
Changes of snowfall under warming in the Tibetan Plateau

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Key Points:

- We use a two-threshold temperature method to model snowfall at 71 stations across the Tibetan Plateau
- Region-wide snowfall increased during 1961-1990 and 1971-2000, but decreased in the more recent period 1981-2010
- Seasonal/elevational contrasts in snow trends can be explained in part referring to mean temperature

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Abstract:

Snowfall is a critical part of the hydrological system in high-altitude regions and strongly impacted by climate change. This study uses a threshold temperature method to estimate spatial and temporal variations of snowfall at 71 stations across the Tibetan Plateau from 1960-2014. Regional air temperature and precipitation have increased by 0.039 °C/yr, and 1.43 mm/yr, respectively. While warming rates have been fairly uniform across the plateau, spatial variations in snowfall trends are large, with decreases in the eastern and north-eastern areas but increases at higher elevations in the centre and west. Region wide snowfall increased during 1961-1990 and 1971-2000, but decreased in 1981-2010 and 1991-2014. Wintertime snowfall has increased but summer snowfall has decreased. These divergent trends can be explained because maximum snowfall is recorded at temperatures between 1 and 2 °C. Above/below this threshold snowfall usually decreases/increases with increased warming. Although maximum snowfall temperature is a key factor to understand future snowfall changes, concurrent influences such as changing moisture sources and atmospheric circulation patterns require further research.

Keywords: Snowfall; Climate change; Mountains; Tibetan Plateau

1 Introduction

Snowfall is a critical component of the hydrological system in mid- and high-altitude regions [Groisman and Easterling, 1994; Groisman *et al.*, 1994] including in mountainous landscapes [Borys *et al.*, 2003]. Snowfall also plays an important role in both energy and water exchange between the atmosphere and the land [Marks and Dozier, 1992]. The presence of snow increases surface albedo resulting in regional cooling over snow-covered regions [Cohen and Rind, 1991]. Snow is also an important reservoir of water, and acts as a buffer in the hydrological system controlling river discharge and associated environmental processes and hazards [Räisänen, 2008].

A warmer climate will influence snowfall and in turn snowpack development, and the timing of snowmelt [Barnett *et al.*, 2005; Burns *et al.*, 2014; Kapnick and Delworth, 2013]. With rising temperatures there is a shift from snow to rain [Berghuijs *et al.*, 2014] and earlier melt [Barnett *et al.*, 2005]. Foster *et al.* [2008] have shown a transition towards melt around 4-7 days earlier since the 1980s over Arctic regions. Higher temperatures have decreased snowfall in winter and early spring and led to earlier snowmelt in the west of North America [Knowles *et al.*, 2006]. In the Tianshan Mountains, melt trended 15.33 days earlier over the period 1960-2010 [Li *et al.*, 2013], and snow cover area (SCA) has decreased between 2002 and 2013 by -0.17%/yr [Chen *et al.*, 2016]. The decrease in snowfall and snowpack is however not ubiquitous. North-eastern China has seen an increase in snowfall [Wang and He, 2013] which is possibly due to a weakening of the dry East Asian winter monsoon. Snowfall increased in northeastern China over 1951-2010 based on 160 stations. In the Tibetan Plateau, snow depth also increased by 2.3%/yr during 1957-1992 based on analysis of the Scanning Multichannel Microwave Radiometer 6-day snow depth product [Qin *et al.*, 2006].

There are strong reasons why the rate of warming in high-mountains can be more than in lower elevation regions [Mountain Research Initiative, 2015] one of which is due to snow albedo feedback around a retreating snowline [Pepin and Lundquist, 2008]. Thus it is of critical importance to see whether snowfall trends also change with elevation and across the

plateau. Most of the Tibetan Plateau has trended to both a warmer and wetter climate during past decades [Xu *et al.*, 2008; Kang *et al.*, 2010] and the warming has been particularly rapid where current temperatures are near freezing [You *et al.*, 2010]. Precipitation however shows a more varied pattern and has decreased in some parts of the plateau but increased in others [Kang *et al.*, 2010; You *et al.*, 2014]. A series of studies have suggested negative effects of warming on snowfall in most areas of the plateau [Bhutiya *et al.*, 2010; Mir *et al.*, 2015; Wang *et al.*, 2016]. However there are regional differences in snow sensitivity to warming because of contrasts in meteorological controls. For example studies indicate a much weaker sensitivity of snowfall to warming in the Karakoram than in the Himalayas [Yao *et al.*, 2012, Kapnick *et al.*, 2014]. Over the plateau as a whole there are still some key unknowns which include (1) the air temperature range over which snowfalls occur; and (2) the temperature at which maximum snowfall occurs. Both these are critical to understand the resulting patterns of snowfall change which will result from future warming.

