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Interdecadal Variability of the East Asian Summer Monsoon in an AGCM

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ABSTRACT

It is well known that significant interdecadal variation of the East Asian summer monsoon (EASM) occurred around the end of the 1970s. Whether these variations can be attributed to the evolution of global sea surface temperature (SST) and sea ice concentration distribution is investigated with an atmospheric general circulation model (AGCM). The model is forced with observed monthly global SST and sea ice evolution through 1958–1999. A total of four integrations starting from different initial conditions are carried out. It is found that only one of these reproduces the observed interdecadal changes of the EASM after the 1970s, including weakened low-level meridional wind, decreased surface air temperature and increased sea level pressure in central China, as well as the southwestward shift of the western Pacific subtropical high ridge and the strengthened 200-hPa westerlies. This discrepancy among these simulated results suggests that the interdecadal variation of the EASM cannot be accounted for by historical global SST and sea ice evolution. Thus, the possibility that the interdecadal timescale change of monsoon is a natural variability of the coupled climate system evolution cannot be excluded.

Key words: global sea surface temperature, sea ice, East Asian summer monsoon, interdecadal change

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1. Introduction

East Asia is one of the most significant monsoon regions, and its summer climate is sensitive to the advancement and retreat of the summer monsoon. The onset and retreat of the summer monsoon of East Asia (EASM) represent a zonal-oriented rain belt moving northward and southward, and determine the summer seasonal cycle in this region. Anomalous activity of the EASM is usually severe for agriculture production and exerts significant influence on the economy of this region. Therefore, exploring the reason for EASM activity has been an important topic for the climate research community.

Recently, many studies have revealed that the EASM underwent a substantial decadal weakening since the late 1970s (Huang et al., 1999). As a result, summer rainfall decreased sharply in both North and South China but increased greatly in the mid-low reaches of the Yangtze River valley (Chen, 1999; Huang et al., 1999; Lu, 2002; Fong et al., 2005; Zhao and Zhou, 2006). Along with these rainfall variations, surface air temperature also exhibited remarkable variations, in which the northern part of China became warmer than normal. In contrast with the warming trend in North China, surface air temperature in a region downstream of the Tibetan Plateau experienced significant cooling (Yu and Zhou, 2004). Such a summer monsoon weakening is also exhibited as northerly wind anomalies over the eastern part of China (Wang, 2001).

A number of studies have been conducted to investigate the factors responsible for this interdecadal EASM change. Since an observed significant increase in aerosol release associated with rapid industrialization in China occurred alongside this interdecadal EASM change, researchers naturally speculated that the two trends were linked (Xu et al., 2001). To ex-

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plore this association, Menon et al. (2002) investigated the climate effect of absorbing black carbon aerosol on precipitation and temperature in China and India via alteration of the aerosol optical depth in a climate model, and found that the simulated changes in both precipitation and temperature were similar to observations. Thus they concluded that the increased black carbon aerosol is responsible for the interdecadal EASM change. However, other factors are also documented to have contributed to this interdecadal EASM change. Huang et al. (1999) found that this interdecadal rainfall decrease in North China is related to a decadal warming in the tropical Pacific, which looks like an ENSO pattern. They affirmed that this decadal timescale warming is responsible for this EASM variation. This affirmation is supported by the below study. SST anomalies influence the EASM by modulating the western Pacific subtropical high via Hadley circulation (Zhang et al., 2004). Accordingly, SST changes in the Pacific, especially the tropical Pacific, may be responsible for EASM change. Since the association between the EASM and ENSO on the interannual timescale is unstable (Wang, 2001, 2002), this argument is thus not persuasive. The SST decadal timescale variation over the Pacific, called Pacific Decadal Oscillation (PDO), is also considered as a factor affecting rainfall decrease in North China. In the warm phase of PDO, North China is controlled by strong westerlies so that the temperature is higher than normal and rainfall lower than normal, and vice versa (Yang et al., 2005). Besides SST anomalies, snow cover and sea ice have changed substantially since the second half of last century. Increased spring snow depth over the Tibetan Plateau is closely related with rainfall in the Yangtze River valley. More solar radiation being consumed to melt excessive snow cover leads to cooling and high pressure over the Plateau and adjacent areas so that the western Pacific subtropical high extends northwestward and strengthens. Associated more east-migrating vortexes developed in the eastern flank of the Plateau, rainfall increases remarkably in the Yangtze River valley (Zhang et al., 2004). The IPCC's (2001) third assessment reported that snow cover and spring sea ice over the Arctic has decreased about 10% and 10%–15% respectively. Wu et al. (1999) found that the sea-ice over north hemisphere has influence on winter monsoon. Based on observation and model experiments, Zhao et al. (2004) investigated the relationship between EASM and spring sea ice extent over the Bering Sea and the Sea of Okhotsk. There is positive correlation between them. When sea ice extent is heavy, a greater cold air mass moves southward, strengthens the mei-yu front, and favors more summer rainfall in South China. Other studies found close relationships between snow over the Tibetan Plateau and Asian monsoon (Bamzai and Shukla, 1999; Chen and Wu, 2000; Wu and Qian, 2003).

Therefore, various factors have been proposed to play roles in the formation of the interdecadal EASM variation of the late 1970s. However, which factor plays the dominant role is unclear and remains an open question. Although previous studies have explored the roles of regional SST and sea ice on the EASM, the influence of global SST and sea ice on EASM decadal change is less well studied. This motivates the present model study. The model and experimental design in this study are described in section 2. The simulated interdecadal changes of EASM in four integrations, together with a discussion of typical features revealed by observational data, are presented in section 3. Section 4 is a summary and discussion.

