Article ID: 1006-8775(2009) 01-0020-08

EFFECTS OF THE ATMOSPHERIC COLD SOURCE OVER THE TIBETAN PLATEAU ON THE QUASI 4-YEAR OSCILLATION OF OCEAN-ATMOSPHERIC-LAND INTERACTION

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Abstract: Using correlation analyses, composite analyses, and singular value decomposition, the relationship between the atmospheric cold source over the eastern Tibetan Plateau and atmospheric/ocean circulation is discussed. In winter, the anomaly of the strong (weak) atmospheric cold source over the eastern plateau causes low-level anomalous north (south) winds to appear in eastern China and low-level anomaly zonal west (east) winds to prevail in the equatorial Pacific from spring to autumn. This contributes to the anomalous warm (cold) sea surface temperature the following autumn and winter. In addition, the anomalous variation of sea surface temperature over the equatorial middle and eastern Pacific in winter can influence the snow depth and intensity of the cold source over the plateau in the following winter due to variation of the summer west Pacific subtropical high.

Key words: climatology; atmospheric cold source; diagnostic analysis; Tibetan Plateau

CLC number: P434 **Document code:** A doi: 10.3969/j.issn.1006-8775.2009.01.003

1 INTRODUCTION

The significance of the Tibetan Plateau as an elevated heat source for the onset of the Asian summer monsoon circulation has been discussed by many authors. Ye et al. [1] suggested that the atmosphere over the plateau in the upper troposphere acts as a heat source in summer and as a heat sink in winter. Chen et al. [2] indicated that this heat source controls the Tibetan Plateau from April to August, and then the heat sink is the dominant force from August to the subsequent February. Many authors [3-10] have studied the thermal conditions of the Tibetan Plateau. Flohn [11] discovered that the seasonal variation of the elevated heat source over this region and the counter-phase relationship of the temperature and pressure gradient between the Tibetan Plateau and the region south of it trigger the variation of the general circulation in Asia and encourage the onset of a monsoon on the Indian subcontinent. Zhang and Qian [12] found that the longitudinal and latitudinal thermal contrasts caused by the middle-latitude Plateau-sensible heating trigger the onset of the Asian summer monsoon.

Regarding the winter cold source over the Tibetan Plateau, many authors have performed research on this topic. Philip [13] analyzed the heat source/sink at the lower troposphere in winter and found that the horizontal scale of the heat source corresponds to that of the temperature field at the lower troposphere. Yanai and Tomita [14] indicated that the major heat sources in winter are located in the region extending from the tropical Indian Ocean through Indonesia to the southern Pacific, Congo, Amazon Basin, Asia, and North America and its east coast. The results of Cen et al.[15] showed that the intensity of the heat source over the eastern plateau is greater than that over the western plateau in summer and winter. Hua [16] discovered that the heat source plays an important role in the formation and maintenance of the general circulation. Using a direct method, Zhao and Chen [17] studied the cold source over the Tibetan Plateau and found that it can influence the Sea Surface Temperature (SST) over the equatorial central and eastern Pacific. Then, Chen et al. [18] studied the effect of the cold source over the Tibetan Plateau on the zonal wind anomaly over the equatorial

Received date: 2009-03-08; **revised date:** 2009-05-20

Foundation item: Natural Science Foundation of China (90711003, 40633018)

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Pacific. The temporal and spatial variation of snow cover showed that the key areas for the interannual variation are located in the Tibetan Plateau, the Mongolian Plateau, the Alps, and the central and western regions of North America; furthermore, the Tibetan Plateau is the most important region with the largest interannual variability (Yang and Zhang [19]). Snow anomalies over the Tibetan Plateau change the soil moisture and the surface temperature first through the melting process of snow, which alters the heat, moisture, and radiation fluxes from the surface to the atmosphere accordingly. Abnormal circulation conditions induced by changes of surface fluxes may affect the underlying surface properties in turn (Qian et al. [20]).

Many authors have studied the effect of the winter cold sources over the Tibetan plateau on SST in the equatorial Pacific; however, there have been few discussions about the effect of SST on the cold source over the plateau. Zhao and Chen [17] considered that the variation of SST over the equatorial Pacific can change the location of the East Asian trough in winter and the intensity of the winter cold source over the plateau by ocean-atmosphere interaction at mid- and low-latitudes. Since snow in the plateau and its adjacent areas can change the thermal conditions from the surface to the atmosphere, it is reasonable to wonder whether SST over the equatorial Pacific can affect the cold source over the plateau through snow. In this paper, the authors will discuss the relationships among the cold source over the plateau, SST in the equatorial Pacific, the west Pacific subtropical high (WPSH), and snow in the plateau.

