Long-Term Variations of Broad-Scale Asian Summer Monsoon Circulation and Possible Causes

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ABSTRACT

The widely applied Webster–Yang index (WYI), which measures the broad-scale dynamical features of the Asian summer monsoon (ASM), has experienced robust interannual and interdecadal variations and a decreasing tendency, with apparent shifts in 1972. The WYI exhibits moderate variability and frequent positive phases before 1972, intensive interannual variability during 1972–98, and an obvious decreasing tendency and mainly negative phase afterward. The vertical shear easterly anomalies over the tropics/subtropics and the anomalous vertical shear anticyclonic circulation over Eurasia (Eu) are the background for the decreasing WYI, associated with reduced summer precipitation around the Bay of Bengal and Sumatra. On interdecadal time scales, the negative (positive) Atlantic multidecadal oscillation (AMO) is characterized by cooling (warming) in Eurasian tropospheric temperature (TT) via the North Atlantic Oscillation. Global warming manipulates the increasing tendency and the interannual variability of TT over the Indian Ocean (IO). The mutual effects of AMO on Eurasian TT and global warming on Indian Ocean TT correspond to the similar decreasing tendency and interdecadal shift of the difference in TT between Eurasia and the Indian Ocean (EuTT—IOTT) with those of the ASM. Thus, the AMO and global warming seem to cause the interdecadal variability of ASM. Although the interannual relationship between Niño-3 SST and ASM weakens recently as a result of the weakening tendency of ASM, the Niño-3 SST still plays an important role in ASM variability via EuTT—IOTT anomalies. In addition, the WYI in the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis shows a larger decreasing tendency for 1999–2010 compared to other reanalysis products, a plausible reason for the inconsistent variations between land–sea thermal contrast and the NCEP–NCAR WYI during that period.

1. Introduction

The Asian summer monsoon (ASM), characterized by the heaviest seasonal precipitation and the most powerful heat source of the atmospheric circulation, is one of the most prominent climate systems affecting the global climate and weather (Webster et al. 1998). It has long been recognized that the ASM varies appreciably with the underlying thermal condition over the Eurasian (Eu) continent (e.g., Blandford 1884; Verma 1980; Liu and Yanai 2001; Yu et al. 2004; Zhou and Zhang 2009; Yang et al. 2011). For example, lighter (heavier) snow over western (eastern) Eurasia during winter is conducive for more (less) summer monsoon rainfall over India (Kripalani and Kulkarni 1999). In general, the warmer (cooler) Eurasian continent is associated with stronger (weaker) monsoon circulation. Other major advances in the monsoon study include finding the strong link between the monsoon and El Niño–Southern Oscillation (ENSO; e.g., Walker 1924; Rasmusson and Carpenter 1983; Zhang et al. 1999). A weak (strong) ASM is related to a warm (cold) ENSO event through an anomalous Walker circulation driven by the tropical eastern Pacific sea surface temperature (SST) anomalies (Webster and Palmer 1997). However, the impact of ENSO events is also modulated by the decadal variability in Indian monsoon rainfall (Kripalani and Kulkarni 1997).

Over the past half century, global warming has been underway (e.g., Hoerling et al. 2008) and decadal abrupt...
These effects lead to anomalies of meridional thermal wave trains and the North Atlantic Oscillation (NAO). The cooling of the Atlantic Ocean SST influenced the thermal condition over Eurasia via associated atmospheric changes such as Rossby wave trains and the North Atlantic Oscillation (NAO). These effects lead to anomalies of meridional thermal gradient, causing changes in the Asian monsoon (Goswami et al. 2006; Lu et al. 2006). In addition, the impact of aerosols on ASM has received research interest in the last decade (Menon et al. 2002; Ramanathan et al. 2005; Lau and Kim 2006; Massimo et al. 2011). However, these studies reported that the plausible scenarios of aerosol impacts strongly depended on aerosol distribution and characteristics and showed large discrepancies among different analyses.

