Why Is There an Early Spring Cooling Shift Downstream of the Tibetan Plateau?

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ABSTRACT

The temperature shift over the eastern flank of the Tibetan Plateau is examined using the last 50 yr of Chinese surface station observations. It was found that a strong cooling shift occurs in early spring (March and April) and late summer (July, August, and September) in contrast to the warming shift in other seasons. The cause of the March–April (MA) cooling is investigated in this study.

The MA cooling shift on the lee side of the Tibetan Plateau is found to be not a local phenomenon, but rather it is associated with an eastward extension of a cooling signal originating from North Africa that is related to the North Atlantic Oscillation (NAO) in the previous winter. The midtropospheric westerlies over the North Atlantic and North Africa tend to intensify during positive NAO phases. The enhanced westerlies, after passing over the Tibetan Plateau, result in strengthened ascending motion against the lee side of the plateau, which favors the formation of midlevel stratiform clouds. The increased amount of stratus clouds induces a negative net cloud–radiative forcing, which thereby cools the surface air and triggers a positive cloud–temperature feedback. In this way, the cooling signal from the upstream could "jump" over the Tibetan Plateau and leave a footprint on its lee side. The continental stratiform cloud–climate feedback plays a significant role in the amplification of the cooling shift downstream of the Tibetan Plateau.

1. Introduction

Prominent continental stratus clouds were observed downstream of the Tibetan Plateau (Yu et al. 2004). These middle-level stratus clouds play important roles in shaping the local climate, because these clouds result in a maximum annual mean cloud optical depth and strong negative net cloud–radiative forcing at the top of the atmosphere in the midlatitudes (Yu et al. 2004). The negative net cloud–radiative forcing indicates that the reflection of solar radiation by clouds exceeds their greenhouse effects; thus, these clouds have a cooling effect (Ramanathan et al. 1989). The unique geographic and topographic conditions and the associated circulations might favor a distinct cloud–climate feedback (Yu et al. 2004).

In opposition to the global warming trend, a moderate surface cooling over some regions of Asia over the past several decades has attracted much attention among climatic scientists (e.g., Folland et al. 2001; Houghton et al. 2001; Hu et al. 2003). Immediately downstream of the Tibetan Plateau, a cooling center can be found in spring (Hu et al. 2003). Some of previous studies have shown that local aerosol radiative forcing might contribute to the cooling over southwest China (Qian and Giorgi 2000; Menon et al. 2002; Qian

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et al. 2003). Yu et al. (2004) suggested that the distinctive local deep continental stratus cloud–climate feedback could play an important role in intensifying the climate variation downstream of the Tibetan Plateau. However, the influence of the changes in the large-scale circulation were not fully discussed.

In addition to the local effects, the climate variability in this region has a global linkage to other major modes of climate variability. Wang et al. (2000) have addressed the physical processes by which El Niño affects the "upstream" East Asian summer monsoon in the El Niño decaying phase. A summertime teleconnection between the East Asia–western Pacific monsoon and North America has been documented recently (Wang et al. 2001; Lau and Weng 2002). Kinter et al. (2002) proposed that the Asian monsoon affects ENSO by first affecting the heat content in the western off-equatorial Pacific and then the westerly winds over the western equatorial Pacific.

A number of studies have suggested that the Asian–Pacific climate is not only related to the variations in the local circulation systems, but also significantly related to remote climatic variability sources in the Atlantic, Europe, and the Southern Hemisphere such as the Arctic Oscillation (AO), the North Atlantic Oscillation (NAO), and the Antarctic Oscillation (AAO) (Chang et al. 2001; Gong et al. 2001; Yang et al. 2002; Gong and Ho 2003; Xue et al. 2003; Nan and Li 2003). Chang et al. (2001) showed that the weakening of the correlation between El Niño and deficient Indian monsoon rainfall might be related to changes in the western European surface air temperature in recent years. Yang et al. (2002) concluded that the subtropical jet stream (SJS) is coupled to a teleconnection pattern spanning the entire Asian–Pacific–American region with the strongest signals over East Asia and the western Pacific and the pattern differs significantly from that associated with ENSO. Yu and Zhou (2004) suggested that the winter NAO–generated cooling signal could extend eastward and affect subtropical Eurasia in March.

According to Yu and Zhou (2004), the surface of the Tibetan Plateau exhibits significant warming in March in recent decades. How does the cooling signal jump over the plateau and extend eastward from its upstream region downstream? Does the cooling on the lee side of the plateau really relate to the cooling on its upside? What kind of circulation changes should be responsible for the cooling shift? In a word, the mechanism for this cooling remains an open issue. In this study, we present evidence that the cooling signal originates from upstream and that the local cloud–radiative feedback plays an important role. Section 2 describes the datasets. Section 3 describes the seasonal features of the surface cooling shift downstream of the Tibetan Plateau, in particular the strongest cooling occurring in early spring. The physical processes that may give rise to the cooling shift are discussed in section 4. Section 5 presents a brief summary.