In this study, we focus on snowfall changes under warming in the Tibetan Plateau from 1960-2014. First we describe the temporal and spatial variations of snowfall changes under warming in the Tibetan Plateau based on the development of a cross-quadrant diagram (see methods). Second we determine the temperature at which maximum snowfall occurs to help explain the patterns of snowfall change. Section 2 describes the study area, data, and methods. Section 3 focuses on recorded temperature, precipitation and snowfall changes before the mechanisms explaining the changes are examined in section 4. Broader implications of these findings are discussed in section 5 and Section 6 presents our conclusions.

2 Materials and Methods

2.1 Study area

The Tibetan Plateau lies between Central and Eastern Asia and can be delimited by 25°~39°N, 70°~106°E (Figure 1 a). The average elevation of the Tibetan Plateau exceeds 4000 m,

meaning that this is the largest extensive area of alpine habitat in the world (referred to as “the Third Pole”). The plateau holds one of the largest areas of permanent snow and ice in mid-latitudes [Yao *et al.*, 2012], and all of the rivers in the surrounding region originate on the plateau [Immerzeel *et al.*, 2010].

The Tibetan Plateau has a typical high-elevation climate. Long-term mean average temperatures are about 4.3 °C based on 87 meteorological stations (Figure 1 a). Mean temperatures over 15 °C occur in July but they fall below -10 °C in January (Figure 1 b). Region-wide annual average precipitation is about 450 mm, most of which is concentrated between June and September (Figure 1 b), due to the influence of the Indian monsoon in the Himalayan ranges (south-west plateau) and the East Asian monsoon in the east of the plateau [Yao *et al.*, 2012]. Both of these large scale flows are modulated by the heating of the plateau itself in summer [Tang *et al.* 1984]. In the winter the mid-latitude westerly circulation is more dominant giving enhanced precipitation at high elevations and in the northern and western regions such as the Pamir and Karakorum [Kääb *et al.* 2012, Gardelle *et al.* 2013]. More recently the westerlies have been shown to be important further south in contributing moisture to parts of the Himalayan ranges and in seasons other than winter [Mölg *et al.*, 2014]. Thus the dominant sources of moisture vary both by season and across contrasting parts of the plateau.

In this study, the Tibetan Plateau was divided into three sub-regions (Figure 1 a) based on drainage basin hydrology. Region A drains to the south and south-east, region B covers basins draining to the east, and region C drains to the north-west (Figure 1 c-e).

[Insert Figure 1 about here]

2.2 Data

87 meteorological stations from the National Meteorological Administration of China (NMAC, <http://ncc.cma.gov.cn>) are used, mainly distributed in the south and east of the Tibetan Plateau (Figure 1 a). Because 16 stations had missing data (black flag symbol in

Figure 1 a), we only selected 71 stations with complete daily data for 1960-2014. Data accuracy was assessed through strict quality control. This included the identification and removal of extremes, checks for internal consistency and the removal of spatial and temporal outliers through comparison between days and between neighbouring stations. The standard normal homogeneity test was also applied to identify breaks in records and adjustments made if required. Full details of the assessments applied are given in the paper by Feng et al.[2004]. Variables examined include daily maximum, mean and minimum temperatures ($^{\circ}\text{C}$), precipitation (mm), and relative humidity (%). We use a combination of relative humidity, precipitation, and temperature to model snowfall.

