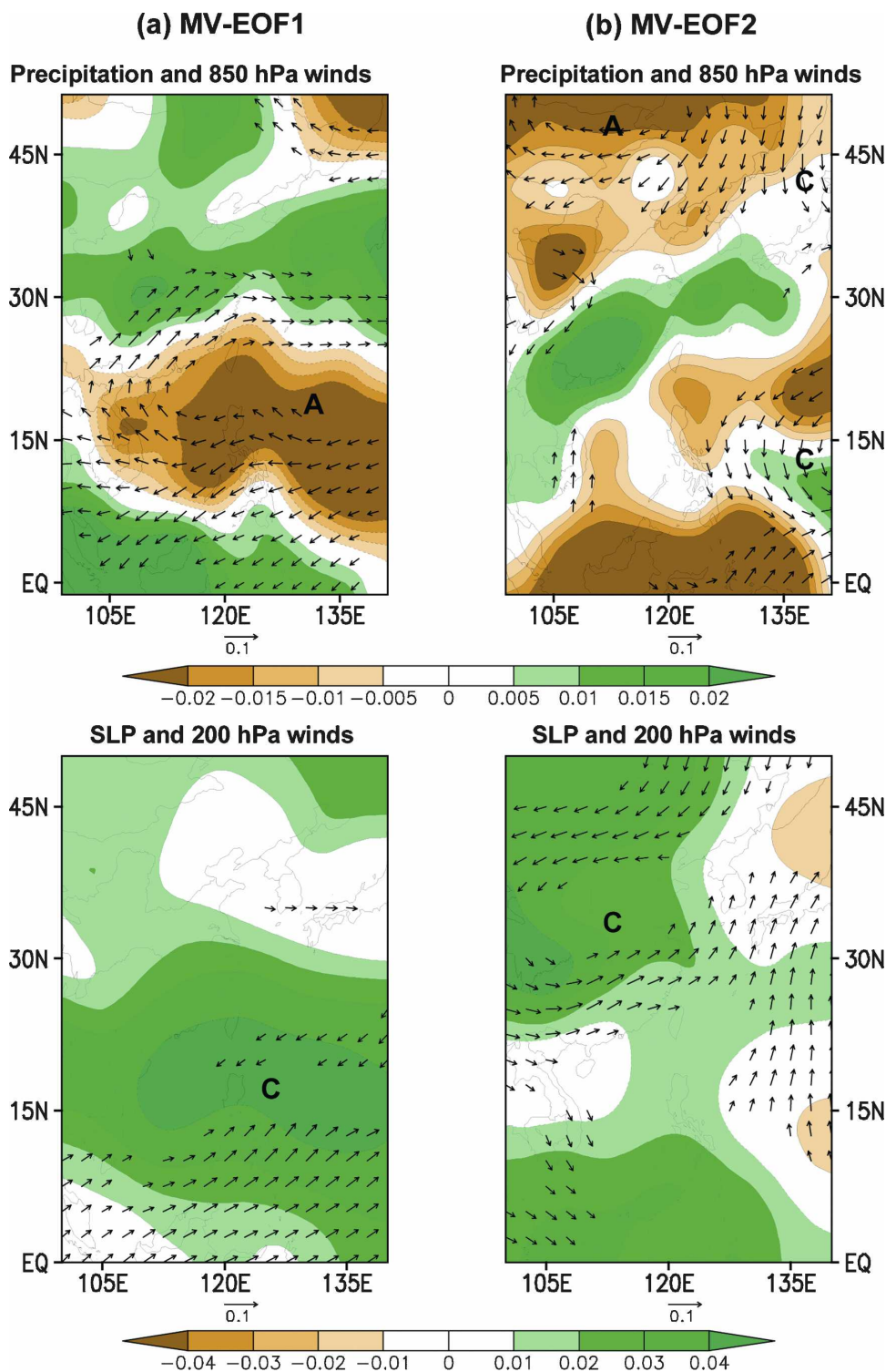


FIG. 2. Spatial patterns of the (a) first and (b) second MV-EOF modes of the East Asian summer precipitation (color shadings in the upper panels), 850-hPa winds (vector in the upper panels), sea level pressure (color shadings in the lower panels), and 200-hPa winds (vector in the lower panels). Anticyclone and cyclone are denoted by “A” and “C,” respectively. All quantities are nondimensional as they were derived from the correlation coefficient matrix.

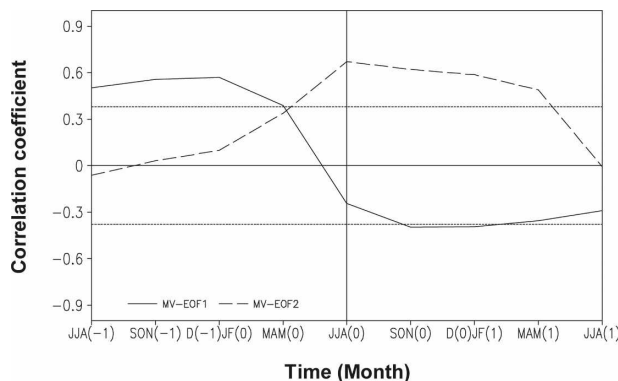


FIG. 3. The lead-lag correlation coefficients between the two leading MV-EOF PCs and the Niño-3.4 SST index from JJA(−1) to JJA(+1). The two black dotted lines represent 95% confidence level based on Student's t test. The vertical line indicates JJA(0), where the simultaneous correlations between the two PCs and Niño-3.4 index are shown.