2. Data

The monthly surface temperature data used in this study come from the temperature data of 160 meteorological stations in China. (Figure 1a shows the distribution of those surface stations.) The data were collected and compiled by the China Meteorological Administration and then were interpolated onto 0.5° × 0.5° horizontal grids in order to facilitate climate studies. The monthly cloud data are derived from the total cloud fraction data of 160 meteorological stations in China and have also been interpolated onto the same 0.5° × 0.5° horizontal grids.

The atmospheric general circulation data are taken from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCAR–NCEP) reanalysis data (Kalnay et al. 1996). The data used in this study include temperature, horizontal and vertical velocity, specific humidity, and relative humidity. All of these variables are gridded at 2.5° latitude × 2.5° longitude resolution and are available from 1951 to 2000. In addition, the time series of Hurrell's winter (DJF) NAO (W-NAO) index (Hurrell 1995) was used.

3. Cooling shift downstream of the Tibetan Plateau

The Sichuan basin (27°–32°N, 103°–108°E) in southwest China is located in the eastern flank of the Tibetan Plateau (Fig. 1a). This region is characterized by the world's strongest cloud–radiative forcing and perhaps the strongest continental stratiform cloud–climate feedback. This region is also the region where the largest cooling shift occurs in the past 50 yr over the entire Eurasia continent (Houghton et al. 2001). And the temperature over this area has experienced a sharp cooling in the mid-1970s, which is significant at the 99% confidence level.

Figure 1b shows the long-term changes (also called shift for simplicity) in the area-averaged surface temperature as measured by the difference between two epoch means (1976–2000 minus 1951–75). The annual mean temperature has dropped 0.15°C from the first to the second epoch. The cooling shift, however, varies with season. There are two major cooling periods in the annual cycle. The strongest cooling occurs in early spring (March and April) and another significant cooling shift is observed in late summer (July–September).

