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Spatiotemporal distribution of snow in eastern Tibet and the response to climate change

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ABSTRACT

The spatial and temporal distribution of snow and its response to changes in climate were investigated from 1979 to 2005 in eastern Tibet. The Lhasa River basin. Nivang River basin and Changdu region cover an area of approximately 15×10^4 km² and ranges in elevation from 2000 to more than 7000 m. This large area necessitates innovative procedures for estimating potential spatial and temporal changes in snow cover. For this analysis we used the microwave long-term snow cover dataset of China with a spatial resolution of 25 km and temporal resolution of 1 day. After data validation between the microwave dataset and MODIS snow product, we defined two parameters for each pixel: (1) median date of the snow-free period (T_m) ; and (2) duration of the snow-free period (ΔT). After removing transient-snow dominated areas, we find that the duration of the snow-free period was inversely correlated with elevation (R = -0.651, p < 0.001). TFPW-MK (Trend-free pre-whitening Mann-Kendall) was then used to examine and highlight the trend of ΔT with time. At lower-elevation sites, the length of the snow-free season increased. In contrast, at higher-elevations, it decreased. Mann-Kendall tests on monthly air temperature and annual precipitation for the period 1979 to 2005 from seven climate stations in the region operated by the China Meteorological Data Sharing Service System show a significant increase in annual precipitation and an increase in monthly air temperatures for the fall through spring months. Thus, the length of the snow covered season appears to be decreasing at lower elevation because of the increase in air temperatures. However, at higher elevations the increase in precipitation appears to compensate for the increase in air temperature such that the snow-free period has decreased.

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1. Introduction

The spatial and temporal distribution of snow is responsive to changes in climate. Changes in air temperature and precipitation timing and seasonality have been shown to affect the melt characteristics of snow and ice (Immerzeel et al., 2010). Therefore, the spatiotemporal distribution of snow and shifts in the timing of snowmelt are expected to be an indicator of climate change.

The extent of annual snow cover has declined by 10% over the Northern Hemisphere since 1970s to 1994 due to the increases in annual temperature over North America, Europe, West Asia and East Asia (Groisman et al., 1994). A shift in the timing of snowmelt runoff towards earlier dates has been observed in many rivers of Western North America since the late 1940s (Stewart et al., 2004). Contrary to the point that 'less snow accumulation in winter and earlier snowmelt in spring due to global warming' (Barnett et al., 2005), growth of the

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Greenland ice sheet $(65^{\circ}-72^{\circ}N, 40^{\circ}-60^{\circ}E)$ was reported during a 548-day period from 1985 to 1986 (Zwally, 1989). Snow depth was found to increase over most of northern Russia $(60^{\circ}-70^{\circ}N, 50^{\circ}-70^{\circ}E)$ and $60^{\circ}-70^{\circ}N, 110^{\circ}-130^{\circ}E)$, while snow depth decreased in southern Russia $(50^{\circ}-55^{\circ}N, 55^{\circ}-65^{\circ}E)$ during the period 1936–1983 (Ye et al., 1998). Northern Canada (zone $55^{\circ}-70^{\circ}N$) has experienced a significant increase in annual snowfall from the 1950s to the 1990s, while a decreasing trend was observed over southern Canada, for the same time period (Karl et al., 1993).

A warmer climate has been shown to result in: (1) snow melting more quickly; and (2) a greater supply of moisture and more precipitation (Ye et al., 1998). If precipitation occurs at air temperatures below 0 °C, the increased precipitation occurs in the solid form and is retain locally as snow, increasing the storage of water as snow and ice. As an example, Zwally (1989) shows that increasing ice thickness of the Greenland Ice Sheet suggests that the precipitation is higher than the long-term average; higher precipitation may be a characteristic of warmer climates in polar regions. In high-latitude regions, from north to south, snow accumulates as long as air temperature remains below freezing (Ye & Mather, 1997). The same affect may occur in mid-latitude mountain regions.