The spatial patterns of the second MV-EOF mode (Fig. 2b) show another dipole pattern with enhanced precipitation over southern China and suppressed precipitation in northern and northeast China (approximately north of 35°N). At 850 hPa an anticyclonic anomaly occupies northern China and northerly wind anomalies occurring over China and Korea (Fig. 2b, upper panel), indicating a weakening monsoon, especially in northern China. In the meantime, rising SLP and the associated large-scale cyclonic anomaly at 200 hPa prevail over central eastern China, while negative SLP anomaly areas are located over the Philippine Sea and Japan, respectively.

Are the first two leading modes related to ENSO? To answer this question we calculate the lead-lag correlation coefficients between the two PCs and the Niño-3.4 SST anomaly (SSTA) from JJA(−1) to JJA(+1) and the results are shown in Fig. 3. The two PCs are used as references (year 0) to denote the strengths of the two major EASM modes in JJA(0). The first mode shows a maximum positive correlation coefficient (0.58) in D(−1)JF(0). Since El Niño events normally mature toward the end of the calendar year (Rasmusson and Carpenter 1982), the significant positive correlation between PC1 in JJA(0) and the Niño-3.4 SSTA in the previous winter D(0)JF(1) indicates that the first MV-EOF mode occurs in the “post-El Niño” summer or the decaying phase of El Niño. Note that EOF1 does show an anticorrelation with Niño-3.4 SST in the El Niño developing phase [JJA(0)], but the correlation coefficient is about −0.2 (not significant) (Fig. 3). Thus, one cannot interpret EOF1 as reflecting the anomalies in the El Niño developing phase.

The mechanisms responsible for this “prolonged” im-

pact of ENSO or a “delayed” response of the EASM to ENSO have been discussed in detail by Wang et al. (2000). They pointed out the critical role of monsoon and warm ocean interaction. This mechanism is characterized by a positive feedback between the off-equatorial moist atmospheric Rossby waves and the underlying SST anomaly in the local monsoon warm pool region. How does the positive feedback between the atmospheric Rossby waves and SST maintain the Philippine Sea anticyclonic (PSAC) anomaly? In the presence of the mean northeasterly trades, the ocean to the east of the PSAC cools as a result of enhanced total wind speed that induces excessive evaporation and entrainment. The cooling, in turn, suppresses convection and reduces latent heating in the atmosphere, which excites westward-propagating, descending Rossby waves that reinforce the PSAC. The moist Rossby wave–SST interaction can maintain both the Philippine Sea anticyclone and the dipolelike SST anomaly in the western Pacific during the decaying El Niño. Thus, even though the SSTA disappears during the summer after the peak El Niño, the EASM remains to be affected by the WNP subtropical anomalies significantly as shown by the leading mode of the EASM.

On the other hand, the observed second mode shows a maximum correlation coefficient (0.63) in JJA(0), suggesting that it concurs with the ENSO development phase. Inspection of the PC time series shown in Fig. 1b confirms this assertion. It can be seen from Fig. 1b that PC2 corresponds to the developing El Niño (1982, 1986, 1987, 1991, 1997, 2002, and 2004). In addition, PC1 is dominated by interannual variation, but PC2 has a large decadal component with a sharp change in 1990–91.

b. How well do the 25 EASM indices capture the above two leading modes?

Over half of the indices are well correlated with the first MV-EOF mode, with their correlation coefficients greater than 0.5 and exceeding the 99% confidence level based on Student's t test (Table 1). Note that the best correlation appears to come from categories 2 through 5, such as the shear vorticity indices (Wang and Fan 1999; Zhang et al. 2003), the southwest monsoon index (Li and Zeng 2002), the SCS monsoon index (Dai et al. 2000), and the north–south thermal contrast index (He et al. 2001), with the absolute value of their correlation coefficients equal to or greater than 0.89. The reason is that the first mode is characterized by a suppressed WNP monsoon trough and monsoon easterly vertical shear in southern SCS, and enhanced southwesterly monsoon over southern China due to the southwestward extension of the WNP subtropical high. Therefore, the Philippine Sea vorticity indices, south-

west monsoon indices, SCS monsoon indices, and north–south thermal contrast indices all capture the leading EOF mode very well. A common feature of these four types of indices is that they all depict a strong mei-yu in China, a strong changma in Korea, and a strong baiu in Japan with a properly defined sign, in other words, a situation where rainfall along the EA subtropical front is enhanced.

The east–west thermal contrast indices, however, do not correlate well with the leading mode, but it reflects the second mode much better. The EASM anomalies during ENSO developing years feature a rising pressure over land and a falling pressure over the WNP (Fig. 2b, lower panel); therefore, it is best represented by the east–west thermal contrast index. Since the other four categories' indices do not reflect the variability of the second mode, the east–west thermal contrast index is complementary to the indices of the other four categories.

