JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 113, D14117, doi:10.1029/2007JD009567, 2008

Arctic Oscillation and the autumn/winter snow depth over the Tibetan Plateau

Jun-Mei Lü,[1,3] Jian-Hua Ju,[2] Seong-Joong Kim,[3] Ju-Zhang Ren,[4] and Yu-Xiang Zhu[2]

Received 4 November 2007; revised 15 April 2008; accepted 1 May 2008; published 22 July 2008.

[1] The present study examines the relationship between the Arctic Oscillation (AO) and the autumn/winter snow depth over the Tibetan Plateau. Results show that there exists significant correlation between the AO and the Tibetan Plateau snow depth on interdecadal timescale. The AO and the snow depth over the Tibetan Plateau experienced interdecadal regime shift in the late 1970s. Before the late 1970s when the AO was in its interdecadal negative phase, the snow depth over the Tibetan Plateau increased in fall and then decreased in the following winter. Conversely, when the AO has entered its interdecadal positive phase since the early 1980s, the snow depth decreased in fall, but increased in winter. The vertical propagation of Rossby waves is proposed to explain the physical process linking the AO with the snow depth. Anomalously excessive fall snow depth over the Tibetan Plateau amplifies orographically forced upward stationary waves. The snow-forced changes in stratosphere are not identified until later in the winter season when Rossby waves propagate into the stratosphere and the AO becomes negative. In winter, when the troposphere and stratosphere are actively coupled, the downward propagation of Rossby waves associated with the positive AO phase modulates the atmospheric circulation in the troposphere, and causes the abnormal increase of snow depth over the Tibetan Plateau.

Citation: Lü, J.-M., J.-H. Ju, S.-J. Kim, J.-Z. Ren, and Y.-X. Zhu (2008), Arctic Oscillation and the autumn/winter snow depth over the Tibetan Plateau, *J. Geophys. Res.*, *113*, D14117, doi:10.1029/2007JD009567.

1. Introduction

[2] The Arctic Oscillation (AO) is a dominant and internal mode of atmospheric variability in the Northern Hemisphere (NH). It plays an important role on the wintertime NH climate in mid to high latitudes [*Thompson and Wallace*, 2000, 2001; *Kerr*, 1999], including East Asia [*Wu and Wang*, 2002; *Gong et al.*, 2001; *Gong and Wang*, 2003; *Wu and Qian*, 2003]. As a barotropic mode in the NH, the AO stretches vertically from the surface into the lower stratosphere through the troposphere, acting as a medium connecting the stratosphere with the troposphere. The feedback can be bidirectional between the troposphere and the stratosphere [*Hartley et al.*, 1998]. *Kodera and Kuroda* [2000] indicated two different types of coupling between the troposphere and the stratosphere. One type occurs in late autumn/early winter with tropospheric origin, while the other preferentially occurs in mid to late winter and originates from the stratosphere. The dynamical mechanism involved in maintaining the AO is still not clear. One possibility is through the internal interaction between the zonal-mean flow and waves in atmosphere [*DeWeaver and Nigam*, 2000; *Chen and Huang*, 2005]. It was demonstrated that the upward propagation of waves originating from the high-latitude troposphere is favored and precedes the poleward and downward shift of negative zonal-mean zonal wind anomalies in the midlatitude upper stratosphere. Following the enhanced upward propagation of waves into the stratosphere, the AO forms in the troposphere associated with an increased poleward propagation of tropospheric waves [*Kuroda and Koder*, 1999].

[3] Eurasian-Tibetan snow cover is an important source of seasonal climate predictability. The influences of the Eurasian-Tibetan snow cover on Indian/Asian summer monsoon circulation and precipitation have been extensively recognized and reported in both observational and numerical simulation studies [*Hahn and Shukla*, 1976; *Dickson*, 1984; *Khandekar*, 1991; *Zhang and Tao*, 2001; *Zhang et al.*, 2004]. In recent years it has been demonstrated that Eurasian fall snow cover is significantly correlated with the NH climate variability during following winter [*Cohen and Entekhabi*, 1999]. The possible physical mechanism is explained by the upward propagation of Rossby waves. Eurasian fall snow cover anomalies not only alter near-surface temperatures but also impact the upward propagation of Rossby waves. Note also that changes forced in the stratosphere by anomalous fall snow cover are not identified until later in winter when the troposphere and stratosphere are actively coupled [*Saito et al.*, 2001]. In addition, simulation results of numerical model have proved that the pathway of vertical-propagating stationary wave provides a physical explanation for how regional land surface snow anomalies, for example Siberian snow anomalies, can

---

[1] Chinese Academy of Meteorological Sciences, Beijing, China.
[2] China Meteorological Administration Training Centre, Beijing, China.
[3] Korea Polar Research Institute, Incheon, South Korea.
[4] Yunnan Institute of Meteorology, Yunnan, China.