2 DATA AND METHODS

2.1 *Data*

The datasets that were used were: (1) the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP/NCAR) reanalysis daily data (of u, v, t, ω, and h) with 2.5° × 2.5° from 1961 to 2001; (2) the 1961–2001 monthly area index of the WPSH from the National Climate Center (NCC) of China; (3) the 1961–1999 monthly SST data of UK/GISST2 at 1° × 1° resolution; and (4) the snow depth data of 86 stations from 1961 to 1998 provided by the National Meteorological Information Center.

2.2 *Calculation of the atmospheric heat source*

From the thermodynamic equation, the atmospheric heat source (AHS) can be denoted as

$$Q_1 = C_p \left(\frac{p}{p_0} \right)^k \left(\frac{\partial q}{\partial t} + V \cdot \nabla q + w \frac{\partial q}{\partial p} \right) , \qquad (1)$$

where Q_1 is the AHS pre unit. Integrating Eq. (1) by mass weight in the entire atmosphere, we get

$$\langle Q_1 \rangle = \frac{1}{g} \int_{P_t}^{P_s} Q_1 dp \quad , \qquad (2)$$

where P_s and P_t are pressures at the surface and at the top (100 hPa), respectively, and $<Q_1>$ is the vertical integration of Q_1 within an air column unit (Yanai et al.[21]).

In the present study, the winter refers to the period from December to the subsequent February, and the summer refers to the period from June to August.

3 RELATIONSHIPS AMONG THE COLD SOURCE IN THE PLATEAU, SST IN THE PACIFIC, THE WPSH, AND SNOW IN THE PLATEAU

3.1 *Distribution and interannual variation of the winter cold source in the plateau*

From the 1961–2000 distribution of the cold source over the plateau in winter (Fig.1), it is seen that the intensity of the cold source in the eastern plateau is stronger than that in the western part. Since the stations of the plateau are mainly concentrated in the eastern part (Fig.2), the eastern plateau (27.5°–35°N, 90°–100°E) is used as the key region in this paper. Fig.3 shows the interannual variation of the winter cold source in the eastern plateau from 1961 to 2000, the strongest cold source with the value of –83.3 W·m^{-2} appears in 1974, and the weakest cooling (–27.6 W·m^{-2}) occurs in 1963. Based on the trend line, we can see that the intensity of the winter cold source has an increasing trend at first, and then weakens later.

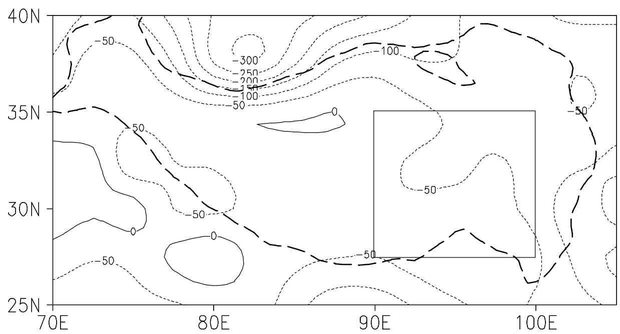


Fig.1 Distribution of the cold source over the Tibetan Plateau in winter during 1961-2000. (unit: W/m^2, the long dashed line is the 3000-m contour of topography and the rectangular box is the key region in this paper)

3.2 *The relationship of the winter cold source in the eastern plateau and 850-hPa wind*

In this part, we define a correlation vector as

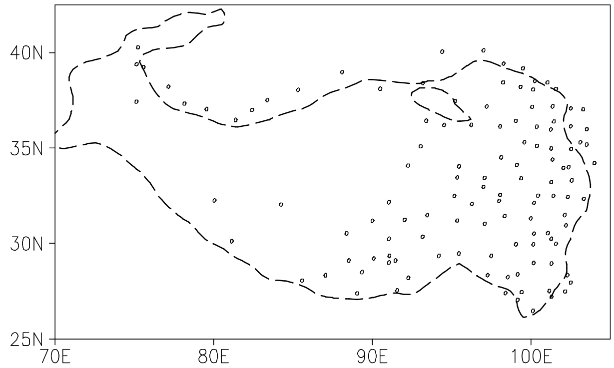


Fig.2 Distribution of stations in the plateau and its adjacent areas.