To measure the overall skill of each of the 25 EASM indices in capturing the first two leading modes of the interannual variation, we used the following formula:

$$I_C = \sum_{i=1}^n E_i R_i,$$

where I_C represents combined skill for the two modes ($n = 2$); E_i denotes fractional variance of the i th MV-EOF mode; and R_i is the correlation coefficient between the EASM index and the i th PC series. It is found that the indices proposed by Wang and Fan (1999), Dai et al. (2000), Li and Zeng (2002), Zhang et al. (2003), and Wang et al. (1998) have the best combined skill (Table 1), which is basically consistent with the results from the first MV-EOF mode.

4. Recommendation

a. A unified EASM index

We have pointed out that the 25 existing circulation indices can be classified into five categories. The indices in the last four categories are able to mirror various aspects of the leading mode of EASM variability, such as the suppressed WNP monsoon trough and southwestward extension of the WNP subtropical high (as depicted by the shear vorticity indices), the enhanced southwest monsoon over subtropical East Asia (southwesterly indices) and over the SCS (SCS monsoon indices), and the reduced vertical wind shear over the tropics (north–south thermal contrast indices). These results suggest that the last four categories of indices can be unified by the PC of the leading MV-EOF mode

of the EASM precipitation and circulation (hereafter the leading PC of EASM).

We, therefore, recommend the leading PC of EASM be used to measure the intensity of the EASM. A high value of this unified index means a strong EASM, which is characterized by an abundant mei-yu/baiu/changma. To support this recommendation, we further demonstrate, in this section, 1) the robustness of the new index, 2) the clear advantage of this index over the existing indices, and 3) the significance of this index in representing the total variance of the EASM.

Is the leading MV-EOF mode (or PC1) sensitive to the choice of EOF analysis domain? To answer this question, five subdomains were tested: (a) a smaller “core” region (20° – 40° N, 105° – 135° E), (b) a northern domain (20° – 50° N, 105° – 135° E), (c) a southern domain (10° – 40° N, 105° – 135° E), (d) a western domain (10° – 40° N, 105° – 125° E), and (e) an eastern domain (10° – 40° N, 120° – 140° E). The results are shown in Fig. 4. For simplicity, the leading MV-EOF modes were obtained by use of precipitation and 850-hPa winds only, so that the spatial pattern can be compared to Fig. 2a. It is shown that the leading modes derived for the five different subdomains all replicate the same leading mode in the original domain (0° – 50° N, 100° – 140° E). Both the spatial patterns and PCs are extremely similar to those of the leading mode shown in Figs. 1 and 2, indicating that the leading PC of EASM is remarkably robust. On the other hand, the second mode (and the rest of the higher mode) is domain dependent (figure not shown), suggesting that the second mode cannot be used as a measure of the intensity for the entire EASM domain. We have made an additional analysis of the 4-month (May–August) summer mean anomalies; the resultant leading mode remains unchanged, confirming that the leading mode is not sensitive to choice of the length of the summer season either.

What is the advantage of the new index over the existing indices? First, the new index reflects the variability of both precipitation and three-dimensional circulation. Different from all existing indices, the new index represents variations of the EASM circulation system, which consists of the lower-level western Pacific subtropical high, tropical monsoon trough, subtropical front, and the upper-level South Asian high and the associated westerly jet to its north and easterly jet to its south. The cohesive spatial structure offers a clear physical meaning to the index. Second, the new index tells us how much percent of total variance it accounts for. This piece of information is desirable for any quantitative measure. Third, more than 80% of the existing indices (categories 2 through 5) can be represented by the new index because each of these indices

