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Changes in the East Asian Cold Season since 2000

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ABSTRACT

Using NCEP–NCAR reanalysis data and observational data from meteorological stations in China, the evolution of the East Asian cold season (EACS) and its long-term changes after 2000 were studied. A monsoon tendency index (MTI), defined as the temporal difference of the East Asian monsoon index, indicates that the winter monsoon setup has been postponed in autumn, while the setup has quickened in early winter. In mid winter, the EACS breakdown process has accelerated, while it has lingered in late winter. The authors suggest that the postponement of monsoon setup in autumn may be caused by strong global warming at the lower levels, which further limits the setup time period and leads to the quickening of the setup process in early winter. Meanwhile, a north–south seesaw of temperature tendency change in China can be observed in December and February, which may be related to large-scale circulation changes in the stratosphere, characterized by a polar warming in mid winter and polar cooling in early spring. This linkage is possibly caused by the dynamical coupling between stratosphere and troposphere, via the variation of planetary wave activities. In spring, the speed of the EACS breakdown has decreased, which favors the revival of the EACS in East Asia.

Key words: seasonal evolution, East Asian winter monsoon, stratosphere-troposphere interaction

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1. Introduction

In January and February 2008, an anomaly of the East Asian winter monsoon (EAWM) caused extremely low temperatures, blizzards, and freezing rain in South China. This disastrous weather episode was a very rare event, consistent with unusual winter monsoon circulation configuration (Gu et al., 2008; Zhou et al., 2009). As pointed in various studies, the unusual circulation may be the result of global adjustment to long-term climate change. Under the global warming scenario, the long-term change of the EAWM shows a strong tendency toward the weaker phase, which is characterized by extreme warming and weaker northerlies in East Asia (Xu et al., 1999; Wang, 2001; Kang et al., 2006; Shi et al., 2007). A number of studies have also suggested a regional cooling of the surface temperature in South China (Yu and Zhou, 2004; Li et al., 2005). As an important climate component, the variation of the EAWM exerts a strong impact on remote larger-scale circulation, not to mention the tremendous economic and social impact on populous East Asian countries. For example, an anomalous East Asian monsoon can influence the convection and sea surface temperatures near the Maritime Continent (Chang et al., 1979; Bueh and Ji, 1999). The winter monsoon is also closely related to other climate systems, such as the East Asian trough and the East Asian westerly jet stream, which influence the climate in both local and remote regions, including North America (Cohen et al., 2001; Yang et al., 2002). Several studies (Sun and Sun, 1994; Ji et al., 1997; Chen and Sun, 1999; Chen et al., 2000) have suggested that an anomaly of the winter monsoon usually

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precedes anomalous circulation in the summer season. Furthermore, the monsoon is believed to play an important role in the monsoon–ENSO relationship and may influence ENSO evolution, intensity and periodicity (Lau and Chang, 1987; Li et al., 2001; Huang et al., 2004). The main focus of the above and current studies is the strength, or intensity, of the EAWM. However, the seasonal evolution of the East Asian cold season (EACS) may be significant in terms of its long-term changes, and this aspect is less well studied.

The variation of the EAWM is complicated, probably due to multiple influencing factors. Previous studies have revealed that the EAWM can be influenced by external forcing, such as the ENSO (Li, 1990; Chen et al., 2000; Wu et al., 2003; Zhou et al., 2007) and SST over the tropical western Pacific warm pool (Huang and sun 1992; Chen et al., 2000). Other external factors include the dynamical and thermal effects of the Tibetan Plateau (Ye and Gao, 1979), Eurasian snow cover (Wu and Zhang, 1998; Jhun and Lee, 2004), and so on. The changes of internal factors also influence the variation of the EAWM. These factors include remote atmospheric oscillations, such as the Arctic Oscillation (AO), the Northern Atlantic Oscillation (NAO), and the Antarctic Oscillation (AAO) (Gong et al., 2001; Wu and Wang, 2002; Gong and Ho, 2003). Recently, Chen et al. (2005) suggested that oscillations in stationary planetary wave activity can also have great impact on the East Asian winter climate. Also, studies have revealed that stratospheric circulation anomalies can have a downward influence on low-level circulation (Baldwin and Dunkerton, 1999, 2001), and the influence is estimated to be comparable to the influence of the ENSO (Baldwin et al., 2003). An important question, therefore, is does the variation of the EAWM relate to upper stratospheric circulation? Furthermore, if the change of seasonal evolution is considered, how and why does the change happen?

In this study, the authors present a new perspective of the EAWM and focus on the feature of its seasonal evolution and its changes after the new millennium. A new index, focusing mainly on the monsoon seasonal evolution feature, is defined in this paper. Evidence that change in the seasonal evolution of the EAWM is influenced by global warming is presented, as is its close association with changes in stratospheric circulation. The importance of the role played by the wave-mean flow interaction is also demonstrated. Section 2 describes the datasets and analysis methods applied in this study. Section 3 describes the climatology of the EAWM seasonal evolution. Change in the seasonal evolution of the EAWM is analyzed in section 4. A discussion of the possible mechanism linking stratospheric circulation and the seasonal evolution of the EAWM is presented in section 5, with a brief summary of the results provided in section 6.