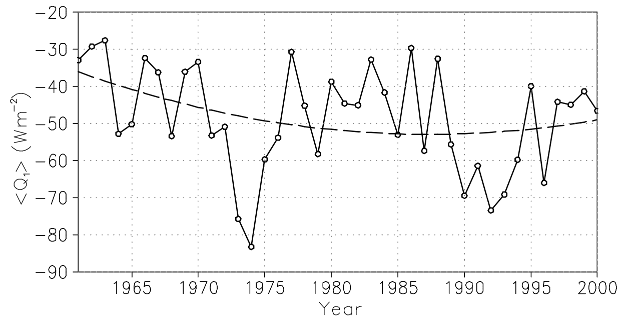


Fig.3 Interannual variation of the winter cold source in the eastern plateau from 1961 to 2000 (unit: W/m^2, the long dashed line is the trend line).

follows:

$$\vec{R} = (-1)\times R_{<Q_1>\&u}\vec{i} + (-1)\times R_{<Q_1>\&v}\vec{j} \quad , \qquad (3)$$

where $R_{<Q1>\&u}$ ($R_{<Q1>\&v}$) is the correlation coefficient between the cold source and the 850-hPa zonal (meridional) wind. Because the atmospheric thermal condition is cooling over the eastern plateau in winter (i.e., the value of $<Q_1>$ is less than 0), $R_{<Q1>\&u}$ and $R_{<Q1>\&v}$ are multiplied by –1 in Eq. (3). In this paper, we describe the correlation vector in accordance with the wind field.

Fig.4 shows the correlation vectors between the winter cold source in the eastern plateau and the subsequent 850-hPa wind field in March, June, September, and December. In March (Fig.4a), the northeasterly wind controls most parts of eastern China; it passes through the Indo-China Peninsula, and turns into the equatorial westerly wind around Kalimantan Island. In the area between Kalimantan Island and the date line, the westerly wind appears in the equatorial Pacific; in the region east of 180°E, the westerly wind is located near 10°N, and the northwesterly wind prevails over the equatorial eastern Pacific. In June (Fig.4b), the scope of the northerly wind over east of China expands from northeast China to the northern part of the Indo-China Peninsula. The westerly wind from the western, southern, and northeastern Pacific covers the equatorial central and eastern Pacific. In September (Fig.4c), the intensity of the northerly wind significantly decreases in eastern China, and the strong westerly wind stays in the equatorial central and eastern Pacific. Until December (Fig.4d), in the region from the Philippine Islands through the South China Sea and the southern part of the Sino-Indian Peninsula to the equatorial Indian Ocean, the easterly wind is present; during this time, the weak southerly wind is located around the Taiwan Strait.