0148-0227/08/2007JD009567$09.00

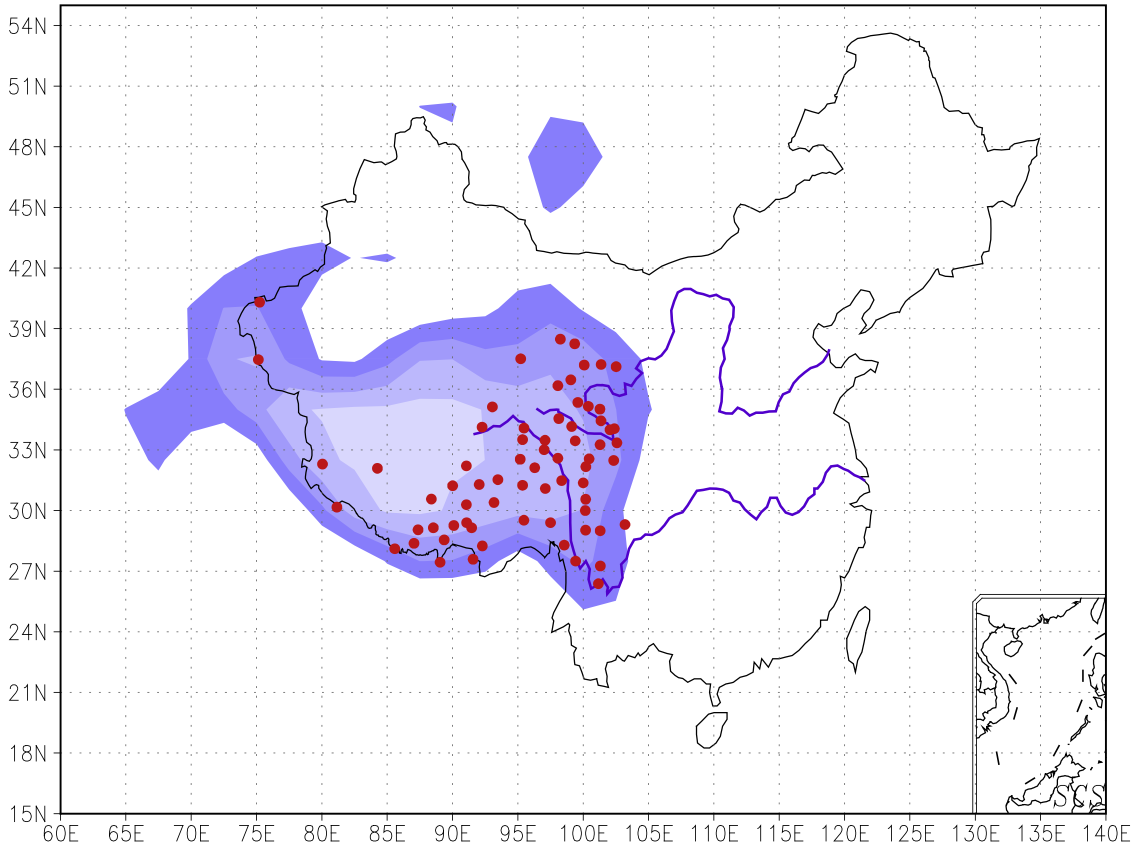

Figure 1. Distributions of 69 observation stations over Tibetan Plateau for snow depth (closed circles). The shading indicates the height of the Tibetan Plateau.

influence winter climate on a hemispheric scale [*Gong et al.*, 2003]. On the other hand, winter season AO and winter/spring season Eurasian snow cover are significantly correlated [*Bamzai*, 2003]. It was proposed that NAO type atmospheric circulation influences the extent of Eurasian snow cover in winter and early spring [*Bojariu and Gimeno*, 2003].

[4] Under global warming, the Eurasian snow cover decreases [*Liu and Yanai*, 2002], whereas the Tibetan Plateau snow cover is observed to increase [*Chen and Wu*, 2000; *Qing et al.*, 2006]. In particular, the spring snow depth of the Tibetan Plateau has exhibited a sharp increase since the late 1970s [*Zhang et al.*, 2004]. It was found that the weakening of the East Asian winter monsoon and the westerlies perturbation on the southern side of the Tibetan Plateau caused the excessive snow over the plateau in spring and winter. *Ju et al.* [2005] proposed that the trend in the AO toward its high-index polarity during the past 2 decades leads to the decadal increase of wintertime precipitation over the Tibetan Plateau, but the associated physical processes have not been clarified yet.

[5] These previous studies imply but do not conclusively prove a causal relationship between snow anomalies over the Tibetan Plateau and the AO. In fact, two important questions that have not been convincingly answered remain: What are the relationships between the anomalous variability of Tibetan Plateau snow cover and the AO throughout the autumn and winter season? Whether may the vertical propagation of Rossby waves provide an explanation for the physical linkage between the AO and the fall/winter snow depth over the Tibetan Plateau? The purpose of this paper is to explore the relation of the AO to autumn and winter snow depth over the Tibetan Plateau, and to propose the possible physical process involved in the linkage.

2. Data and Methods

[6] Monthly snow depth data from January 1954 to December 2003 are calculated from the daily snow depth archived by National Meteorological Information Center of China Meteorological Administration. Sixty-nine station observations with an altitude greater than 3000 m are selected. The geographical locations of the station are presented in Figure 1. It is notable that the numbers of observation stations for the snow depth over the Tibetan Plateau increased quickly, especially in the early years from 12 in 1954 to 40 in 1957. The numbers were also variable along with the years, and the missing rate was less than 5% until 1973. During the period from 1978 to 1988, all of the 69 stations have complete records, but these data are not long enough to discuss the relationship between the AO and the snow depth on interdecadal timescale. Therefore, we decided to use the data from 1954 to 2003.

[7] Air temperature, zonal and meridianal wind, and geopotential height from 1000 to 10 hPa were obtained from the monthly mean National Centers for Environmental Prediction/National Center for Atmospheric Research

Table 1. Correlation Coefficients Between the SDI and SDI6 During 1956–2003

Jan	Feb	March	April	Sept	Oct	Nov	Dec
0.52	0.34	0.29	0.47	0.24	0.60	0.59	0.51

(NCEP/NCAR) reanalysis [*Kalnay et al.*, 1996] for the period 1954–2002.

[8] In order to reveal the relationships between the Tibetan Plateau snow depth and the AO, we first defined a snow depth index (SDI) by averaging the individual 69 station observations. The missing data were omitted. When there was not visible snow at observation station, the snow depth was regarded as zero. The SDI was calculated as the area-weighted mean over the Tibetan Plateau to represent the extent of snow cover. To investigate whether the SDI defined in this paper is reasonable, we selected 6 stations (i.e., Dulan, Yushu, Maduo, Changdu, Lhasa, Rikaze) which have complete records from 1956 to 2003 and defined another snow depth index (SDI6). Correlation coefficients between the SDI and SDI6 during 1956–2003 listed in Table 1. show that there exists significant correlations between the SDI and the SDI6 in 8 months, especially in winter months. This suggests that the SDI safely demonstrates the variability of the snow over the Tibetan Plateau for the entire period in a reasonable degree. AO index was downloaded at web page http://jisao.washington.edu/data/ao/ derived from work by *Thompson and Wallace* [1998].

[9] The data above are analyzed by several methods, such as correlation analysis, regression analysis, and 11-year running mean. The 11-year running mean is a low-passband filter. Since the degree of freedom (DOF) of filtered time series will be decreased, we used the Monte Carlo method to estimate the critical correlation coefficients of the 11-year running mean time series at different confidential levels [*Yan et al.*, 2003].

3. Relation Between the Tibetan Plateau Snow Anomalies and the AO

[10] The snow cover over the Tibetan Plateau starts to increase in fall, reaches its maximum in winter, and decreases gradually in next spring, whereas the AO is a major mode in the NH wintertime circulation. The month-to-month correlation coefficients between the SDI and the AO index from September to following April during the period 1954/1955–2000/2001 were calculated (Table 2). The months for the SDI are shown in the first row, and that for the AO index are shown in the first column. The correlation coefficients indicate the relation between the AO index and SDI in corresponding months. The variability of the AO that leads the snow depth is shown in upper right triangle, while the opposite case is in lower left triangle. According to correlation coefficients that reach the confidence level in Table 2, there exists significant negative correlation between the Tibetan Plateau autumn snow depth and the AO in following winter when the variance of the snow leads the AO. For example, September (October) snow is negatively correlated with December (February) AO. This suggests that increased fall snow depth precedes a negative AO in following winter and decreased fall snow depth precedes a positive AO.