Large fluctuations of the Asian monsoon are often associated with floods, droughts, and other extreme climate events. Agricultural harvests, water management, and fishery yields, among others, are directly affected by the ASM. These pronounced influences and many other issues urge the development of monsoon indices that provide a simple characterization of the state of the monsoon (Miyakoda et al. 2003). The fundamental mechanism for driving ASM is the land–sea thermal contrast (Webster 1987). Based on the relationship between the heating over Asia and regional vertical wind shear (Webster 1972; Gill 1980), Webster and Yang (1992) defined an ASM index as the vertical shear of the June–August (JJA) 850- and 200-hPa zonal winds averaged over 0°–20°N, 40°–110°E [hereafter the Webster–Yang index (WYI)]. It depicts the thermally driven nature of the monsoon and has been widely used in both research and operational prediction of the ASM. The ASM system is very complicated, with distinct spatial features over tropical South Asia and subtropical East Asia. In addition, different indices need to be defined for different fields and purposes (e.g., Parthasarothy et al. 1992; Webster and Yang 1992; Wang and Fan 1999; Goswami et al. 1999; Lau et al. 2000; Hao et al. 2005). Therefore, it is impossible to derive a universal index to measure all monsoon features over the entirety of Asia. As a measure of the large-scale monsoon circulation over Asia and the Indo-Pacific Ocean basin including the integrated heat source over Asia, the WYI provides a general description of the broad-scale characteristics and showed large discrepancies among different analyses.

As discussed above, under the background of global warming, the ASM has experienced a complex change. Its relationships with the Eurasian continent, ENSO, the Indian and Atlantic Ocean SSTs, and aerosols have also become more complicated in recent decades. Although previous studies have discussed some of these aforementioned aspects, the variability of large-scale ASM circulation on multiple time scales in recent decades and the possible causes still need to be fully documented. The ultimate purpose of this study is to explain the interannual and interdecadal variations and decadal tendency.
of ASM in recent decades from a thermodynamic argument. The datasets used in this study are described in section 2, and the variation of ASM and possible causes are presented in sections 3 and 4, respectively. Conclusions and discussion are provided in section 5.

2. Data description

For the development of monsoon circulation indices and associated characteristics, we use the winds for 61 years (1950–2010) from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996). The monthly, 32-yr (1979–2010) data from the NCEP Climate Forecast System Reanalysis (CFSR; Saha et al. 2010), the Modern-Era Retrospective Analysis for Research and Applications (MERRA; Rienecker et al. 2011) from the National Aeronautics and Space Administration, and the Japanese 25-year Reanalysis (JRA-25; Onogi et al. 2007) from the Japan Meteorological Agency are also used to demonstrate the robustness of the results obtained from the NCEP–NCAR reanalysis.

We also use the 61-yr (1950–2010) National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) gauge-based monthly global land precipitation data (Chen et al. 2002). The 32-yr (1979–2010) CPC Merged Analysis of Precipitation (CMAP) data (Xie and Arkin 1997) are used to investigate rainfall variability over oceans. The two precipitation datasets were derived from multiple sources including gauge observations, satellite infrared and microwave estimates, and the NCEP–NCAR reanalysis. Other datasets used include the improved extended reconstructed sea surface temperature, version 2 (Smith and Reynolds 2004), and the CPC monthly land surface air temperature analysis (Fan and van den Dool 2008).

Several indices of large-scale atmospheric and oceanic variations are also used in this study. They are the Atlantic multidecadal oscillation (AMO; available at http://www.esrl.noaa.gov/psd/data/timeseries/AMO), the North Atlantic Oscillation (NAO; available at ftp://ftp.cpc.ncep.noaa.gov/wd52dg/data/indexes/nao_index.tim), and the global-mean land and sea temperature indices (from http://www.esrl.noaa.gov/psd/data/correlation/gmsst.data).

3. Variability of the Asian summer monsoon

Figure 1a presents the time series of WYI for the period of 1950–2010. Clearly, besides the negative trend discussed in Zuo et al. (2012), an apparent feature is that the WYI exhibits different variations in different periods of time. A moving t test technique (Xiao and Li 2007b) indicates a significant decadal abrupt change of the mean value in 1972. The index before 1972 was generally higher than that afterward and exhibited moderate variability, with the smallest value about 24.0 m s⁻¹. For 1972–98, the monsoon variation became more intensive and the smallest value was below 23.0 m s⁻¹. The averages (standard deviations) of WYI for 1950–71 and 1972–98 were 26.24 (0.99) and 25.39 (2.56), respectively. The differences in both mean values and standard deviations between 1950–71 and 1972–98 exceed the 95% confidence level (the Student’s t test used in this study). The WYI in the last decade (1999–2010) experienced a rapid decreasing tendency, which was different from the variability in previous decades. The linear trend for this period exceeded the 99.9% confidence level (F test). In addition, the mean value is 24.29 m s⁻¹, less than the values in 1950–71 and 1972–98.

In the following discussion, we focus on the difference between 1972–98 and 1950–71 in terms of the mean value and the standard deviation of WYI, as well the difference in the mean value between 1999–2010 and 1950–71 and the tendency in 1999–2010.