2.3 Methods

2.3.1 Snowfall calculation

Precipitation can fall as solid (snow), liquid (rain) or a mixture (sleet). Based on previous studies [Bocchieri, 1980; Stewart, 1992] the solid-liquid mixed phase can be challenging to classify and numerous studies examine how to determine precipitation type using observed data [Ding et al., 2014]. The main methods used to discriminate include

(1) the temperature profile method [Bocchieri, 1980; Schuur et al., 2012; Thériault et al., 2006]. This uses vertical temperature and wet-bulb temperature profiles and calculates the elevation of the 0°C isotherm to discriminate precipitation types, and requires radiosonde or satellite data. Bourgoïn [2000] enhanced the temperature profile method by developing an area method to diagnose precipitation type.

(2) temperature threshold methods [Clark et al., 2006; Gustafsson et al., 2001; Stewart, 1992; Wigmosta et al., 1994; Yang et al., 1997; Zhang et al., 2013]. These methods use temperature or wet-bulb temperature (usually at screen level) to define thresholds which demarcate precipitation types using surface observations only. Ding et al. [2014] developed a new temperature threshold method to determine precipitation types based on wet-bulb temperature and relative humidity. Upper and lower threshold temperatures at which snowfall could occur were calculated based on

wet-bulb temperatures. The wet-bulb temperature is more useful than dry-bulb alone since it accounts for evaporative cooling effects [Bocchieri, 1980]. In dry air enhanced evaporative cooling can mean that snowfall can occur at higher dry-bulb temperatures. Under high relative humidities the transition from solid to liquid is more rapid whereas in dry air with a large wet-bulb depression the opposite is the case.

In this study we calculate threshold temperatures to divide precipitation types using wet bulb temperatures based on *Ding et al.* [2014]:

$$P_{type} = \begin{cases} \text{snow,} & T_w \leq T_{min} \\ \text{sleet,} & T_{min} < T_w < T_{max} \\ \text{rain,} & T_w \geq T_{max} \end{cases} \quad (1)$$

where P_{type} is the daily precipitation types (snow, rain, or sleet), T_w is daily wet-bulb temperature, and T_{min} and T_{max} are threshold temperatures. T_{min} represents the temperature where snow and sleet are equally likely, and T_{max} the temperature where sleet and rain are equally likely. Both T_{min} and T_{max} depend on relative humidity (RH). There is a smaller/larger temperature range for sleet if RH is low/high. Detailed calculations are given in Appendix A. Figure 2 plots the resultant amounts of rain, and snow modelled for different temperature ranges based on all stations. Sleet and mixed precipitation can fall at temperatures as high as 5°C. The original station data lists mixed precipitation as a separate category which amounts to up to 5% of the total original precipitation depending on location, so it is important to model the contribution of sleet/mixed precipitation to the final snowfall in a realistic way.

[Insert Figure 2 about here]

According to the method of *Vehviläinen* [1992] the fraction of precipitation falling as snowfall on any individual day is calculated as:

$$f_s = \begin{cases} 1, & T_w \leq T_{min} \\ \frac{T_{max} - T_w}{T_{max} - T_{min}}, & T_{min} < T_w < T_{max} \\ 0, & T_w \geq T_{max} \end{cases} \quad (2)$$

where f_s is the fraction of precipitation recorded as snowfall.

Thus sleet contributes fractionally to snowfall in our model, but in proportion to where in the transition between rain and snow it falls. In dry air ($RH < 78\%$) there is an instantaneous change between rainfall and snowfall because $T_{max} = T_{min}$ but in wet humid environments the transition becomes more gradual, sleet can fall at a wider range of temperatures ($T_{max} > T_{min}$), and the transitional state in equation 2 becomes more important.

The amount of snowfall in daily precipitation is calculated by

$$P_s = P * f_s \quad (3)$$

where P is the total daily observed precipitation (mm), and f_s is calculated by Eq.(2).

Daily snowfall totals were aggregated for each station and mean annual values were determined. We also calculated a weighted annual average of the lower quartile (Q1), median (Q2), and upper quartile (Q3) of daily snow amounts to avoid the excessive influence of outliers.

$$\bar{x} = Q1 * 0.25 + Q2 * 0.5 + Q3 * 0.25 \quad (4)$$

where $Q1$, $Q2$, and $Q3$ are lower quartile, median, and upper quartile, respectively.