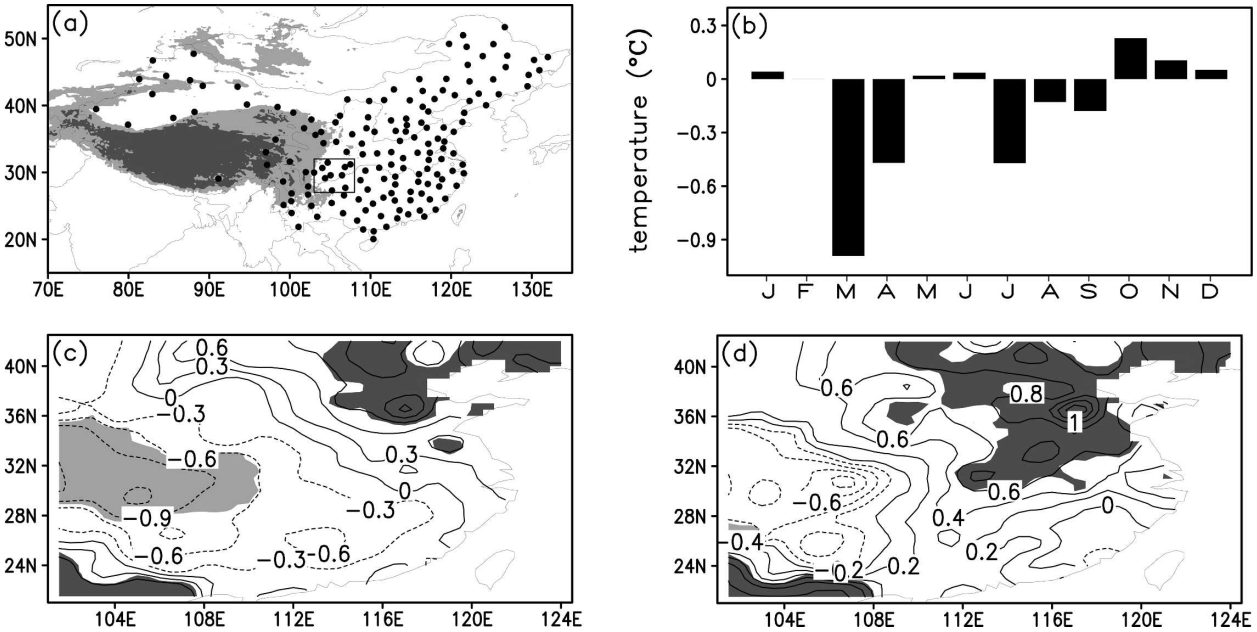

FIG. 1. (a) Heavy dots demonstrate the locations of 160 meteorological stations. The mountain heights are denoted by shading. The edges of light and dark shading denote the contours of 2000 and 4500 m, respectively. The rectangle indicates the location of the Sichuan basin. (b) Annual cycle of climatic mean changes (1976–2000 minus 1951–75) in the surface air temperatures of the Sichuan basin (27°–32°N, 103°–108°E) in units of °C. Climatic mean changes in (c) Mar and (d) Apr surface air temperature in units of °C. Shading in (c) and (d) represents significant changes at the confidence level of 95% using a Student's *t* test. The surface air temperature data are taken from a database of 160 meteorological stations in China.

Figures 1c and 1d present the geographic distributions of the cooling changes in March and April, respectively. The cooling shift is obviously seasonally and geographically dependent. In early spring, the cooling signal is confined to the eastern periphery of the Tibetan Plateau and then weakens after April. In late summer, the cooling changes dominate all of the Yangtze River valley and the strongest cooling center is located away from the periphery of the plateau, which agrees with the previous analysis of Hu et al. (2003) (not shown).

East China has a typical monsoon climate and exhibits a pronounced seasonal variation of the large-scale circulation. The cooling shifts observed in the early spring and late summer could be regulated by different circulation changes.

The primary focus of the present study is the early spring cooling in the Sichuan basin. The vertical structure of the cooling shift in March is shown in Fig. 2. In the longitude–height cross section of temperature change between 1976–2000 and 1951–75 (Fig. 2a), the cooling is seen beneath the middle troposphere over the Sichuan basin, and a northward tilt with the height of the cooling shift is evident. In the zonal–height cross sections of temperature change (Fig. 2b), a large area of cooling shift can be detected over North Africa and subtropical Eurasia. The maximum cooling downstream of the Tibetan Plateau is found right over the Sichuan basin at the 850-hPa level. A notable difference between the cooling over northern Africa and that over southwestern China is that the former penetrates into the upper troposphere, but the latter is restricted in the lower troposphere beneath 600 hPa.

Figure 3 further shows the 925–600-hPa vertical mean temperature change in March. A cooling shift extends from North Africa to East Asia. This pattern is consistent with the spatial distribution of the cooling in the surface air temperature, which is shown in Yu and Zhou (2004). The coherent cooling shifts between the periphery of the Tibetan Plateau and the upstream Eurasian continent suggest that the cooling on the lee side of the plateau might result from the cooling signal from above the plateau. However, a question arises as to why the cooling is intensified and persists until April in the eastern periphery of the Tibetan Plateau. The answer, as will be shown, is that the positive cloud–temperature feedback enlarges the cooling signal passing over the plateau.

4. Discussion: Causes of the cooling shift

Yu et al. (2004) suggested an important role played by a positive feedback between the continental stratus