2. Data and methodology

2.1 *Data*

The atmospheric general circulation data were taken from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCAR/NCEP) reanalysis dataset (Kalnay et al., 1996), available from January 1948 to date. The monthly and daily mean temperature, geopotential height, and horizontal velocity were used. These variables have a $2.5°\times2.5°$ latitude/longitude resolution and extend from 1000 hPa to 10 hPa with 17 pressure levels in the vertical. The reanalyzed surface air temperature (SAT) and meridional wind were also examined. Additionally, use was made of monthly surface temperature data from 160 meteorological stations across China, collected and compiled by the China Meteorological Administration (CMA) from January 1951 to date. In addition, to determine the occurrence of cold surge events, CMA daily SAT data from 740 Chinese stations and reanalyzed NCEP–NCAR sea level pressure (SLP) data were also employed. The total time period studied was from January 1951 to December 2009.

2.2 *Methodology*

East Asia is characterized by strong surface northeasterlies in winter, which are quasigeostrophically linked to the surface Siberian High and the Aleutian Low. Therefore, the regionally averaged meridional wind along the coast of East Asia (Chen and Graf, 1998; Chen et al., 2000) and the pressure difference between the Siberian High and the Aleutian Low are frequently used as indicators of the strength of the winter monsoon (Guo, 1994; Shi and Lu, 1996; Shi and Zhu, 1996; Wu and Wang, 2002). Here, the authors define the monthly East Asian monsoon index (EAMI) as the mean zonal SLP difference (110°E minus 160°E) over 20°–50°N using the NCEP–NCAR reanalysis dataset, i.e. $\mathrm{EAWI} = \overline{\mathrm{SLP}}_{110°\mathrm{E}} - \overline{\mathrm{SLP}}_{160°\mathrm{E}}$. This index represents the strength of the zonal pressure gradient; therefore, the larger the index, the stronger the SLP difference between the Siberian High over land and the Aleutian Low over the ocean, and the stronger the northerly along the East Asian coastal region.

To study the seasonal evolution, tendency is defined as the difference of a parameter between one lagged month and one leading month. For example, the tendency of temperature in October would be the temperature difference between November and

September. Therefore, the developing rate of a parameter can be prescribed by its tendency. The development of the EAWM can therefore be presented by the monsoon tendency index (MTI), which is defined as the tendency of EAMI:

$$\mathrm{MTI} = \mathrm{EAMI}(t+1) - \mathrm{EAMI}(t-1)\ , \qquad (1)$$

where the EAMI is the East Asian monsoon index, and t+1 means one month after and t-1 means one month before. Therefore, this index represents the time evolution speed of the East Asian monsoon.

Cooling in East Asia is often caused by an outbreak of cold air, i.e. cold surges or cold waves. The more frequent cold surges are, the lower the surface temperature will be. Cold surges were selected according to the criteria of Jeong and Ho (2005), which include the following. Firstly, an anticyclonic center should be identified near the south of Siberia (40°–50°N, 95°–110°E) prior to the breakout of cold surges. The criteria of Zhang et al. (1997) were used to identify an anticyclone. Secondly, during the movement of this surface anticyclone, the 24–48 h SAT drop should exceed 1.5 sd (sd being the standard deviation of the SAT anomaly for 44 winters) in mid China (18 stations; 32.5°–37.5°N, 110°–115°E) or South China (22 stations; 22.5°–27.5°N, 112.5°–117.5°E). The difference between the present study and that of Jeong and Ho (2005) is that, here, daily-mean surface temperatures from 740 Chinese stations, instead of their 172 stations, were used; additionally, the area defined as South China in the present study was 2.5° further west, owing to the fact that the area defined by Jeong and Ho (2005) was half located over the ocean, where no meteorological stations exist. The South China area of the current study is, however, the same as that defined in Zhang and Wang (1997). On average, 8.15 cold surge events were found to have occurred per year from 1979–2007, which is close to the number suggested by Jeong and Ho (2005) and Wang and Ding (2006), but smaller than that of Zhang et al. (1997) and Chen et al. (2004), indicating a slightly stricter criteria by this procedure.