Mann-Kendall nonparametric tests [e.g., *Hamed*, 2008; *Hamed and Ramachandra Rao*, 1998; *Hirsch and Slack*, 1984], were applied to assess trend significance for air temperature, precipitation, and snowfall over the past 55 years. The slope of a trend is estimated by Sen's nonparametric trend estimator [*Sen*, 1968].

2.3.2 Cross-Quadrant diagram

[Insert Figure 3 about here]

After all trend Z values for temperature and snowfall have been calculated (Figure 3 a) the significance can be calculated using the M-K trend test. P values less than 0.05 are used to identify significant trends. The quadrant diagram (Figure 3 b) plots temperature trend Z value (y axis) against snowfall trend Z value (x axis). The dotted lines separate significant from

insignificant trends. Detailed interpretation of each section of the diagram is given in Table 1.

[Insert Table 1 about here]

3 Results

3.1 Temperature and precipitation trends

[Insert Figure 4 about here]

The whole Tibetan Plateau shows obvious warming during the past 55 years [Chen *et al.*, 2013]. However this is not consistent. From 1960-1969 the temperature decreases by 0.026 °C/yr, but the trend is statistically insignificant (Figure 4 a). Beginning in 1970 there is a significant increase of 0.039 °C/yr. Precipitation shows a concurrent increase of 1.43 mm/yr from 1970 to 2014 (Figure 4 b). The wettest year is 1998 (530 mm). Since 1998 there appears to be a slight decrease but this is not significant. The mean inter-site variance in temperature is about 4.25 °C (Figure 4 c), and with warming this variance across the plateau has decreased, suggesting that the coldest areas are warming faster than warmer areas. Inter-site variance in precipitation is large (Figure 4 d), ranging from 240 to 300 mm and this has also increased over the record.

[Insert Figure 5 about here]

Figure 5 shows the trend magnitudes over the past 55 years. Temperature is significantly increasing at nearly all stations (Figure 5 a). There is a large amount of variance in rates but overall Region A has lower rates (0.02~0.06 °C/yr) than regions B and C (0.04~0.08 °C/yr). There is only one station with a decreasing trend (-0.0143 °C/yr), but this is not significant. Precipitation (Figure 5 b) and rainfall (Figure 5 c) have 87.3% (62) and 85.9% (61) stations with increasing trends, respectively. Thus there is an overall tendency towards wetter conditions at the vast majority of stations.

Despite relatively uniform temperature and precipitation trends, snowfall trends show big spatial differences. Figure 5 d shows that about 40.9% (29) of stations have a decreasing trend but 59.1% (42) stations an increase. However not all of these trends are significant with

only 8/7 stations showing significant decreases/increases respectively. Thus the pattern is more complex than may be expected. Decrease of snowfall is focused in the southern and eastern parts of regions A and B. Increasing snowfall is concentrated in the west of region A and region C.

[Insert Figure 6 about here]

[Insert Table 2 about here]

Comparing the spatial distributions of each parameter (temperature, rain, snow and total precipitation observed at each of the 71 stations) for different sub-periods will help illustrate observed changes over time. For a shifting inter-decadal perspective we choose the periods 1961-1990, 1971-2000, 1981-2010 and 1991-2014 based mostly on standard 30 year periods recommended by the World Meteorological Organisation for climate analysis [WMO, 2009]. The most recent period is of necessity slightly shorter. Temperature curves for the various periods have a marked upward shift, particularly in the later two periods (Figure 6 a). The mean temperature (based on all stations) in 1991-2014 increased ~ 0.9 °C above that of 1961-1990 (Table 2). Both total precipitation and rain also have smaller upward shifts (Figure 6 b and c). Mean precipitation increased by 17 mm in 1991-2014 compared to 1961-1990 and mean rain increased by the same amount using the same periods for comparison. There is on average a very small increase of mean snowfall (Figure 6 d) of 0.38 mm from 1961-1990 to 1991-2014 (Table 2) but the variance across the plateau has also increased meaning that the 95% confidence interval for plateau-wide mean snowfall has widened. Therefore, snowfall can be seen to have complex changes even under fairly consistent warming conditions and a trend to increased precipitation across the Plateau.

