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Observed changes in snow depth and number of snow days in the eastern and central Tibetan Plateau

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ABSTRACT: The Tibetan Plateau (TP) has the largest area of snow in the mid-latitude regions, and is strongly affected by the climate change. We examine the temporal variability of winter snow depth and number of days of snow cover at 69 Chinese Meteorological Administration stations above 2000 m a.s.l. in the eastern and central TP during 1961–2005. Snow depth is positively correlated with the number of snow days ($R = 0.89$, $p < 0.0001$). Regional mean winter (DJF) depth and days of snow cover increase at rates of 0.32 mm decade^{-1} and 0.40 d decade^{-1} from 1961 to 1990, but at rates of –1.80 mm decade^{-1} and –1.59 d decade^{-1} (i.e. decrease) between 1991 and 2005. The long term trends are weakly positive, but unrepresentative of shorter time periods. Thus snow depth and cover change depends on the timescale examined and cannot be attributed solely to increased greenhouse gas forcing. The decreasing snow depth in recent years will influence hydrological processes and water resources on the plateau and downstream. Both snow depth and duration have positive correlations with the winter Arctic Oscillation/North Atlantic Oscillation (AO/NAO) index and Niño-3 region (5° N–5° S, 150°–90° W) sea surface temperature (SST). During high AO/NAO index years, both a deeper India–Burma trough and an intensified cyclonic circulation near Lake Baikal bring more snowfall to the TP, consistent with a higher water vapor flux. The opposite is true in low AO/NAO years. Thus secular changes of snow depth and duration in the TP are not independent of changes in the macro-scale atmospheric circulation.

KEY WORDS: Snow depth · Snow days · Tibetan Plateau · Arctic Oscillation · North Atlantic Oscillation

1. INTRODUCTION

Averaging over 4000 m above sea level (a.s.l.) and an area of approximately 2.5×10^6 km^2, the Tibetan Plateau (TP) has the largest area of semi-permanent snow and ice in the mid-latitudes, and is regarded as the Asian water tower (Yeh & Gao 1979, Immerzeel et al. 2010). Changes in both snow cover duration and snow pack (including depth and water content) are of great importance, in particular due to their influence on water resources on the plateau and downstream. Snow and glacial melt are also important hydrological processes in the TP, and changes in temperature and precipitation are expected to seriously affect the melt characteristics (Immerzeel et al. 2010). Water originating from the plateau accounts for nearly half of the total surface water resources in China, supporting nearly 40% of the world's population (Xu et al. 2008). Yanai & Li (1994) revealed the mechanism of heating and the structure of the boundary layer in the TP, and found that the sensible heat flux at the surface is the major source of heating in the TP in summer. The greatest temperature increase in the TP is found in winter (0.40°C decade^{-1}), which is the main component of the increase in annual mean temperature (You et al. 2010a). Snow cover in the TP influences surface mois-

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ture and energy fluxes, and therefore weather over a large area. For example, it has been suggested as a potential predictor of East Asian summer monsoon rainfall (Zhang et al. 2004).

Despite such importance, there are limited surface snow cover and depth data at high elevations on the TP which makes knowledge patchy about cryospheric trends. Surface observations should be supplemented by remotely-sensed data, and are dependent on the period studied; consequently, different data sources have given different results (Kang et al. 2010). Previous studies using surface data from 60 stations suggested that snow cover increased during 1957–1992, which is inconsistent with the reduced extent of snow cover in other extra-tropical areas in the late 1980s, but coincides with recent increases in Antarctic and Greenland snow accumulation (Li 1996, Chen & Wu 2000). Based on Scanning Multichannel Microwave Radiometer (SMMR) NOAA snow-extent charts, mean snow depth and the number of snow-cover days have both increased in western China between 1951 and 1997 (Qin et al. 2006). However, the snow-cover fraction from the Moderate Resolution Imaging Spectroradiometer (MODIS) in the TP during 2000–2006 shows a slight decreasing trend (Pu & Xu 2009).

The Arctic Oscillation/North Atlantic Oscillation (AO/NAO) is one of the dominant patterns of Northern Hemisphere climate variability, and most prevalent in winter and in the mid and high latitudes (Hurrell 1995, Thompson & Wallace 1998). The oscillation has important effects on the weather and climate of the North Atlantic region and surrounding continents, especially Europe. The NAO is a measure of the strength and position of the storm tracks and depressions across the North Atlantic and into Europe, and of the strength of the prevailing westerly winds associated with the storm track (Hurrell 1995). The NAO is usually regarded as the regional manifestation of the global scale AO in the Atlantic sector. Thus the NAO is part of the AO (Wallace 2000). Recent studies have shown that winter season AO and snow cover are significantly correlated (Bamzai 2003), and that the winter AO/NAO index influences Eurasian winter snow-cover extent (Saito & Cohen 2003). The NAO is a factor in the weakening of the Indian monsoon (Chang et al. 2001). Furthermore, the role of winter NAO in March cooling trends over subtropical Eurasia continent has been demonstrated by Yu & Zhou (2004), who also found that positive winter NAO triggers cooling signals — occurring between the surface and the tropopause — in northern Africa. In the TP, there also exists significant correlation between the AO and snow depth at an inter-decadal time scale (Lü et al. 2008). In addition, El Niño/Southern Oscillation (ENSO) interacts with Asian monsoon rainfall and Eurasian snowpack. But there has been little investigation of AO and ENSO influence on long-term snow trends in the TP.

The TP has undergone pronounced warming, and the environmental consequences of warming are already evident (Zhang 2007). Therefore, more attention should be paid to the cryospheric changes, which may result from global warming. In this study, we investigate changes in snow depth and snow cover duration, based on recent datasets from the China Meteorological Administration. As well as examining broad trends over differing periods, we examine the role of the AO/NAO and ENSO in influencing both snow depth and snow cover duration in the TP.

2. DATASET AND METHOD

Monthly mean snow depth and days of snow cover at 69 stations (Fig. 1) are provided by the National Climate Center, part of the China Meteorological Administration. A 'snow day' is characterized by an accumulated snow depth over 1 cm. Since coverage in the western TP is extremely patchy, we confine our analysis to the eastern and central TP and between the years 1961 and 2005. We concentrate on winter (DJF) variation of snow depth and duration because of higher mean values in this season. Stations were selected

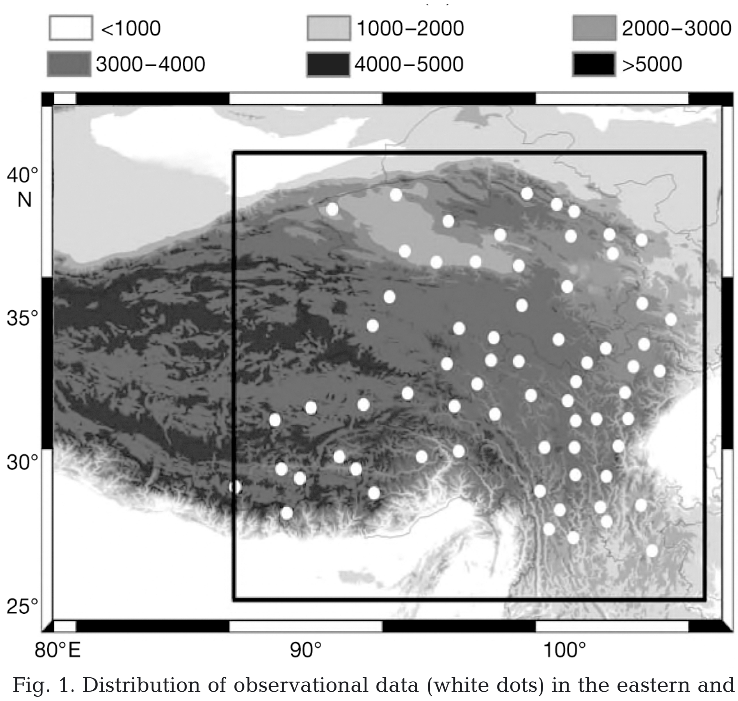

Fig. 1. Distribution of observational data (white dots) in the eastern and central Tibetan Plateau

according to procedures described in our recent papers (You et al. 2008a,b, 2010a,b). The NAO index was available from www.cru.uea.ac.uk/cru/climon/data/nao, updated from works by Hurrell (1995). AO index (AOI) is defined as the difference in the normalized monthly zonal-mean sea level pressure (SLP) between 35° and 65° N (Li & Wang 2003), derived from http://web.lasg.ac.cn/staff/ljp/data-NAM-SAM-NAO/NAM-AO.htm. The Niño-3 region (5° N–5° S, 150°–90° W) was selected as the ENSO index, which is the eastern tropical Pacific sea surface temperature (SST) from NOAA Climate Prediction Center (CPC) (www.cpc.ncep.noaa.gov/data/indices/).