[11] On the other hand, the winter AO is positively correlated with the snow depth in simultaneous months and 1–2 months later when the variability of the AO leads the snow. The simultaneous correlation appears significantly from January to March, especially in February (0.45). It should be noted that the positive relation persists 1–2 months, implying that the AO plays a role on the snow depth simultaneously in January to March, and in the following months from February to April.

[12] Figure 2 shows the variation of the normalized winter/early spring AO index and SDI for the period 1954/1955–2000/2001. It is clear that the two series appear very coherent. Two curves are correlated with a Pearson correlation coefficient of $r = 0.34$, with significance at the $\alpha = 0.05$ level. This suggests that their relation is statistically robust.

[13] To investigate the variability of the AO and the snow depth over the Tibetan Plateau on interdecadal timescale, we performed 11-year running mean on autumn and winter/early spring SDI and AO index, respectively (Figure 3). Both the snow depth over the Tibetan Plateau and the AO underwent the distinct interdecadal regime shift during the late 1970s. In winter/early spring, the snow depth tends to vary consistently with the AO. The AO showed a negative phase and the snow depth decreased from the mid 1950s to the late 1970s, whereas the snow depth has increased since the early 1980s with the AO in its positive phase. The correlation coefficient between 11-year running mean of DJFM SDI and that of DJFM AO was also calculated. In this case, the critical correlation coefficients at the $\alpha = 0.05$ level is 0.71 derived from Monte Carlo simulations. The two 11-year running mean curves are correlated at $r1 = 0.95$, indicating that the snow depth over the Tibetan Plateau has the positive relation with the AO on interdecadal timescale in winter/early spring.

[14] On the contrary, there exist inverse trends in the interdecadal variation of the autumn snow depth, and in that of the winter/early spring snow depth and the AO (Figure 3). The Tibetan Plateau snow depth anomalously increased in preceding autumn, however it decreased quickly in following winter/early spring before the end of 1970s when the AO was in its negative phase. Since the early 1980s, the snow depth has declined in fall, but it has

Table 2. Correlation Coefficients Between AOI and SDI From September to Following April[a]

	Snow							
AO Index	Sept	Oct	Nov	Dec	Jan	Feb	March	April
Sept	0.17[b]	−0.16	0.10	*0.28*	0.15	0.10	0.03	0.16
Oct	0.02	0.10[b]	0.17	0.06	0.08	0.01	*−0.25*	0.05
Nov	0.08	*−0.24*	0.08[b]	0.04	0.15	**0.30**	0.23	**0.35**
Dec	*−0.26*	−0.03	−0.19	−0.02[b]	*0.24*	0.09	0.11	−0.03
Jan	−0.05	0.11	0.02	0.00	**0.42**[b]	**0.33**	*0.26*	0.22
Feb	−0.06	**−0.29**	−0.03	−0.02	0.19	**0.45**[b]	**0.31**	0.03
March	0.04	−0.04	0.16	0.00	0.17	**0.31**	**0.29**[b]	*0.27*
April	0.10	0.03	−0.17	−0.01	−0.22	−0.10	*−0.25*	0.08[b]

[a] The boldface values indicate significance above 95%; the italic values indicate significance above 90%. The variability of AO leads the snow depth in upper right triangle, while it is opposite in lower left triangle.

[b] The simultaneous correlations.

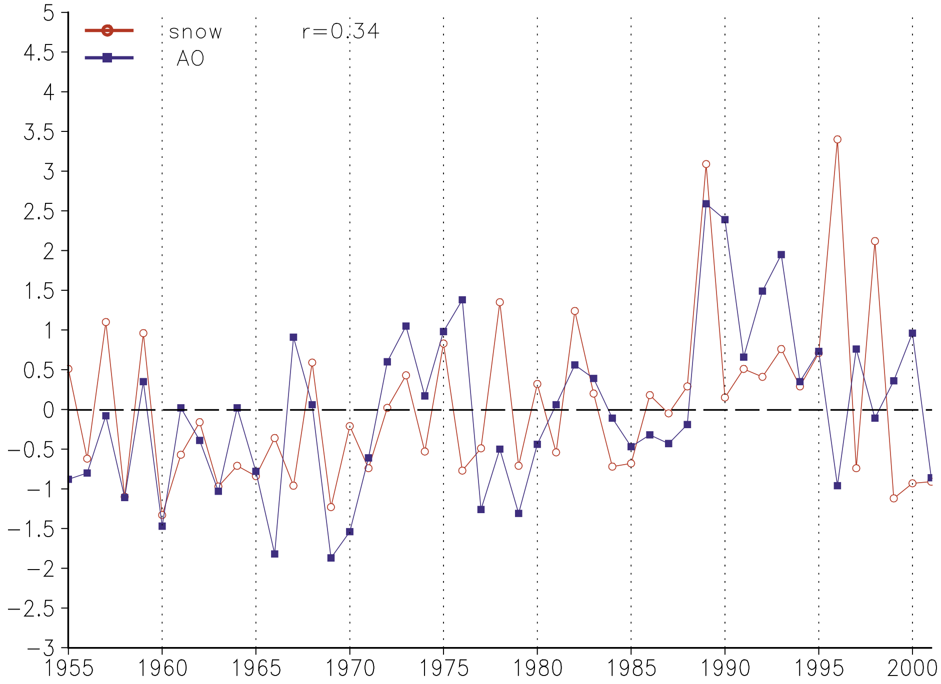

Figure 2. Time series of the normalized AO index and SDI for December–March.

increased quickly in following winter/early spring when the AO entered its positive phase. Two of the curves, the 11-year running mean of autumn snow depth and winter/early spring AO, are significantly correlated with a correlation coefficient of r2 = −0.84, implying that the correlation between autumn snow depth over the Tibetan Plateau and winter/early spring AO is highly negative on interdecadal timescale. The results may potentially lead to improved predictions of the winter AO mode, on the basis of the Tibetan Plateau snow conditions during the preceding autumn.