2. Model description and experimental design

The IAP 9L AGCM is a global grid-point atmospheric general circulation model, with a resolution of 4° (lat) $\times 5^\circ$ (lon) and nine unequal sigma levels in the vertical direction. The subtraction of a standard stratification in the dynamical framework reduces the error in computation. A detailed description of this model is presented in Bi (1993) and Liang (1996). The aerosol's effect on climate described in the AGCM is in a simple manner so that it is not proper to measure the contribution of aerosols to climate change unless improving the related parts in the model, and it will not be discussed in this study. The model could be used to measure CO_2 effect on climate. Bi (1993), Wang and Bi (1996) and Wang et al. (1997) evaluated the performance and variability of the model in simulating the present climate, such as statistical features, seasonal cycle and interannual variations. The model has also been used in simulation of interdecadal variations of atmospheric circulation (Li et al., 2000), extraseasonal short-term climate prediction (Lang et al., 2003; Lang and Wang, 2005) and paleoclimate simulation (Wang, 1999; Jiang et al., 2003, 2005).

After 13 months of climatological integration with monthly SST and sea ice prescribed in Alexander and Mobley (1976), the model is forced by observed month-varying global sea surface temperature and sea ice from the Hadley Centre covering 1958–1999 (Rayner et al., 2003). Because of large differences, four integrations with different initial conditions are conducted and their results are shown respectively. The observational dataset used in the study to compare with the simulations is ERA-40 reanalysis (Gibson et al., 1996, 1997) during 1958–1999, which is $2.5^\circ \times 2.5^\circ$, and

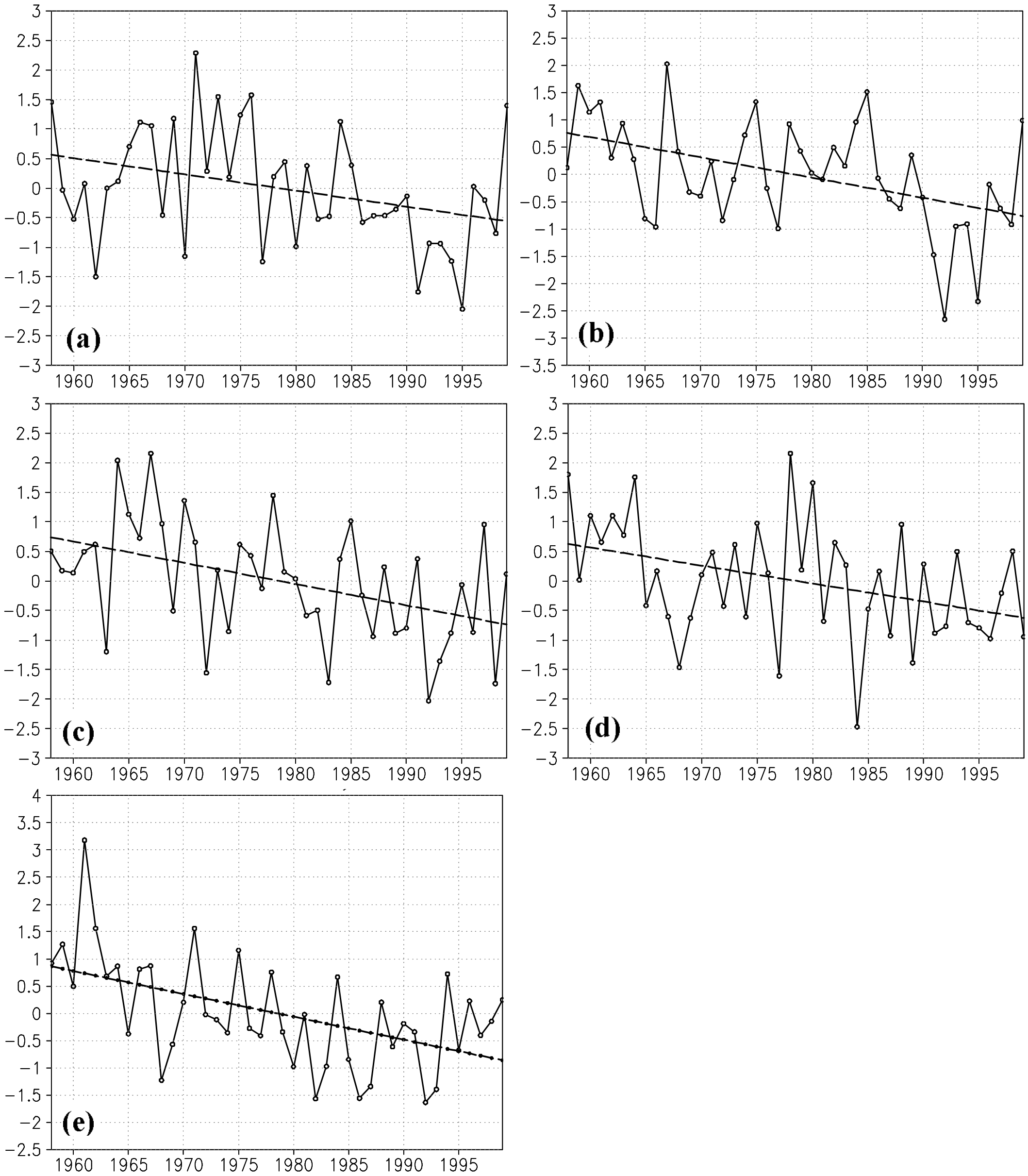

Fig. 1. (a)–(d) EASM intensity index based on model simulations and (e) ERA reanalysis, defined as normalized zonal wind shear between 850 hPa and 200 hPa over (20°–40°N, 110°–140°E) during June–August (JJA). The open circle lines are linear trends of indices.

provided by the European Centre for Medium-Range Weather Forecasts (ECMWF).