To quantify changes in large scale atmospheric circulation, monthly mean geopotential height and zonal and meridional wind at 500 hPa were obtained from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al. 1996). Water vapor flux was also calculated from the NCEP/NCAR reanalysis data, which is defined as follows:

$$Q_u(\lambda,\phi,t) = \int_{Ps}^{Pt} q(\lambda,\phi,p,t)\,u(\lambda,\phi,p,t)\,\mathrm{d}p$$
$$Q_v(\lambda,\phi,t) = \int_{Ps}^{Pt} q(\lambda,\phi,p,t)\,v(\lambda,\phi,p,t)\,\mathrm{d}p$$

where Q is water vapour flux, t is time period, p is pressure, λ is longitude, ϕ is latitude, q is specific humidity, u is latitudinal wind speed and v is meridional wind speed, respectively. Ps is surface pressure, and Pt is pressure at 300 hPa.

The Mann-Kendall test for trends and Sen's slope estimates are used to detect and estimate trends in annual and seasonal reanalysis wind speeds (Sen 1968), with significance defined as $p < 0.05$.

3. RESULTS: SNOW DEPTH AND DURATION CLIMATOLOGY

The mean snow depth and number of snow days during 1961–2005 in the eastern and central Tibetan Plateau is shown in Fig. 2. Fig. 3 is snow depth and snow day differences between 1983–2005 and 1961–1982 in the study region. For snow depth, the largest mean values occur in the middle of the TP (Fig. 2A), where snow depth has increased in recent decades compared with 1961–1982 (Fig. 3A). For snow days, the south-eastern TP has large mean values, corresponding to larger trend magnitudes (Figs. 2B & 3B). Stations having a higher mean depth and a higher number of days of snow cover are situated on the western and central TP, which is consistent with the higher elevations.

Fig. 4 shows the regional mean winter snow depth and number of snow days at surface stations in the eastern and central TP during 1961–2005, together with a 9 yr smoothing average curve. Both snow depth and duration have similar inter-decadal variability, with increasing (decreasing) trends before (after) the late 1980s. Snow depth has a good positive correlation with the number of snow days during the study period ($R = 0.89$, $p < 0.0001$) (Table 1 and Fig. 5). The peak in the late 1980s means that during 1961–1990, the mean winter snow depth and days of snow cover increase at rates of 0.32 mm decade^{-1} and 0.40 d decade^{-1} respectively. However, between 1991 and 2005 the rates are –1.80 mm decade^{-1} and –1.59 d decade^{-1} (i.e. decreases) respectively. Thus the long term trends (1961–2005), although weakly positive, are somewhat misleading.

Time series plots of the NAO and AO indices also show similar broad trends, with increasing (decreasing) trends before (after) 1990 (Fig. 4). Thus there are positive correlations between the NAO and AO circulation indices ($R = 0.80$, $p < 0.001$) (Fig. 5). Because of the similar trend for NAO and AO, we only used the AO index in this study.

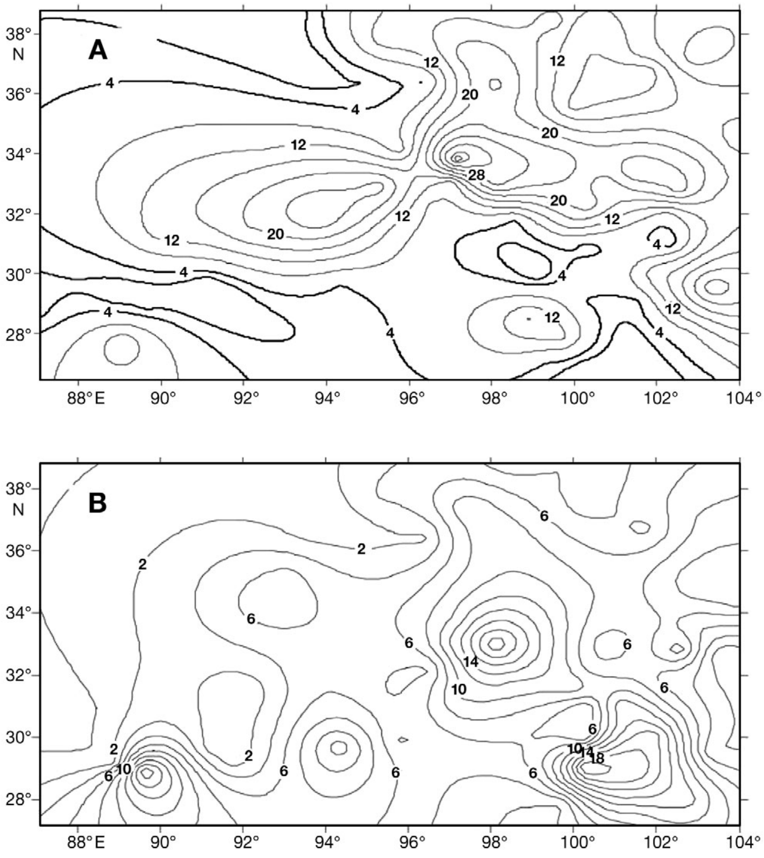

Fig. 2. Mean (A) snow depth (cm) and (B) snow days during 1961–2005 in the eastern and central Tibetan Plateau

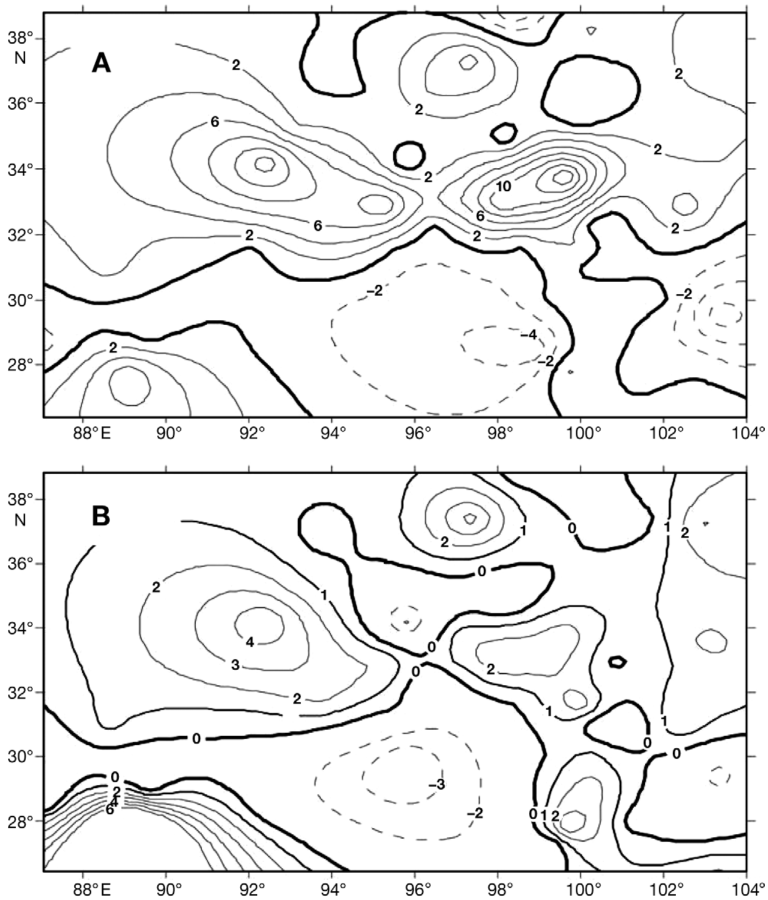

Fig. 3. (A) Snow depth and (B) snow day differences between the periods 1983–2005 and 1961–1982 in the eastern and central Tibetan Plateau

4. RELATIONSHIP BETWEEN SNOW DEPTH AND THE AO/ENSO

4.1. Snow depth–AO relationship

Fig. 6 shows the relationship between the AO and snow depth/snow days in winter in the eastern and central TP during 1961–2005. Taking the TP as a whole, snow depth has a good positive correlation with the AO index during the study period ($R = 0.40$, $p = 0.007$), and a similar positive correlation is also found with snow days. Table 1 provides further evidence that AO anomalies are correlated with snow depth and duration on the regional scale ($R > 0.3$, $p < 0.05$). Fig. A1 in the Appendix shows the spatial correlation coefficients between the AO index and snow depth and snow days. In most regions of the TP, both snow depth and snow days have positive correlation with the AO index, and the larger correlation coefficients occur in the northern TP, although there also exists negative correlation in the southern TP. Fig. A2 shows the differences in snow depth and snow days between strongly positive and strongly negative AO/NAO years during 1961–2005, and it is clear that snow depth and snow days are higher (lower) in the TP when the AO/NAO index is positive(negative).