3.2 Relationship between temperature and snowfall trends

[Insert Figure 7 about here]

Most stations (56) show insignificant trends in snowfall despite significant warming (Figure 7 a1 and a2). 11.3% (8) stations have significant decreasing snowfall under warming (Figure 7 a2). 4 of these are in region A (Figure 7 b2) and 4 in region B (Figure 7 c2). There are 6

stations with increasing snowfall under warming, 3 of which are in region A, 2 in region B and 1 in region C (Figure 7 d2). One station shows increasing snowfall but has no significant warming.

[Insert Figure 8 about here]

[Insert Table 3 about here]

Although over the whole record there are relatively few stations showing significant snowfall change, an examination of sub-periods shows an interesting story. Snowfall has tended to change in different directions in different sub-periods, thus often cancelling out over the period as a whole. Figure 8 shows maps of snow trends for four sub-periods. During 1961-1990, snowfall increases at 76% (54) stations (Figure 8 a) although this is significant at only 20% (11). 1971-2000 is broadly similar with a tendency toward snow increase but insignificantly so at the majority of stations (Figure 8 b). By 1981-2010 however there is a rapid change with a reduction in snowfall at 76.1% (54) stations (Figure 8 c), mainly in region B. This does not last and from 1991-2014 most snowfall trends are insignificant (Figure 8 d).

Table 3 summarises the significance of temperature and snowfall trends using a contingency table, where the letters represent the areas of the cross-quadrant diagram, and the number of stations with each snowfall/temperature trend combination are listed. For an interpretation of the letters refer to Table 1. It is immediately obvious that at most stations although temperature trends are positive (significant warming) snowfall trends are insignificant (category i). However there are subtle changes between periods. The black bold cells identify interesting tendencies at sub-sections of stations. These include category f (warmer, more snow) in 1961-1990, category l (no temperature change, more snow) in 1971-2000, and category e (warmer, less snow) in 1981-2010 and to a lesser extent in 1991-2014. Most of the significant temperature increase has been since 1981. As time has progressed there has been a subtle switch from an increase in snowfall (f) to a decrease (e) even though temperatures have continued to increase throughout.

[Insert Figure 9 about here]

[Insert Table4 about here]

Table 4 shows a comparable summary to Table 3 but for individual seasons. Significant frequencies are highlighted in bold. Over the whole period (1961-2014) there is a large contrast between snowfall changes in different seasons. Winter and spring show broadly similar patterns. In DJFM (Figure 9 a), 77.5% (55) stations show an increase in snowfall and at 34.6% (19) stations this trend is significant. Trends are insignificant at 48 stations and only 2 stations show a significant decrease in snow. In AM (Figure 9 b) patterns are broadly similar, but there are only 12 stations with significant increases in snow. There are quite a lot of stations with insignificant decreases, mostly on the fringe of the plateau. In summer (JJAS) (Figure 9 c) however most stations (68) show reductions in snowfall and at 15 stations this trend is significant. Unlike winter and spring no stations show a significant increase. Snowfall changes in ON (Figure 9 d) are similar to JJAS but to a lesser extent, with 11 stations showing a significant decrease in snowfall. The opposing trends in the cold and warm part of the year at a minority of stations are a significant finding suggesting that a seasonal breakdown is critical to understand long term trends.

Table 4 also shows that as time has progressed there have been subtle changes in the trend combinations for each season. The most common category is usually i (significant warming but no snow trend) and this has remained so in recent decades. Most of the significant snow increases (category f) are focused in winter in earlier periods and over time category e (warming, less snow) has become more common, but mostly in summer and autumn. Some of the snow increase in winter and spring in the earlier periods was associated with no temperature change (category l), suggesting precipitation changes to be an influence earlier on.