3. Interdecadal changes of the EASM

Much different from the Indian summer tropical monsoon, the EASM includes tropical and subtropical monsoon (Chen et al., 1991; Zhao et al., 2006; Zhao and Zhou, 2006) and its intensity index is far more complicated. There are many monsoon index definitions that have been proposed (Guo, 1983; Zhou, 1983; Zhang et al., 2003; Zhao and Zhou, 2005). In Indian monsoon studies, Webster and Yang (1992) defined a monsoon intensity index as zonal wind shear between 850 hPa and 200 hPa. This definition is also used in East Asian monsoon studies (Li and Ji, 1997; Wu and Ni, 1997). In this study, the East Asian summer monsoon index is defined similar to Webster and Yang (1992) definition: zonal wind shear between 850 hPa and 200 hPa over (20°–40°N, 110°–140°E) during June to August (JJA) (Fig. 1). Besides the interannual variation, decadal scale variation exists that both

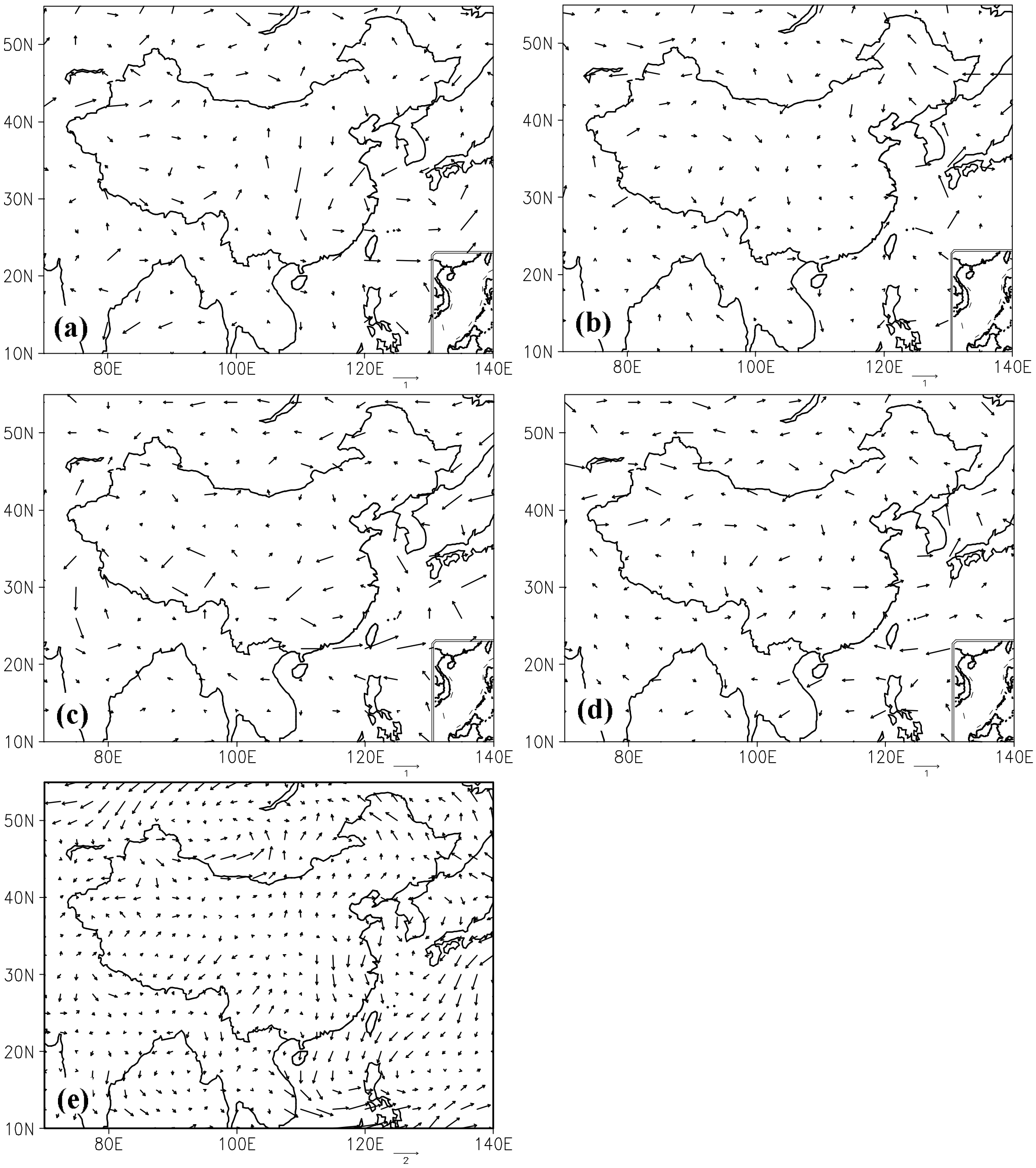

Fig. 2. Differences in JJA winds at 969-hPa vector between P1 and P2 (P2−P1) in (a)–(d) model simulations and at 925 hPa in (e) ERA (m s^{-1}).

the simulations and observation show, in which the EASM begins to weaken from the 1960s and becomes even weaker after the late 1970s. The year 1980 is a turning point because mean index values in preceding and succeeding years have different signs, and linear trends of the serial change from positive to negative occur from this year. Therefore, 1966–1976 is taken as a strong period of the EASM (P1) and 1981–1991 as a weak period (P2). The following discussion will be divided into two aspects: horizontal circulation and meridional cell.