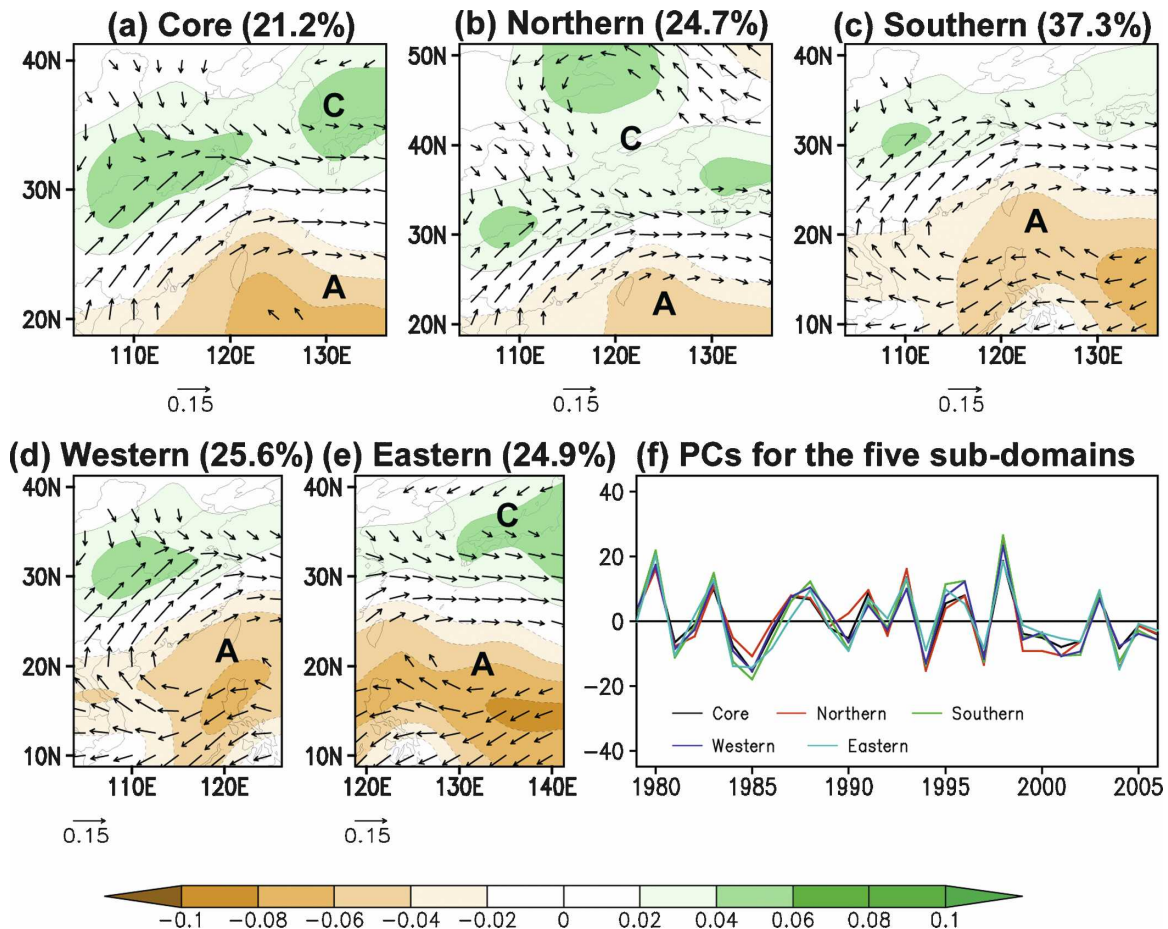


FIG. 4. Spatial patterns of the leading MV-EOF mode of the EASM precipitation (color shading) and 850-hPa winds (vector) obtained for the following five subdomains: (a) the core region (20° – 40° N, 105° – 135° E), (b) the northern region (20° – 50° N, 105° – 135° E), (c) the southern region (10° – 40° N, 105° – 135° E), (d) the western region (10° – 40° N, 105° – 125° E), and (e) the eastern region (10° – 40° N, 120° – 140° E). (f) The principal components derived for the five subdomains are compared. The fractional variance for each domain is indicated at the title of each panel. Anticyclone and cyclone are denoted by “A” and “C,” respectively. All quantities are nondimensional as they were derived from the correlation coefficient matrix.

reflects some specific aspects of the variations of the EASM system.

A possible concern is the significance of this new index: the leading mode presented in Fig. 1 seems to account for only about 21% of the total variance. As we mentioned in section 2, the MV-EOF analysis shown in Figs. 1 and 2 was based on correlation coefficient matrix (CCM) and the eigenvectors are dimensionless. The CCM analysis yields eigenvectors that correctly describe spatial distribution of the anomalies but not the amplitude of the anomalies. Therefore, the fractional variance computed by the CCM method cannot be used. To obtain accurate fractional variance and amplitude information, we repeated the MV-EOF analysis using the covariance matrix method. Figure 5 shows the spatial structure of the leading mode. The corresponding principal component was not shown because it is

identical to that shown in Fig. 1b. Compared with Fig. 2a, it is clear that the amplitude of the leading mode decreases with latitude. The leading mode accounts for 50% of the total variance. This fractional variance is comparable to that of the leading mode of the tropical Pacific SST anomalies (the El Niño/La Niña mode). Thus, the leading PC of EASM captures a substantial amount of the variance and is suitable to depict the variability of the entire EASM system.

b. A simple EASM index

While the leading PC of EASM provides a robust unified index, weaknesses exist. This index is difficult for making real-time monitoring of the EASM variation. This is an analog to ENSO monitoring: no one uses the PC of the leading EOF mode of Pacific SSTA, rather, a very simple index, such as Niño-3 (5° N– 5° S,