3. Climatology

The 3D EACS circulation was considered as the combined configuration of the East Asian jet, the broad East Asian trough, and the Siberian high at the upper, middle and lower troposphere, respectively. This monsoon circulation has been documented in many studies (Tao, 1956; Tao et al., 1965; Boyle and Chen, 1987; Lau and Chang, 1987; Zhang et al., 1997; Chan and Li, 2004). The seasonal evolution of the winter monsoon climate in East Asia is synchronous

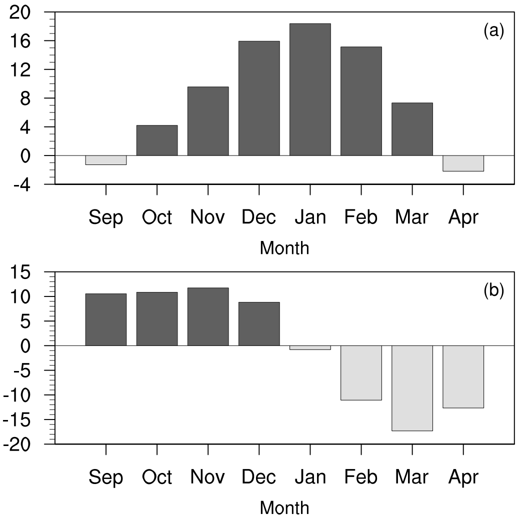

Fig. 1. The climatology (30-yr mean from 1971 to 2000) of (a) East Asian monsoon index (EAMI) and (b) monsoon tendency index (MTI) in winter months from September to April. Dark and light bars indicate positive and negative values respectively.

with the seasonal development of the above components. The setup (collapse) of the EAWM is characterized by the reconstruction (destruction) of the East Asian trough, the Siberian high, and the Aleutian low. Here, the climatology of winter monthly EAMI (Fig. 1a) can reveal the main monsoon evolution features. During the seasonal evolution, the EACS begins in October and a positive SLP gradient is constructed, which is directly associated with the setup of the Siberian high and Aleutian Low. The EAMI increases in autumn and early winter and reaches its maximum value of around 18 hPa in January, which corresponds quasigeostrophically to northerlies of about 4.7 m s^{-1}, comparable to the average northerlies along coastal East Asia in January. The EAMI decreases in late winter and spring. The opposite SLP gradient with high pressure over the ocean and low pressure over the continent is established firstly in April, which means the final withdrawal of the continental cold season. The MTI can be used as a measure of the rate of evolution of the EAWM (Fig. 1b). Positive values can be observed in autumn and early winter and negative value in late winter and spring. The largest positive value occurs in November, indicating an acceleration of the seasonal evolution in this month; therefore, November is considered as the main EAWM buildup period in this paper. The MTI is very small in January, which characterizes the stabilization of the EAWM in mid winter. The largest negative value occurs in March, corresponding to a quick withdrawal of the winter season; hence, March is considered as the main EAWM withdrawal period during this time. In the winter mon-

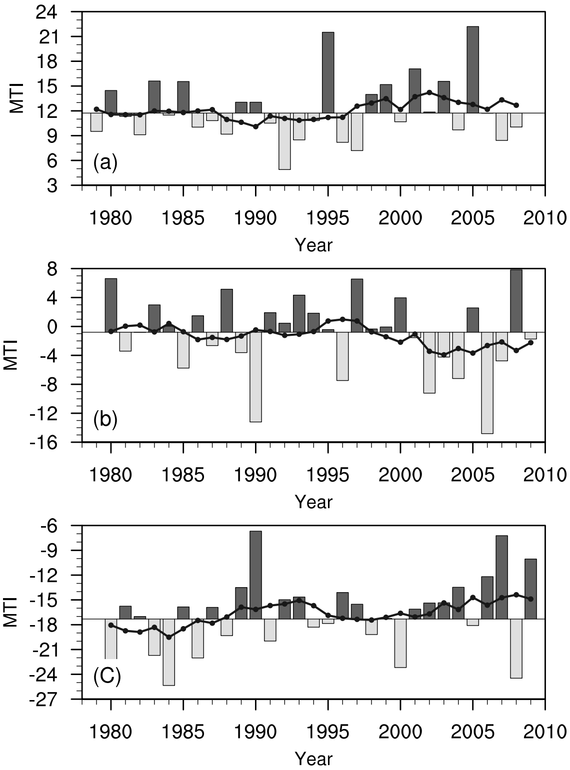

Fig. 2. Time series of the MTI index in (a) November, (b) January, and (c) March. The curves indicate the 9-year running mean. Dark and light bars indicate values above and below the long-term mean values.

soon developing period from September to December, the MTI is about 10 hPa and has a small difference among the four months; while the values in late winter months are greater than 10 hPa and have a large difference. Therefore, the buildup of the EAWM is gradual and lasts for a longer time relatively, while the EAWM decays in a shorter time and mainly collapses in March.