4. Physical Processes Linking the Tibetan Plateau Snow Cover and the AO

[15] Results presented in the previous section showed that a statistically significant relation exists between the fall/wintertime snow depth anomalies over the Tibetan Plateau and the wintertime AO. Even then, no cause-effect relationship is illustrated by statistical means alone. We provide additional analysis to investigate a possible physical mechanism linking the snow depth and the AO. As mentioned in the introduction, the vertical propagation of stationary Rossby waves provides a physical explanation for the influence of the regional land surface fall snow anomalies on the following winter at a hemispheric scale. A hypothesized teleconnection pathway between the Siberian snow forcing and the hemispheric climate response involves the upward propagation of snow-forced stationary wave activity from the (local) surface to the stratosphere. *Saito et al.* [2001] and *Gong et al.* [2003] described the general features of this pathway. In this section, we further pursue this idea as a possible physical mechanism linking the fall/winter snow depth over the Tibetan Plateau and the winter AO.

[16] The propagation and sources of wave activity may be studied in terms of the wave activity flux (WAF), which is a three-dimensional extension of Eliassen-Palm (E-P) flux. The wave activity vectors were calculated as defined by *Plumb* [1985] to diagnose the vertical propagation of the stationary Rossby waves. According to the correlation coefficients shown in Table 2, the months when the fall snow depth is significantly correlated with winter AO are selected, and then December (February) zonal-mean WAF and zonal wind are regressed onto September (October) SDI for 1954/1955–2001/2002 period (Figure 4). The anomalous increase of the Tibetan Plateau snow depth in fall induces perturbed thermal forcing of lower troposphere, amplifying orographically forced upward stationary waves. *Ringler and Cook* [1999] demonstrated that the presence of low-level cooling above the Tibetan Plateau tends to amplify both the mechanical forcing and the far-field stationary wave response. It is also plausible that latitudinal temperature gradient changes forced by snow cover lead to anomalous potential vorticity gradients in the stratosphere, which then alter the wave refraction index [*Saito et al.*, 2001]. As a result, an upward WAF anomaly responds almost immediately to the snow-forced thermal anomaly and appears in troposphere (not shown). Once this pathway is initiated in late autumn, it continues into the winter, maintained by the poleward refraction of ambient vertical wave activity. While the Rossby waves initiating in midlatitudes (40°N–55°N) propagate upward into the mid to upper troposphere in December and February (Figures 4a and 4b), they were divided into two branches. The two branches propagate poleward fist, afterward one branch descends in high latitudes, and another branch propagates upward into the stratosphere. Along with the upward propagation of stationary wave, the negative zonal-mean zonal wind anomalies appear in the stratosphere and shift downward into the troposphere. That is, the increased upward wave activity in the stratosphere weakens the

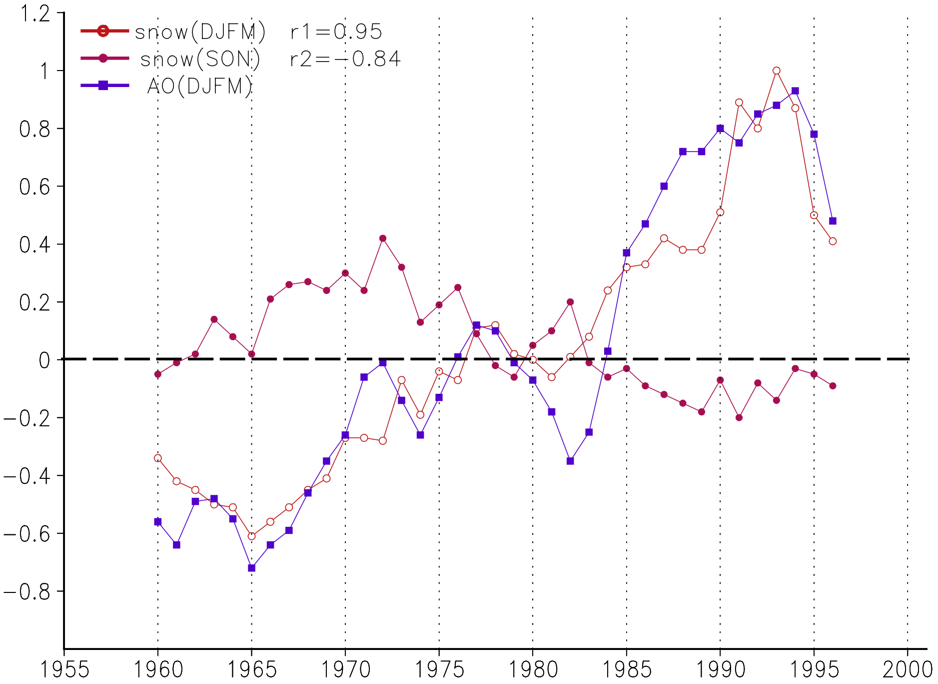

Figure 3. The 11-year running mean of the SDI and the AO index for September–November and December–March, respectively.

stratospheric polar vortex and then the tropospheric AO. The AO is in its negative phase in this case [*Baldwin et al.*, 2003], suggesting that preceding positive snow anomalies over the Tibetan Plateau lead to negative AO response in following winter. It shows that the response of the variability in the stratosphere to the early seasonal (fall) snow anomalies cannot be recognized until winter. These results are consistent with previous studies [*Saito et al.*, 2001; *Gong et al.*, 2003].

[17] In order to improve our understanding of the role played by fall snow over the Tibetan Plateau, Figure 5 shows the horizontal distribution of the vertical WAF component difference between the six heaviest fall snow depth years (1961, 1967, 1971, 1977, 1985, 1997) and the six lightest fall snow depth years (1954, 1964, 1975, 1991, 1994, 2001) at the 100 hPa level. In December anomalous upward fluxes are observed over a wide area of the eastern Tibetan Plateau and the middle China, and a maximum is found over the eastern Tibetan Plateau. Another maximum upward flux anomaly is also found over the Gulf of Alaska. In contrast, anomalous downward fluxes are observed over high-latitude northern Atlantic, eastern Siberia and the Bering Sea (Figure 5a). In midwinter some remnant anomalous upward flux initiated by the anomalous snow is still seen over eastern Tibetan Plateau, but the major flux anomalies over Eurasia are downward (Figure 5b). Our results, presented in Figure 5, are quit different from *Saito et al.* [2001], who obtained that the anomalous upward fluxes originated by Eurasian snow cover were located from the eastern Siberia to the Gulf of Alaska. Analysis in this paper indicates that the anomalous upward flux in early winter may also be derived from the Tibetan Plateau snow anomalies in previous fall.