[Insert Figure 10 about here]

Figure 10a plots snowfall and temperature trends for all stations classified by elevation band. Mean temperature will decrease with elevation, and lower temperatures will enhance snowpack. It is not immediately clear how snowfall or its trend would be influenced by

elevation. Although there is little correlation between temperature trend and elevation in this dataset, some elevation dependency is illustrated in snowfall trends (Figure 10b). Higher level stations (>3500 m) nearly all show increasing snow fall, whereas moderate elevation stations (2500m-3500 m) mostly show a decrease. The lowest stations (<2500 m) have very weak snowfall trends, probably because there is not much snow below 2500 m. Using a simple linear model based on all stations ($r^2 = 0.23$), the transition elevation at which the trend changes from negative to positive is 3000 m. However trend magnitudes below 3000 m are more constrained towards zero because of the absence of snow at lower elevations. Thus applying a linear model to sites above 3000 m only ($r^2 = 0.29$) yields a slightly higher transition elevation of 3396 m.

3.3 Analysis of the mechanisms behind snowfall changes under warming

Temporal and spatial changes of snowfall are complex under warming. In order to account for the contrasts in snowfall changes observed despite fairly universal warming, we calculated the multi-year mean snowfall at different air temperatures for each station and combined all the results. Figure 2 (based on all stations) shows that snowfall mainly occurs when air temperatures range between $-12\text{ }^{\circ}\text{C}$ and $4\text{ }^{\circ}\text{C}$, and the maximum fall is for temperatures between $1\text{ }^{\circ}\text{C}$ and $2\text{ }^{\circ}\text{C}$. Thus at lower temperatures air temperature increase should result in increased snowfall but at higher temperatures the same increase should result in rapid decrease in snowfall.

[Insert Figure 11 about here]

[Insert Table 5 about here]

During 1961-1990 mean snowfall has an increasing trend over the plateau as a whole and in all three subregions (Figure 11). Table 5 show that region C in particular shows a strong significant increase ($p < 0.05$). The high elevation means that the air temperature was well below $1\text{ }^{\circ}\text{C}$ in this period, which would explain why snowfall increases under warming in this

region. By 1971-2000, snowfall has an increasing trend in 3 out of 4 cases but these trends are insignificant. By 1981-2000, snowfall shows a significant decrease in the Tibetan Plateau as a whole and in region B. The later region has mean air temperatures higher than 1 °C in this period, explaining these decrease. Region C remains cold enough to show a weak increasing trend in this period (Figure 11 d3), albeit insignificant (Table 5). Finally by 1991-2014, snowfall has a decreasing trend in all regions except region C (Figure 11 d4).

[Insert Figure 12 about here]

[Insert Table 6 about here]

Over the past 55 years, there are again strong differences between seasons, with significant snowfall increases in winter (when mean temperatures are well below freezing) and significant decreases in summer and fall (Figure 12 and Table 6). Mean annual snowfall has significantly increased in Region C which is the coldest and has mean annual temperatures below 1 °C. On a seasonal basis winter snowfall has significantly increased in all regions. This is also true to a lesser extent in spring, but changes are not always insignificant. Summer and fall on the other hand show significant decrease in snowfall in all regions except the coldest area (region C). These changes together would lead to a more concentrated (but shorter) snow season across most of the plateau.

[Insert Figure 13 about here]

Much of the spatial contrast in snow trends can be explained by the relationship between snow trend and mean annual/seasonal temperature (Figure 13). Particularly in winter (and to a lesser extent in spring) there is a significant negative correlation between the two with colder stations showing increases and warmer stations the opposite. On an annual basis the relationship does become a little more complex.

The exact proportion of snowfall change which is a result of temperature change can be assessed using the elasticity method of Shaake [1990] and Zheng et al. [2009]. This quantifies

the percentage change in snowfall which results per degree change in temperature, averaged over the whole period (1961-2014) and over the whole year or an individual season. Figure 14 shows the snowfall elasticity for the year as a whole and for winter alone. On an annual basis (Figure 14a), sensitivity is greatest in the south-east of the plateau, reaching over 10% decrease in snowfall per degree change in temperature at some stations, and quite widely values are between 1 and 10% decrease. In winter many more stations show an increase in snow per degree increase in temperature (because temperatures are lower) but this is still not the case in the far south-east of the plateau.