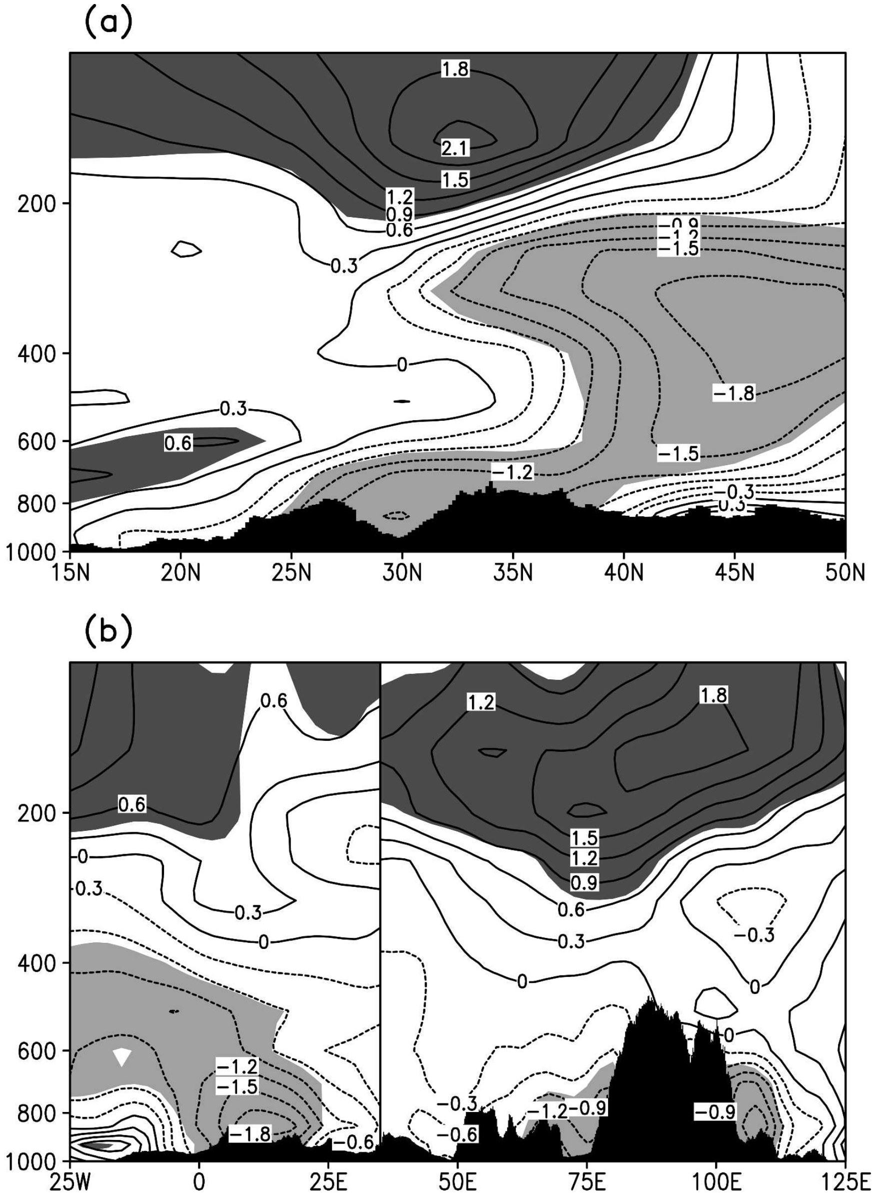

FIG. 2. (a) Meridional–height cross section of temperature changes in Mar between 1976–2000 and 1951–75, from 15° to 50°N meridionally and from 1000 to 100 hPa vertically. The zonal average of the cross section is taken from 102.5° to 107.5°E. (b) Zonal–height cross section of temperature changes in Mar between 1976–2000 and 1951–75, from 15°W to 125°E zonally and from 1000 to 100 hPa vertically. The longitude average of the cross section is taken from 27.5° to 32.5°N in the range of 35°–125°E. From 25°W to 35°E, the longitude average is taken from 20.5° to 25.5°N. Shaded areas are at the 95% confidence level. Temperature data are derived from NCEP–NCAR reanalysis dataset.

clouds and surface temperature through changing lower-tropospheric relative humidity and stratification downstream of the Tibetan Plateau. The positive cloud–temperature feedback could intensify the perturbation temperature signal and retain it for another month. To test this hypothesis, we examine the cloud changes in March and April.

Figures 4a and 4b show that the patterns of the positive cloud changes resemble, to some extent, the cooling changes shown in Figs. 1c and 1d. In March, the total cloud amount increases significantly over the Sichuan basin. The central locations where the cloud amount increases move southward in April with decreasing intensity. The significant negative correlation