[18] In winter, when the strong tropospheric-stratospheric coupling occurs [*Kuroda and Koder*, 1999], the process linking the AO and the snow depth over the Tibetan Plateau is completely different. Figure 6 indicates latitude-pressure cross section of winter zonal-mean WAF (arrow) and zonal wind (contour), both regressed onto the simultaneous SDI. It is obvious that the anomalous westerlies prevail from the stratosphere through the troposphere. This pattern is indicative of the strengthening of the stratospheric polar vortex and the positive winter AO mode. Along with the occurrence of the positive zonal-mean zonal wind anomalies, the stationary wave originating in the stratosphere propagates downward. The downward WAF and the upward WAF converge at the top of troposphere. Then the confluent WAF propagates equatorward and descends to the surface in midlatitudes (30°N–50°N). *Chen and Kang* [2006] discovered that the quasi-stationary planetary wave activity may play a role as a bridge in the relationship between the AO and the climate anomalies over East Asia. The AO influences the strength of westerlies in the mid- to high-latitude lower stratosphere first, then the vertical propagation of planetary waves is affected, leading to the variations in the amplitudes of planetary waves in the mid- to high-latitude lower troposphere. So the signal of the positive AO phase appears as the weakening of the Siberian High and the activity of India-Burma trough in the atmospheric circulation of the troposphere [*Wu and Wang*, 2002; *Ju et al.*, 2005]. The formation of anomalous atmospheric circulation in troposphere associated with the positive AO phase causes the abundant snowfall over the Tibetan Plateau in winter. It suggests that the atmospheric teleconnection pathway links Tibetan Plateau snow anomalies to the AO mode via vertical stationary wave propagation and stratospheric-tropospheric coupling in winter.

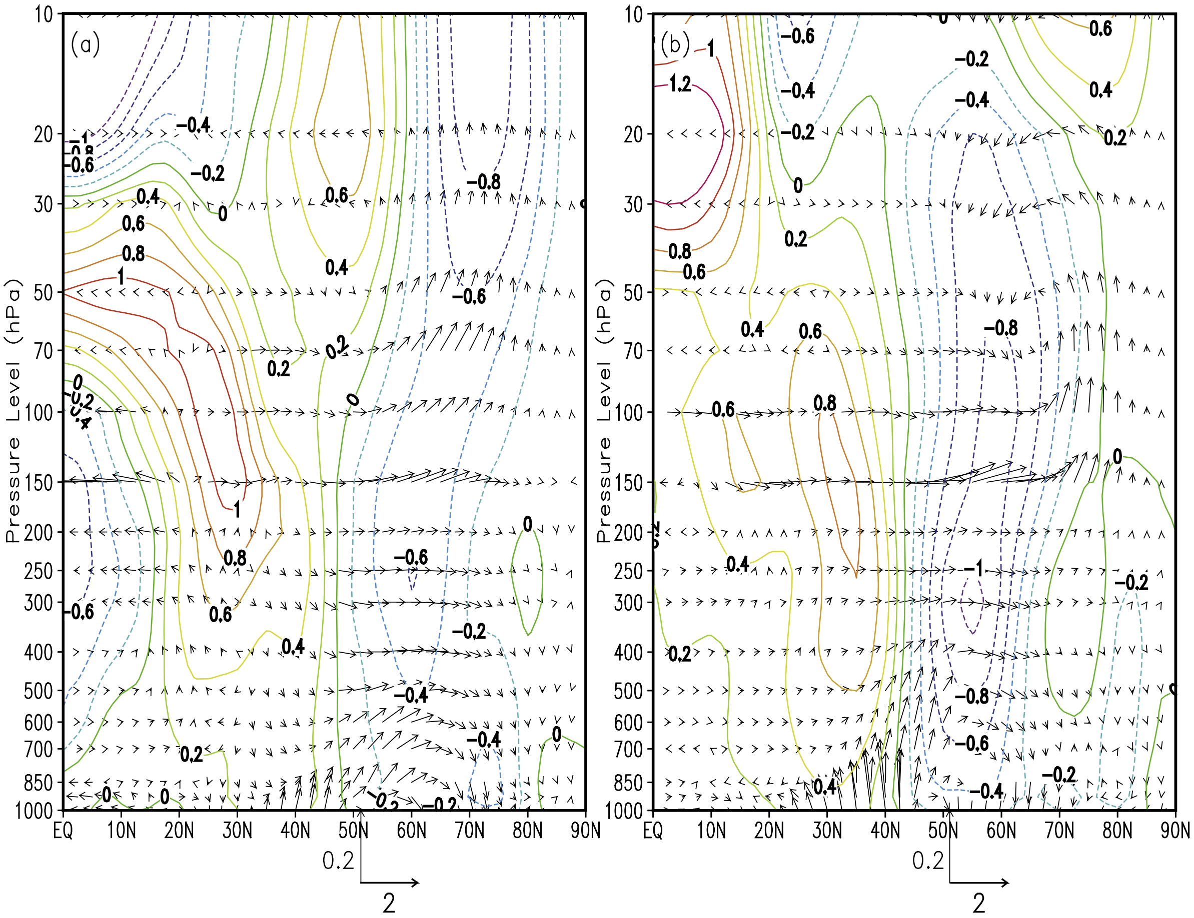

Figure 4. Latitude-pressure cross section of the regression coefficients of winter zonal-mean WAF (arrow) and zonal wind (contour) upon Tibetan Plateau autumn snow depth. The regression coefficients of (a) December WAF and zonal wind upon September snow depth and (b) February WAF and zonal wind upon October snow depth. Contour interval is 0.2 m s^{-1}. Horizontal (vertical) scale of arrows is shown at bottom of panel and represents 2.0 (0.2) m^{2} s^{-2}.