To investigate the association between AO/NAO and snow depth and duration, Fig. 7 shows the mean differences (positive minus negative years) of (A) geopotential height, (B) wind field (m s^{-1}), and (C) zonal winds at 500hPa in circulation between high and low AO years, based on composite maps for strongly positive (1982, 1988, 1989, 1991, 1994) minus strongly negative (1962, 1964, 1968, 1976, 1978) AO/NAO years. To be defined as strongly positive/negative, index anomalies had to exceed ±1σ. The selected region covers the domain 0° to 70° N and 0° to 160° E. The largest differences in geopotential height are approximately 50 to 60 geopotential metres (gpm). As a result, enhanced anticyclonic circulation is centered over the Alps (45° N and 10° E) and cyclonic circulation over the Urals (focused near 55° N and 60° E). This generates an anomalous north-westerly flow over eastern Europe which carries relatively warm maritime air over the cold Eurasian wintertime land mass. When this enhanced flow reaches the higher terrain of central Asia and western China it bifurcates, with the northern axis flowing north-east towards Russia, and the southern axis entering India. A surface counterflow develops over northern China, which decreases the zonal wind at 500 hPa. This in turn deepens the cyclonic circulation and brings more south-easterly winds with increased water vapor into the northern TP. The southern axis strengthens the India-Burma trough, and an increased south-westerly flow on the rising limb of the trough in southern and eastern China weakens the northwesterly winter monsoon, limiting its southward extension. Thus incursions of colder air into southern China tend to decrease when the westerlies are strong (Thompson & Wallace 2001). Overall, when the AO/NAO index is in its high phase, the atmospheric circulation pattern is beneficial to transportation of water vapor to the TP, and this will increase snow cover and duration. Differences in mean water vapor flux between strongly positive and negative AO/NAO years (Fig. 8) support this idea.

At present, there are different possible mechanisms for the variation of AO/NAO, which could be classified into 3 types: atmospheric processes, ocean influences, and sea ice and land snow cover, although there is no consensus on the mechanisms (Hurrell et al. 2001, Hurrell et al. 2006, Hurrell & Deser 2010). The external forcing of the strength of the atmospheric circulation in the lower stratosphere on long time scales by reductions in stratospheric ozone and increases in greenhouse gas concentration will probably influence the recent trend in winter AO/NAO (Hurrell et al. 2006). Most climate models simulate a strengthening of the westerly circulation in response to increasing greenhouse gas concentrations (Osborn 2004), which results in an underlying shift towards higher values of the NAO index. The magnitude of this simulated change, however, is much smaller than the rate of change observed between the 1960s and the 1990s, so that other external forcing must also be considered. Stratospheric ozone depletion and natural changes in volcanic activity and solar irradiance might also have contributed to increases in the NAO index over this period, though the evidence from modeling the response to these factors remains equivocal (Osborn 2004, 2006). The downward trend in the winter AO/NAO index since 1990 cannot be attributed solely to the monotonically-increasing greenhouse gas forc-

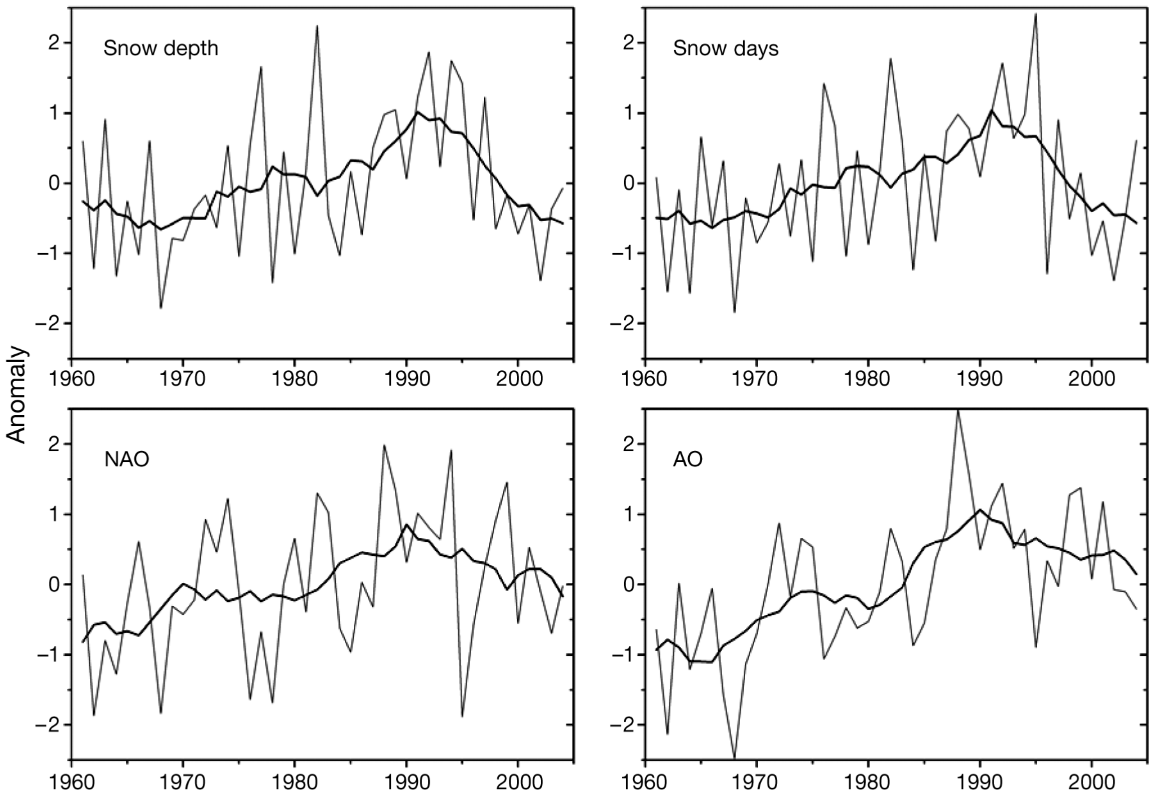

Fig. 4. Anomalies of snow depth, snow days and AO/NAO index in winter in the eastern and central TP during the 1961–2005 period. The smoother curve is the 9 yr smoothing average

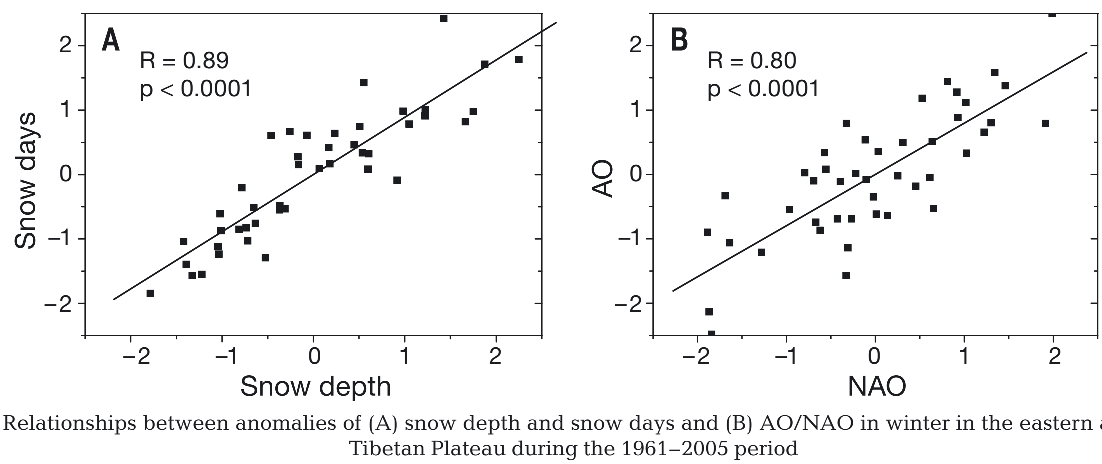

Fig. 5. Relationships between anomalies of (A) snow depth and snow days and (B) AO/NAO in winter in the eastern and central Tibetan Plateau during the 1961–2005 period