4. Changes in the seasonal evolution of the EAWM

Recent changes in the seasonal evolution of the EAWM are shown in Fig. 2, which shows the time series of the MTI index in November, January and March and their nine-year running mean. All the indices are characterized by obvious interannual and interdecadal fluctuations, with increasing trends in November and March and decreasing trends in January. A clear transition of the MTI index around 2000 can be observed in January and March. In November, the index shows a weak increasing trend from the 1980s to the 2000s. Strong positive MTI index values can be observed as having occurred more frequently since the late 1990s. In January, there is a strong decreasing trend, with a clear transition around 2000. Based on the student's $t-$test, the mean MTI difference between 1991–2000 and 2001–2009 exceeds the 90% confidence level, indicating an early collapse of the winter monsoon compared to the previous decade. The MTI index in March, however, shows a strong increasing trend. As the climate MTI is negative in March, the increasing trend indicates a weakening of the winter monsoon withdrawal and a lingering of the cold season. A transition can also be observed around 2000, although the statistical significance between the two decades is not very high (about 85%). To study recent changes in the seasonal evolution of the East Asian monsoon, the average anomalies from 2000/01 to 2008/09 were calculated, with the long-term climatology subtracted. Recent changes (i.e. in the last decade) in the evolution of the EAWM are shown in Fig. 3, which shows the average of MTI (Fig. 3a) and its anomaly (Fig. 3b) from 2000/01 to 2007/08. The main changes can be summarized as follows: (1) The MTI is negative in the autumn months of September and October, indicating a slowdown of the EAWM buildup and a delayed EAWM during this decade; (2) positive values occur in the early winter months of November and December, which characterizes a quickening of the EAWM setup. Particularly in November, the MTI anomaly is about 1.3 hPa, corresponding to an approximate 12% increase in the MTI; (3) negative values occur in January and February. Particularly in January, the negative value reaches -2.9 hPa, about 3.6 times larger than the climatological value of about -0.8 hPa. The negative values lead to a quickened withdrawal of the EAWM in late winter; (4) January

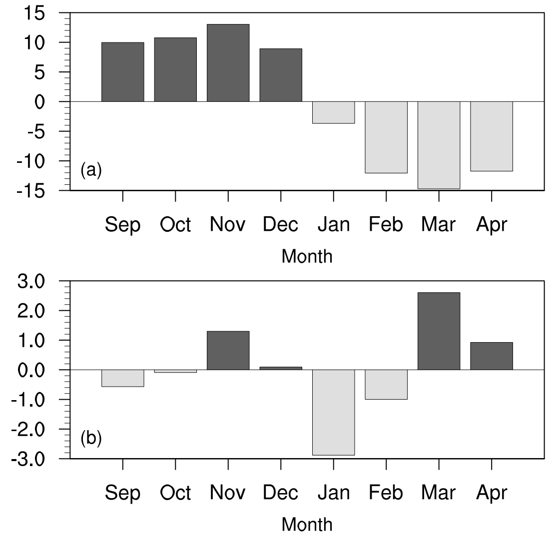

Fig. 3. (a) The monsoon tendency index (MTI) from September to April averaged for the 2000s (Jan 2000–Sep 2008) and (b) its anomalies from the climatology.