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Mountains are the water tower of the world (Immerzeel et al., 2010; Viviroli et al., 2007). They take important roles in the water supply of downstream areas. More than one sixth of the global population relies on water supplied by mountains (Barnett et al., 2005; Immerzeel et al., 2009). Meanwhile, mountain areas yield a large spatial variation in climate zones due to large differences in elevation over small horizontal distances (Beniston et al., 1997; Immerzeel et al., 2009). Therefore, "where" (what elevation) snow falls largely determines the snow-free/snow-covered regime in mountain areas.

Runoff from the Tibetan plateau and adjacent mountain ranges are the source of many major Asian rivers including the Indus, Ganges, Brahmaputra, Irrawaddy, Salween, Mekong, Yellow, and Yangtze rivers (Immerzeel et al., 2009). Glacier and snow melt are important hydrological processes in these areas (Bookhagen & Burbank., 2010; Immerzeel et al., 2009). Understanding the spatial and temporal distribution of snow is particularly important because during the latter half of 20th century, climate observations in high-altitude regions have shown more sensitivity to climate change (Beniston, 2003; Diaz & Bradley, 1997; Rangwala et al., 2009b). Moreover, within these mountain regions there is often an elevational dependence, with greater warming rates at higher altitudes (Diaz & Bradley, 1997). Using an extensive selection of weather stations (178) within the Tibetan Plateau and its vicinity, Liu and Chen (2000) reported a differential increase in the rate of surface warming, dependent primarily on the elevation of the observing station during the 1960-1990 period. Previous modeling studies have suggested the snow-albedo feedback to be one of the primary factors in causing an elevation dependent warming in high mountain regions (Chen et al., 2003; Giorgi et al., 1997). Rangwala et al. (2009b) reported that the Tibetan Plateau has warmed most rapidly during the winter months in the latter half of the 20th century, with higher rates of warming at higher elevations. Rangwala et al. (2009a) examined trends in climate variables and their interrelationships over the Tibetan Plateau using global climate model simulations and found elevation-dependent increases in air temperature over the plateau between 1950 and 2100, accompanied by an elevation-based decrease in snow cover. A time series of changes in snow cover with elevation are needed to verify these modeling results.

In contrast to the results of Rangwala et al. (2009a), Qin et al. (2006) revealed that annual snow depth may be increasing in western China. A confounding factor in understanding the response of snow cover to changes in climate in the region of the Tibetan Plateau is that long-term variability of western Chinese snow cover is characterized by a large interannual variation superimposed on a small increasing trend (Qin et al., 2006). For all of these researches, knowledge on snow cover dynamics is of great importance, however, regional studies on this topic are lacking (Immerzeel et al., 2009). The focus of this study is to quantify spatial-temporal patterns of snow cover with elevation in the southeastern portion of the Tibetan Plateau, an area for which there has been little research on this topic in the past.

2. Study area

Qin et al. (2006) have shown that only about 59% of the Tibetan Plateau is snow-covered in the winter, with the southeastern portion having the deepest and most persistent snow. Thus, our area of interest is the headwater region of the Brahmaputra River in southeastern Tibet where the Yarlung Zangbo River originates. The spatial domain of the study area ranges from 90°3′E, 28°24′N to 99°4′E, 32°34′N and covers an area of 1.53×10^5 km² (Fig. 1). The elevation of the study area ranges from 2073 m to 7093 m. The study area is bisected by the low-elevation Yarlung Zangbo Grand Canyon, where there is little accumulation of seasonal snow (Qin et al., 2006).

We divided our study area into three regions. On the western side, the Lhasa River basin covers an area of 3.23×10^4 km², ranging in elevation from 3535 m to 7093 m. The neighboring Niyang River basin drains an area of 1.81×10^4 km², with elevation ranging from 2905 m to

6588 m. Annual precipitation is 635 mm (Zheng et al., 1979). On the eastern side of the Yarlung Zangbo Grand Canyon is the Changdu Region of the Hengduan Mountains, with an area of 10.30×10^4 km² and elevation ranging from 2073 m to 6644 m, with annual precipitation ranging from 500 mm to 1000 mm (Zheng et al., 1979).