Figure 1b shows the normalized WYI and the zonal winds at 850 and 200 hPa averaged over 0°–20°N, 40°–110°E (hereafter WYI.U850 and WYI.U200). As expected, the WYI was generally positive before 1972 and mainly negative after 1998, with robust variations during
By definition, the variability of WYI.U200 is opposite that of WYI. Although mainly positive WYI.U850 anomalies occurred before 1972 and frequently negative WYI.U850 anomalies appeared after 1998, the WYI.U850 in 1972–98 exhibited a weakening tendency, different from the WYI. That is, the WYI.U850 showed a persistently decreasing tendency in the past decades, whereas the WYI exhibited a more obvious three-period characteristic due to the variability of WYI.U200.

Figure 2 shows the mean horizontal winds at 850 and 200 hPa for 1950–71 and their differences from 1972–98 and 1999–2010. The largest lower-tropospheric westerly and upper-tropospheric easterly flows appear in the domain where the WYI is defined (Figs. 2a,d). Nevertheless, the most intensive differences between 1950–71 and both 1972–98 and 1999–2010 do not occur in the WYI domain. For the lower troposphere, an abnormal anticyclonic circulation pattern occurred over the Asian landmass, and the easterly anomalies over the WYI domain seemed to be part of the pattern (Fig. 2b). Compared with 1972–98, the anomalous anticyclonic circulation became stronger, and the easterly anomalies became more significant over the whole WYI domain for 1999–2010 (Fig. 2c). However, the differences in variance between 1950–71 and both 1972–98 and 1999–2010, as well the tendency for 1999–2010, were not significant for the whole WYI domain (Figs. 2b,c).

For the upper troposphere, the prominent variability of the zonal wind is the significant weakening of easterly wind between 15°S and 10°N, which contributes to the variations of WYI.U200. The difference between 1950–71 and 1999–2010 was significant in most of the WYI domain. Moreover, the difference in variance between 1972–98 and 1950–71 was significant over the WYI domain (Fig. 2e). However, the tendency in 1999–2010 was insignificant (Fig. 2f). Additionally, the extratropical westerly jet stream intensified significantly from 1950–71 to 1972–98 (Fig. 2e). Notable meridional wind anomalies appeared over the midlatitudes, represented by an anomalous cyclonic circulation over eastern Europe and
an anomalous anticyclonic circulation along the east coast of the Eurasian continent in 1999–2010 (Fig. 2f).

Figure 3a shows the JJA vertical shears defined by the difference between 850 and 200 hPa horizontal winds, $\mathbf{U}_{850} - \mathbf{U}_{200}$, for 1950–71. Eurasia is dominated by a vertical shear cyclonic pattern, with maximum westerly shear exceeding 25 m s$^{-1}$ over $0^\circ$–$20^\circ$N, $40^\circ$–$110^\circ$E. The westerly vertical shear provides a first-order approximation of the strength of the gravest baroclinic mode, which represents the latent heat release in the midtroposphere and dominates the monsoon circulation. Thus, the definition of WYI is an appropriate measure of the variability of ASM circulation. Several other dynamical monsoon indices, for example, the South Asian monsoon index (Goswami et al. 1999), the Southeast Asian monsoon index (Wang and Fan 1999), and the East Asian monsoon index (Lau et al. 2000), have also been defined. Although these indices may show a stronger correlation with rainfall over particular regions, the WYI presents the intrinsic thermally driven nature of the monsoon and depicts the features of the monsoon over a much broader scale.

Compared with the zonal wind shears for 1950–71, significant easterly shear anomalies occurred in most of the Indian Ocean and the western Pacific, except part of the Arabian Sea and the Bay of Bengal, in 1972–98 (Fig. 3b). This feature presents a decreasing WYI and thus weakening ASM. In addition, the anomalous westerly shear over the midlatitude Eurasian continent...
and northerly shear along eastern China indicate a weakened cyclonic shear circulation. The U&V850 – U&V200 anomalies in 1999–2010 were similar to those in 1972–98 except that the cross-equatorial flow anomaly, the easterly shear over the tropics, and the vertical shear anticyclone over the Eurasian landmass were larger (Fig. 3c). Moreover, as expected, the difference in standard deviation between 1950–71 and 1972–98, as well the tendency in 1999–2010, were significant in much of the WYI domain.