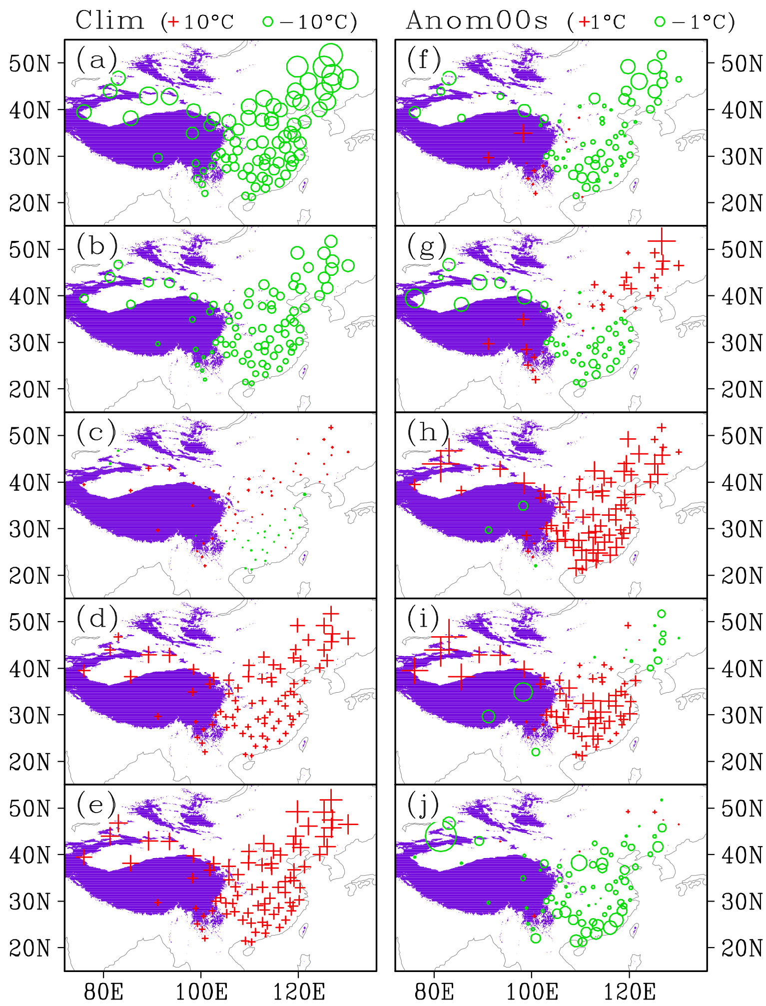

Fig. 4. Left: the climatology (30-yr mean from 1971 to 2000) of temperature tendency (the difference of temperature between lagged one month and leading one month, refers to the text for the exact definition and example) in China using station observational data. Right: the change of winter monthly temperature tendency after 2000; (a) and (f): November, (b) and (g): December, (c) and (h): January, (d) and (i): February, (e) and (j): March. Positive values are shown with plus sign and negative values with open circle. The size of mark is proportional to the temperature tendency value as indicated by the scale on the top of each column. For the purpose of clarity, the marks are drawn every other station.

is usually the month when the EAWM reaches its mature phase. However, the strong negative value indicates a strong early collapse of the winter monsoon, and that the mature and stable period is slightly shifted ahead; (5) there are usually strong negative MTI values in March, indicating a quick final collapse of the winter monsoon. Also, strong positive anomalies are seen in the 2000s, which amount to about a 15% decrease in the MTI, favorable to an early spring prolongation and revival of the winter monsoon.

Figure 4 shows the tendency in surface temperature anomalies since 2000 using observational data from meteorological stations in China, together with temperature tendency climatology during 1971–2000. Climatologically, the temperature evolution during the cold season is characterized by temperature drops at all the stations in early winter months (Figs. 4a and 4b) and temperature rises in the late winter months (Figs. 4d and 4e), with the largest temperature drop in November and the largest temperature rise in March, respectively. These large values are consistent with the largest MTI index in November and March

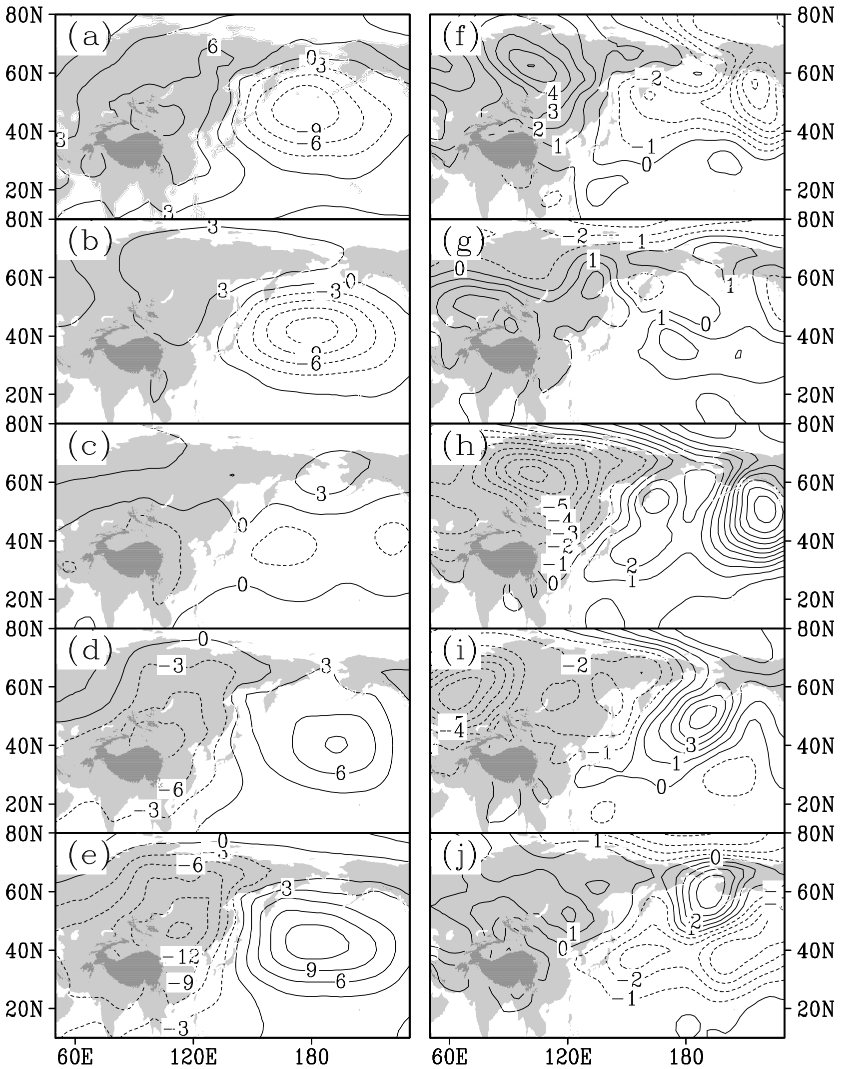

Fig. 5. Left: the climatology (30 yr mean from 1971 to 2000) of sea level pressure (SLP) tendency. Right: The change of winter monthly SLP tendency after 2000; (a) and (f): November, (b) and (g): December, (c) and (h): January, (d) and (i): February, (e) and (j): March. The contour interval is 3 hPa in the left column and 1 hPa in the right column. The edges of the dark shading denote the contour at 2000 m.