3.1 *Horizontal pattern of variation*

It is in the wet season for the East Asian area during summer that airflows at low levels bring plenty of water vapor to the land, leading to rainfall, which causes the regional climate to be highly sensitive to EASM variability. Figure 2 shows the differences in the wind vector in the lower troposphere between P2 and P1: 969 hPa for the integrations (the lowest level

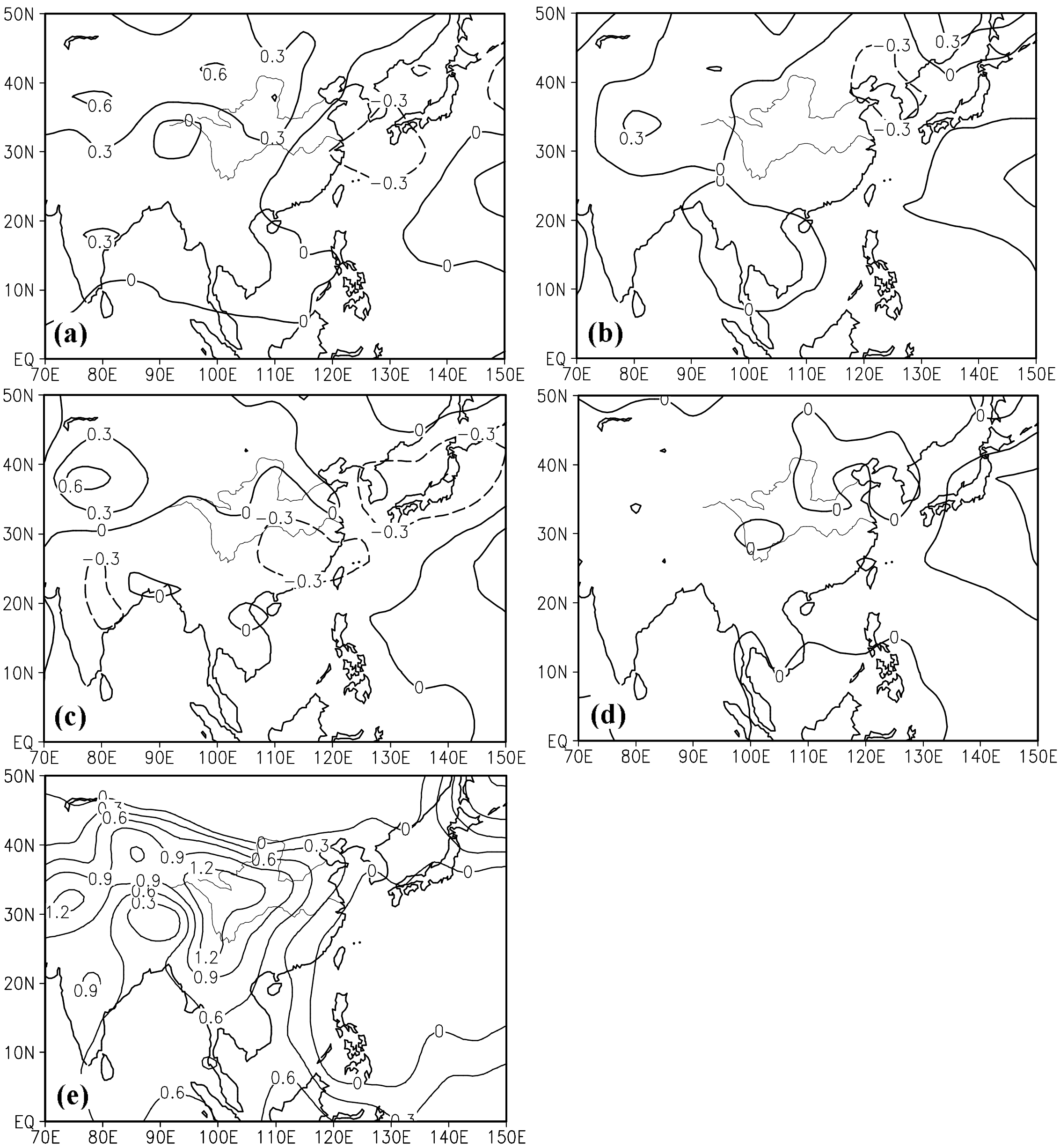

Fig. 3. Differences in JJA sea level pressure (SLP) between P1 and P2 (P2−P1) (hPa) in (a)–(d) model simulations and in (e) ERA.

in the model) and 925 for the ERA reanalysis. Four integrations show highly disagreeable results. Two integrations (Figs. 2a and 2c) demonstrate the northern wind anomaly over mid China. There are anomalous north winds over 110°–120°E to the south of 40°N extending to the South China Sea in the first integration (Fig. 2a). Figure 2b shows little wind change in the latter decade. Except for Fig. 2b, three integrations present the northern wind anomaly to the north of 30°N and the western wind anomaly in an area around 20°N, so there seems to be a cyclone over the eastern part of China and the adjacent East China Sea. The anomalous north wind means that south winds, which are the prevailing meridional winds in the summer, are weakened and the amount of water vapor transported from the ocean reduces significantly. Strengthening south winds over South Japan in Fig. 2b and over the eastern Korean Peninsular in Fig. 2d does not show in the observation (Fig. 2e). The cyclonic anomaly in the wind field in the observation locates in the South China Sea (Fig. 2e), similar to that demonstrated by Zhao and Zhou (2006). It seems that model integrations running from different initial conditions produce quite different results.