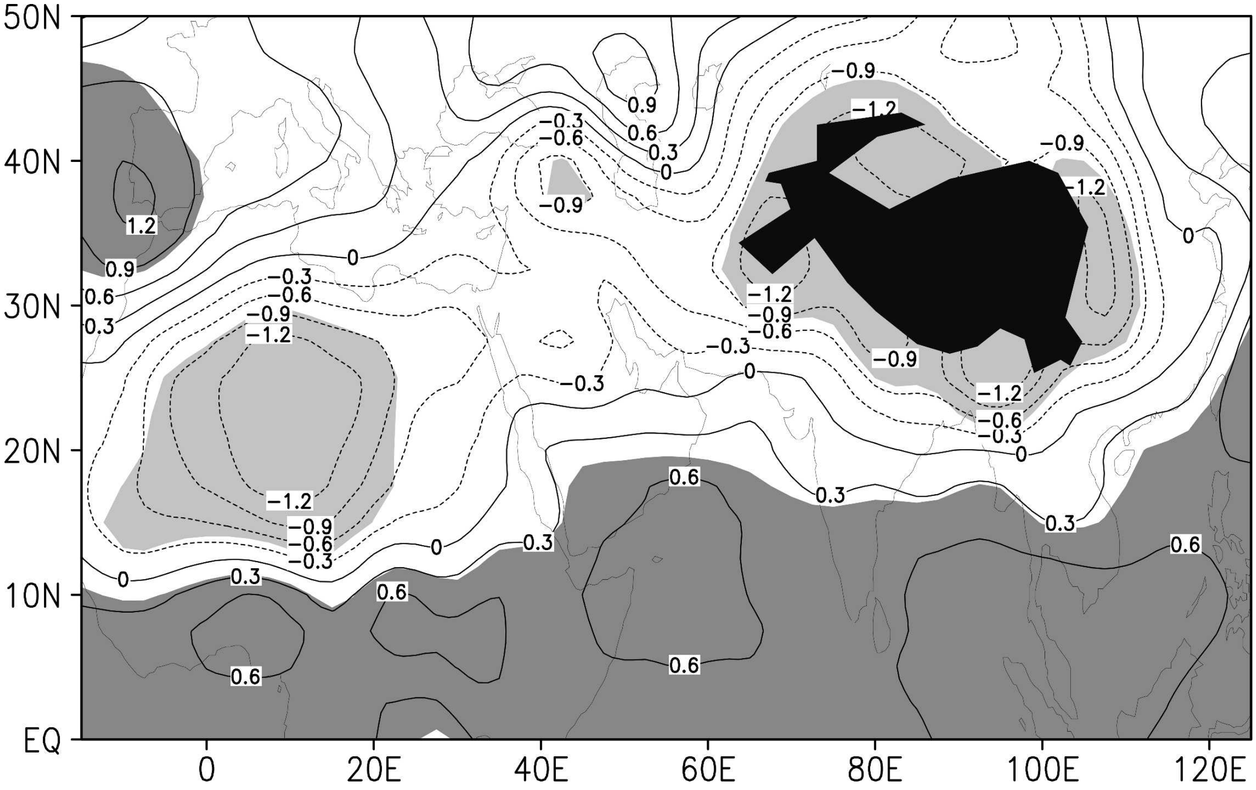

FIG. 3. Climatic mean changes (1976–2000 minus 1951–75) in temperature in Mar averaged between 925 and 600 hPa in units of °C. Temperature data are derived from NCEP–NCAR reanalysis dataset.

between the cloud amount and surface temperature in March is shown in Fig. 5. The mean cloudiness and surface temperature vary in phase and the correlation coefficient between them reaches 0.68 for the period from 1951 to 2000. It is conceivable that the positive cloud trends are responsible for the surface cooling shifts.

An interesting feature of the season-dependent trend of relative humidity and specific humidity over the Sichuan basin (Fig. 6) is that in March and April (and only in these two months) the relative humidity and specific humidity have opposite trends. Although the specific humidity over southwestern China decreases in March, the relative humidity and total cloud amount increase. This implies that the temperature change plays an important role in the adjustment of the relative humidity. In the Sichuan basin, the correlation coefficient between the surface temperature anomaly and relative humidity anomaly is about −0.59 in March for the period from 1951 to 2000, indicating that accompanying the surface temperature decrease, the relative humidity increases, although the air has less water vapor. The surface temperature exerts a dominant influence upon the static stability of the lower troposphere. The correlation coefficient between the surface temperature and the difference between the potential temperature at 850 and 500 hPa reaches −0.70 in March for the period from 1951 to 2000. So, after decades of surface cooling in early spring, the air below the middle troposphere becomes more stable. The positive trends in both the relative humidity and static stability caused by the surface cooling favor the formation of midlevel stratiform clouds. Consequently, the increased optical depth of the stratus clouds over southwestern China further blocks the downward solar radiation flux and cools the surface air by intensifying the negative net cloud–radiative forcing.

After showing evidence that there is a strong feedback between the cloud and temperature happens in early spring over the Sichuan basin, one question left is why the cloud amount has a positive trend. Yu et al. (2004) proposed that the cloud downstream of the Tibetan Plateau is primarily generated and maintained by the frictional and blocking effects of the Tibetan Plateau. The plateau slows down the overflow westerly, inducing downstream midlevel divergence; meanwhile, it forces the low-level circumplateau flow to converge downstream, generating sustained large-scale lifting. The topographically forced upward motion and the stable stratification maintain the thick stratus clouds.