(Fig. 1b). It is worth noting that the amplitude of temperature tendency is apparently larger at higher latitudes, especially in Manchuria, where a tendency of −20°C and 20°C per month can be observed in November and March, respectively. The changes in temperature tendency after 2000 are shown in Figs. 4f–4j. In November, there are negative temperature tendency anomalies over most stations in eastern China after 2000, adding to the usual negative values and causing acceleration of monsoon evolution. In December, however, negative temperature tendency anomalies mainly occur in southern China, while Northeast China is dominated by positive anomalies. Therefore, the evolution of the EAWM accelerated in the southern China and lingered in Northeast China. In January, a uniform warming can be observed at all the stations except several on the Tibetan Plateau. This warming adds to the small climate temperature tendency in January and leads to an early and strong decrease in the EACS. February witnesses strong warming in the southern China and Northwest China, while the change at most stations in Northeast China is negative, revealing a quickened EAWM withdrawal in southern China and Northwest China and a weakened withdrawal in Northeast China. In March, the temperature tendency is almost uniformly negative over the whole of China; therefore, early spring is characterized by a revival and lingering of the EAWM.

The associated lower level circulation change is presented with the SLP tendency anomalies in Fig. 5. In the early winter months (Figs. 5a–5b), the SLP increases over the Asian continent (positive SLP tendency) and decreases in the northern Pacific Ocean (negative SLP tendency). Therefore, a zonal SLP gradient develops, which causes the northerly wind along the East Asian coastal region. The distribution of the November SLP tendency anomalies since 2000 (Fig. 5f) presents strong positive values over the Asian continent and negative values over the northern Pacific Ocean. Therefore, the west–east SLP gradient is increased greatly and the winter monsoon evolution quickened during this month. Strong cold northerly anomalies are favored, curving southward between the strengthened Siberian high and Aleutian low.

In December, there are still positive SLP tendency anomalies over the mid latitudes of the Asian continent, indicating an overall quickened seasonal evolution. The positive anomaly center is located around Lake Balkhash. According to the quasi-geostrophical wind relationship, a westerly wind anomaly is favored in mid and high latitudes over the Asian continent and a northerly wind anomaly in South China, which corresponds to the warming tendency in Northeast China and cooling tendency in southern China (Fig. 5b).

Usually, the winter monsoon stabilizes in January, characterized by a small SLP tendency over both the continent and the ocean (Fig. 5c). The January SLP tendency anomalies since 2000, however, show strong negative values over Siberia and positive values over the central northern Pacific. This distribution helps to weaken the winter monsoon and leads to an early collapse of the winter circulation system.

In the late winter months (Figs. 5d–5e), the SLP decreases over the Asian continent (negative SLP tendency) and increases in the northern Pacific Ocean (positive SLP tendency). Therefore, a zonal SLP gradient weakens, favoring the withdrawal of the northerly along the eastern coast of the Asian continent. The February SLP tendency since 2000, still with negative values over the Asian continent and positive values over the central northern Pacific, quickened the winter monsoon withdrawal processes in this most recent decade.

In March, a positive SLP tendency appears over the Asian continent and a negative tendency over the Pacific Ocean. This large-scale circulation change helps to delay the withdrawal of the winter season and witnesses a revival of the EAWM.

5. Discussion: Causes of the EAWM change

It has been pointed out that the strongest warming due to global warming during the last 100 years occurred in the interior region of Asia (Hu et al., 2000; Bueh et al., 2003; Hu et al., 2003; Hulme et al., 2006; IPCC, 2007). The composite analysis using station observational temperature data indicated that the warming is obvious after 2000 (figure not shown). This kind of warming can, therefore, increase the surface temperature and cause the delay in the buildup of the winter monsoon in autumn (negative MTI in September and October, as shown in Fig. 3b) and the early collapse of the winter monsoon (strong negative MTI value in January, as shown in Fig. 3b). The warming also helped to cause the speedup of the EAWM in early winter (strong positive MTI value in November, as shown in Fig. 3b) because the warming in autumn shortened the transition time between the summer and winter seasons (Dong et al., 2010).