Figure 3d displayed the JJA precipitation in 1950–71. Heavy rainfall centers exceeding 80 mm day$^{-1}$ were mainly located around the Bay of Bengal and over the western Pacific Ocean, a portion of India, the Indochina peninsula, and Indonesia. The precipitation over these centers generally decreased from 1950–71 to 1972–98 and the difference between 1950–71 and 1999–2010 became more obvious (Figs. 3e,f). The difference was significant between 1950–71 and 1999–2010 surrounding the Bay of Bengal, which is the center of latent heat release. The decreasing rainfall is consistent with the weakening monsoon circulation shown in the WYI. However, the change in East Asian monsoon precipitation is different. Although the JJA rainfall over East Asia decreased from 1950–71 to 1972–98 as a whole, positive precipitation anomalies appeared to the south of 40°N in China, Korea, and Japan and negative anomalies appeared over northeastern China after 1998. Li et al. (2010) reported that the East Asian summer monsoon experienced a change in position, instead of intensity, for 1958–2008.

To investigate the decadal variation of monsoon precipitation over both land and oceans, the difference in composite CMAP rainfall between 1999–2010 and 1979–98 are also analyzed (Fig. 4). Reduced rainfall appears over the India subcontinent, the Bay of Bengal, the South China Sea, and the Philippines, indicating a weakening Asian summer monsoon. Over East Asia, the reduced rainfall over northeastern China and enhanced rainfall to the south of 40°N in China and Korea are consistent with the feature illustrated by Fig. 3f. Additionally, negative rainfall anomalies are shown over the equatorial central Pacific and the southern Indian Ocean, whereas positive precipitation anomalies appear to both the north and the south of the equatorial western Pacific.

4. Possible causes of ASM variability
a. Ocean and land surface temperatures

To investigate the possible causes of monsoon decadal variations, the composite differences of 1950–71, 1972–98, and 1999–2010 from the climatology in JJA 2-m air temperature over land (T2m) and SST are shown in Figs. 5a–c. As expected, the principal feature is the global warming signal. Except the cooling tendency from 1950–71 to 1972–98 in the North Pacific, most of the globe has been warming in recent decades, with the largest warming over the North African landmass and the Indian Ocean. Overall, the amplitude of enhanced surface temperature over land is larger than that over the oceans because the thermal capacity of soil is much less than that of water.

The correlation between WYI and SST and T2m changes with global warming, and the most prominent feature is the weakening of the long-recognized negative relationship between ENSO and ASM. The significant correlation between WYI and the Pacific SST covered the whole eastern-central tropical and northern Pacific in 1950–71, which withdrew eastward to the eastern tropical Pacific in 1972–98 and totally disappeared afterward. Previous studies have reported that the relationship between ENSO and the Indian monsoon was weakening in the greenhouse warming scenario (e.g., Kumar et al. 1999; Ashrit et al. 2001). Here, we further illustrate that the relationship between ENSO and the broad-scale ASM, not only the Indian monsoon, is also weakening. Additionally, the negative relationship between WYI and the southwestern Pacific SST weakened from 1950–71 to 1972–98 and reversed after 1998. Generally, the correlation between WYI and the land surface temperature is positive. The correlation also exhibits an interdecadal variation, but the variation is much smaller than that shown in the correlation between WYI and SST.
On the contrary, the link between WYI and SST over the IO becomes stronger. A strong negative relationship was only limited to a small part of the intertropical western IO along the East African coastal regions in 1950–71 but extended to the Arabian Sea in 1972–98. For 1999–2010, significant relationships appeared over the Bay of Bengal, the Arabian Sea, and the northern and southwestern IO. Based on the significant relationships between WYI and SST over IO (5°S–20°N, 40°–100°E) and Niño-3 (5°S–5°N, 150°–90°W), we show the time series of normalized SSTs over the Niño-3 region and IO and minus WYI for convenience (Fig. 6). Clearly, the interannual relationship between WYI and Niño-3 SST is strong, with high Niño-3 SST generally corresponding with low WYI. For 1999–2010, Niño-3 SST was not associated with WYI due to the rapid decrease in WYI. Although the interannual relationship between IO SST and WYI is not as significant as that between Niño-3 SST and WYI in general, the IO SST–WYI relationship was stronger than the Niño-3 SST–WYI relationship for 1999–2010 because a strong tendency appeared in both WYI and IO SST but not in Niño-3 SST. The large sensitivity of western IO SST to the above atmospheric anomalies may be the reason for the obvious increasing tendency in IO SST (Wu and Kirtman 2007).