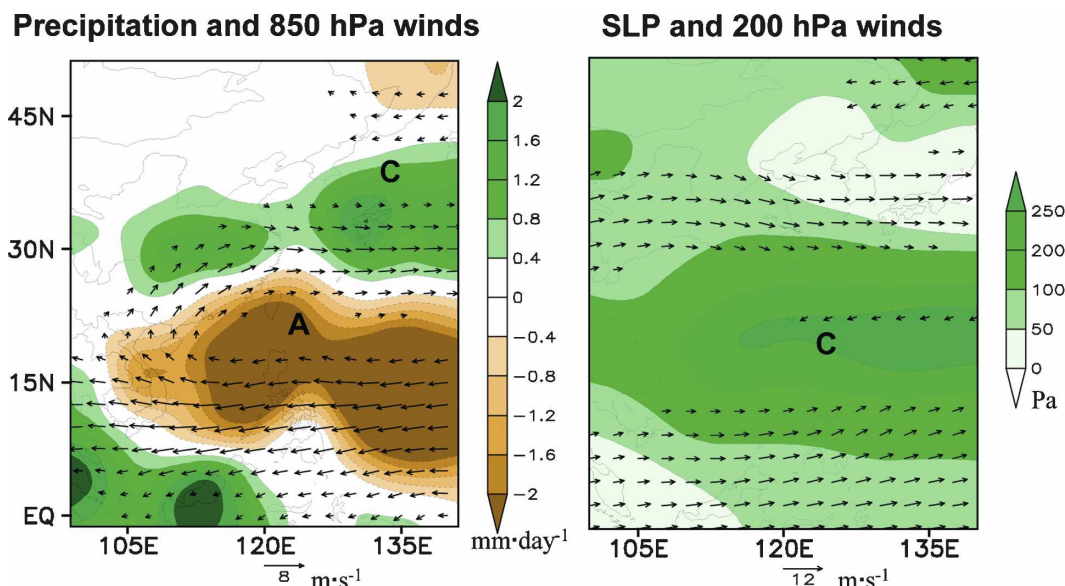


FIG. 5. The same as in Fig. 2a except that the leading mode was derived by use of the covariance matrix analysis, which yields accurate amplitude of the anomalies and accurate fractional variance. The corresponding PC1 is not shown because it is identical to that shown in Fig. 1b.

150°–90°W) or Niño-3.4 index is conveniently used for ENSO monitoring. The reason is simple: the Niño-3.4 index represents the variation in the activity center of the variability and it is highly correlated with the leading SST mode.

In a similar manner, we can define a simple yet highly relevant EASM index, which facilitates real-time monitoring. Among the 25 indices, the shear vorticity index defined by Wang and Fan (1999) (WF index hereafter) is best correlated with the leading PC of EASM (correlation coefficient is -0.97) and may be considered as a potential candidate. The WF index was defined by the U_{850} in (5° – 15° N, 90° – 130° E) minus U_{850} in (22.5° – 32.5° N, 110° – 140° E). This index was originally designed as a circulation index to quantify the variability of the WNP summer monsoon. Then, what is the physical basis for the WF index to measure EASM intensity? To what extent can the WF index represent the EASM variation?

Physically, the WF shear vorticity index reflects the variations in both the WNP monsoon trough and subtropical high. The two subsystems are the key elements of the EASM circulation system (Tao and Chen 1987). The WF index was originally designed to signify the rainfall variation over the northern SCS and Philippine Sea (Wang and Fan 1999). This function is confirmed here by the high correlation coefficient between the WF index and the GPCP precipitation anomaly averaged over that northern SCS–Philippine Sea region (10° – 20° N, 110° – 140° E), which is 0.80 for the 28-yr pe-

riod of 1979–2006. Thus, the WF index offers an excellent measure for the latent heat source over the Philippine Sea, which exerts fundamental influences to EASM (Nitta 1987; Huang and Wu 1989). As such, the WF index could potentially provide a useful measure of the subtropical and extratropical EASM. Indeed, Lee et al. (2005) has focused on the JJA precipitation variations over a large subtropical and extratropical region (20° – 50° N, 100° E– 180°). They found the leading EOF mode of the interannual variation in precipitation is closely linked to the WF index with a correlation coefficient of 0.71 for the period of 1979–2004.

The WF index not only represents well the leading modes of tropical and subtropical–extratropical rainfall variability but also represents extremely well the low-level monsoon wind variability. B. Wang et al. (2001) have shown that the correlation between the WF index and the leading PC of the 850-hPa wind anomalies in a large domain (5° N– 45° N, 100° – 170° E) reaches 0.88 for a 50-yr period (1948–97). More inspiring, as found in the present study, the WF index has the highest correlation, among all 25 circulation indices, with the leading MV-EOF mode of the EASM system; the correlation coefficient is -0.97 (Table 1). That means that PC1 of the EASM is nearly identical to the negative WF index. All these facts indicate that the WF index represents the leading modes of the large-scale EASM variations with high fidelity. In addition, the WF index can be conveniently monitored on a variety of time scales ranging from daily to seasonal. In fact, a pentad mean

WF index has been shown to be an excellent index for describing the year-to-year variation of the SCS summer monsoon onset (Wang et al. 2004). Therefore, we recommend that the WF index with a reversed sign be used as a simple index for monitoring EASM.

5. Discussion

Our aim is not to add another index to the already crowded world of EASM indices; rather we aim at clarifying the controversial issues associated with the EASM indices. There are a number of outstanding issues that deserve further clarification and discussion.

a. Can we construct a unified EASM index?

Why have there been such a large number of indices proposed? One of the major reasons is the complexity of the EASM variability. As mentioned in the introduction, the Indian monsoon occurs within the South Asian monsoon trough and this uniformity allows the average of all Indian rainfall to be used to measure its variability. Unlike the tropical ISM, the EASM encompasses the tropics, subtropics, and midlatitude. As a result, the climate variability of the EASM precipitation is highly variable in the meridional direction. For this reason, southerly or southwesterly winds were preferably used to construct monsoon indices. However, the wind anomalies in the EASM are also inhomogeneous (Fig. 2a), because the circulation is tightly coupled with precipitation that provides a heat source for driving the circulation.