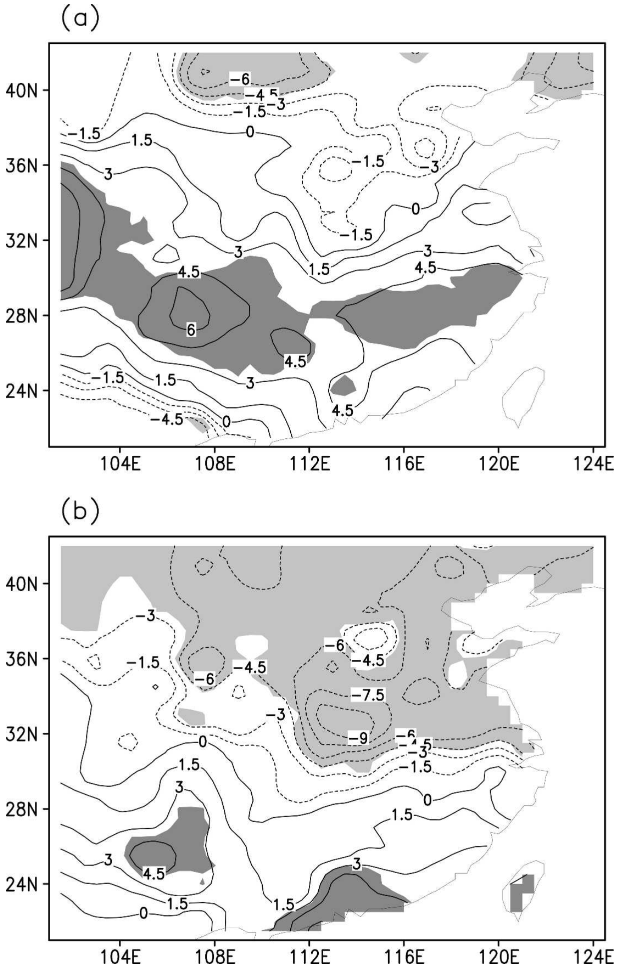

FIG. 4. Climatic mean changes (1976–2000 minus 1951–75) in (a) Mar and (b) Apr total cloud amount in percent. Shaded regions are at the 95% confidence level by using a Student's *t* test. Cloud fraction data come from a dataset of 160 meteorological stations in China.

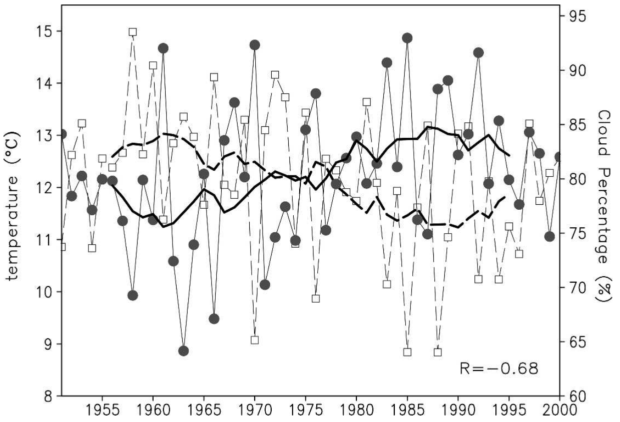

FIG. 5. The 1951–2000 time series of Mar surface air temperature (thin-dashed line with open squares) and total cloud amount (thin solid line with closed circles) averaged over the Sichuan basin (27°–32°N, 103°–108°E). The 11-yr running mean is shown by the thick line. Data are taken from a dataset of 160 meteorological stations in China.

This mechanism also works here as shown in Figs. 7a and 7b. Figure 7a shows the 1951–2000 mean winds at 850 hPa in March. As stated by Trenberth and Chen (1988), the component of the flow going around the Tibetan Plateau dominates that over the plateau. The southern branch of the low-level westerly flows curves its way around the edge of the plateau and heads northwest over the Sichuan basin. Blocked by the plateau, the southeasterly wind converges and goes upward against the elevated land, just as shown in Fig. 7b. The local westerly in the middle troposphere plays an important role in generating the cloud downstream of the plateau. Strong local zonal wind can take the air ascending over the Sichuan basin farther downstream and limit the lifting against the lee side of the Tibetan Plateau within the lower troposphere. The fact that during the recent half century the correlation coefficient of 0.35 between the March zonal wind at 500 hPa averaged over (25°–35°N, 105°–115°E) and the total cloud amount over the Sichuan basin is significant at the 95% confidence level, suggesting an important role for the local westerly wind in the formation of clouds in southwestern China. From the zonal–height cross section of climatic changes of March zonal and vertical streamline (Fig. 7c), we can see that the steady lifting over the Sichuan basin is reinforced by the intensified middle-tropospheric westerly wind and the corresponding strengthened low-level easterly wind. This local circulation change favors the generation of cloud in the recent decades.

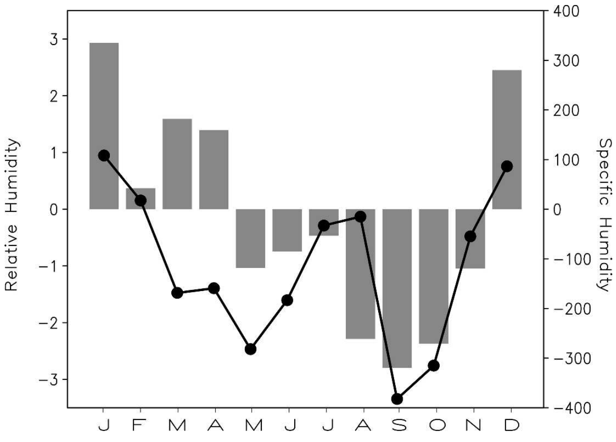

FIG. 6. Annual cycle of climatic mean changes (1976–2000 minus 1951–75) in relative humidity at 850 hPa (gray bars) and specific humidity vertically integrated from 1000 to 500 hPa (black line with closed circles). Both are area averaged over the Sichuan basin (27°–32°N, 103°–108°E).