![Fig. 5. Composite differences of T2m and SST between (a) 1950–71, (b) 1972–98, and (c) 1999–2010 and climatology (K). Correlation coefficients between difference of T2m and SST and WYI during (d) 1950–71, (e) 1972–98, and (f) 1999–2010. Dark brown (green) shaded coloring denotes the correlation coefficients exceeding the 95% confidence level.](http://journals.ametsoc.org/jcli/article-pdf/26/22/8947/4014281/jcli-d-12-00691_1.pdf)
b. Lower-tropospheric air temperature

Land–sea thermal contrast anomalies and the Coriolis force dominate the variability of the westerly monsoon flow. As discussed above, the amplitude of surface temperature enhancement over land is larger than that over the oceans, indicating an increased land–sea temperature gradient. However, because of the different thermal capacity of water and soil, the lower-tropospheric temperature, which can also depict the surface thermal condition, may be a better measure for depicting the variations in land–sea thermal contrast than surface temperature. Figure 7 shows the composite differences between 1950–71 and both 1972–98 and 1999–2010 for 700-hPa temperature (TMP700), surface level pressure (SLP), and 500-hPa geopotential high (HGT500). Compared with the warming across most of the globe, a significantly prominent cooling is found over the Eurasian landmass, with the coldest period in 1972–98 (Figs. 7a,d). Thus, the meridional and zonal thermal contrasts between the Eurasian continent and its adjacent regions have become relatively smaller, corresponding to weaker westerly wind over the northern IO and weaker southerly wind along the east of China, as shown in Fig. 2. The significant positive SLP and HGT500 anomalies over the Eurasian continent reflect the effect of cooling in Eurasia (Figs. 7b–c and 7e–f). That is, the warm low over the Eurasian landmass in summer weakens, generally representing a weakened ASM. Zuo et al. (2012) reported that the relatively smaller warming in Eurasia compared to the surrounding regions could be a plausible reason for the weakening ASM. This study confirms that the weakening ASM is correlated with the cooling in the

**FIG. 6.** Time series of normalized minus WYI (black solid line), Niño-3 (blue solid line), and IO (green dash line) SST.

**FIG. 7.** Composite differences between 1950–71 and both (a)–(c) 1972–98 and (d)–(f) 1999–2010 in JJA (left) TMP700, (middle) SLP, and (right) HGT500. Shading denotes the composite differences exceeding the 95% confidence level and color bar denote the $t$ test values.
lower-tropospheric atmosphere. The differences between 1950–71 and both 1972–98 and 1999–2010, in terms of the variances of TMP700, SLP, and HGT500, are significant over Eurasia, a feature consistent with the variability of WYI. In addition, the larger positive SLP and HGT500 anomalies in 1999–2010 than those in 1972–98 are consistent with the decrease in WYI from 1972–98 to 1999–2010 (Figs. 7e–f). However, the warmer Eurasian TMP700 in 1999–2010 than 1972–98 (Figs. 7b–c), corresponding to the larger land–sea thermal contrast, conflicts with the decrease in WYI from 1972–98 to 1999–2010.

The different warming amplitude between different latitudes may play an important role in the variability of low-level tropospheric westerly wind under the global warming scenario. Figure 8 shows the zonal mean (30°–120°E) JJA TMP700 in 1950–71 and its composite differences of 1972–98 and 1999–2010 from 1950–71. As expected, the maximum TMP700 appears around 30°N. Compared with 1950–71, TMP700 increased over the south of 30°N and decreased over the north of 30°N in 1972–98. For 1999–2010, TMP700 increased over most of the latitudes except in 36°–48°N. Furthermore, the maximum enhancement occurred over IO in both 1972–98 and 1999–2010. Thus, the land–sea temperature gradient was smaller than that in 1950–71, presenting a weakening ASM. Nevertheless, comparing the land–sea thermal contrast anomaly in 1999–2010 with the anomaly in 1972–98 indicates that the former was smaller than the latter. That is, the land–sea thermal contrast was larger in 1999–2010 than that in 1972–98, inconsistent with the decrease in WYI from 1972–98 to 1999–2010.

Thus, the following two questions should be answered. What causes the tropospheric cooling in the Eurasian landmass and why is the change in the land–sea temperature gradient anomaly inconsistent with that in WYI for 1999–2010?

c. Tropospheric temperature, AMO, and global warming

Above, we have documented that the lower-tropospheric temperature has decreased from 1950–71 to 1972–98 over Eurasia, although the surface temperature over the Eurasian landmass increases under the global warming background. Moreover, as discussed in the introduction, the thermal condition over Eurasia is an important factor for the ASM anomalies and the AMO can produce tropospheric temperature anomalies over Eurasia by modulating the frequency of the occurrence of strong NAO events (e.g., Knight et al. 2005). Hence, we show the time series of the normalized vertically integrated temperature from 850 to 200 hPa over Eurasia (15°–40°N, 30°–120°E; hereafter Eurasian TT), AMO, and the NAO index for 1950–2010 (Fig. 9a). Generally, the positive (negative) unsmoothed AMO index is significantly associated with warm (cold) Eurasian TT, with a correlation coefficient exceeding the 99.9% confidence level ($r = 0.56$). For the interdecadal
time scale, the variability of Eurasian TT is significantly related to the phase of AMO, with the warm (cold) AMO phase corresponding to above (below) normal Eurasian TT except around the mid-1980s. The mean values of Eurasian TT in positive and negative AMO phases are $-35.047$ and $-36.051$, respectively, and the composite difference exceeds the 98% confidence level. Additionally, the Eurasian TMP700 anomalies are also positive (negative) in the positive (negative) AMO phase, similar to the case for Eurasian TT ($r = 0.80$ between Eurasian TT and Eurasian TMP700). Because the negative AMO phase (1965–94) covers most of the period of 1972–98, the AMO also explains the intensive negative TMP700 anomalies in 1972–98 in Fig. 7b.