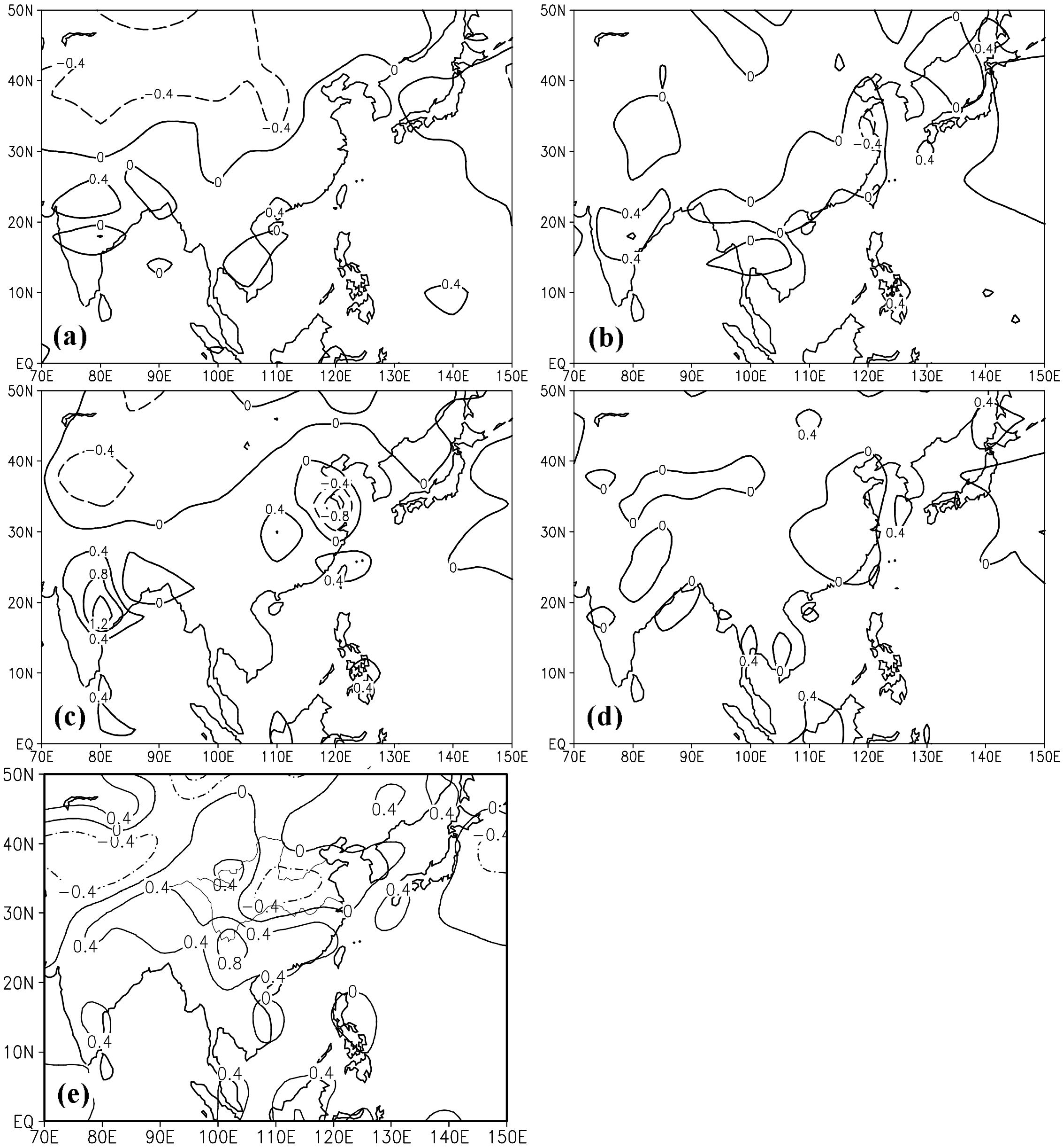

Fig. 4. Differences in JJA surface air temperature (SAT) between P1 and P2 (P2−P1) (°C) in (a)–(d) model simulations and (e) ERA.

Generally, land-sea thermal contrast is considered fundamental factor in monsoon formation. In the summer hemisphere, surface air temperature over land rises faster than that over the ocean, leading to a pressure gradient between land and ocean so that a quantity of air from the ocean taking an abundance of water vapor flows on to the land at low levels. Figure 3 and Fig. 4 show changes in sea level pressure (SLP) and surface air temperature (SAT) during two periods respectively. In the observed data, maximum SLP change is located over west-central China and most of the middle latitudes are positive (Fig. 3e) (Zhao and Zhou, 2006). Unfortunately, only one of the four integrations could reproduce the positive anomaly of SLP in west-central China (Fig. 3a). In Fig. 3a, contrasts of SLP between land and ocean are reduced due to a positive anomaly over central China and a negative anomaly in the northwestern Pacific. In Figs. 3b and 3c, southwest-northeast oriented bands of negative anomaly are located over the Yangtze River valley and South China. There is no significant change in Fig. 3d. The SAT decreases in the later decade over the area between the Yellow River and the Yangtze River in the observed data (Fig. 4e). In Fig. 4a, the distribution