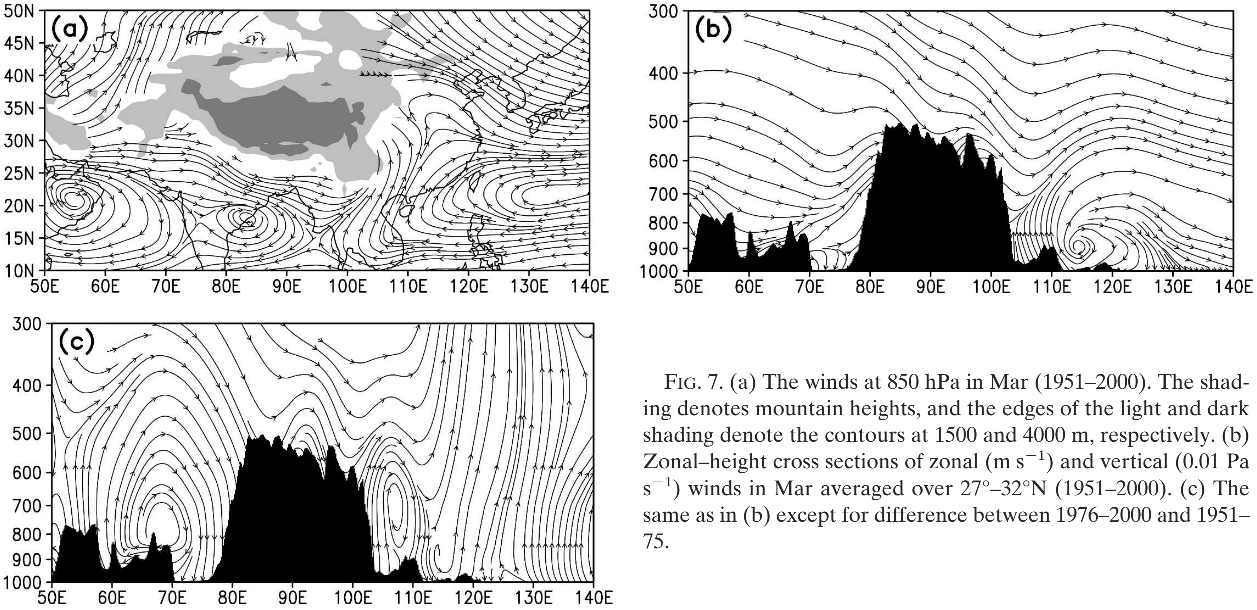

FIG. 7. (a) The winds at 850 hPa in Mar (1951–2000). The shading denotes mountain heights, and the edges of the light and dark shading denote the contours at 1500 and 4000 m, respectively. (b) Zonal–height cross sections of zonal (m s^{-1}) and vertical (0.01 Pa s^{-1}) winds in Mar averaged over 27°–32°N (1951–2000). (c) The same as in (b) except for difference between 1976–2000 and 1951–75.

The March local zonal wind at 500 hPa is linked with the winter NAO. Figure 8a shows the trend of the March 500-hPa zonal wind. A coherent speedup of the 500-hPa zonal wind occurs in North Africa and subtropical Eurasia in March. Over the North Atlantic, the changes in the zonal wind field at the middle troposphere present a positive NAO phase pattern. This is consistent with the persistent positive phase of the NAO in the past two decades (Marshall et al. 2001). Figure 8b indicates that in the past half century the distribution of the correlations between the winter NAO index and the March zonal wind at 500 hPa has a similar pattern over North Atlantic and subtropical Eurasia. Hoskins and Ambrizzi (1993) and Branstator (2002) have shown that the midlatitude jet that stretches across South Asia acts as a waveguide and variability tends to be extended downstream along this jet. The regional average of the correlation coefficient between the winter NAO index and the March zonal wind at 500 hPa over southwest China (25°–35°N, 105°–115°E) is 0.34, implying that the change of the zonal wind over this area is partly associated with the previous winter NAO. We speculate that the strengthened westerlies at 500 hPa over North Africa and the Middle East associated with the positive phases of the winter NAO might extend their influence downstream and lead to stronger March zonal wind in the middle troposphere over southwestern China. The correlation between the total cloud anomaly and the winter NAO index is 0.34 for the period from 1951 to 2000. By altering the local zonal wind downstream of the plateau, the winter NAO can have a delayed influence on the amount of cloud over southwestern China.