Long-term changes in temperature and precipitation for the study area were evaluated from climate records provided by the China Meteorological Data Sharing Service System (http://cdc.cma.gov.cn/) (Fig. 1). We selected climate stations based on two criteria: (1) continuous and complete long-term records of 27 years, from 1979 to 2005; and (2) representation of an elevation gradient from about 3000 m to more than 4000 m (Table 1). We evaluated the statistical significance of temperature and precipitation trends and determined the linear slopes for trends using the Mann–Kendall non-parametric test as shown in Racoviteanu et al. (2008) for the mountainous Cordillera Blance region of Peru.

3. Methods

3.1. Datasets

The baseline dataset for this study was the snow cover area product generated from passive microwave sensors, provided by the Environmental & Ecological Science Data Center for West China, National Natural Science Foundation of China (http://westdc.westgis.ac.cn). The long-term snow dataset of China provides information from 1978 to 2005 and has a spatial resolution of 25 km (Che, 2006). Thus, the passive microwave results allow us to evaluate long-term trends in snow cover area (SCA) for our study area. The study area is covered by 246 pixels, with 80 from the Lhasa River and Niyang River basins, and 166 pixels for the Changdu Region (Fig. 1). Passive microwave sensors provide snow observations that are not affected by cloudiness or darkness and thus provide daily values of SCA, at the expense of a large pixel size (Foster et al., 2005).

Snow-covered area from 1978 to 1987 was from the SMMR instrument on the NIMBUS-7 platform, when SSM/I was launched (Clifford, 2010). The following algorithms were used to calculate average snow depth (SD) for each pixel, based on the results of Chang et al. (1987) and Che (2006)

 $SD = 0.78 (T_{b,18H} - T_{b,37H})$ for SMMR from 1978to1987

$$SD = 0.66 (T_{b,19H} - T_{b,37H})$$
 for SSM /I from 1987to2005

where $T_{b,n}$ is brightness temperature at *n* frequency. However, snow cover on the Tibetan Plateau is often shallow and patchy, and is frequently of short duration (Qin et al., 2006; Robinson & Dewey, 1990). Thus, we report snow-covered area rather than snow depth from the microwave products. After validation with ground stations, the standard deviation of snow depth retrieval from SMMR was 6.2 cm and the standard deviation for SSM/I retrieval was 6.0 cm. Pixels with snow depths greater than one standard deviation were considered to be snow-covered.

We compared the microwave snow product with the MODIS/Aqua SCA product, which is available starting in 2002. We use the 8-day MYD10A2 MODIS data product at a spatial resolution of 500 m, which has been shown by Li and Williams (2008) to work well for Central Asia. Compared with the large spatial extent $(23^{\circ}-41^{\circ}N, 68^{\circ}-106^{\circ}E)$ and coarse spatial resolution (5-km) applied by Immerzeel et al. (2009), our study area equals 1.53×10^{5} km² and that means if higher resolution version at 500-m would be used the entire area is covered by about 6.12×10^{5} pixels, which seems acceptable and also allows a more accurate analysis. For the period from 2002185 to 2005361, snow cover derived from MODIS was based on a band rationing of MODIS band 4 (green) (0.545–0.565 µm) and band 6 (near-infrared)



Fig. 1. The study area is located in southeastern Tibet. Snow-covered area is estimated for three high-elevation areas near Yarlung-Zangbo river basin: Lhasa River basin, Niyang River basin, and Changdu region. The black boxes are 25-km pixels for which snow covered area was estimated from the passive microwave sensors SMMR and then SSM/I. Meteorological stations are shown as red dots.