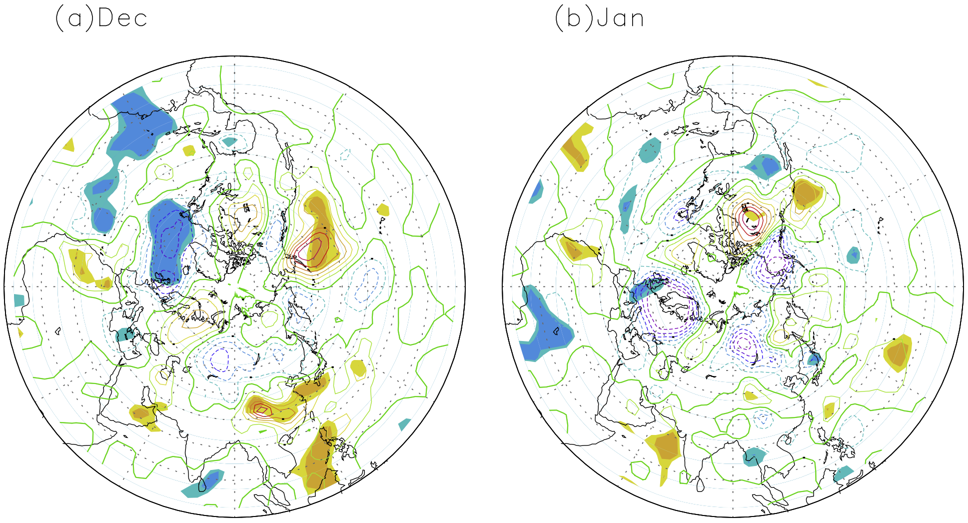

Figure 5. Composite difference of vertical component of WAF at 100-hPa surface between six heaviest and lightest fall snow depth years of (a) December and (b) January. Contour interval is 2.0 × 10^{-2} m^{2} s^{-2}. Zero line is thickened. Shaded areas indicate significant changes at 90% and 95% level, estimated by a local student t-test.

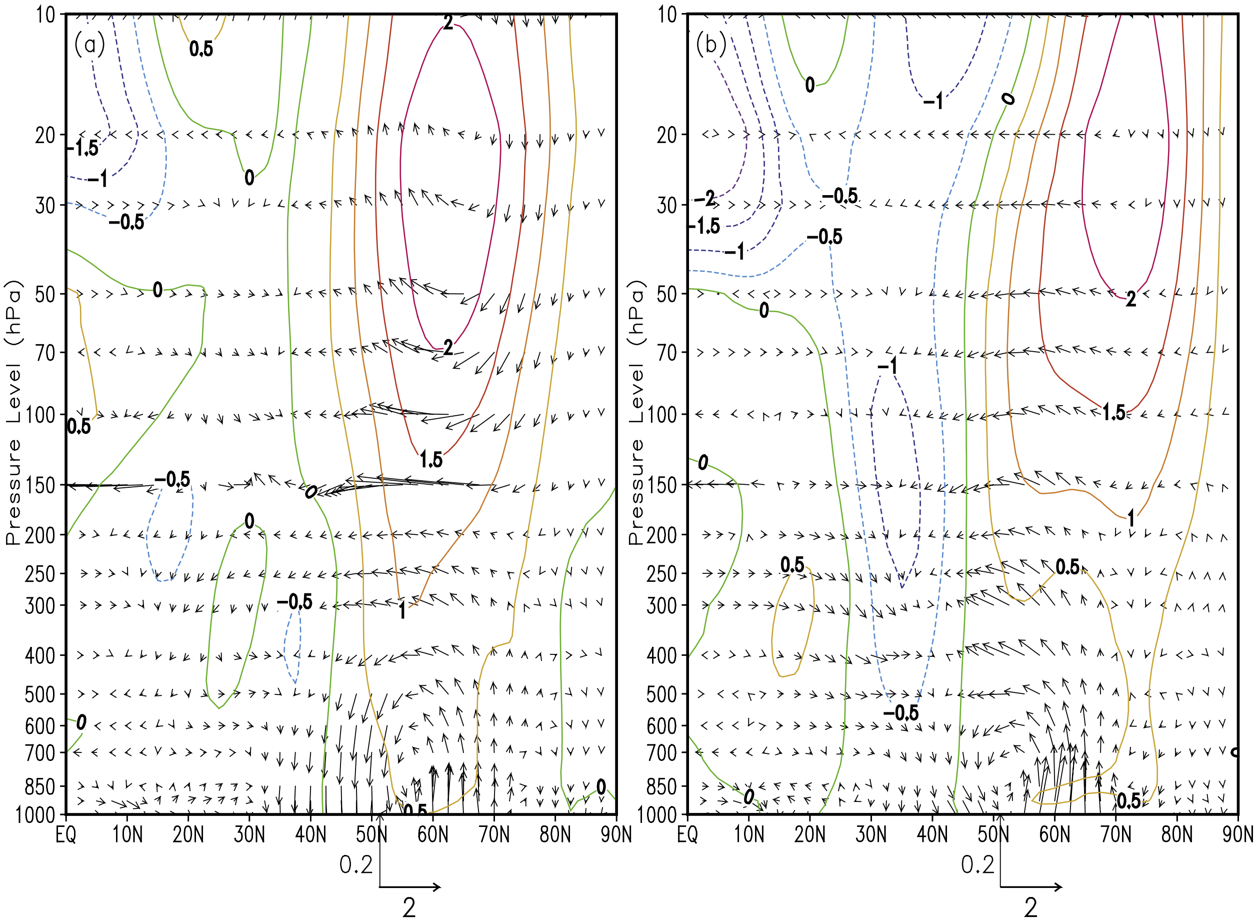

Figure 6. Latitude-pressure cross section of the regression coefficients of winter zonal-mean WAF (arrow) and zonal wind (contour) upon simultaneous Tibetan Plateau snow depth. The regression coefficients of (a) January WAF and zonal wind upon January snow depth and (b) February WAF and zonal wind upon February snow depth. Contour interval is 0.5 m s^{-1}. Horizontal (vertical) scale of arrows is shown at bottom of panel and represents 2.0 (0.2) m^2 s^{-2}.

[19] According to results obtained in this study, one may reasonably conclude that the bidirectional vertical propagation of Rossby waves is the physical process involved in the influence of preceding fall snow anomalies on the winter AO, and at the other direction, the influence of the winter AO on the simultaneous snow anomalies over the Tibetan Plateau.

5. Summary and Concluding Remarks

[20] In this research the connection of the autumn and winter snow depth over the Tibetan Plateau to the AO are investigated using the station observational snow depth. The vertical propagation of Rossby waves is proposed to explain the possible physical process linking the snow depth with the AO. The major conclusions are summarized below.

[21] Fall snow depth over the Tibetan Plateau is negatively correlated with the winter AO. In other words, when the snow increases in fall, the AO enters its negative phase in following winter, and vice versa. This significant lead-lag relationship between the Tibetan Plateau fall snow anomalies and the AO provides a leading indicator for the prediction of the AO.

[22] There exists a significant positive correlation between the winter/early spring AO and simultaneous snow depth over the Tibetan Plateau. That is to say, there is extensive snow over the Tibetan Plateau while the AO is in its positive phase in winter. Moreover, the influence of the AO on the snow depth should persist till spring.