The out-of-phase relationship between AMO and NAO on interdecadal time scales in Fig. 9a indicates the possible modulation of AMO on NAO. The out-of-phase ratio between the unsmoothed AMO and NAO is 69% for 1950–2010, and the composite difference in NAO between positive and negative AMO phases exceeds the 99% confidence level. Furthermore, the correlation coefficient between AMO and NAO is $-0.40$ for 1950–2010, exceeding the 99.8% confidence level. Indeed, previous studies have reported that when the AMO is in warm (cold) phase, the summer NAO tends to be in negative (positive) phase (e.g., Knight et al. 2005; Folland et al. 2009). The NAO is associated with significant surface temperature anomalies over Eurasia (Hurrell 1995; Chang et al. 2001; Yang et al. 2004), and the vertical structure of NAO anomalies is equivalently barotropic (Thompson and Wallace 1998), indicating that the AMO may be associated with Eurasian TT via NAO.

The vertically integrated air temperature from 850 to 200 hPa over IO ($15^\circ$S–$10^\circ$N, $30^\circ$–$120^\circ$E, hereafter IO TT) exhibits a persistently increasing tendency (Fig. 9b), consistent with the IO SST. In addition, the IO TT and the IO SST display similar interannual variability, with a correlation coefficient of 0.83 for 1950–2010. The mechanism of tropical TT response to SST forcing disclosed by Su et al. (2003) may explain the link of IO SST and IO TT. In fact, the global-mean land–sea temperature index shows significant correlation with IO TT and IO SST ($r = 0.80$ and 0.76, respectively) for 1950–2010.

As seen from Fig. 9, the variations of Eurasian TT exhibit similar variations to those of AMO, while the change in IO TT is a reflection of global warming. Clearly, neither the IO TT nor the Eurasian TT can present the variability of ASM solely. Because the ASM is driven primarily by the meridional gradient of deep-tropospheric heating, we further analyze the change in the difference between Eurasian TT and IO TT (hereafter EuTT – IOTT; Fig. 10). On the interannual time scale, EuTT – IOTT exhibits consistent variations with the WYI. It shows a significant positive correlation with WYI, with a correlation coefficient of 0.70 for 1950–2010, exceeding the 99.9% confidence level. On the interdecadal time scale, EuTT – IOTT is associated with the decreasing tendency, and the abrupt fluctuation of meridional thermal contrast in 1972 was consistent with the interdecadal variability of the ASM. In 1950–71, the EuTT – IOTT was generally higher than that afterward, and the amplitude of variability was smaller than that in 1972–98. Moreover, the difference in EuTT – IOTT between 1950–71 and 1972–98 exceeds the 99.9% confidence level. Thus, the EuTT – IOTT generally explains the variability of WYI on both interannual and interdecadal time scales, except the intensive decreasing trend for 1999–2010.

Because of the significant relationships between AMO and Eurasian TT (and between the global-mean temperature and IO TT), the difference of AMO minus global-mean temperature (hereafter AMO – Glb. Tmp) exhibits generally similar variability with that in EuTT – IOTT on both interannual and interdecadal time scales. The AMO – Glb. Tmp anomalies in 1950–71 are generally positive, and the difference between 1950–71 and 1972–98 is significant, consistent with that in
EuTT – IO TT (Fig. 10b). The Niño-3 SST, on the one hand, influences the Eurasian TT via stationary waves (Goswami and Xavier 2005) and, on the other hand, affects IO TT through modulating the IO SST. On interannual time scales, Niño-3 SST is more similar to EuTT – IO TT in both variability and amplitude compared with AMO – Glb.Tmp, although they exhibit quite different interdecadal variations (Fig. 10b). The correlation between Niño-3 SST and the EuTT – IO TT is significant, with positive (negative) Niño-3 SST anomalies corresponding to negative (positive) EuTT – IO TT anomalies in general. These features indicate that the mutual effect of AMO and global warming is associated with the variability of meridional land–sea thermal contrast on the interdecadal time scale, and the tropical central-eastern Pacific SST play a major role on the interannual variability.