Table 1. Correlation coefficients between the AO/NAO index, snow depth or snow days during 1961–2005 in the eastern and central TP

	Snow depth	Snow days	NAO index	AO index
Snow depth	1			
Snow days	0.89	1		
NAO index	0.39	0.34	1	
AO index	0.40	0.36	0.80	1

ing (Visbec et al. 2000). Interactions between oceans and atmosphere, by the changes in surface sea temperature, are also important for understanding the variation of the AO/NAO (Hurrel et al. 2006). Some studies have suggested that snow cover producing large changes in sensible and latent heat flux partially modulates the winter AO/NAO (Saito & Cohen 2003).

4.2. Snow depth–ENSO relationship

Fig. 9 shows the change of winter SST in the Niño-3 region during 1961–2005, when SST varied greatly. We select strongly positive and strongly negative Niño-3 region SST years during 1961–2005, which are those with index anomalies exceeding ±1σ. Composites of strongly positive (1965, 1972, 1982, 1991, 1997) and strongly negative (1967, 1970, 1973, 1975, 1984, 1988, 1999) Niño-3 region SST years are analyzed. Fig. A3 shows the spatial correlation coefficient between snow depth and snow days in the study region and Niño-3 region SST in winter during 1961–2005. Both snow depth and snow days have positive correlations with the Niño-3 region SST, meaning that higher (lower) Niño-3 region SST will result in more (less) snow depth and snow days in the TP.

The differences of snow depth and snow days in the TP between winter with strongly positive and strongly negative SST in the Niño-3 region during 1961–2005 are shown in Fig. A4. Groisman et al. (1994) used a similar approach to check the relationship between snow cover and ENSO in the Northern Hemisphere. It is clear that strongly positive and strongly negative winter Niño-3 region SST years correspond to a substantially higher (lower) snow depth and number of snow days. This is confirmed by the correlation coefficients between snow depth/cover and Niño-3 region SST (Fig. A3). Yin et al. (2000) showed that the concurrent ENSO events had a certain influence on temperature in the TP. Snow cover tended to be higher (lower) than normal, and temperature tended to be lower (higher) than normal during the El Niño (La Niña) events. Note that the relationship between snow cover and temperature is more complex in the context of global warming (Fig. A5).

To check the relationship between ENSO and snow depth and duration, we show the mean difference (positive minus negative years) of geopotential height, wind field (m s^{-1}) and zonal winds at 500hPa between high and low Niño-3 region SST years (Fig. 10), based on composite maps for strongly positive minus strongly negative winter Niño-3 region SST years. The differences in geopotential height show negative values in the extreme northeast of China, and positive values in most other parts. This generates an anomalous northwesterly flow in northern China, preventing cold winter air masses from reaching the TP, and also produces warmer and wetter flows from the ocean to the TP.

There is no consensus on the relationship between ENSO and snow cover in the TP. Groisman et al. (1994) found that El Niño events are generally accompanied by increased spring snow cover over northern hemisphere extratropical land areas during the first half of the hydrological year, while they are associated with a

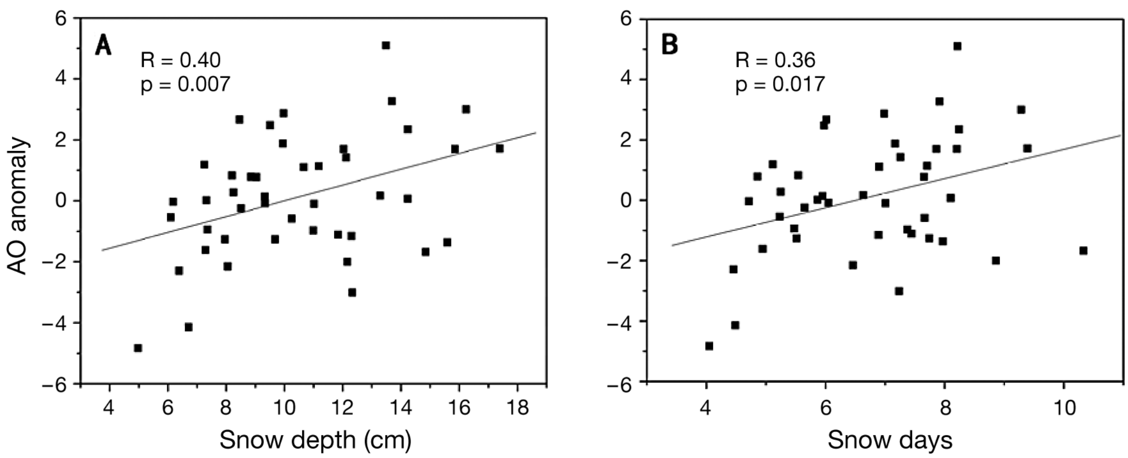

Fig. 6. Relationship to AO anomaly for (A) snow depth and (B) snow days in winter in the eastern and central Tibetan Plateau during the 1961–2005 period