Note that while the traditional monsoon definition used solely winds (Ramage 1972) due to historical reasons, the contrast between rainy summer and dry winter is a fundamental character of monsoon climate (Webster 1987). Further, changes in precipitation are far more relevant for food production and water supply than a wind change. The precipitation heating plays the most important role in determining atmospheric general circulation and hydrological cycle. Therefore, a useful monsoon index must have clear implications on rainfall intensity.

Given the high level of the inhomogeneity and complexity, how can one construct an appropriate index that is more broadly applicable to measure the EASM intensity? What is the physical basis for constructing such an index? In the present study, we have made considerable effort to address these questions. We used the leading mode of EASM variability as the meteorological basis for measuring the monsoon strength and assessing the adequacy of the various indices. The leading mode analysis is a powerful tool. The same ap-

proach was applied to ENSO study (Weare et al. 1976) and recently to Indian summer monsoon study (Straus and Krishnamurthy 2007). In conclusion, we have recommended the leading PC of EASM as a unified measure for the intensity of EASM, because this new index is extremely robust, captures a large portion (50%) of the total variance of the precipitation and three-dimensional circulation, and has unique advantages over all the existing indices (see previous section). We also recommend a simple index, the reversed WF index, simply because it is nearly identical to the leading PC of the EASM and it greatly facilitates real-time monitoring.

b. How should we define a strong Chinese summer monsoon?

A positive value of the new unified index features a strong mei-yu/baiu/changma. The correlation coefficient between the unified index and the rainfall anomaly averaged over the mei-yu region (27° – 35° N, 105° – 125° E) is 0.64 for the period of 1979–2006, indicating that the PC1 reflects reasonably well the variation of the total amount of mei-yu rainfall. Note also that the rainfall pattern of the leading mode of EASM (Fig. 2a) in the continental China region is extremely similar to that of the leading EOF mode of the rainfall anomalies derived based on Chinese rain gauge data for the period of 1951–90 (Ye and Huang 1996, p. 32), indicating that the new index captures the leading mode of rainfall variability in China very well.

However, the conventional concept of a strong summer monsoon used by Chinese meteorologists implies a weak mei-yu. Traditionally, a strong summer monsoon in China has been perceived as extensive southerlies penetrating inland to northern China, which corresponds to increased rainfall in northern China and a deficient mei-yu. Thus, the traditional notion on the strength of the Chinese summer monsoon emphasizes the northern China rainfall and has an opposite meaning to the unified index.

While this conventional notion has been dominant and repetitively occurring in Chinese monsoon literatures, we could not locate the earliest literature that originally articulates this notion. One of the possible reasons for this is that the Chinese term for the word “monsoon” means exclusively “seasonal wind” (e.g., Ding 1994) and it does not contain any connotation of rainfall. Thus, the northern limit of the southerly penetration was emphasized. To some extent, this conventional wisdom might also be motivated by the spectacular seasonal advance of the rainy season, which starts from 20° N in mid-May all the way to 45° N in mid-July (Guo and Wang 1981; Tao and Chen 1987; Ding 1994;

Wang and LinHo 2002). This unique phenomenon has drawn enormous attention since the classic work of Tu and Huang (1944). From the point of view of seasonal march, the rainy season in northern China may be a possible indication of the monsoon intensity. However, the monsoon index is designed to depict the year-to-year variation in summer monsoon intensity. If the year-to-year variability is not controlled by seasonal march, then taking northern China as a focus might be inappropriate. Unfortunately, this is indeed the case as shown in the next paragraph.

To reveal the behavior of the seasonally evolving (interannual) anomalies and in particular to see whether the year-to-year variation also features northward migration, we have applied “season-reliant EOF” (S-EOF) analysis. The details of the S-EOF are described in Wang and An (2005). In summary, in the S-EOF analysis, the principal component remains a yearly time series but the corresponding eigenvectors consist of a sequence of anomaly patterns that varies with season. To reduce the influence of the intraseasonal variation, we took a sequence of three bimonthly anomalies for each year, May–June (MJ), June–July (JJ), and July–August (JA), as a yearly “block” for the S-EOF analysis. The leading S-EOF structure and the corresponding PC are presented in Fig. 6. The PC (Fig. 6b) is very similar to the leading PC of the MV-EOF analysis (Fig. 1b). The eigenvector shows seasonally evolving spatial patterns for MJ, JJ, and JA. It is seen that the anomalies for MJ, JJ, and JA are similar, although not the same; all resemble the spatial pattern shown in Fig. 2a. The result here suggests that while the EASM rainbelt migrates northward with season, the year-to-year variability of the EASM rainfall is a “persistent” mode. This result demonstrates that the interannual anomaly is not a modification to the seasonal march; rather, it has its own pattern that persists through the May–August season. Why? It is important to recognize that the causes responsible for the seasonal march and for the interannual variation are fundamentally different. The former is driven by the external solar forcing, while the latter is primarily determined by the internal feedback processes within the coupled climate system. The fundamental causes for the interannual variation of the East Asian monsoon system are the impacts of ENSO and the monsoon–warm pool ocean interaction (Wang et al. 2000). These two processes occur in a time scale longer than the seasonal march, resulting in a persistent anomaly pattern through the entire summer.