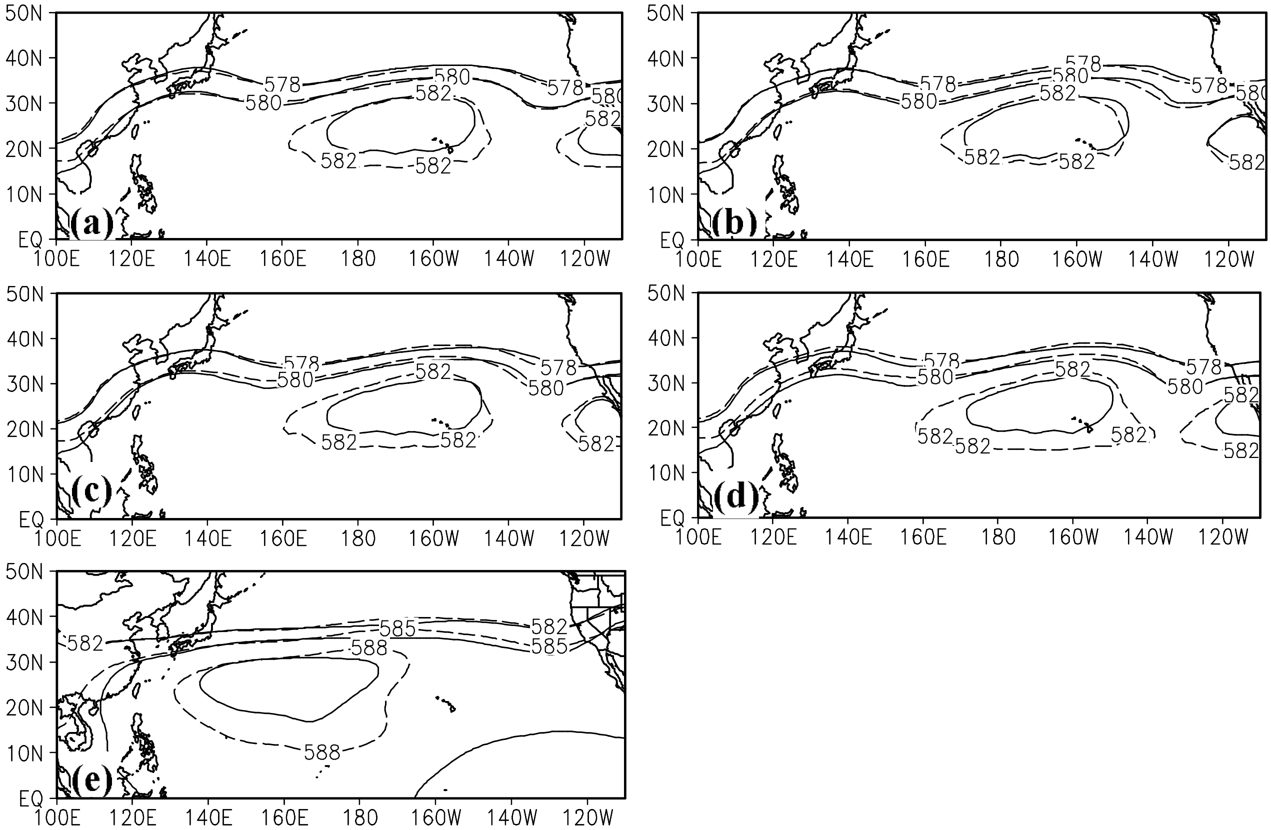

Fig. 5. Mean JJA geopential height at 500 hPa (10 gpm) in (a)–(d) model simulations and in (e) ERA. Only isolines of 578, 580, and 582 in simulations and 582, 585, and 588 in ERA-40 are shown. Solid lines for P1 and dashed lines for P2.

pattern is negative over the mid latitudes and positive over the lower reaches of the Yangtze River valley, South China and the western Pacific. The other three integrations present negative anomalies over the area between the lower reaches of the Yellow River and the Yangtze River. The anomalous pattern in East China in Fig. 4b is almost the opposite to Fig. 4a. The rising SATs in later period similar to the observed data are reproduced in Figs. 4c and 4d.

Subtropical high over the northwestern Pacific plays a key role in the EASM because the monsoon rain belt locates to the northwest of it, jumps northward and retreats southward in phase with it during the wet summer. Figure 5 shows the JJA mean 500-hPa geopotential height with only three isolates of 578, 580, 582 in integrations and 582, 585, 588 in ERA-40 in two decadal years separately, in units of 10 gpm. It is consistent in both the integrations and observation data that the location of the western Pacific subtropical high ridge shifts toward the southwest while its northern boundary moves very little. This explains the pattern of interdecadal change at the 500-hPa geopotential height: the geopotential height in the tropics grows to a significant level (Zhao and Zhou, 2006) and those in the mid latitudes change very little. Three integrations can present the positive anomalies at the 500-hPa geopotential height (figures omitted). As a result, water vapor transported by southern flow favors reaching lower latitudes in the latter period than in the former.

In monsoon circulation, air mass transported from the winter hemisphere into the summer part, collecting moisture from tropical oceans, forms the wet monsoon at low levels. There are compensating return flows back to the winter hemisphere at high levels, and the 200-hPa level is often considered as a level to demonstrate these flows. The differences in the wind vector at 200 hPa are shown in Fig. 6. In the observed data, an area over 33°–40°N is controlled by anomalous westerlies and the low latitudes, especially in the South China Sea, and experiences anomalous southern flows. These features denote that the return flows of the EASM are weakened. Two integrations partly reproduce the strengthening westerlies over the mid latitudes (Figs. 5a and 5c). The other two integrations show little change in East China (Figs. 5b and 5d).