[Insert Figure 14 about here]

4 Discussion

Over the past 55 years, both air temperature and precipitation have increased over the Tibetan Plateau, particularly since the late 1960s (Figure 4 a, b). Most recently, the precipitation trend has reversed or at least stalled from 1998 to 2014 (Figure 4 b). The mean warming rate reported over the plateau (0.39 °C/decade, Figure 4 a) is faster than for other lower elevation region's [Li *et al.*, 2012; Pepin and Lundquist, 2008].

However there are also spatial variations across the plateau particularly when examining snowfall trends. 40.9% (29) stations show decreasing snowfall and 59.1% (42) stations show increasing snowfall. Although some studies have showed that snowfall is reducing in the Tibetan Plateau [Zhu *et al.*, 2017], other high-resolution model simulations exhibit future snowfall increase in the northwestern Himalaya region [Kapnick and Delworth, 2013]. Thus contrasting snow trends have also been obtained in past studies. We also show a change over time from a predominant snowfall increase during 1961-1990 and 1971-2000, to a decrease during 1981-2010 and 1991-2014.

The main explanation consistent with these patterns is an increase over time in mean temperature [Ye, 2008]. Air temperatures at which snowfall is reported range from -12 °C to 4 °C from observed data (Figure 2) but the critical threshold for maximum snowfall is around

1-2 °C for the Tibetan plateau. Seasonal contrasts in snowfall changes, namely a winter increase and summer decrease, further support evidence that mean seasonal temperature is an important control of snow trends (Figure 9). Changes of snowfall are significantly influenced by whether the temperature is above or below the snow-rain transition threshold [Krasting *et al.*, 2013; Ye, 2008]. Therefore, the variability in daily air temperature must be considered for making predictions about future snowfall change [O'Gorman, 2014; Räisänen, 2008]. Snowfall changes in region C appear to be in contrast to the other regions. Although it is reasonable that lower mean temperatures are partly responsible for this, we must bear in mind that there are only four stations in this region, and that there is a higher level of uncertainty in results.

Temperature is however unlikely to be the sole forcing factor on snow trends. Temperature increases are strongest at higher elevations in the centre of the plateau (Figure 5 a), yet precipitation increases are strongest around the periphery of the plateau in the east and south (Figure 5 b). The combined effect is that changes in snowfall are extremely complex. Lower elevation areas in the north and east of the plateau mostly show decreases in snowfall but higher elevations in the central areas show increases (Figure 5 d). Although differences in temperature are a contributory factor, precipitation dynamics on the plateau may also relate to changes in dominant synoptic circumstances [Yao *et al.*, 2012, 2013], in particular the North Atlantic oscillation (NAO) and mid-latitude westerlies [Liu and Yin, 2001]. There is evidence that the Indian monsoon is weakening [Wu 2005] and the westerlies strengthening [Li and Wang 2003], which would encourage reduced snowfall in the north and east of the plateau and increased snowfall in higher western regions respectively.

The effect of elevation on snowfall trends is also significant with most high/low elevation stations increasing/decreasing respectively. A range of 3000-3400 m currently appears to be the critical threshold elevation dependent on the data and model used to estimate the transition level (Figure 10b). In the Tien Shan Mountains, the snowfall/liquid precipitation ratio also has been shown to have a decreasing trend below 2500 m during 1961-2010 [Guo

and Li, 2015] based on station observations. This is a slightly lower threshold elevation than we report for the whole plateau in this study.

Although elevation is strongly correlated with mean temperature, there may be other confounding factors which create some elevation-dependency in snow trends this case. Most of the highest elevation stations are in the west of the plateau, whereas the lower stations are in the south and east. Thus spatial contrasts in precipitation which could arise if the East Asian Summer monsoon was to weaken [Wu 2005, Zhou et al. 2009, Liu et al. 2012, An et al. 2015] and the mid-latitude westerlies strengthen would also preferentially increase snowfall at the higher elevation stations while reducing it at lower stations in our dataset [Yao et al., 2012]. Many high elevation areas in the west of the plateau receive most of their moisture from westerly storms [Kapnick et al. 2014]. Different precipitation regimes could also therefore be a contributory factor to the snow changes [Berghuijs et al., 2014; Maussion et al., 2014]. Orographic enhancement of precipitation is also known to be stronger in mid-latitude westerly climates because of stronger wind speeds at high elevations, whereas in the Asian monsoonal circulation dominated by convective precipitation, orographic enhancement is reduced [Barry, 2008]. Thus strengthened westerlies over the plateau would also encourage a tendency towards an enhanced snowfall elevational gradient similar to our observations. Further work on the synoptic drivers of our observed changes, and whether they work in conjunction with temperature changes, is required.