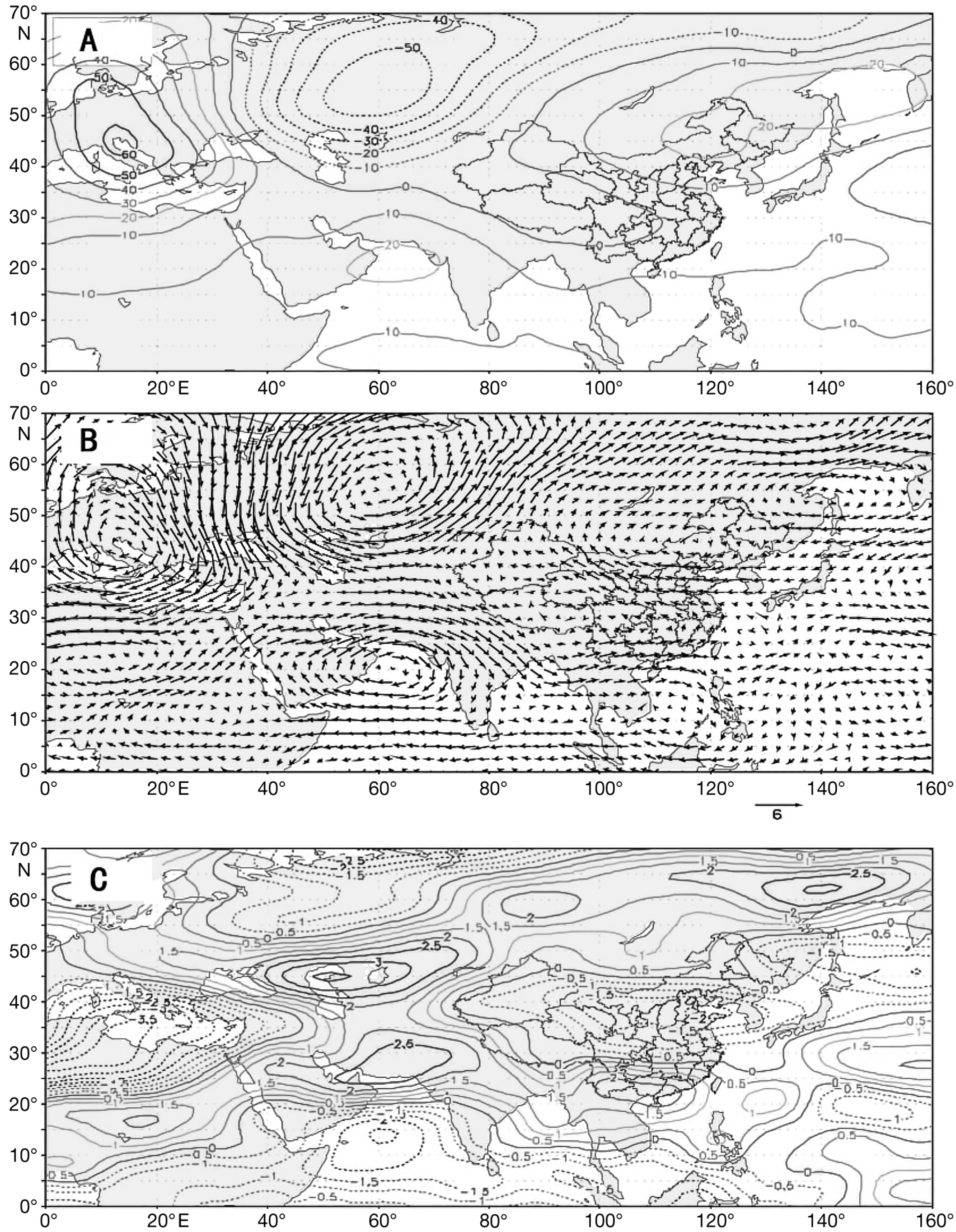

Fig. 7. (A) Mean geopotential height, (B) wind field (m s^{-1}) and (C) zonal wind (m s^{-1}) differences between strongly positive and strongly negative AO/NAO years during 1961–2005. Strongly positive and strongly negative AO/NAO years are the same as for Fig. A2

global retreat of snow cover extent in the second half of the hydrological year (spring and summer). Shaman & Tziperman (2005) show TP snow and ENSO to be related by Rossby wave activity, as winter SST anomalies associated with ENSO produce stationary Rossby waves, which modify winter storm activity and thus TP snowfall. Yuan et al. (2009) found that winter TP snow cover is not related to the influences of ENSO, but has a significant positive correlation with the Indian Ocean Dipole (IOD). In the positive IOD years, negative geopotential height in the north of India is developed, which supports more warm and humid southwesterlies from Bay of Bengal into TP where it enhances the snow cover.

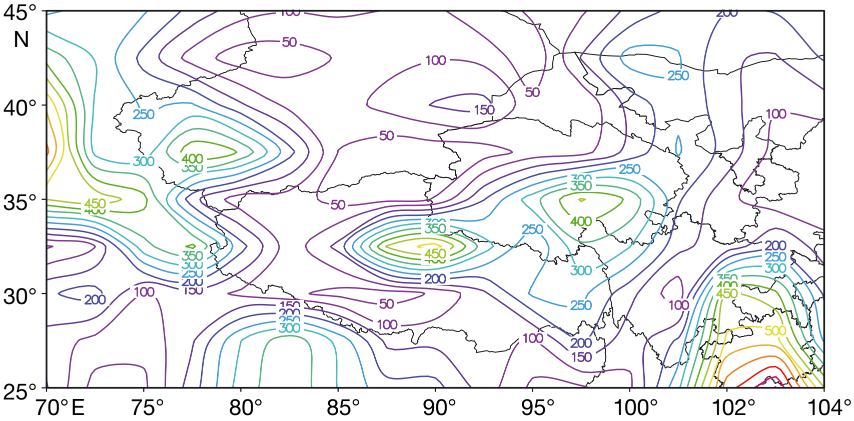

Fig. 8. Differences between mean water vapor fluxes in strongly positive and strongly negative AO/NAO years during 1961–2005

5. DISCUSSION AND CONCLUSIONS

We have examined observed winter snow depth and duration and their trends based on 69 surface stations with elevations above 2000 m a.s.l. in the eastern and central TP during 1961–2005. On a regional scale, both winter snow depth and the number of snow days show general increasing (decreasing) trends before (after) the late 1980s. The increasing trends until the late 1980s were reported by previous researchers (Li 1996, Chen & Wu 2000, Qin et al. 2006), and appear to be in contradiction with the pronounced warming observed in the region (Zhang 2007, You et al. 2010a). However snow cover and duration are not solely dependent on temperature. Precipitation in the snow season is equally important. Previous studies (Li 1996, Chen & Wu 2000, Qin et al. 2006) show that >50 % of the total snow cover variance can be explained by the linear variations of snowfall and snow season temperature. Since 1980 there has been rapid warming, at a rate of 0.38° decade^{-1} in the TP (You et al. 2008a). Consistent with this warming trend, the winter snow depth and duration have decreased since the 1980s. This decreasing trend is likely to continue in the future because of the projected warming for the TP (IPCC 2007) which may affect availability of water resources in the TP and its adjacent regions.

The AO/NAO is the main mode of climate variability in the northern hemisphere in winter. The AO/NAO index has significant positive correlations with winter snow depth and snow duration in the TP, which can be explained by the vertical Rossby wave propagation (Lü et al. 2008). When the troposphere and stratosphere are actively coupled, the downward propagation of Rossby waves (associated with the positive AO phase) modulates the tropospheric winter circulation causing abnormal snow depth growth in the TP (Lü et al. 2008). Large atmospheric circulations have influenced the variation of snow cover in the TP. For example, Zhang et al. (2004) showed that after the mid-1970s, increased snow depth in the TP was concurrent with a deeper India-Burma trough and an intensified subtropical westerly jet with enhanced ascending motion and more moisture supply to the TP, which intensified the southerly moisture flow from the Bay of Bengal with additional humidity originating from over the Indian Ocean (Zhang et al. 2004).

The winter SST in the Niño-3 region has positive correlation with snow depth and snow days in the eastern and central TP, that is, higher (lower) Niño-3 region SST leads to higher (lower) snow depth and snow days. But there is no consensus on the relationships. It is well documented that there was a climate shift in the northern hemisphere in the mid-1970s (Zhang et al. 2004, Duan &Wu 2008, Lü et al. 2008, You et al. 2010c). The variation in snow depth and snow days may be part of this shift and, if so, the shift would have affected the TP snow cover, thus affecting temperature, precipitation, water vapor flux and so on. If interactions are 2-way, there are feedback processes which would encourage persistence or breakdown. That is, more attention should be directed to the interaction between snow depth and snow cover and AO/NAO as well as ENSO. It was not possible to investigate the respective influences of AO/NAO and ENSO on winter snow depth and snow cover in the TP using observational data. To investigate such feedback processes on hemispheric scale, numerical climate model experiments may be a more appropriate method to uncover the dynamics of this than a purely observational one.