We argue that there are several downsides with the traditional definition that emphasizes northern China rainfall over the mei-yu rainfall. First, during summer the total amount of rainfall in northern China is only a

fraction of that in the mei-yu region (Ding 1992; Chang et al. 2000a). Second, the largest rainfall variability (anomaly) is also located in the mei-yu region rather than in northern China. Third, mei-yu better represents the variability of the large-scale EA subtropical monsoon rainfall and associated subtropical southwesterly over the EA. Fourth, the intensity of mei-yu reflects closely the variation of the EASM circulation system, which consists of the WNP subtropical high, the WNP monsoon trough, and the subtropical trough (Fig. 2a). Conversely, variations in northern China rainfall do not correspond as well to the changes in EASM system. Increase in northern China rainfall is often caused by teleconnection with the Indian monsoon through “silk-road” teleconnection (Enomoto et al. 2003) or circum-global teleconnection (Ding and Wang 2005) rather than a change in the major EASM circulation system.

In all aspects, one has reasons to believe that mei-yu/baiu/changma rainfall, which is produced in the primary rain-bearing system, the EA subtropical front, is a good indicator of the EASM strength. Yet, as mentioned earlier, the traditional Chinese definition of a strong summer monsoon means a weak mei-yu. The latter is the opposite of the definition used by other monsoon communities. In all other monsoon regions around the world, abundant rainfall within the major local rain-bearing monsoon system is considered to be a strong monsoon. For example, abundant rainfall in Ganges River valley within the Indian monsoon trough means a strong Indian monsoon. Also a strong baiu or changma is never considered to be a weak monsoon in Japan and Korea. Since the mei-yu front is the major rain-bearing system for EASM, it is more meaningful to call a season of an abundant mei-yu a strong monsoon season. In this way, the new definition will be consistent with the worldwide convention. More importantly, the new definition will be consistent with the fact that the mei-yu system is the most important rainfall-producing agent of the EASM and therefore the most important provider of the heat source that drives the EASM.

In summary, we propose that the Chinese monsoon research community consider adopting our suggestion of using the first principal component of the MV-EOF mode or a negative WF index as an objective measure of the EASM strength, so that a strong Chinese summer monsoon means an abundant mei-yu, which would have the same universal meaning as all other regional monsoons worldwide.

c. Understanding of the potential limitations of the new unified index

Like any index, the new EASM index has its potential limitations. One of them (lack of monitoring capa-