The lingering of the winter monsoon in the late winter month (March) can be attributed to the cold surge activities in the East Asian region. The monthly frequency distributions of cold surges are shown in Fig. 6, both climatology and average for the 2000s. Climatologically, the maximum frequency usually occurs in early and late winter, i.e. November and March, and the cold surge activities are calm in mid winter. This monthly distribution of cold surges is similar to previous studies (Zhang et al., 1997; Wang and Ding, 2006). For the average of the 200s, the increased cold surge activities in March can lead to the revival of the cold winter monsoon in the late winter month. This increase of cold surges may be attributed to both dynamic and thermodynamic factors. A cold surge is formed when the 24–48 h decrease in surface air temperature exceeds a certain level. When the surface air temperature is relatively cold, it will take a very strong anticyclone to induce a substantial temperature

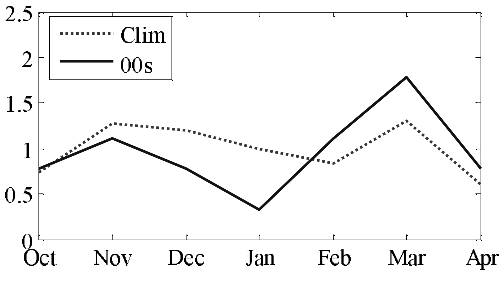

Fig. 6. The frequency of cold surges in winter months: (a) climatology (solid line), and (b) average after 2000 (dashed line).

decrease, whereas when the surface air temperature is relatively warm, cold air and an anticyclone with modest intensity can significantly reduce the temperature. Therefore, the increase in early spring temperature due to global warming is favorable to the revival of cold surges in the transition period. On the other hand, as shown in Fig. 5j, March is characterized by an increased zonal SLP gradient, with a positive SLP tendency on the continent and negative SLP tendency over the ocean. This change in zonal contrast is favorable for the development of low-index flow, and therefore increases the activity of cold surges.

The monthly distribution also shows a minimum cold surge frequency in January after 2000. Less disturbance of mid-winter circulation by the cold surges thus helps maintain a weak winter monsoon and a strong regional warming. This decrease is also related to both dynamic and thermodynamic factors. Under the background of global warming after 2000, the intensity of the EAWM largely decreased during mid winter, which then decreased the zonal SLP contrast, favoring high-index flow (stronger westerly in mid-latitudes). Therefore, the low zonal SLP gradient and stronger westerly are unfavorable for the occurrence of cold surges because the short waves tend to move rapidly through such a pattern without causing any disturbances (Boyle, 1986), and also stronger westerlies help to suppress baroclinic wave activity (Nakamura, 1992). On the other hand, greater warming and a decrease in the Siberian high in the interior Asian continent is unfavorable for the accumulation of cold air.

It is worthwhile noting that there is a systematic distribution pattern of the temperature tendency change in December and March: temperature seesaws between northeastern China and southern China (Figs. 4g and 4i). Chen et al. (2005) pointed out that the north–south temperature seesaw in eastern China is related to the equatorward propagation of planetary waves in the middle and upper troposphere, which is further modulated by the strength of the stratospheric polar vortex (Chen and Kang, 2006). A stronger equatorward wave activity in the middle and upper troposphere can be related to the positive phase of the AO, which favors warming in Northeast China and cooling in southern China. Figure 7 shows the anomalies of the monthly mean zonal-mean temperatures at 50 hPa in the Northern Hemisphere as a function of latitude and months. Generally speaking, cooling can be observed in the stratosphere over the tropics and extratropics after 2000, and is probably related to the radiative effects of increases in greenhouse gases. However, an exceptional warming occurs in mid-winter, with the maximum value of about 6 K occurring in the polar region in February. Strong cooling occurs thereafter in early spring, with the maximum value of about −4 K in April. Accordingly, the polar temperature tendency shows strong warming in early winter and strong cooling in late winter. The stratospheric warming is coupled dynamically with the upward planetary wave activities (Matsuno, 1971; Kodera et al., 2000; Limpasuvan et al., 2004). In December, there is stronger upward Elissen-Palmer (EP) flux from the troposphere to the stratosphere (Fig. 8a). Through interaction with the zonal flow, the anomalous upward flux can cause a warming polar region and a weaker polar vortex. In the troposphere, the equatorward EP flux is increased. According to the work by Chen et al. (2005), this wave activity is related to the positive phase of the AO in the Northern Hemisphere, which leads to warming in Northeast China and cooling in southern China, and is consistent with the temperature tendency change distribution in December (Fig. 4g). In late winter, the polar region switches to a cold domain. This transition is associated with weakened upward EP flux (Fig. 8b) into the stratosphere and decreased equatorward EP flux in the troposphere. According to the planetary-waves–monsoon relationship demonstrated by Chen et al. (2005), this can lead to cooling in Northeast China and warming in southern China (Fig. 4i), favorable to the lingering of the EAWM in northern China in early spring.