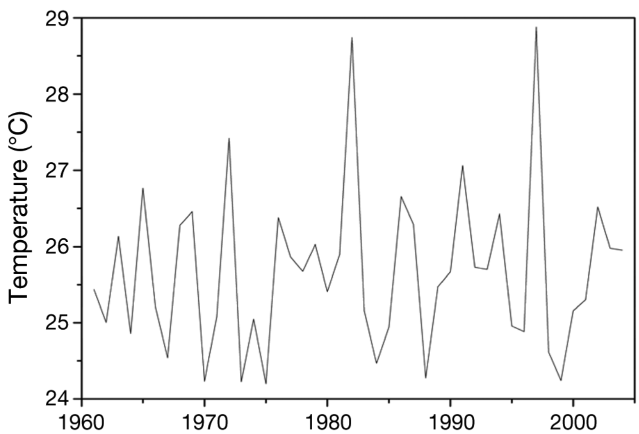

Fig. 9. Average winter SST in Niño-3 region, 1961–2005

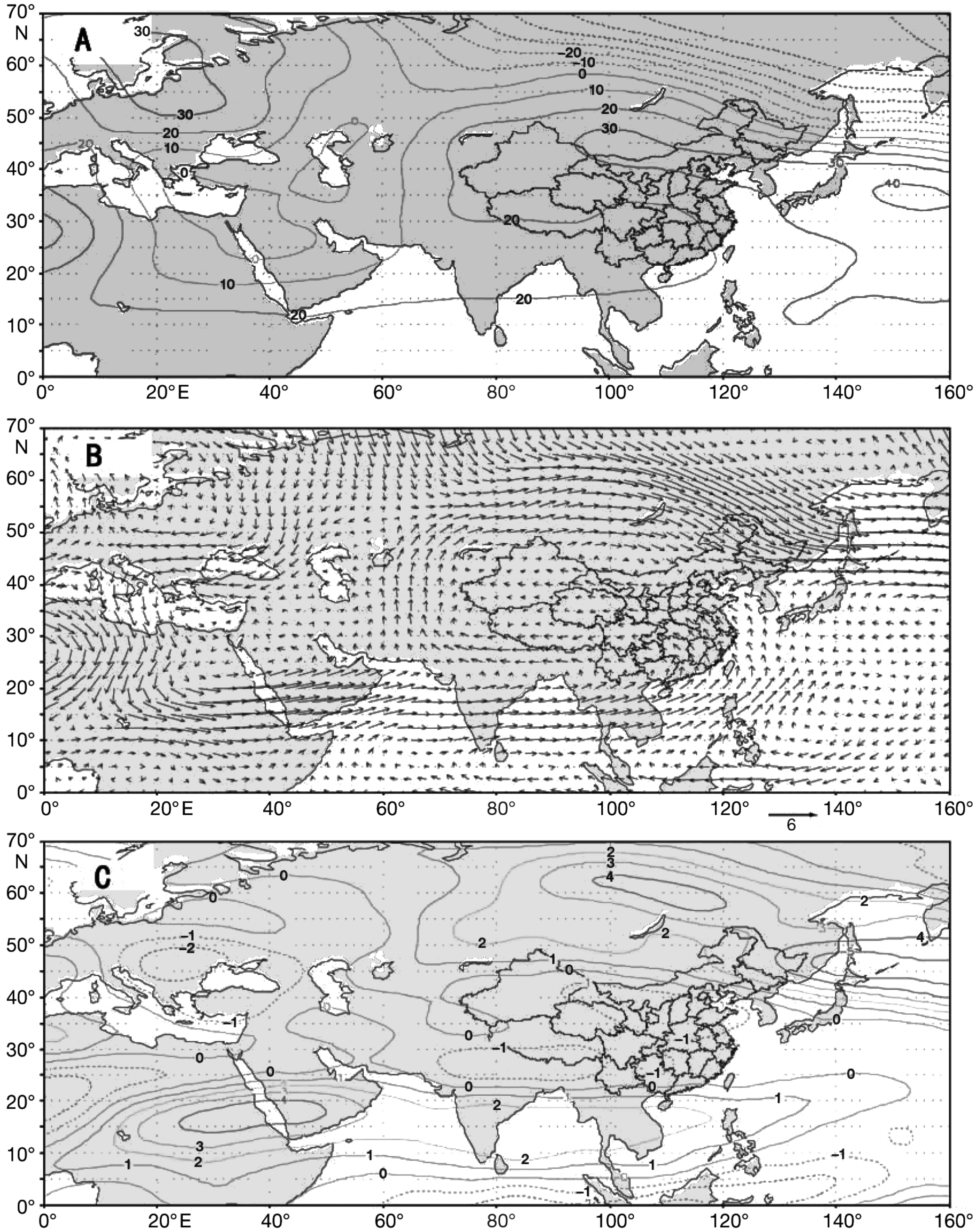

Fig. 10. (A) mean geopotential height, (B) wind field (m s^{-1}), and (C) 500 hPa zonal wind (m s^{-1}) differences between winters with strongly positive and strongly negative SST in the Niño-3 region during 1961–2005. Winters with strongly positive and strongly negative SST in the Niño-3 region years are the same as for Fig. A4

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LITERATURE CITED

Bamzai AS (2003) Relationship between snow cover variability and Arctic oscillation index on a hierarchy of time scales. Int J Climatol 23:131–142

Chang CP, Harr P, Ju JH (2001) Possible roles of Atlantic circulations on the weakening Indian monsoon rainfall–ENSO relationship. J Clim 14:2376–2380

Chen LT, Wu RG (2000) Interannual and decadal variations of snow cover over Qinghai-Xizang Plateau and their relationships to summer monsoon rainfall in China. Adv Atmos Sci

17:18–30

Duan AM, Wu GX (2008) Weakening trend in the atmospheric heat source over the Tibetan plateau during recent decades. I. Observations. J Clim 21:3149–3164

Groisman PY, Karl TR, Knight RW, Stenchikov GL (1994) Changes in snow cover, temperature and radiative heat balance over the northern hemisphere. J Clim 7:1633–1656

Hurrell JW (1995) Decadal trends in the North Atlantic Oscillation—regional temperatures and precipitation. Science 269:676–679

Hurrell JW, Kushnir Y, Visbeck M (2001) Climate—the north Atlantic oscillation. Science 291:603–605

Hurrell JW, Visbeck M, Busalacchi A, Clarke RA and others (2006) Atlantic climate variability and predictability: A CLIVAR perspective. J Clim 19:5100–5121

Hurrell JW, Deser C (2010) North Atlantic climate variability: the role of the North Atlantic Oscillation. J Mar Syst 78: 28–41

Immerzeel WW, van Beek LPH, Bierkens MFP (2010) Climate change will affect the Asian water towers. Science 328: 1382–1385

IPCC (2007), Summary for policymakers. Climate change 2007: the physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge

Kalnay E, Kanamitsu M, Kistler R, Collins W and others (1996) The NCEP/NCAR 40-year reanalysis project. Bull Am Meteorol Soc 77:437–471

Kang S, Xu YW, You QL, Fluegel WA, Pepin N, Yao TD (2010) Review of climate and cryospheric change in the Tibetan Plateau. Environ Res Lett 5:015101 doi: 10.1029/2002GL 016341

Li PJ (1996) Response of Tibetan snow cover to global warming. Acta Geogr Sin 51:260–265 (in Chinese)

Li J, Wang J (2003) A modified zonal index and its physical sense. Geophys Res Lett 30:1632. doi:10.1029/2003GL017441

Lü JM, Ju JH, Kim SJ, Ren JZ, Zhu YX (2008) Arctic Oscillation and the autumn/winter snow depth over the Tibetan Plateau. J Geophys Res 113:D14117 doi:1029/2007JD009567

Osborn TJ (2004) Simulating the winter North Atlantic Oscillation: the roles of internal variability and greenhouse gas forcing. Clim Dyn 22:605–623

Osborn TJ (2006) Recent variations in the winter North Atlantic Oscillation. Weather 61:353–355

Pu ZX, Xu L (2009) MODIS/Terra observed snow cover over the Tibet Plateau: distribution, variation and possible connection with the East Asian Summer Monsoon (EASM). Theor Appl Climatol 97:265–278

Qin DH, Liu SY, Li PJ (2006) Snow cover distribution, variability, and response to climate change in western China. J Clim 19:1820–1833

Saito K, Cohen J (2003) The potential role of snow cover in forcing interannual variability of the major Northern Hemisphere mode. Geophys Res Lett 30:1302 doi:10.1029/2002 GL016341

Sen PK (1968) Estimates of regression coefficient based on Kendall's tau. J Am Stat Assoc 63:1379–1389

Shaman J, Tziperman E (2005) The effect of ENSO on Tibetan plateau snow depth: a stationary wave teleconnection mechanism and implications for the south Asian monsoons. J Clim 18:2067–2079

Thompson DWJ, Wallace JM (1998) The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. Geophys Res Lett 25:1297–1300

Thompson DWJ, Wallace JM (2001) Regional climate impacts of the Northern Hemisphere annular mode. Science 293:85–89

Visbeck MH, Hurrell JW, Lorenzo P, Cullen HM (2001) The North Atlantic Oscillation: past, present, and future. Proc Nat Acad Sci 98:12876–12877

Wallace JM (2000) North Atlantic Oscillation/annular mode: two paradigms—one phenomenon. QJR Meteorol Soc 126: 791–805

Xu XD, Lu CG, Shi XH, Gao ST (2008) World water tower: an atmospheric perspective. Geophys Res Lett 35:L20815 doi: 10.1029/2008GL035867

Yanai M, Li CF (1994) Mechanism of heating and the boundary layer over the Tibetan Plateau. Mon Weather Rev 122: 305–323

Yeh TC, Gao YX (1979) Meteorology of the Qinghai-Xizang (Tibet) Plateau (in Chinese). Science Press, Beijing

Yin ZY, Lin ZY, Zhao XY (2000) Temperature anomalies in central and eastern Tibetan Plateau in relation to general circulation patterns during 1951–1993. Int J Climatol 20: 1431–1449