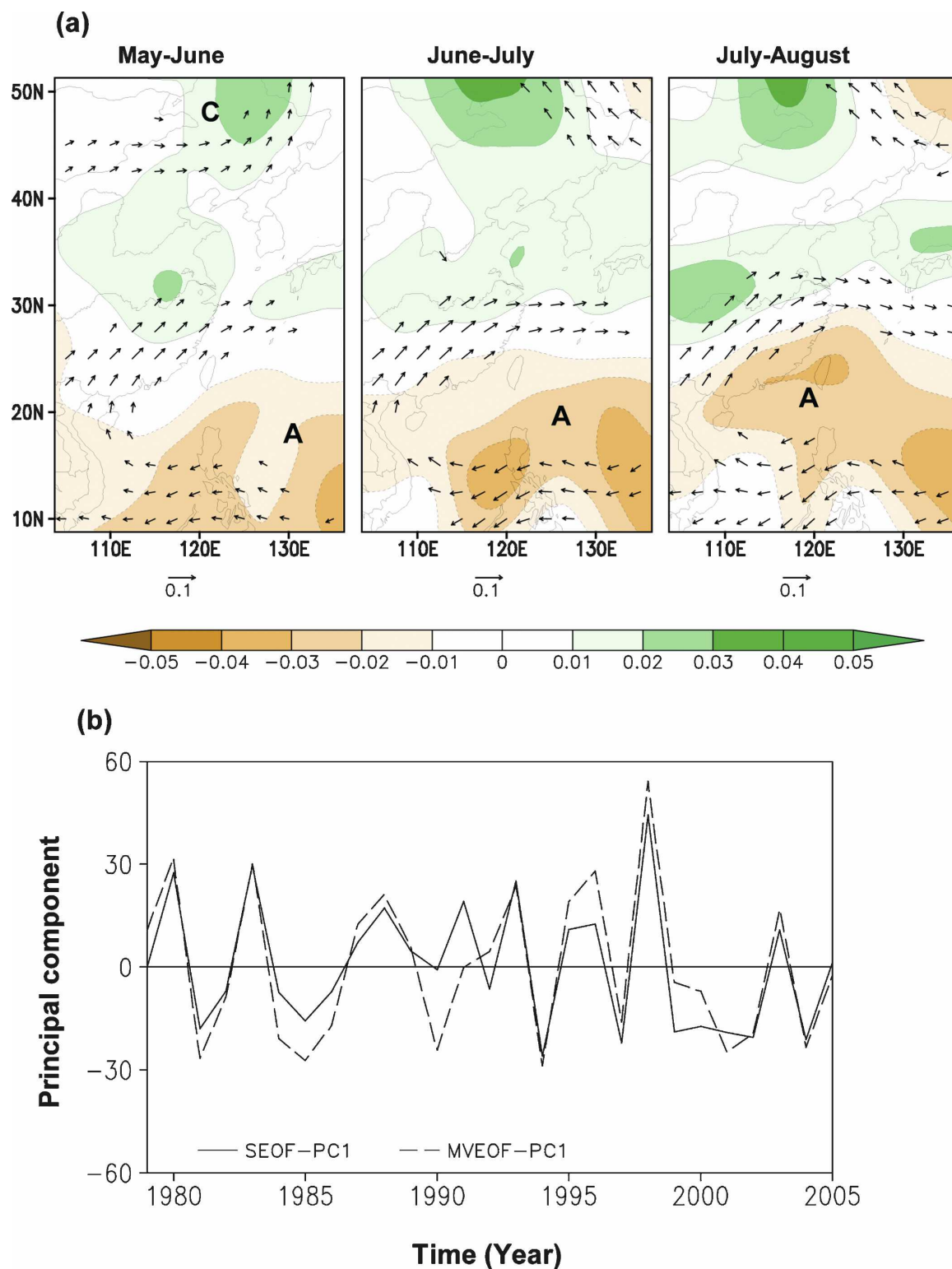


FIG. 6. (a) Spatial structure of the leading seasonal-reliant MV-EOF analysis of the EASM precipitation and 850 hPa-wind anomalies, which describes seasonally evolving patterns for MJ, JJ, and JA. Anticyclone and cyclone are denoted by “A” and “C,” respectively. All quantities are nondimensional as they were derived from the correlation coefficient matrix. (b) The corresponding principal component in comparison with the PC obtained by MV-EOF analysis (Fig. 1b).

bility) has been discussed in the previous section. The second is associated with its representativeness of the rainfall over the entire EASM region. A strong mei-yu often means less rainfall in northern China (e.g., Huang 2004) and southern China. The rainfall anomalies associated with the dominant mode are primarily dipole-enhanced mei-yu/baiu and suppressed rainfall in the northern SCS and PS, a pattern reflecting the robust Pacific–Japan pattern (Nitta 1987; Huang and Lu 1989). The decrease of rainfall in northern China and southern China shown in the spatial pattern of the leading mode is suggestive but not significant. This weakness is rooted in nature, and one cannot do much about it. Recall the Pacific SST analog we have made in the previous section: when eastern equatorial Pacific warms, the western Pacific experiences a complementary cooling. It is the inhomogeneity in rainfall anomaly that does not allow any single index to represent the entire EASM rainfall anomaly. This weakness simply reflects the fact that the rainfall variations in northern China (36°–42°N), the mei-yu region (27°N–35°N), and southern China (20°–26°N) do not completely share the same interannual variability, partly because the northern China rainfall is affected by midlatitude processes while southern China is affected by tropical cyclones. These regional differences should not be viewed as a depreciation of the significance of the dominant mode; rather, it suggests that some complementary regional indices may be necessary.

It is worth mentioning that the new index measures the EASM variations on interannual to interdecadal time scales. On geological time scale or orbital scale, whether it applies or not remains to be seen. In the mountain lift experiment, for instance, the uplift of the Tibetan Plateau may extend EASM domain poleward (Kitoh 2002; Wu et al. 2005). In that case, an alternative index, which describes the northward advance of the monsoon climate, might be relevant. On the orbital time scale, the northern edge of the southerly monsoon responds to the solar forcing, thus the focus on northern China might be an appropriate idea. However, as far as the monsoon intensity changes on interannual to decadal time scales are concerned, the northern China variation is much less important than the changes in the Yangtze–Huai River region as demonstrated in the present analysis. Again, we reiterate that the causes responsible for the variations on geological and orbital time scale and for the variations on interannual time scales are fundamentally different. The former is driven by external forcing (changes in solar radiation for a given land–sea configuration) (Clemens 2006), while the latter is primarily determined by the internal feedback processes within the coupled climate system—the

ENSO and monsoon–warm pool ocean interaction (Wang et al. 2000, 2003). Therefore, the importance of the northern China variation perceived in the conventional concept might be suitable for quantifying paleomonsoon variations but not for the interannual–interdecadal variation.

Finally, the EASM is not a stationary system. For example, Chang et al. (2000b) have shown that the phase relationship between the mei-yu rainfall along the Yangtze River region and that in the southeast coast region of China can vary between in phase, out of phase, and uncorrelated due to interdecadal changes of sea surface temperatures. Due to the limitation in the precipitation record over the ocean, we have only been able to evaluate the EASM interannual variation over the last 28 yr. The leading modes might change if the time scale becomes longer. The long-term variations of the leading mode of EASM on the multidecadal, centennial, orbital, and geological time scales remain elusive, but the unique importance of mei-yu rainfall in EASM most likely remains.

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