6. Conclusions

In this study, the authors have investigated the seasonal evolution of the EAWM and its changes after 2000. The EAMI revealed that the EAWM usually increases its strength in early winter, reaches its maximum in January, and decreases thereafter. A MTI,

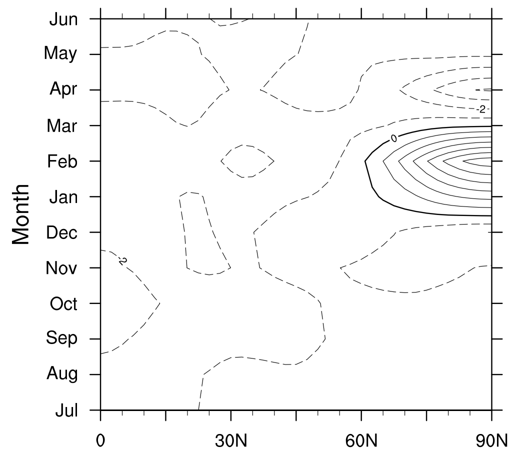

Fig. 7. Northern hemisphere zonal-mean temperature anomalies averaged from Feb 2001 to Aug 2007 at 50 hPa. Contour interval is 1.0 K.

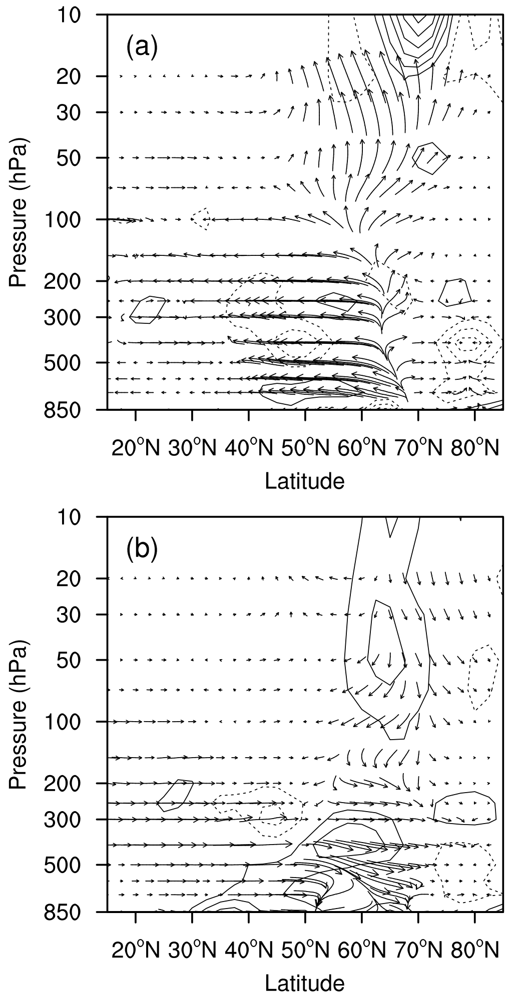

Fig. 8. The change of EP flux (vector) and its divergence (contour) after 2000: (a) December, and (b) March. EP flux divergence is contoured at ±0.5, ±1.0, ±2.0, ±4.0, ±6.0. . . .. Units: m s^{-1} d^{-1}.

defined as the temporal difference of the EAMI, can, and was, used to describe the seasonal evolution. The MTI was found to be characterized by positive and negative values before and after January, respectively. Using data from meteorological stations in China, the temperature tendency revealed that in early winter a uniform cooling is observed at all stations, while in late winter a uniform warming is evident, again at all stations.

As for the long-term variation, the MTI anomalies revealed that the buildup of the EAWM is delayed in late autumn, while it speeds up in early winter. The delay is thought to be caused by strong warming, which prolongs the autumn into the usual early winter days. Meanwhile, the usual stable period in mid winter moves slightly ahead. This will shorten the buildup period in early winter and cause a speedup of the EAWM in early winter. A stronger and earlier collapse of the EAWM in mid winter was observed after 2000, which may also be related to the warming in this region. However, the early spring witnesses a revival of the EAWM, which is also shown with negative temperature tendency in China. The authors suggest that this revival may be related to circulation adjustment and cold surge activity in this region.

The temperature tendency changes showed a south–north contrast in eastern China, with cooling in the south and warming in the north in December. This change is suggested to be related to circulation changes in the stratosphere, which leads to polar warming in mid winter and strong cooling in early spring. The peculiar warming is accompanied by stronger upward planetary wave activity in early winter, and followed by weaker upward planetary wave activity in late winter. Via stratosphere–troposphere dynamical coupling, probably the refraction of planetary wave activity (Perlwitz and Harnik, 2003; Chen and Kang, 2006; Li et al., 2007), this can cause anomalous wave activity in the troposphere. The equatorward wave flux was found to be increased in December, which can cause the temperature contrast in eastern China, i.e. warming in Northeast China and cooling in southern China (Chen et al., 2005). The strong stratospheric polar cooling in early spring is coupled with weaker upward planetary wave activity from the troposphere. Meanwhile, the troposphere equatorward wave flux becomes decreased, which can then lead to cooling in Northeast China and warming in southern China via the wave–EAWM relationship (Chen et al., 2005). The causes of stratospheric mid-winter warming and early spring cooling are still not clear. Hu et al. (2005) suggested that the causes may include the effects of increased greenhouse gases or tropical warming in the mid latitudes and ozone depletion in early spring. However, further studies are needed to understand the exact mechanism.

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