(1.628–1.652 µm). These bands are used to calculate the Normalized Difference Snow Index (NDSI), defined as (Hall et al., 2002)

NDSI = (band4-band6)/(band4+band6)

Compared snow-cover maps of North America derived from Moderate Resolution Imaging Spectroradiometer (MODIS) with maps derived from SSM/I for the period 23 October–25 December 2000. They reported that MODIS maps of SCA were superior to those from SSM/I. Therefore, we compared SCA between MODIS and the microwave sensors for the total study area over the time period where the two sensors overlapped, from 2002 to 2005. For each microwave grid, there were 25 5-km MODIS pixels and about 2500 500-m sub-pixels (Fig. 2). The MODIS SCA product thus provides an estimate of sub-pixel variability in the microwave SCA product for the four years of overlap.

The remote sensing imagery was overlaid on a Digital Elevation Model (DEM) based on the Shuttle Radar Topography Mission (SRTM) Version 2 with a resolution of about 90 m, released by USGS (https:// wist.echo.nasa.gov/). The elevation of seasonal snowlines was determined using the MODIS maps of SCA overlaid on the SRTM DEM, following the protocols in Immerzeel et al. (2009). Variations in SCA with elevation over the time span of 27 years were quantified using the microwave SCA product at a spatial resolution of 25 km. The mean and variance of elevation and aspect in each microwave pixel were calculated from all the 90×90 m SRTM pixels in each microwave pixel to represent the topographic characteristics of each 25×25 km microwave pixel.

3.2. Changes in SCA with season and elevation

We evaluated the changes in SCA over time as a function of elevation using the concept of a fluctuating snow-covered elevation with time (FSCE). For a given elevation *H*, there exists two timing points that we

Table 1

Location and description of meteorological stations. All stations have a continuous record of monthly air temperatures and precipitation from 1979 to 2005.

Station	Elevation (m)	Air	Annual	ΔT	p value	ΔP	p value
		temp (°C)	precipitation (mm)	°C y ⁻¹		$mm y^{-1}$	
Jiali	4488	-0.5	732	0.056	0.0003	5.80	0.0097
Dangxiong	4200	1.8	482	0.035	0.0097	7.60	0.0016
Dingqing	3873	3.5	653	0.034	0.0031	4.66	0.0371
Zuogong	3780	4.6	451	0.043	0.0023	4.68	0.0196
Lhasa	3648	8.2	446	0.056	0.0000	6.90	0.0052
Changdu	3306	7.6	492	0.017	0.1039	3.90	0.1624
Linzhi	2991	8.9	709	0.040	0.0005	3.04	0.3170



Fig. 2. Remote sensing data of study area at three different scales. There are 246 grids of microwave remote sensing data with a spatial resolution of 25 km × 25 km. Each grid covers about 25 0.05° MODIS pixels and approximately 2500 500-m MODIS sub-pixels.

can determine from the remote sensing imagery, the onset (T_1) and end (T_2) dates of the snow-free period (Fig. 3). The area above *H* is covered by snow, while it is snow-free for the area below *H*. Transient snow events occurred where there were numerous T_1 and T_2 values on the horizontal line and were considered to be the snow-free period for the entire year. In the region between the maximum and minimum of FSCE, seasonal snow dominates. The seasonal-snow dominant period within FSCE is from T_1 to T_2 , which is the snow-fit zone and temperature-sensitive.

For each year and for each elevation, the median date of the snow-free period ($T_{\rm m}$) and duration of the snow-free period (ΔT) were selected to characterize the snow-free period.

$$\begin{cases} T_m = \frac{T_1 + T_2}{2} \\ \Delta T = T_2 - T_1 - 1 \end{cases}$$



Fig. 3. Schematic view of T_m and ΔT in FSCE.

The length of the snow-covered season for a given pixel for any year is thus $1 - \Delta T$. To avoid the effect of transient snowfall events we used a backward temporal filter based on the approach of Li and Williams (2008). From Jan 1 to Dec 31, the filter calculates the frequency of snow occurrence within the next days of year and snow-free season commences (T_1) when snow-free days last for 45 days. Meanwhile, from Dec 31 to Jan 1, such filter counts the frequency of snow occurrence within the past days of that year and snow-free season ends (T_2) if snow-free days last for 45 days. Executed shown in Fig. 4.