[23] The significant relation between the AO and the Tibetan Plateau snow depth also exists on interdecadal timescale. There is an inverse relation in the interdecadal variability of fall snow depth and that of winter/early spring AO and snow depth. During the period from the mid 1950s to the late 1970s, the snow over the Tibetan Plateau increased in fall, however it decreased in the following winter/early spring as the AO entered its interdecadal negative phase. On the contrary, the Tibetan Plateau snow has underwent decrease in fall and increase in winter/early spring since the early 1980s when the AO was in its interdecadal positive phase.

[24] The vertical propagation of Rossby waves and the interaction between the troposphere and the stratosphere are the possible physical processes linking the early and simultaneous season snow anomalies over the Tibetan Plateau and the AO. Figure 7 presents schematic illustration for the

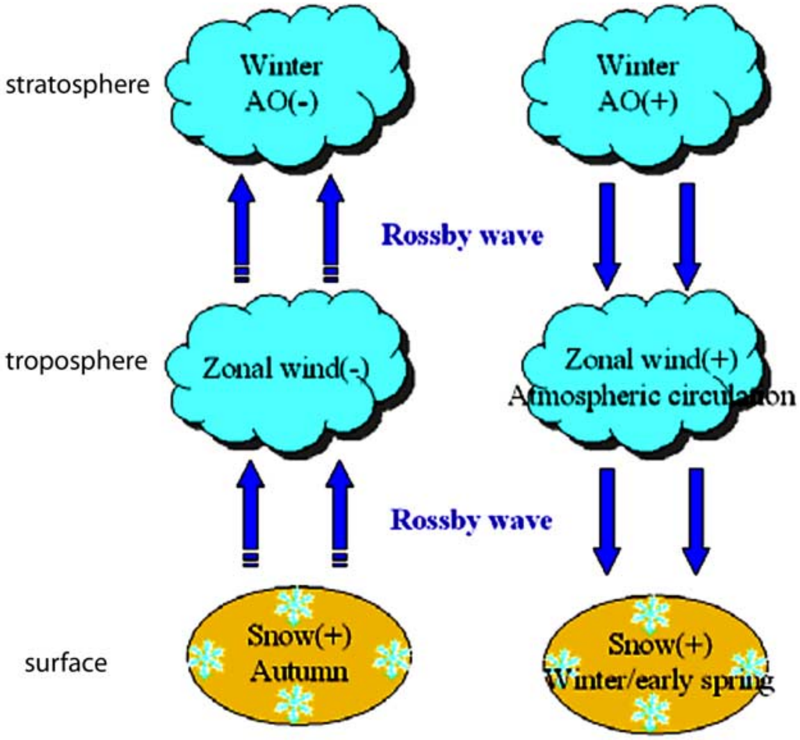

Figure 7. Schematic illustration for the physical process linking the AO and snow depth over the Tibetan Plateau from autumn to winter.

vertical propagation of Rossby waves. It can be found that there are different teleconnection pathways linking Tibetan Plateau snow anomalies to the AO from fall to winter. In preceding fall, the abnormally excessive snow alters the thermal regime of underlying surface. The upward-propagating stationary Rossby waves can be amplified by snow-forced upward WAF anomalies that originate in midlatitude troposphere. The variation in the stratosphere forced by the fall snow anomalies cannot be identified until winter, when the Rossby waves propagate upward into the stratosphere. Meantime, the anomalous easterlies appear in the stratosphere and shift downward into the troposphere, indicating the weakening of stratospheric polar vortex and the negative AO phase.

[25] In winter, on the other hand, with the strong coupling of the troposphere and stratosphere, the anomalous stationary waves originating in high-latitude stratosphere propagate down into the troposphere while the AO is in its positive phase. The anomalous westerlies prevail through the stratosphere and troposphere along with the anomalous downward-propagating Rossby waves. The atmospheric circulation in the troposphere is modulated and plays a role in the increase of Tibetan Plateau snow. It is shown that the anomalous downward propagation of Rossby waves is a reason that causes the increased snow over the Tibetan Plateau when the AO is abnormally positive in winter.

[26] It should be noted that a numerical climate model of adequate temporal and spatial resolution and refined physical/dynamical components can be used to simulate the relative role of autumn snow cover on the interannual variability of NH wintertime circulation, to deepen our understanding of the dynamical mechanism in future work.

[27] **Acknowledgments.** This investigation was jointly supported by The Climate Change Special Project of China Meteorological Administration (grant CCSF2006–14), Project of Integrated Research on the Composition of Polar Atmosphere and Climate Change (COMPAC) of Korea Polar Research Institute (PE08030), and the National Natural Science Foundation of China under grants 40705026 and 40675056. We would like to thank Ke Wei for his assistance with the analysis of the WAF and three anonymous reviewers for helpful suggestions and comments on the original manuscript.

References

Baldwin, M. P., et al. (2003), Stratospheric memory and skill of extended-range weather forecasts, *Science*, *301*, 636–640, doi:10.1126/science.1087143.

Bamzai, A. S. (2003), Relationship between snow cover variability and Arctic Oscillation index on a hierarchy of time scales, *Int. J. Climatol.*, *23*, 131–142, doi:10.1002/joc.854.

Bojariu, R., and L. Gimeno (2003), The role of snow cover fluctuations in multiannual NAO persistence, *Geophys. Res. Lett.*, *30*(4), 1156, doi:10.1029/2002GL015651.

Chen, L. T., and R. G. Wu (2000), Interannual and decadal variations of the snow cover over Qinghai-Xizang Plateau and their relationship to summer monsoon rainfall in China, *Adv. Atmos. Sci.*, *17*(1), 18–30, doi:10.1007/s00376-000-0040-7.

Chen, W., and R. H. Huang (2005), The three-dimensional propagation of quasi-stationary planetary waves in the Northern Hemisphere winter and its interannual variations (in Chinese), *Chin. J. Atmos. Sci.*, *29*(1), 137–146.

Chen, W., and L. H. Kang (2006), Linkage between the Arctic Oscillation and winter climate over East Asia on the interannual timescale: Roles of Quasi-stationary planetary waves (in Chinese), *Chin. J. Atmos. Sci.*, *30*(5), 863–870.

Cohen, J., and D. Entekhabi (1999), Eurasian snow cover variability and Northern Hemisphere climate predictability, *Geophys. Res. Lett.*, *26*, 345–348, doi:10.1029/1998GL900321.

DeWeaver, E., and S. Nigam (2000), Zonal-eddy dynamics of the North Atlantic oscillation, *J. Clim.*, *13*, 3893–3914, doi:10.1175/1520-0442(2000)013<3893:ZEDOTN>2.0.CO;2.