Based on the above considerations, we calculate AMO minus global-mean temperature and minus Niño-3 SST (hereafter AMO – Glb.Tmp – Niño-3; black line in Fig. 10b). The AMO – Glb.Tmp – Niño-3 exhibits more similar variability with EuTT – IO TT than with AMO – Glb.Tmp and Niño-3 SST. The AMO – Glb.Tmp – Niño-3 anomalies are generally positive in 1950–71, and the difference between 1950–71 and 1972–98 is significant, similar to those in EuTT – IO TT. The correlation coefficients of EuTT – IO TT with AMO – Glb.Tmp – Niño-3, AMO – Glb.Tmp, and Niño-3 SST are 0.67, 0.30, and −0.59, respectively, indicating that the mutual effects of AMO, global warming, and the tropical central-eastern Pacific SST can better represent the variability of meridional land–sea thermal contrast than their individual effects.

In brief, the meridional land–sea thermal contrast, EuTT – IO TT, generally represents the variability of ASM, except the intensive decreasing trend during 1999–2010. The interdecadal variation of EuTT – IO TT may be the consequence of the mutual effects of the modulation of AMO on Eurasian TT and the influence of global warming on IO TT, whereas the interannual variability is mainly modulated by the tropical central-eastern Pacific SST. The AMO, global warming, and the tropical central-eastern Pacific SST may be the three key factors of ASM variability via exerting influence on meridional land–sea thermal contrast. We also examine EuTT – IO TT in the other three reanalysis products: CFSR, MERRA, and JRA-25. Although the values in the four products are not identical, the variations in all the datasets are almost the same on both interannual and interdecadal time scales. All the correlation coefficients among different reanalysis products exceed 0.90 for 1979–2010, which validates the quality of the NCEP–NCAR tropospheric temperature.

For 1999–2010, the variability in EuTT – IO TT did not exhibit a rapid decreasing tendency as appeared in WYI. We have not found any evidence that can explain the apparent decrease in WYI after 1999 so far, which may lead to questions about the quality of the NCEP–NCAR reanalysis for 1999–2010. Because there exists several reanalysis products after 1979, we also calculate the WYI using the CFSR, MERRA, and JRA-25 reanalysis products to test the NCEP–NCAR WYI for 1979–2010.

d. WYI in other reanalysis products

Figure 11 shows the time series of WYI, WYI.U850, and WYI.U200 in the NCEP–NCAR, CFSR, MERRA, and JRA-25 reanalysis products, respectively. The prominent discrepancy between the NCEP–NCAR WYI and the other three WYIs occurred in 2009 and 2010. The abrupt decrease from 2008 to 2009 in the NCEP–NCAR WYI did not appear in the others. The NCEP–NCAR WYI in 2009 and 2010 was 20.24 and 21.76 m s⁻¹, respectively, whereas the averaged values of CFSR WYI, MERRA WYI, and JRA-25 WYI in 2009 and 2010 were 23.07 and 24.46 m s⁻¹, respectively. Note that the NCEP–NCAR WYI in 1979–2008 was 25.26 m s⁻¹, whereas the averaged value of the other
three was 24.24 m s\(^{-1}\). The differences between the NCEP–NCAR WYI and the other three WYIs in 2009 and 2010 were 3 times of those in 1979–2008. The increase in WYI.U850 in the other three products and the decrease in the NCEP–NCAR WYI.U850 from 2008 to 2009 are the major causes for the difference between the NCEP–NCAR WYI and the others. The larger increasing tendency in the NCEP–NCAR WYI.U200 than the WYI.U200 in the other products also contributes to the difference in WYI.

Because the decreasing tendency in the NCEP–NCAR WYI in 1999–2010 is more intensive than that in the other products, the authenticity of the rapid tendency in the NCEP–NCAR WYI can be questionable. This feature may explain why no associated change in the response of the meridional thermal contrast to the intensive decreasing tendency can be observed.

5. Conclusions and discussion

Westerly vertical shears represent the release of latent heat in the troposphere, which dominates the monsoon circulation. The maximum westerly vertical shears in boreal summer appear over \(0^\circ\text{N}–20^\circ\text{N}, 40^\circ\text{E}–110^\circ\text{E}\), a domain used to define the broad-scale ASM index, the WYI. In the present study, we have described the characteristics of broad-scale ASM circulation on multiple time scales in recent decades using the WYI. Overall, the ASM exhibits a decreasing tendency for 1950–2010, with interdecadal shifts in 1972. In 1950–71, WYI anomalies were frequently positive and their interannual variability was only moderate. The interannual fluctuation became intensive in 1972–98, and the values were generally smaller than those in 1950–71. The difference in WYI between 1950–71 and 1972–98 significantly exceeds the 99.9% confidence level. After 1998, the WYI shows a large decreasing tendency.