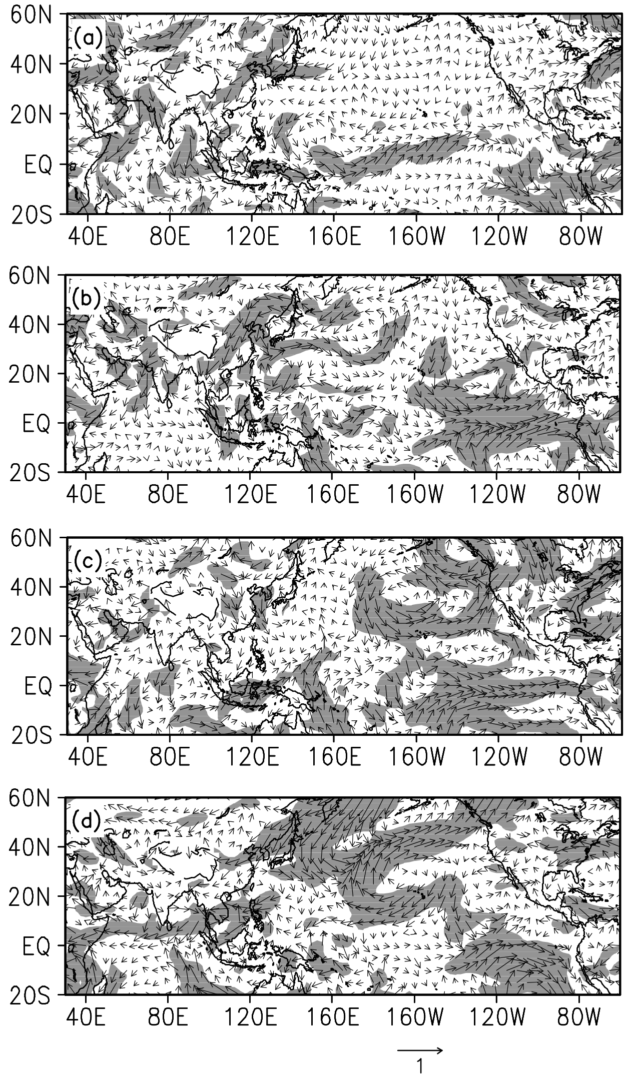


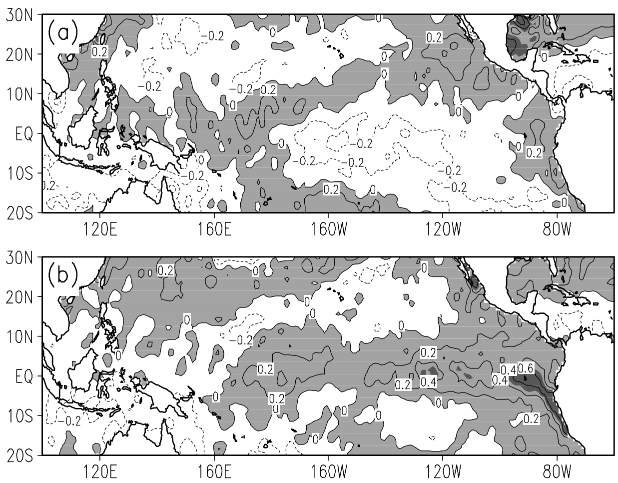
Fig.4 Distributions of the correlation vector between the winter cold source in the eastern plateau and the subsequent 850-hPa wind field in (a) March, (b) June, (c) September, and (d) December. (The shaded regions pass the test at 90% significance and the long dashed line is the 3000-m contour of topography)

In order to explain the relationship between the winter cold source over the eastern plateau and the 850-hPa wind, the lag correlation coefficients are calculated. The largest lag correlation coefficients between the winter cold source and meridional and zonal wind at 850 hPa in east of China appears in May and August and surpasses the 90% significance mark.

3.3 *The relationship of the winter cold source in the eastern plateau and SST in the equatorial Pacific*

We defined the winters when the normalized cold source in the eastern plateau was greater than 1.0 to be the strong cold source years and those to be the weak cold source years when the normalized cold source was less than –1.0. The strong years were 1973, 1974, 1990, 1992, 1993, and 1996; the weak years were 1961, 1962, 1963, 1966, 1970, 1983, 1986, and 1988.

We analyzed the distributions of the SST anomaly (SSTA) in the strong and weak years and discussed the response of the subsequent SST in the equatorial Pacific to the cold source. In the strong years, in March (Fig.5a), the positive SSTA regions appear in the equatorial western Pacific, while the negative SSTA regions are in the equatorial central and eastern Pacific. In June (Fig.5b), the positive SSTA region extends westward from Peru to 140°W, and the original positive SSTA region in the equatorial western Pacific expands to the equatorial central Pacific. Therefore, the positive SSTA covers all of the equatorial central and eastern Pacific with the central value of 0.8°C. In September (Fig.5c), the positive SSTA region in the equatorial central and eastern Pacific expands further. In December (Fig.5d), the value of the positive SSTA in the equatorial central Pacific reaches its peak, and the regional SSTA value in the Niño3 region exceeds 0.5°C. In the weak years, in March (Fig.6a), the negative SSTA regions appear in the equatorial central and eastern Pacific, and the value of SSTA around the date line reaches –0.6°C. In June (Fig.6b), there is an abnormally strong cold tongue in the coast of Peru. In September (Fig.6c), the region of the negative SSTA in the equatorial central and eastern Pacific is reduced. However, in November (Fig.6d), the intensity of the negative SSTA increases and the central value reaches –0.4ºC.



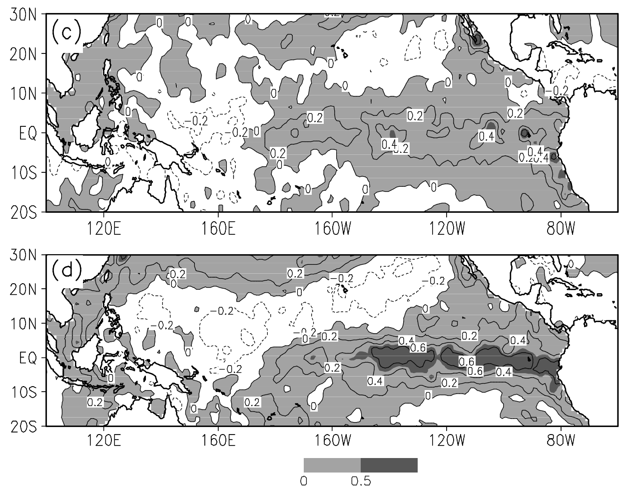


Fig.5 Composite SSTA fields of (a) March, (b) June, (c) September, and (d) December based on the strong winter cold source in the eastern plateau. (unit: °C)