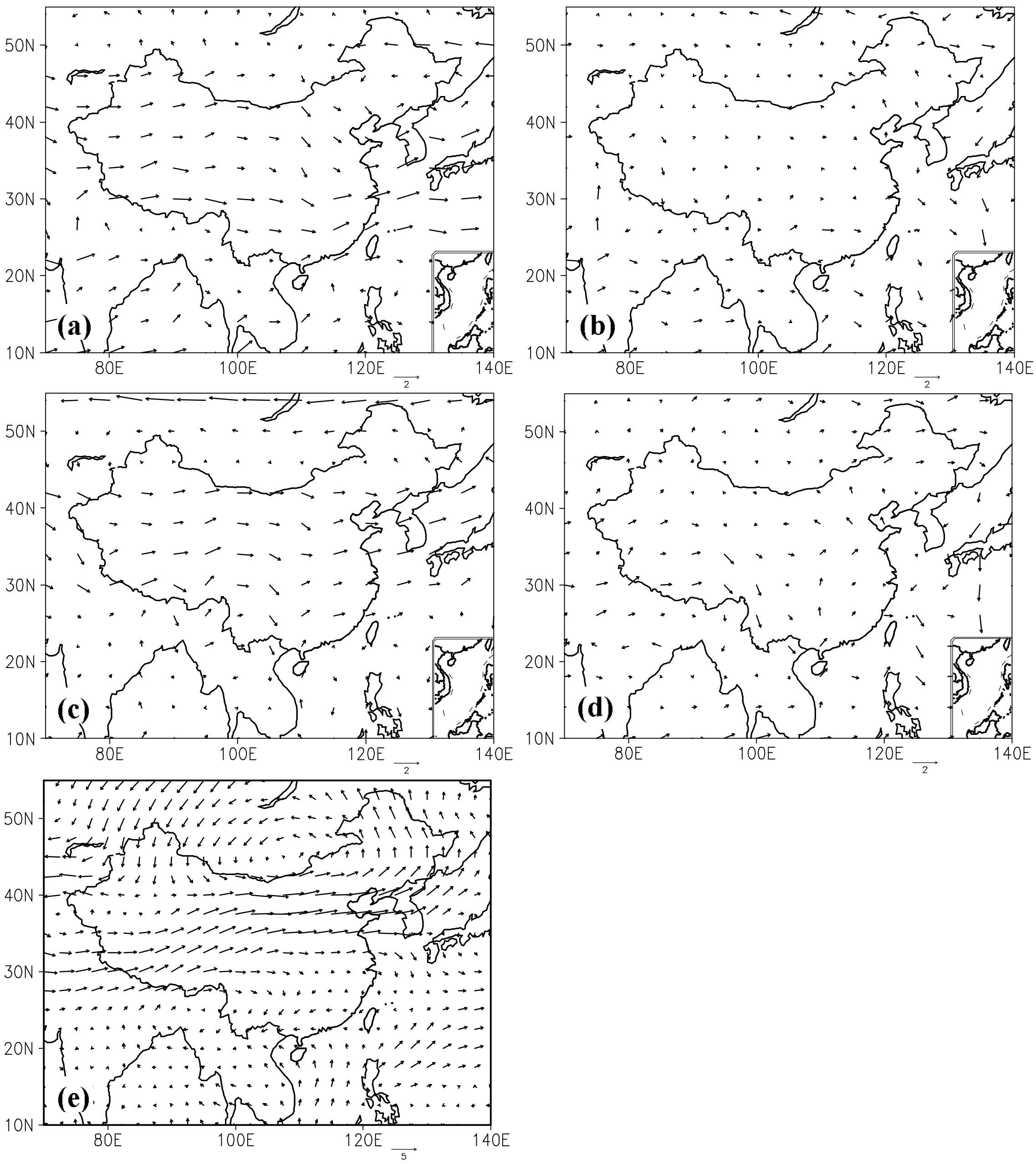

Fig. 6. Differences in JJA wind vector at 200 hPa between P1 and P2 (P2−P1) (m s^{-1}) in (a)–(d) model simulations and in (e) ERA.

3.2 *Vertical structure of the variation*

As shown in Fig. 2e, the obvious interdecadal changes are located over 110°–120°E. Meridional circulation for the longitudinal range 110°–120°E is appropriate to show its interdecadal variations. In the observed data (Fig. 7e), there are south winds under 300 hPa and north winds at upper levels over the tropics. Winds over 10°–40°N are opposite to those over the tropics under 200 hPa, e.g., negative anomaly, and there are anomalous southerlies at upper levels over the subtropics. Anomalous south winds exist in the whole troposphere and stratosphere over the mid latitudes. In four model integrations, only one of them reproduces similar changes (Fig. 7a). However, the magnitude and location of the anomaly are slightly different from ERA-40. For example, the anomalous north wind is northward over the subtropics and the anomalous velocity is a little small in the simulation. The other three integrations fail to reproduce these