The intensive anomalies of the zonal winds at both the lower and upper troposphere are not limited to the WYI domain. The prominent variability in the lower-tropospheric horizontal wind is the anomalous anticyclonic circulation over Eurasia. The easterly anomalies over the WYI domain seem to be part of the anomalous circulation pattern. For the upper troposphere, large-scale westerly anomalies appear over \(15^\circ\text{S}–10^\circ\text{N}\), contributing to much of the variability in WYI.U200 and the difference in WYI variance between 1950–71 and 1972–98. Thus, the vertical shear easterly anomalies in the tropical and subtropical regions and the vertical shear anticyclonic circulation anomalies over Eurasia provide a background for the decreasing tendency of WYI.

Because of the decreasing tendency of ASM in recent decades, the relationship between tropical Pacific SST and ASM changes with time. The domain of significant negative correlation withdrew from the entire eastern-central tropical Pacific in 1950–71 to the eastern tropical Pacific in 1972–98 and totally disappeared afterward. On the contrary, the negative relationship between IO SST and ASM became stronger because of the increasing tendency of the IO SST. The domain of significant correlation extended from a small region of the tropical western IO in 1950–71 to the Arabian Sea in 1972–98 and covered the entire northern and southwestern IO after 1998.

Under the global warming background, warming in land is larger than that in the oceans. Thus, the land–sea thermal contrast measured by changes in surface temperature cannot explain the weakening of ASM. The lower-tropospheric temperature can also reflect the atmospheric thermal condition, and may be superior to the surface temperature for presenting the land–sea temperature gradient. The lower-tropospheric temperature showed a cooling over the Eurasian landmass in 1972–98 and exhibited a persistently increasing tendency over the IO, indicating a plausible weakened land–sea thermal contrast. The SLP and 500-hPa geopotential high showed positive anomalies over Eurasia in 1972–98, and they became more apparent after 1998. The anomalies in 1972–98 were consistent with the cooling Eurasia, indicating the weakened heat low over Eurasia in summer.

Furthermore, we have investigated the deep-tropospheric heating using the vertically integrated temperature from 850 to 200 hPa. On interdecadal time scales, the Eurasian TT may be modulated by the AMO, with a negative (positive) AMO phase characterized by a warming (cooling) Eurasian continent. The negative AMO phase in 1972–95 also explained the negative lower-tropospheric temperature anomalies over Eurasia in 1972–98. The obvious increasing tendency and interannual variation in the IO TT are similar to those in the IO SST and are also manipulated by global warming. Overall, the mutual effects of AMO on Eurasia TT and global warming on TT over the Indian Ocean correspond to the decreasing tendency and the interdecadal shifts of EuTT – IOTT in 1972, consistent with those in WYI. The Niño-3 SST influences the ASM via EuTT – IOTT anomalies, primarily on interannual time scales. In addition, we have calculated the EuTT – IOTT using data in the CFSR, MERRA, and JRA-25 products. All correlation coefficients of EuTT – IOTT between the NCEP–NCAR reanalysis and other three products exceed 0.90, providing supporting evidence for the quality of the NCEP–NCAR tropospheric temperature.

Nevertheless, while the change in the meridional thermal contrast is applied to explain the variability of
WYI in the NCEP–NCAR reanalysis, it is difficult to explain the rapid decreasing tendency in 1999–2010. Given this consideration, we have calculated the WYI in the other reanalysis products (CFSR, MERRA, and JRA-25) for 1979–2010. The result shows that the decreasing tendency in the other products is smaller than that in the NCEP–NCAR WYI for 1999–2010.

It should also be pointed out that the recent studies by Zhao et al. (2011, 2012) have demonstrated that for summer, the Asian land heating plays a more important role than the Pacific SST in causing variations of the most dominant atmospheric modes over both the Asian–Pacific sector and the European–Atlantic sector. The current study also indicates the importance of Eurasian tropospheric temperature for the change in the Asian summer monsoon. In addition, although previous studies have reported relationships between aerosols and the weakening of ASM using numerical models (e.g., Menon et al. 2002; Ramanathan et al. 2005), they have also suggested that the aerosol effects on the monsoon are extremely complex. Thus, the increasing aerosols represented by the Total Ozone Mapping Spectrometer (TOMS) aerosol index (not shown) may play a role on the weakening ASM, but more evidence of support needs to be explored.

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