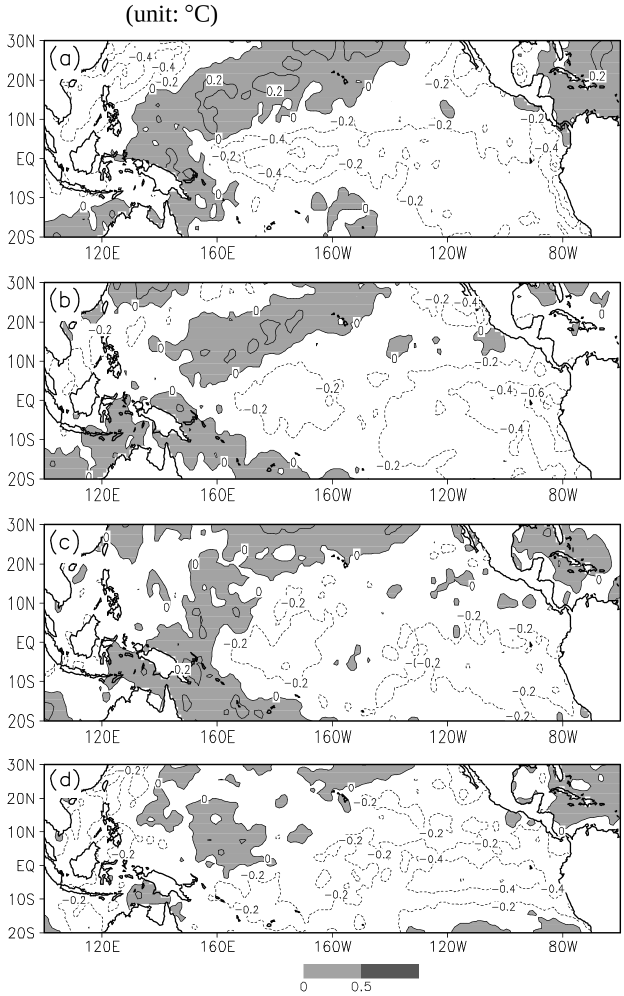


Fig.6 Same as Fig.5, but based on the weak winter cold source in the eastern plateau.

Using the Automatic Weather Stations (AWS) for thermal balance observations through the Sino-Japanese Asian Monsoon Mechanism Co-operative Project, Zhao and Chen [17] calculated the atmospheric heat source in the plateau by a direct method, and determined that the variation of the winter cold source in the plateau can influence SST in the equatorial central and eastern Pacific. In this paper, using the NCEP/NCAR reanalysis datasets and reverse method, we verify the conclusions of Zhao and Chen [17].

3.4 *The relationship of the WPSH and the snow depth in the eastern Tibetan plateau*

From the above analysis, it can be seen that the variation of the winter cold source in the eastern plateau affects the subsequent SST in the equatorial central and eastern Pacific. We should next determine whether the variation of the SST can affect the winter cold source in the eastern plateau in turn.

Fig.7 shows the time series of the mean SST from October to December 1961–1999 in the Niño3 region and those of the area index of the 1962–2000 summer WPSH. The correlation coefficient of the two time series is 0.64, exceeding the significance level of α = 0.01, which means that they have a strong correlative relationship. The high (low) SST favors the WPSH to strengthen (weaken). This result is consistent with that of Long and Li [22]; they indicated that during the summer after El Niño, the subtropical high over the western Pacific is intensified and shows northward and westward displacement. Wang and Wu et al. [23] demonstrated that the anomalous Philippine Sea anticyclone results from a Rossby-wave response to suppressed convective heating in the western Pacific.

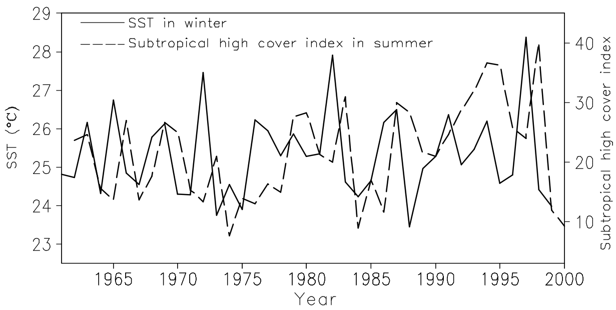


Fig.7 Time series of the mean SST from October to December in Niño3 region during 1961-1999 and those of the area index of the summer WPSH during 1962-2000.

It is important to determine whether the WPSH can affect the winter cold source over the eastern plateau. Firstly, Fig.8 shows the distribution of the correlation coefficients of the summer WPSH with the mean 1961–1998 snow depth from November to December in the plateau. It is seen that the remarkable negative correlation emerges in the eastern plateau, indicating that if the WPSH is strong (weak), there is less (more) snow over the eastern plateau from November to December. This characteristic can be reflected in the difference fields of 500-hPa geopotential height between more and less snow years over the eastern plateau (more years minus less years) (Fig.9). On the basis of the normalized snow depth over the eastern plateau from November to December 1961–1998, the more (or less) snow years can be determined. The more snow years are 1961, 1963, 1967, 1985, and 1997; the less snow years are 1968, 1969, 1980, and 1990. From November to December (Fig.9d), the negative difference fields appear in the plateau and the northwestern Pacific, and positive ones in Siberia with the center over the north side of Lake Baikal, meaning there is a strong Ural high ridge. Because of the deep East Asian trough and the strong Ural high ridge, cold air coming from Siberia can easily affect the plateau. Fig.10 shows the distribution of correlation coefficients in the area index of the summer WPSH and the subsequent 500-hPa wind field from November to December. The northerly flows extend from the mid- and high-latitudes in East Asia to the plateau through northeast China and north China, and induce more snowfall over the plateau. From summer (Fig.9a) to September (Fig.9b), most of the subtropical regions are controlled by the negative differences that control most of East Asia until October (Fig.9c).