Using this approach, the dates between T_1 and T_2 are symmetrical, and we can then define non-snow-covered regions (F_{min}), seasonally snow-covered areas (F(t)), and perpetual snow (F_{max}):

$$FSCE = \begin{cases} F_{\min} = \sup(\text{non snow} - \text{covered}) \text{ for } T < h_1 \text{ or } T > h_2 \\ F(T) = F(2T_m - T) \text{ for } h_1 \le T \le h_2 \\ F_{\max} = \inf(\text{perpetual snow}) \text{ for } T = T_m \end{cases}$$

in which $F(T) = F(T_m + \frac{1}{2}\Delta T) = F(T_m - \frac{1}{2}\Delta T)$ for $T_1 \le T \le T_2$, *T* is date, and h_1 and h_2 are the dates between transient snow and season snow, respectively.

There is thus a pair of T_1 and T_2 each year for each elevation. Thus, we calculated T_m and ΔT every year as a function of elevation. The multiyear mean T_m and ΔT from 1978 to 2005 for each grid were calculated as follows:

$$T_{mi} = \frac{\sum_{j=1}^{27} \left(\frac{T_{1ij} + T_{2ij}}{2}\right)}{27}, \quad \Delta T_i = \frac{\sum_{j=1}^{27} \left(T_{2ij} - T_{1ij} - 1\right)}{27}$$

where subscript *i* denotes the *i*-th pixel (i = 1, 2, ..., 246) and *j* denotes the *j*-th year (j = 1, 2, ..., 27).

We evaluated changes in T_m and ΔT over time using Trend-free pre-whitening Mann–Kendall (TFPW-MK). Non-parametric MK trend test (Mann, 1945) has been widely used for trend assessment of hydro-meteorological variables (Racoviteanu et al., 2008; Yang & Tian, 2009; Zhang et al., 1991) in serially independent series. In addition, Yue et al. (2002) proposed TFPW-MK to deal with serially correlated series. AR(1) is calculated to judge whether observed sample data Y_t are serially correlated. If the lag-1 serial correlation falls within the confidence interval calculated by the significance level of $\alpha = 0.10$ of twotailed test, the observed data are assumed to be serially independent.

$$AR(1) = \frac{\sum_{t=1}^{n-1} (Y_t - \overline{Y_t}) (Y_{t+1} - \overline{Y_{t+1}})}{\sqrt{\sum_{t=1}^{n-1} (Y_t - \overline{Y_t})^2 \cdot \sum_{t=1}^{n-1} (Y_{t+1} - \overline{Y_{t+1}})^2}}{\frac{-1 - 1.645\sqrt{n-2}}{n-1} \le r_1 \le \frac{-1 + 1.645\sqrt{n-2}}{n-1}}$$

Otherwise, for a serially correlated series:Firstly, calculate the Kendall slope β of observed dataset Y_t and get the detrended series Z_t based on slope β .

$$\beta = \text{Median}\left(\frac{Y_j - Y_i}{j - i}\right), \forall i < j$$
$$Z_t = Y_t - \beta \times t$$

Secondly, the residual Z_t' is obtained after removing the AR(1) from Z_t :

$$Z'_t = Z_t - r_1 \times Z_{t-1}$$

Thirdly, blend the trend T_t and the residual Z_t' :

$$Z''_{t} = T_{t} + Z'_{t}$$

Finally, MK trend test is conducted on the newly-created series $Z_{t''}$. The number of samples for each parameter is 27, the number of years for which we have SCA from the microwave sensors.