The above results show that the positive trend of the winter NAO might result in significant intensification of the 500-hPa zonal wind over subtropical North Africa. This anomalous perturbation can extend eastward along the subtropical westerly jet stream, pass over the plateau, and affect southwestern China. The intensification of westerlies over the Sichuan basin favors the formation of middle-level stratus clouds. Then, with the increasing total cloud amount, the cloud–surface temperature feedback may help explain why the cooling is intensified and maintained on the lee side of the plateau.

5. Conclusions

The observed long-term variations of surface temperature in southwest China are analyzed and a possible mechanism responsible for this shift is discussed. The linear trend of surface temperature in China in the past several decades is highly spatially and seasonally dependant. A cooling trend exists in southwest China and the most significant cooling happens in March. Our analyses show that the decrease of the surface temperature over the Sichuan basin is in accord with the increase of the local midlevel stratiform clouds. The significant correlation between them suggests a positive cloud–temperature feedback. The formation of cloud over this area is highly dependent on the circulation over and around the Tibetan Plateau. Strengthened westerly flow in the middle troposphere favors large-scale lifting in the lower troposphere on the lee side of the plateau. This change can lead to the increase of midlevel stratiform clouds that then trigger the positive

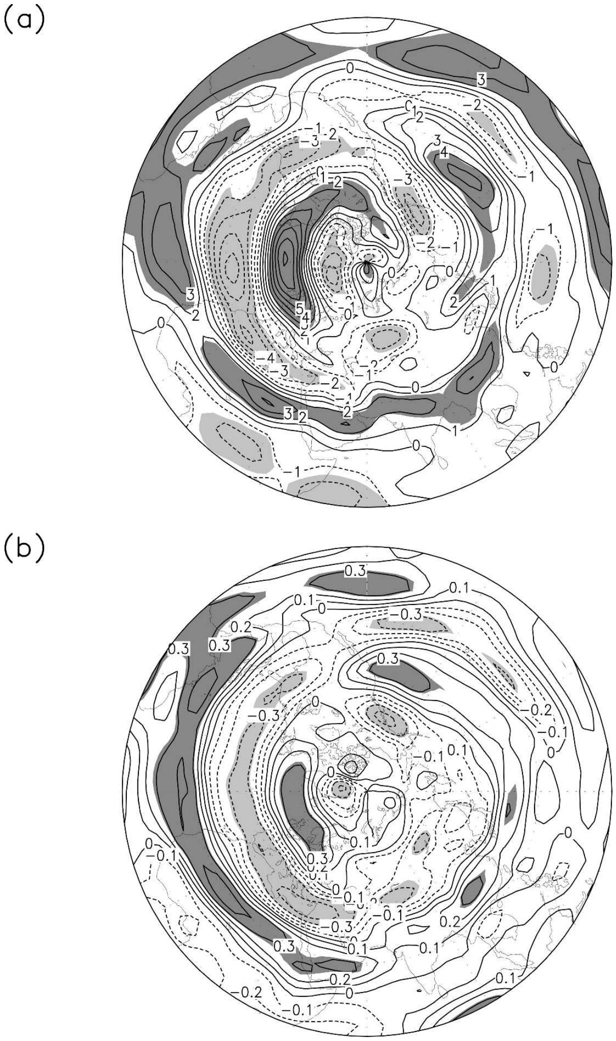

FIG. 8. (a) Climatic mean changes (1976–2000 minus 1951–75) in Mar zonal wind at 500 hPa in units of m s^{-1}. (b) Distribution of correlations between the winter NAO index and the Mar zonal wind at 500 hPa for the period from 1951 to 2000. Shaded area is for the 95% confidence level.

cloud–temperature feedback. Evidence is presented to show that the zonal wind intensification over the plateau may be traced upstream to a positive NAO in the previous winter.

The above conclusion is based on limited data. It leaves a number of questions unaddressed. One of them is why the perturbation of the midtroposphere zonal wind is amplified over East Asia after extending from upstream. In March, a separate cooling center is located in the upper troposphere over the midlatitudes of east China (Fig. 2). This strong cooling appears in the previous winter and is primarily confined to the troposphere and stratosphere above 400 hPa. In the early spring, the cooling penetrates downward and reaches the 600-hPa level. This change in the temperature field might favor the intensification of midtroposphere zonal wind over southwest China. So, the thermodynamic structure of the upper troposphere might play a role in the teleconnection between the winter NAO and the surface air temperature over the lee side of the Tibetan Plateau. Further investigation is needed to clarify this issue in the future.

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