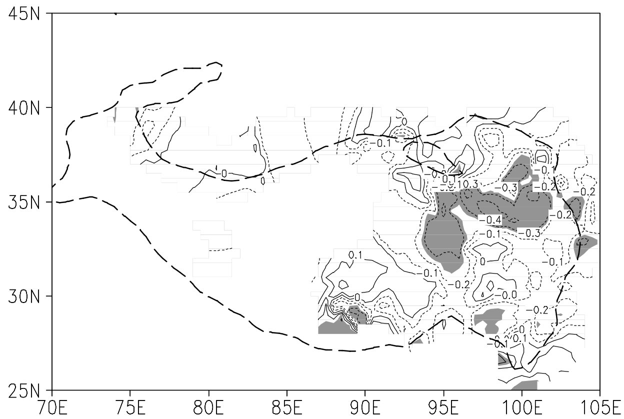


Fig.8 Distribution of correlation coefficients of the summer WPSH with the mean snow depth from November to December in the plateau during 1961-1998. (The shaded regions pass the test at 90% significance and the long dashed line is the 3000-m contour of topography)

3.5 *The relationship of the snow depth in the eastern Tibetan Plateau and the winter cold source*

Through the singular value decomposition analysis (SVD) of the winter cold source in the plateau (the left field) and the snow depth from November to December (the right field), we obtain the heterogeneous correlation maps (Fig.11) and the time coefficients (Fig.12). The variance contributions of the first two modes are 38.6% and 18.0%. The correlation coefficients between the winter cold source and snow depth in the first and second modes are 0.6 and 0.75, respectively, which means that there is a close