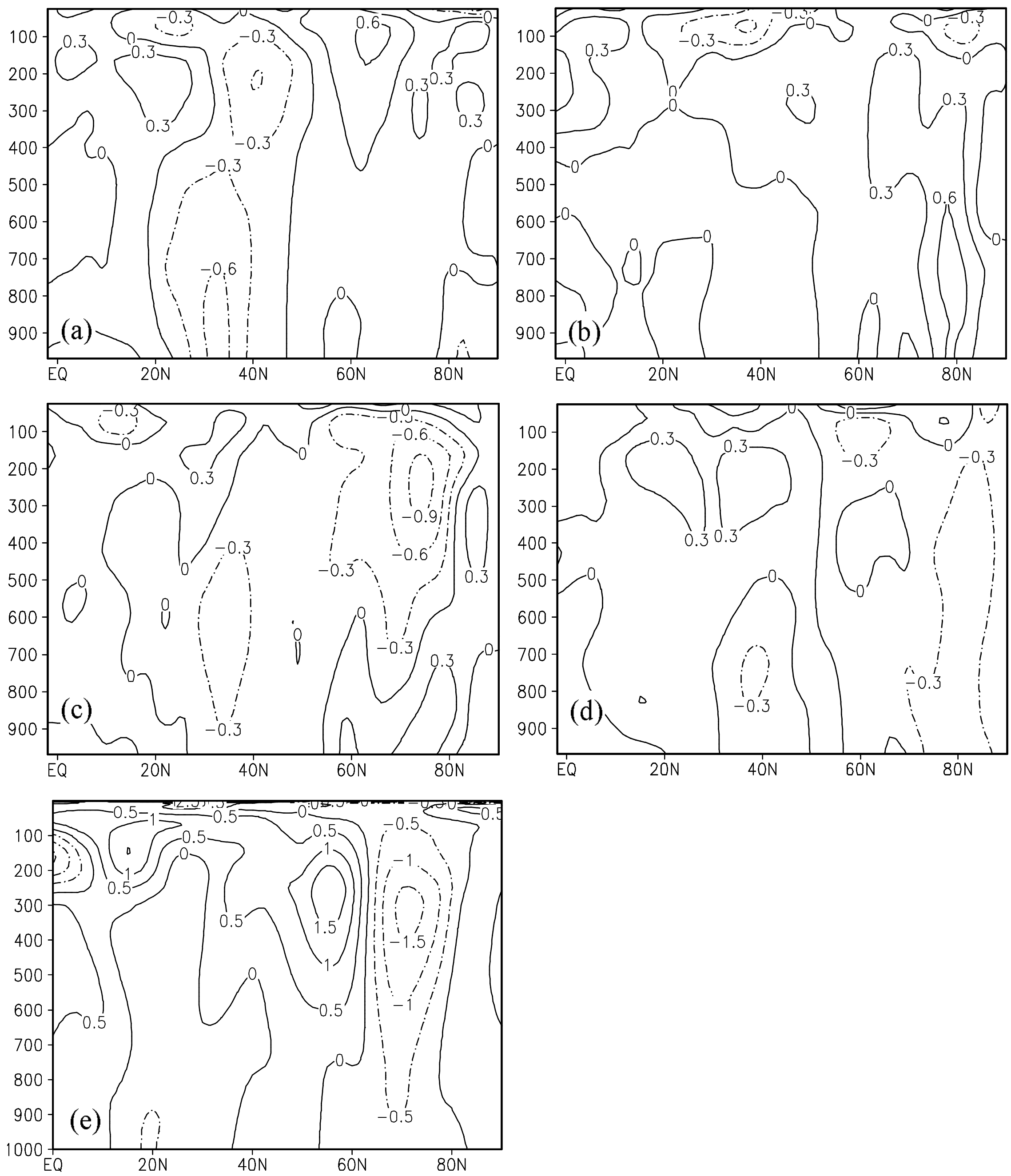

Fig. 7. Differences in JJA meridional winds for the longitudinal range 110°–120°E between P1 and P2 (P2−P1) (m s^{-1}) in (a)–(d) model simulations and in (e) ERA.

features, for either the location or the magnitude of the anomalies, or both.

4. Summary and discussion

To estimate the extent to which historical global SST and sea ice influence interdecadal changes of the EASM, four IAP-9L AGCM integrations forced with observed monthly varying SST and sea ice covering 1958–1999 were carried out. The modeled results suggest a large discrepancy among the four integrations. One integration qualitatively simulates the pronounced observed interdecadal changes in central China, including the weakened southerly at low levels, decreased surface air temperature and increased sea level pressure, the southwestward extension of the ridge of the subtropical high over the northwest Pacific and the resultant decreased water vapor transportation to the mid and high latitudes, as well as the strengthened 200-hPa westerlies over this region. However, the other three integrations failed to reproduce these observed decadal timescale changes. No integration reproduced well the observed interdecadal precipitation anomalies in central China. Such a discrepancy among the integrations indicates that the interdecadal change of the EASM since the end of the 1970s cannot be explained by historical SST and sea ice evolution.

It is well known that global oceans are the primary heat tank of the climate system and can memorize the signals from external forcing. Thus, if the anthropogenic climate forcing signal is stored in global oceans, the present result will suggest that interdecadal climate change may be just a random nature of the coupled climate system, as discussed in Jiang and Wang (2005). This needs further verification by directly forcing the model with anthropogenic forcings, including increased greenhouse gases. Until now, however, the current model has not been capable of performing such a study. Additionally, Deser and Phillips (2006) examined the contribution of tropical sea surface temperature to the 1976/77 climate transition of the winter atmospheric circulation over the North Pacific and found two versions of CCM3 that reproduced the climate shift, but CAM3 failed. Their study suggests that different models can possibly respond quite differently with the same external forces. Thus the results discussed above are possibly model-dependent. Hence, to further study the climate impact of global oceans or anthropogenic forcing with state-of-the-art climate models will be a focus of future work.

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