4. Results and discussion

4.1. Temperature and precipitation trends

Annual air temperatures for the 27 years of record at the seven meteorological sites ranged from +8.9 °C at the lowest elevation site (Linzhi, 2,991 m) to -0.5 °C (Jiali, 4,488 m) (Table 1). Air temperature decreased with elevation (R = -0.94, p < 0.05), a lapse rate of 6.3 °C per 1000 m. This long-term lapse rate is similar to that of 7.0 °C per 1000 m reported for the Colorado Rocky Mountains (Williams et al., 2011). However, these air temperatures are relatively warm when compared to other continental high-elevation sites. For example, Williams et al. (1996) reported an average air temperature of -3.8 °C for the D1 climate station (3720 m) in the Colorado Rocky Mountains of the US.

Air temperature increased significantly from 1978 to 2005 at six of the seven meteorological stations, with only the low-elevation Changdu station (3306 m) not showing a significant increase in annual air temperature (p=0.10). The warming rates of 0.017 to 0.056 °C y⁻¹ are similar to the 0.028 to 0.043 °C y⁻¹ reported by Immerzeel et al. (2009) for the greater Tibetan Plateau from 1972 to 2002. The largest increases in monthly air temperatures were during the winter and weakest during the summer (Fig. 5), similar to the results reported by Immerzeel et al. (2009) and Rangwala et al. (2009b) for the Tibetan Plateau. However, in contrast to those studies, which reported higher rates of warming with increasing elevation, while there was an upward trend in annual air temperature with elevation over the 27 years for the southeastern Tibetan Plateau, it was not significant (R=0.42, p>0.05). Average annual precipitation ranged from 451 mm y⁻¹ to 732 mm

 y^{-1} (Table 1). There was no correlation between elevation and



Fig. 4. Identification of T_1 and T_2 in FSCE.



Fig. 5. Mann-Kendall trend test and slopes of monthly air temperature for the seven meteorological stations. The seven stations are Linzhi, Changdu, Lhasa, Zuogong, Dingqing, Dangxiong and Jiali, left- to- right and low to high elevation, respectively.

precipitation amount (R = 0.074, p > 0.05). All sites showed an increasing trend in annual precipitation that ranged from 3.04 mm y⁻¹ to 7.60 mm y⁻¹, with the trend significant at the α = 0.05 level for five of the seven meteorological stations (Table 1). In contrast to air temperature, the increasing rate of annual precipitation increased with increasing elevation (R = 0.701, p < 0.05), consistent with Rangwala et al. (2009a) who reported increasing winter and spring precipitation with increasing elevation for the Tibetan Plateau.

4.2. Snow cover area: MODIS

The average annual seasonal snow cycle for our study area is similar to that of most mid-latitude mountain ranges in the northern hemisphere (Fig. 6). Minimum SCA occurs in August, ranging from 7 to 17%. Snow then begins to accumulate in fall, with the maximum SCA occurring around mid-February to early-March. There is then a precipitous drop-off in SCA from about mid-May to early July, which we assume to be the main snowmelt season. There are some regional differences in SCA. Peak annual SCA of more than 70% in the Niyang River basin is about 20–30% greater than in either the Lhasa River basin or the Changdu River basin.

4.2.1. Snow cover area: passive microwave

We compared the average SCA snow product from MODIS (MYD10A2 (MODIS/Aqua Snow Cover 8-Day L3 Global 500 m Grid)) from 2002185(July 4 2002) to 2005361(December 27 2005) to that of the passive microwave results. Among them, three days (2002201, 2002209, and 2002217) with bad quality were removed. There were 158–8 day periods where MODIS and SSM/I data products overlapped; note that these averages include snow-free and snow-covered dates. For each SSM/I pixel that was classified as having SCA (light, or "1", in Fig. 7), the mean SCA value for MODIS ranged from 1% to 87%, with a mean SCA area of $35 \pm 20\%$. In contrast, for pixels classified as being snow-free in the SSM/I product (dark, or "0", in Fig. 7), the SCA value for MODIS was $15 \pm 7\%$. Thus, SSM/I SCA product only classified a pixel as snow-covered when the subpixel SCA resolution was at least 15% relative to MODIS values. Similarly, the SSM/I considered a pixel



Fig. 6. Average seasonal snow covered area expressed as percentage of total basin area derived from MYD10A2 (MODIS/Aqua Snow Cover 8-Day L3 Global 500 m Grid data) from 2002185 to 2005361.

snow-covered when on average MODIS indicated a subpixel SCA of about 35%.