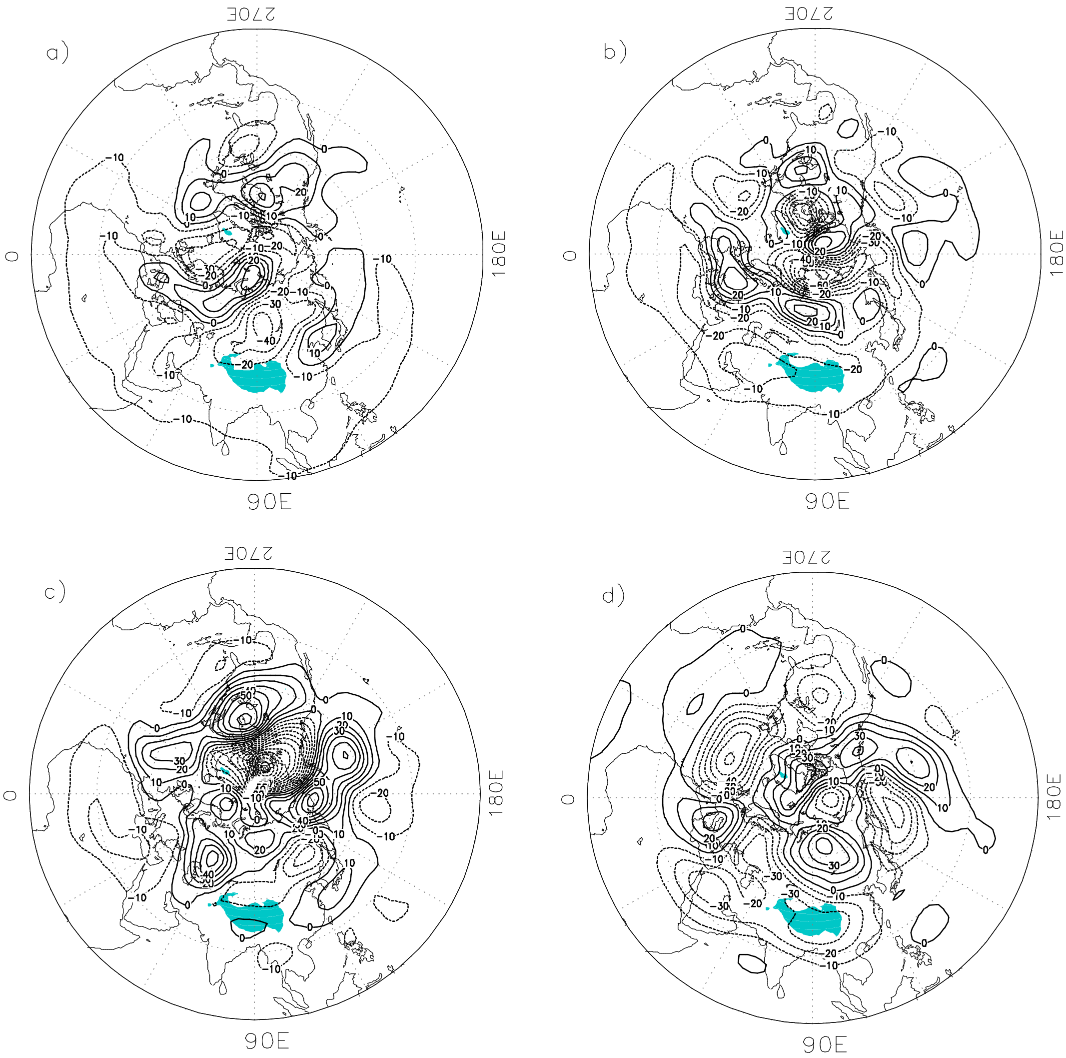


Fig.9 Difference field of 500-hPa geopotential height between more and less snow years over the eastern plateau in (a) summer, (b) September, (c) October, and (d) from November to December. (The shaded region is 3000-m contour of topography)

relationship (Figs.12a and 12b). The distribution of the negative cold source in most of the plateau (Fig.11a) and the positive snow depth over the central plateau (Fig.11b) indicate that if the snow depth from November to December is more (less), there is a strong (weak) cold source in the winter. The most remarkable region is over the eastern plateau in the second left mode, and the best correlation coefficients appear over the same region in the second right mode; it is seen that more (less) snow depth from November to December over the eastern plateau is related to the strong (weak) cold source in winter.

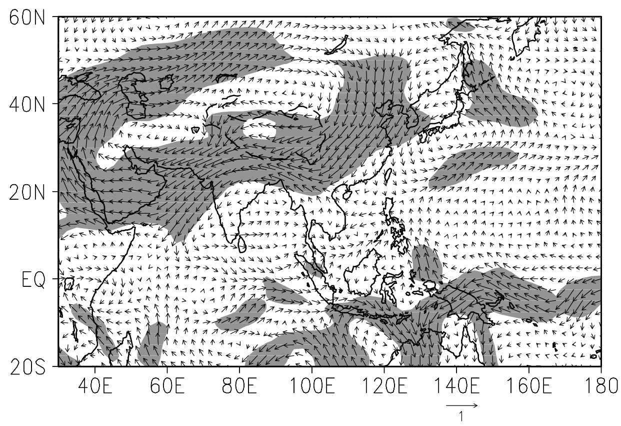


Fig.10 Distribution of correlation coefficients between the area index of the summer WPSH and the subsequent 500-hPa wind field from November to December. (The shaded regions pass the test at 95% significance level, the long dashed line is the 3000-m contour of topography)

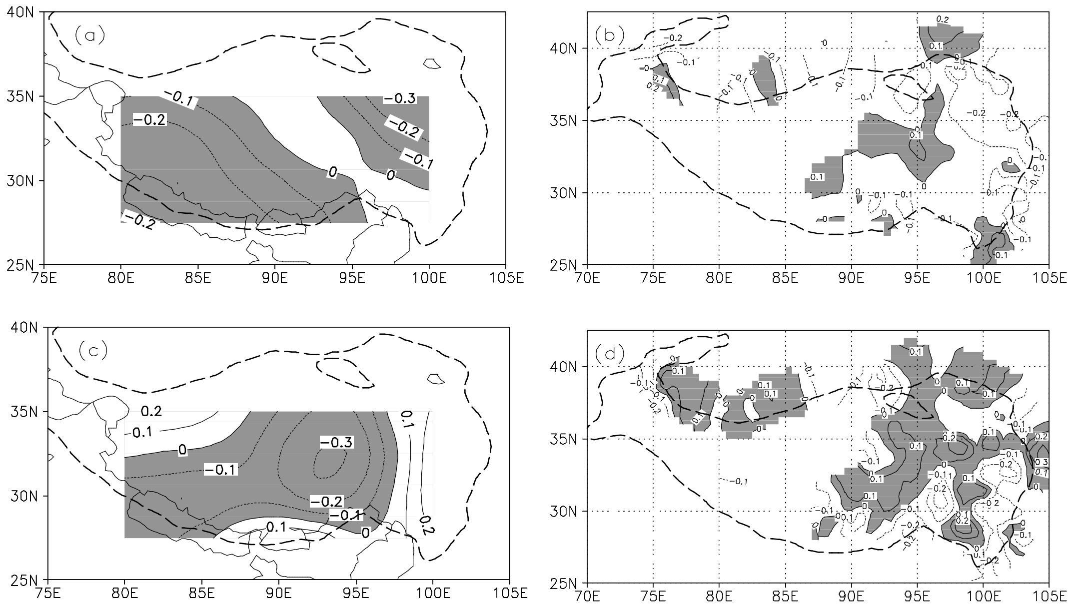


Fig.11 SVD between the winter cold source and the snow depth from November to December. (a) The left heterogeneous correlation map of the first mode; (b) the right heterogeneous correlation map of the first mode; (c) The left heterogeneous correlation map of the second mode; (d) the right heterogeneous correlation map of the second mode. (The long dashed line is the 3000-m contour of topography)