You QL, Kang SC, Aguilar E, Yan YP (2008a), Changes in daily climate extremes in the eastern and central Tibetan Plateau during 1961–2005. J Geophys Res 113:D07101, doi:10.1029/2007JD009389

You QL, Kang SC, Pepin N, Yan YP (2008b) Relationship between trends in temperature extremes and elevation in the eastern and central Tibetan Plateau, 1961–2005. Geophys Res Lett 35 L04704 doi:10.1029/2007GL032669

You QL, Kang SC, Pepin N, Fluegel WA, Sanchez-Lorenzo A, Yan YP, Zhang YJ (2010a) Climate warming and associated changes in atmospheric circulation in the eastern and central Tibetan Plateau from a homogenized dataset. Global Planet Change 72:11–24

You QL, Kang SC, Pepin N, Fluegel WA, Yan YP, Behrawan H, Huang J (2010b) Relationship between temperature trend magnitude, elevation and mean temperature in the Tibetan Plateau from homogenized surface stations and reanalysis data. Global Planet Change 71:124–133

You QL, Kang SC, Fluegel WA, Pepin N, Yan YP, Huang J (2010c) Decreasing wind speed and weakening latitudinal surface pressure gradients in the Tibetan Plateau. Clim Res 42:57–64

Yu RC, Zhou TJ (2004) Impacts of winter-NAO on March cooling trends over subtropical Eurasia continent in the recent half century. Geophys Res Lett 31:L12204 doi:10.1029/2004 GL019814

Yuan CX, Tozuka T, Miyasaka T, Yamagata T (2009) Respective influences of IOD and ENSO on the Tibetan snow cover in early winter. Clim Dyn 33:509–520

Zhang YS, Li T, Wang B (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

Zhang TJ (2007) Perspectives on environmental study of response to climatic and land cover/land use change over the Qinghai-Tibetan plateau: an introduction. Arct Antarct Alp Res 39:631–634

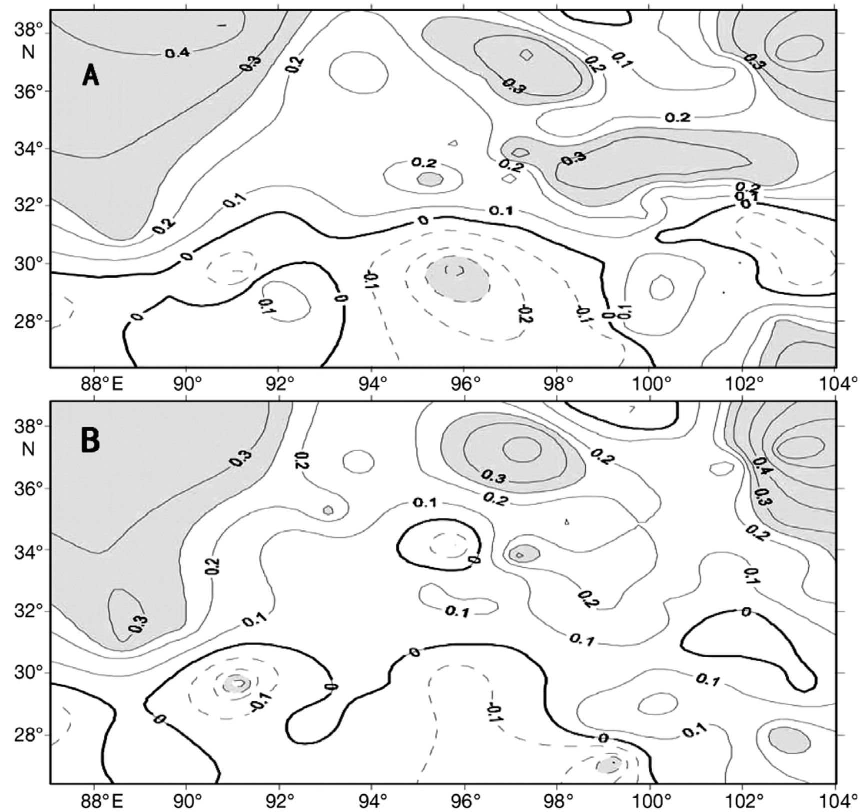

Fig. A1. Spatial correlation coefficient between the AO index and (A) snow depth and (B) snow days in winter in the eastern and central Tibetan Plateau during the 1961–2005 period. Gray shading: significant ($p < 0.05$)

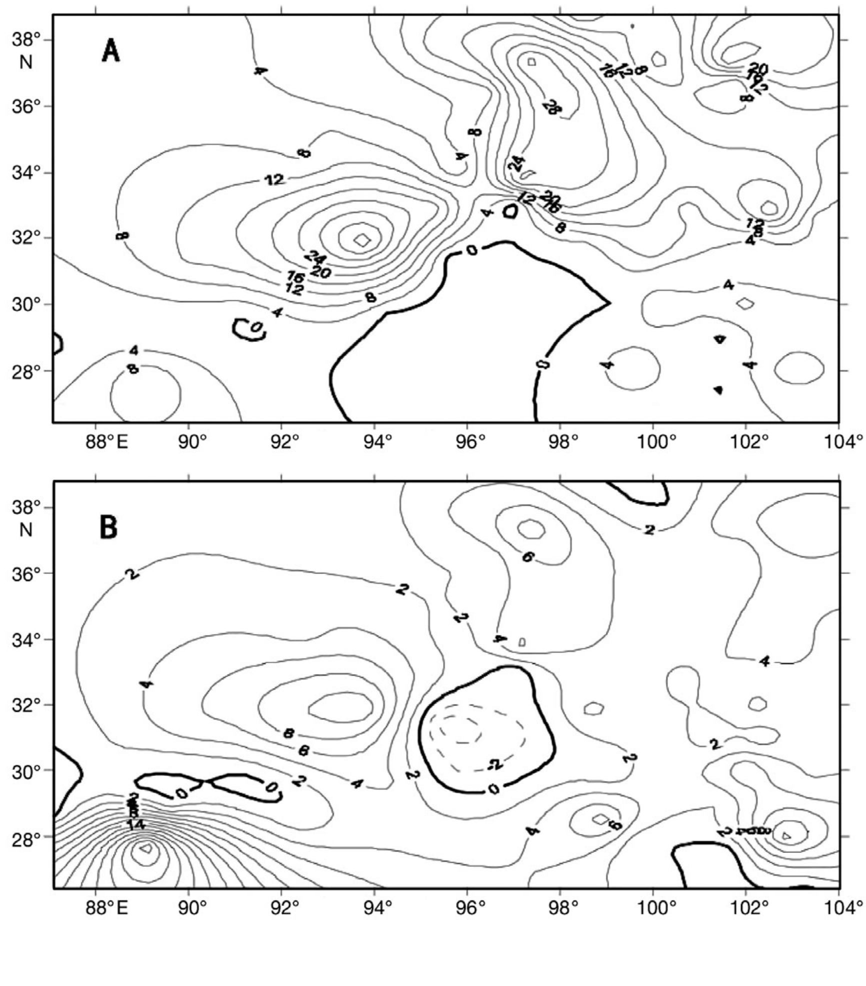

Fig. A2. (A) Snow depth and (B) snow day differences between years with strongly positive and strongly negative AO/NAO during 1961–2005. Strongly positive (1982, 1988, 1989, 1991, 1994) and strongly negative (1962, 1964, 1968, 1976, 1978) AO/NAO years are those with index anomalies exceeding ± 1σ, which correspond to substantially higher/lower snow depths