4.3. Spatial distribution of T_m and ΔT

In order to evaluate only the pixels with seasonal snow cover, the pixels without a consistent snow-covered season (F_{min}) were removed following a protocol similar to Li and Williams (2008). This filtering process reduced the number of microwave pixels from 246 to 223 (Fig. 8). The removed pixels were mainly located in the lower reaches of the Lhasa River basin, which is characterized by a dry and hot valley; north of the Lhasa River basin, where wind erosion and transport play an important role; and the eastern side of Changdu region, which is on the leeward side of Hengduan Mountain.

The 90 m×90 m SRTM elevation estimates were then overlaid on each 25 km×25 km microwave pixel, resulting in an aggregation of elevation from 3854 m to 5488 m. For the time period from 1978 to 2005, the date of the mid-point of the snow-free season (T_m) was DOY 213 (1 August), with a standard deviation of 3.5 days (Fig. 9). Thus, almost 96% of the dates of the midpoint of the snow-free season happened from 25 July to 8 August. These results are consistent with our MODIS results of the changes in snow-covered area with time,



Fig. 7. Comparisons between the microwave and MODIS snow cover products. Stripe denotes whether the microwave product considered the pixel snow-covered (light, or "1") or snow-free (Dark, or "0"). The y-axis shows the percentage of SCA from MODIS for each microwave grid. Values are averages from 158 8-day MODIS averages for each microwave grid from 2002 to 2005, compared to the average microwave value of SCA for that 8-day period over the same time period. Pixel number is arbitrarily ordered from low to high MODIS SCA values.



Fig. 8. Study area after removing pixels with only transient snow. This filtering process reduced the number of microwave pixels from 246 to 223. The removed pixels were mainly located in the lower reaches of the Lhasa River basin, which is characterized by a dry and hot valley; north of Lhasa River basin, where blowing snow plays an important role; and the eastern side of Changdu region, which is on the leeward side of Hengduan Mountains.

which show a minimum in late July and early August (Fig. 6). In contrast to the mid-point of the snow-free season, the length of the snow-free season (ΔT) ranged widely, from 90 to 216 days (Fig. 10). The greater the elevation, the shorter the snow-free season:

 $\Delta T = 400.690 - 0.055 \times elevation(m)$

(R = -0.651, p < 0.001). This value is very similar to the results of Varhola et al. (2010), which was for snow depth and elevation at the meso-scale, while this study focuses on the macro-scale.

We can then quantify the average values of T_m and ΔT from 1978 to 2005 by combining the mean value of T_m with the regression equation above:

$$\begin{cases} |t - T_m| = \frac{\Delta T}{2} \\ \Delta T = b_0 + b_1 H \end{cases} \Rightarrow H = \frac{1}{b_1} (2|t - T_m| - b_0) \\ \text{FSCE} : f(t) = \frac{1}{b_1} (2|t - T_m| - b_0) (H \in [3854, 5488]) \end{cases}$$

when considering the uncertainty of T_m and ΔT , T_m changes in the range of $[T_m - \sigma, T_m + \sigma]$ and ΔT vary from the lower envelope to upper envelope. Here, $E(T_m) = 212.7$, the amplitude of T_m is ± 3.5 , $b_0 = 400.7 \pm 19.8$, and $b_1 = -0.05 \pm 0.004$ for the lower and upper



Fig. 9. For the time period from 1978 to 2005, the date of the mid-point of the snow-free season (T_m) was DOY 213 (1 August), with a standard deviation of 3.5 days. The median of the snow-free season showed little change with elevation or time.

envelops, respectively (Fig. 11). For the average value of T_m and ΔT , snowmelt happened from late-April to mid-June from lower elevation to higher elevation. Snow accumulated from mid-September to mid-November from high to low elevation.