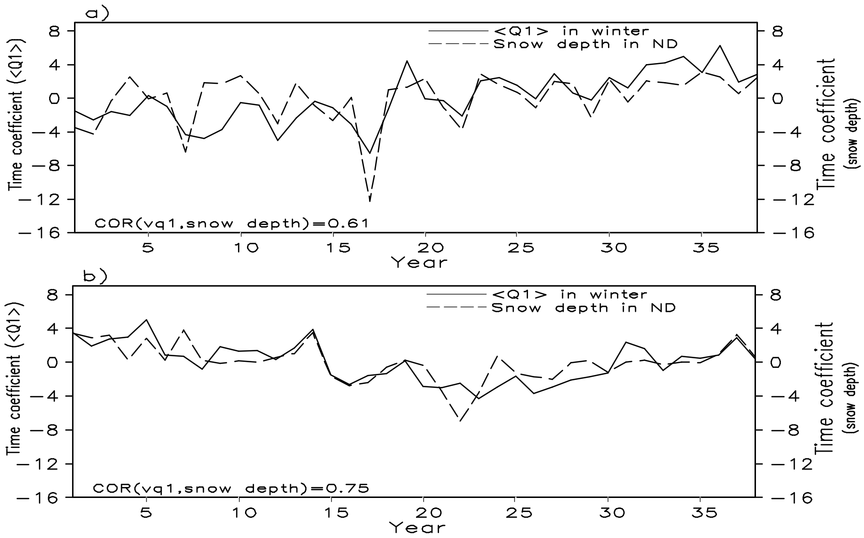


Fig.12 Time coefficients of (a) the first mode; (b) the second mode.

4 CONCLUSIONS AND DISCUSSION

In this paper, we discuss the relationship between the winter cold source over the eastern plateau and the general circulation. It is shown from a quasi-4-year cycle (Fig.13) that the abnormal strong (weak) winter cold source in the eastern plateau induces the abnormal northerly (southerly) winds passing the Chinese mainland and the Indo-China Peninsula, and then turning into the abnormal westerly (easterly) winds near Kalimantan Island. This condition aids the formation of El Niño (La Niña). However, the increase (decrease) of SST in the equatorial Pacific induces the strong (weak) summer WPSH, less (more) snow from November to December, and the weak (strong) cold source in winter over the plateau. Therefore, we conclude that a coupled atmosphere-land-ocean process occurs among the winter cold source in the eastern plateau, the SST in the equatorial Pacific, the WPSH, and snow depth in winter over the plateau. This process exhibits a quasi-4-year oscillation and is consistent with the 3-5 year period of ENSO. In this paper, we discuss the snow in the plateau. The possibility that there is a relationship between snow in the plateau and that in Eurasia is the topic of our next work.

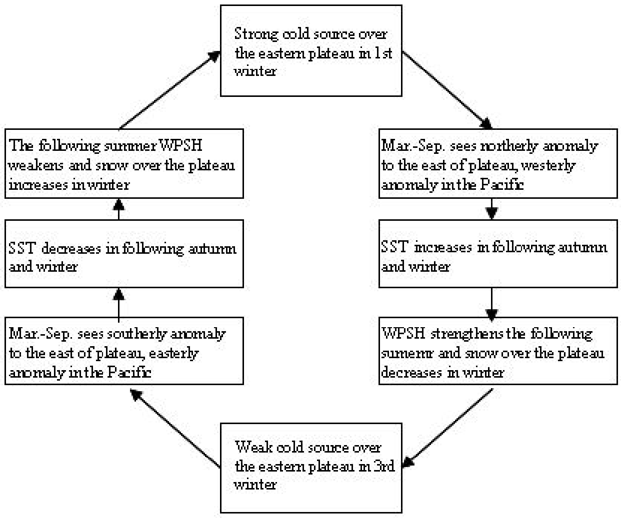


Fig.13 Quasi-4-year cycle among the cold source over the eastern plateau in winter, SST in the equatorial central and eastern Pacific, the WPSH and snow depth in the plateau.

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Citation: ZHANG Bo, CHEN Long-xun and ZHOU Xiu-ji et al. Effects of the atmospheric cold source over the Tibetan Plateau on the quasi 4-year oscillation of ocean-atmospheric-land interaction. *J. Trop. Meteor.*, 2009, 15(1): 20-27.