Dickson, R. E. (1984), Eurasian snow cover versus Indian monsoon rainfall-An extension of the Hahn-Shukla results, *J. Clim. Appl. Meteorol.*, *23*, 171–173, doi:10.1175/1520-0450(1984)023<0171:ESCVIM>2.0.CO;2.

Gong, D. Y., and S. W. Wang (2003), Influence of Arctic Oscillation on winter climate over China, *J. Geogr. Sci.*, *13*, 208–216.

Gong, D. Y., S. W. Wang, and J. H. Zhu (2001), East Asian winter monsoon and Arctic Oscillation, *Geophys. Res. Lett.*, *28*, 2073–2076, doi:10.1029/2000GL012311.

Gong, G., D. Entekhabi, and J. Cohen (2003), Modeled Northern Hemisphere winter climate response to realistic Siberian snow anomalies, *J. Clim.*, *16*, 3917–3931, doi:10.1175/1520-0442(2003)016<3917:MNHWCR>2.0.CO;2.

Hahn, D. G., and J. Shukla (1976), An apparent relationship between Eurasian snow cover and Indian monsoon rainfall, *J. Atmos. Sci.*, *33*, 2461–2462, doi:10.1175/1520-0469(1976)033<2461:AARBES>2.0.CO;2.

Hartley, D. E., et al. (1998), A new perspective on the dynamical link between the stratosphere and troposphere, *Nature*, *391*, 471–474, doi:10.1038/35112.

Ju, J. H., J. M. Lü, J. Cao, and J. Z. Ren (2005), Possible impacts of the Arctic Oscillation on the interdecadal variation of summer monsoon rainfall in east Asia, *Adv. Atmos. Sci.*, *22*(1), 39–48.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year Reanalysis Project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2.

Kerr, R. A. (1999), A new force in high-latitude climate, *Science*, *284*, 241–242, doi:10.1126/science.284.5412.241.

Khandekar, M. L. (1991), Eurasian snow cover, Indian monsoon and El Niño/Southern Oscillation—A synthesis, *Atmos. Ocean*, *29*, 636–647.

Kodera, K., and Y. Kuroda (2000), Tropospheric and stratospheric aspects of the Arctic oscillation, *Geophys. Res. Lett.*, *27*, 3349–3352, doi:10.1029/2000GL012017.

Kuroda, Y., and K. Koder (1999), A role of planetary waves in the stratosphere–troposphere coupled variability in the Northern Hemisphere winter, *Geophys. Res. Lett.*, *26*, 2375–2378, doi:10.1029/1999GL900507.

Liu, X. D., and M. Yanai (2002), Influence of Eurasian spring snow cover on Asian summer rainfall, *Int. J. Climatol.*, *22*, 1075–1089, doi:10.1002/joc.784.

Plumb, R. A. (1985), On the three-dimensional propagation of stationary waves, *J. Atmos. Sci.*, *42*, 217–229, doi:10.1175/1520-0469(1985)042<0217:OTTDPO>2.0.CO;2.

Qing, D. H., S. Y. Liu, and P. J. Li (2006), Snow cover distribution, variability, and response to climate change in western China, *J. Clim.*, *19*, 1820–1833, doi:10.1175/JCLI3694.1.

Ringler, T. D., and K. H. Cook (1999), Understanding the seasonality of orographically forced stationary waves: Interaction between mechanical and thermal forcing, *J. Atmos. Sci.*, *56*, 1154–1174, doi:10.1175/1520-0469(1999)056<1154:UTSOOF>2.0.CO;2.

Saito, K., J. Cohen, and D. Entekhabi (2001), Evolution of atmospheric response to early-season Eurasian snow cover anomalies, *Mon. Weather Rev.*, *129*, 2746–2760, doi:10.1175/1520-0493(2001)129<2746:EOARTE>2.0.CO;2.

Thompson, D. W. J., and J. M. Wallace (1998), The Arctic Oscillation signature in the wintertime geopotential height and temperature fields, *Geophys. Res. Lett.*, *25*, 1297–1300, doi:10.1029/98GL00950.

Thompson, D. W. J., and J. M. Wallace (2000), Annular modes in the extratropical circulation, Part I: Month-to-month variability, *J. Clim.*, *13*, 1000–1016, doi:10.1175/1520-0442(2000)013<1000:AMITEC>2.0.CO;2.

Thompson, D. W. J., and J. M. Wallace (2001), Regional climate impacts of the Northern Hemisphere Annular Mode and associated climate trends, *Science*, *293*, 85–89, doi:10.1126/science.1058958.

Wu, B., and J. Wang (2002), Possible impacts of Winter Arctic Oscillation on Siberian High, the East Asian Winter Monsoon and sea-ice extent, *Adv. Atmos. Sci.*, *19*, 297–320, doi:10.1007/s00376-002-0024-x.

Wu, T., and Z. Qian (2003), The relation between the Tibetan winter snow and the Asian summer monsoon and rainfall: An observational investigation, *J. Clim.*, *16*, 2038–2051, doi:10.1175/1520-0442(2003)016<2038:TRBTTW>2.0.CO;2.

Yan, H. M., M. Zhong, and Y. Z. Zhu (2003), The determination of degrees of freedom for digital filtered time series—An application in the correlation analysis between length of day variation and SOI (in Chinese), *Acta Astron. Sin.*, *44*(3), 324–329.

Zhang, S., and S. Tao (2001), A diagnosis and simulation study of the influences of snow cover over the Tibetan Plateau on Asian summer monsoon (in Chinese), *Chin. J. Atmos. Sci.*, *25*(3), 372–390.

Zhang, Y. S., T. Li, and B. Wang (2004), Decadal change of the spring snow depth over the Tibetan Plateau: The associated circulation and influence on the East Asian summer monsoon, *J. Clim.*, *17*, 2780–2793, doi:10.1175/1520-0442(2004)017<2780:DCOTSS>2.0.CO;2.

J.-H. Ju and Y.-X. Zhu, China Meteorological Administration Training Centre, 46 Zhongguancunnandajie Road, Beijing 100081, China. (jujh@cma.gov.cn; yunboyi@163.com)

S.-J. Kim and J.-M. Lü, Korea Polar Research Institute, Get Pearl Tower, Songdo Techno Park, 7-50 Songdo-dong, Yeonsu-gu, Incheon 406-840, South Korea. (seongjkim@kopri.re.kr; wind-ljm@163.com)

J.-Z. Ren, Yunnan Institute of Meteorology, Kunming, Yunnan 650034, China. (leaf-ren@163.com)