4.4. Trend test of ΔT

We then evaluated whether the length of the snow-free season (ΔT) changed over time as a function of elevation (Fig. 12). For ease of display, the 223 pixels were binned with an elevation interval of 400 m. There were 32 pixels within the elevation band of 3800 m to 4200 m, 87 pixels between 4200 m and 4600 m, 73 pixels between 4600 m and 5000 m, 30 pixels between 5000 m and 5400 m, and only 1 pixel above 5400 m. The ratio of the number of pixels with a significant trend to the total pixel number of that elevation band was then plotted to reveal the elevation distribution of significantly changed pixels (Fig. 12(b)). For the ΔT trend test, an upward trend means the snow-free period is getting longer while a downward trend means that the snow-free season becomes shorter (or the snow covered season is longer). In the results, TFPW-MK trend test on ΔT shows a prolonged snow-free season for low elevation sites but a reduced snow-free season at higher elevations (Fig. 12(b)).

Our results indicate that from 1978 to 2005 the snow-covered season increased in length with increasing elevation, while the snow-



Fig. 10. The length of the snow-free season (ΔT) ranged widely, from 90 to 216 days. The greater the elevation, the shorter the snow-free season.



Fig. 11. The length of the snow-free season (ΔT) was inversely related to elevation, with $\Delta T = 400.690-0.055 \times$ elevation (m) (R = -0.651, p < 0.001). Legend: Significantly decreasing trend, it means a shortend snow-free period for ΔT . Significantly increasing trend, it denotes a prolonged snow-free period for ΔT .

covered season decreased in length at low elevations. These results are not consistent with those of Rangwala et al. (2009a) who reported that their modeling simulations suggested an increasing reduction in snow cover over time with increasing elevation for the Tibetan Plateau. Precipitation will remain as snow as long as it is below freezing (Ye et al., 1998). Moreover, snow on the ground will cool the overlying air, which may counteract a regional warming trend in air temperature (Pepin, 2000). The increasing precipitation with elevation that we report is most likely resulting in more snowfall (Qin et al., 2006). For

a) Spatial distribution of ΔT changing with time





Fig. 12. (a). Spatial distribution of ΔT changing with time. Legend: Significantly decreasing trend, it means a shortened snow-free period for ΔT . Significantly increasing trend, it denotes a prolonged snow-free period for ΔT (b). Elevation distribution of ΔT changing with time.

low elevation sites, the increase in air temperature results in an increase in the number of precipitation events with air temperature above freezing, and thus less snowfall and earlier snowmelt. Our results suggest a precipitation-driven system at high elevation with a temperaturedriven system at low elevation.

5. Conclusion

In this paper, based on a comparison between the MODIS snow product and microwave snow depth dataset and an analysis of local temperature and precipitation, we examined the spatiotemporal distribution of snow in eastern Tibet. The concept of FSCE is proposed based on two parameters, the median date of the snow-free period (T_m) and duration of the snow-free period (ΔT). T_m is quite a stable value for this region, while ΔT changes with elevation. We found that the duration of the snow-free season ΔT was inversely correlated with elevation (R = -0.651). However, the relation was not sufficient to establish a rigid FSCE curve, though our results suggest that the concept of an FSCE has some utility. Firstly, there is a significant influence of elevation on snowmelt leading to lower risk of spring snowmelt flooding in mountainous regions due to differentiated onset melting along the gradient (Varhola et al., 2010). FSCE demonstrates the seasonal variability of ΔT for different elevation bands. Secondly, FSCE enlightens us to examine the interannual variability of ΔT for different elevation band, which indicates the response of snow to climate change from 1978 to 2005. Our study also illustrates that snow distribution in this region is precipitation-driven at high elevation sites but temperature-driven in the low ones.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at doi:10.1016/j.rse.2012.01.006. These data include Google maps of the most important areas described in this article.

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