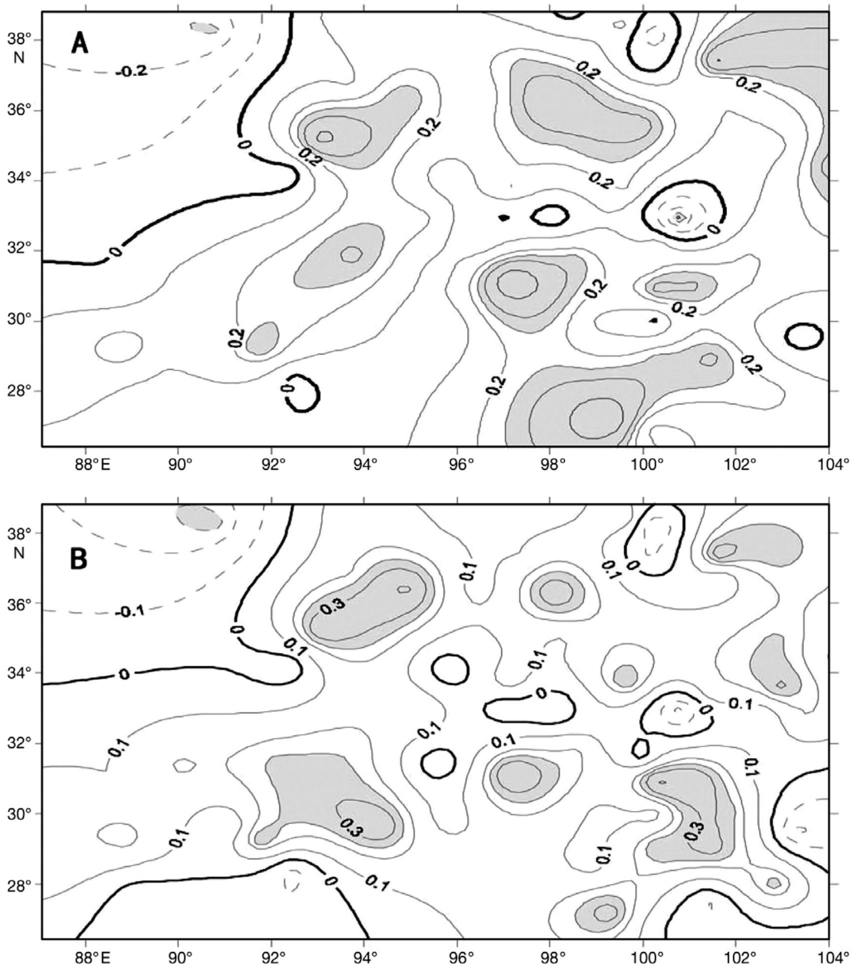

Fig. A3. Spatial correlation coefficient between Niño-3 region SST and (A) snow depth and (B) snow days in the study region in winter during the 1961–2005 period. Gray shading: significant ($p < 0.05$)

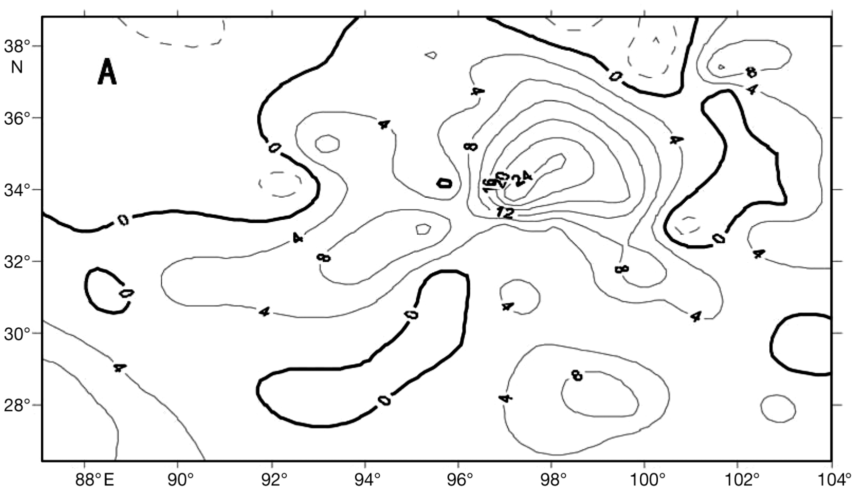

Fig. A4. (A) Snow depth and (B) snow day differences between winters with strongly positive and strongly negative SST in the Niño-3 region during 1961–2005. Winters with strongly positive (1965, 1972, 1982, 1991, 1997) and strongly negative (1967, 1970, 1973, 1975, 1984, 1988, 1999) SST in the Niño-3 region are those with index anomalies exceeding ± 1σ, which correspond to substantially higher/lower snow depth and snow days

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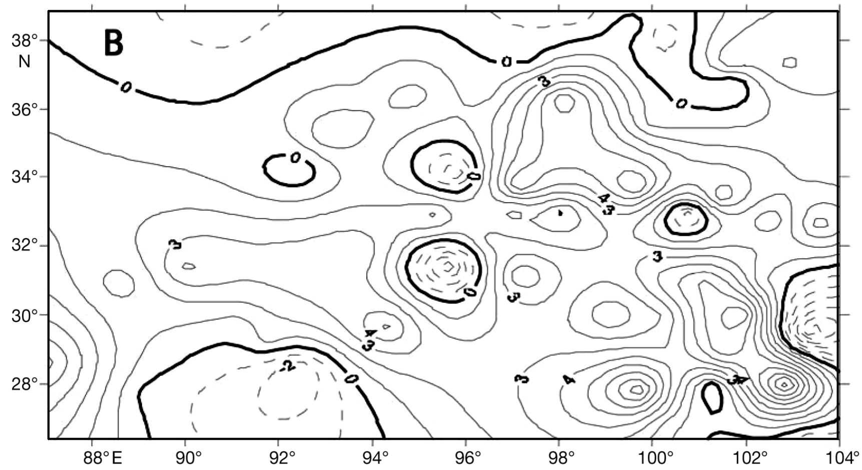

Fig. 4 (continued)

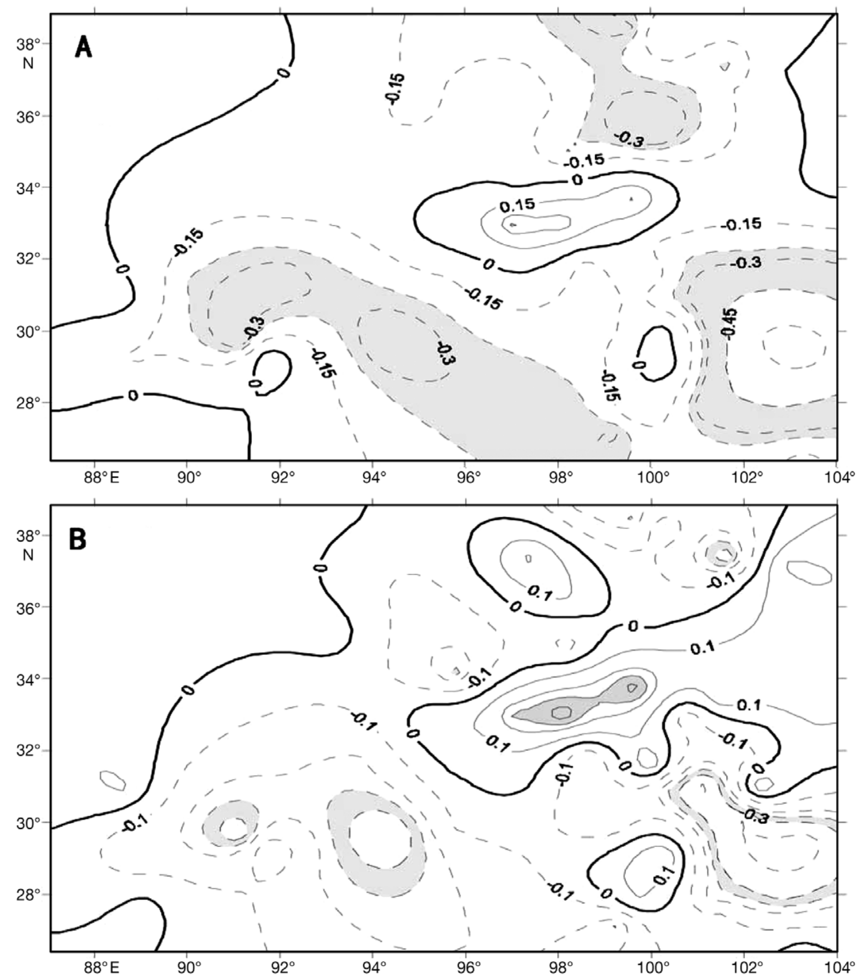

Fig. A5. Spatial correlation coefficient between mean winter temperature and (A) snow depth and (B) snow days in the study region during 1961–2005. Gray shading: significant ($p < 0.05$)

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