Elevation also has a mechanical impact on the landing speed of snowfall droplets [Ding et al., 2014]. At high elevations the reduced air density provides reduced frictional resistance to droplet terminal velocity and as such all droplets have an increased likelihood of reaching the ground as snow at slightly higher temperatures than at low elevations. However no detailed research has been performed on how this factor may influence future elevational contrasts in snowfall in a warming world.

A caveat is that our method of using temperature to define the snowfall/rain transition means that changes in modelled snowfall will by definition be in part influenced by temperature and

thus it is not surprising that some relationship between snow trends and mean temperature is observed. Despite this the snow trends are still derived from modified precipitation data and not temperature data alone.

We have not analysed how future snowfall will change in this paper, but many model simulations predict large decreases. For example, in the Tibetan Plateau, snowfall is predicted to decrease by 5 cm/decade in the transition seasons (spring and autumn) but increase by 4 cm/decade in midwinter during 2006-2100 based on a mean of 60 model simulation results under the AR5 RCP4.5 scenario [Krasting *et al.*, 2013]. Meanwhile, in the northwestern Himalaya, snowfall is forecast to increase from CM2.5 model simulation results [Kapnick and Delworth, 2013]. Thus future uncertainty is large. It is therefore critical to understand the reasons behind past contrasts in observed trends to allow more detailed testing and calibration of model simulations.

We have modelled snowfall which is not the same as snow water equivalent (SWE), snow depth, or snowpack. SWE can be recorded by the weight of the overlying snowpack (e.g. from snow pillow data) [Serreze *et al.*, 2001] and is required to quantify moisture content available for runoff. Snow depth is not the same as SWE because the density of snow varies [Mudryk *et al.*, 2015]. However snowfall will have a knock on effect on snowpack which in turn has an important effect on vegetation cover [Wigmosta *et al.*, 1994] and water resources [Berghuijs *et al.*, 2014; Singh and Bengtsson, 2004] on the plateau. Snowfall changes may also amplify future warming trends [Mountain Research Initiative, 2015], particularly where snow cover shows rapid retreat (i.e. at the current critical threshold). Thus future work needs to examine snowfall change in more detail, along with its effects on other related variables such as snow cover, snow depth and SWE.

This paper has concentrated on in situ data from weather stations, in a similar way to numerous other studies [Ding *et al.* 2014, Guo and Li 2015, Zhu *et al.* 2016]. Other sources of data such as satellite records are now available and are potentially useful for examining snow changes on the plateau. However, there are a whole host of issues involved with satellite/in

situ data comparisons, particularly at high elevations [Pepin et al. 2016], including differing spatial resolutions; cloud contamination and conflation with snow cover; differences between snow cover presence/absence (binary), snow covered area (%), SWE and snowfall; point versus pixel comparisons, inter- and intra-pixel variance in elevation; topographic effects; and timing issues (local solar time versus time of in situ data recording).

5 Summary

Our study has summarised changes of snowfall under warming in the Tibetan Plateau between 1960 and 2014 based on surface station data. Despite relatively uniform temperature change, snowfall trends show large spatial variations with about 40.9% of stations showing a decrease and 59.1% of stations showing an increase. The stations with significant decreases (8/71) are in the south of region A and east of region B. Most stations with significant increases (7/71) are in the central plateau. There are differences in snowfall trends between sub-periods. Most increases occur in the early period of 1961-1990 but in later periods (especially 1981-2010) most stations show a decrease in snowfall. Meanwhile, winter (and to a lesser extent spring) tend to show increase in snowfall, but summer (and to a lesser extent fall) tend to show decreases. Temperature is shown to be an important factor controlling these snowfall patterns, because maximum snowfall occurs when air temperatures range between 1 and 2 °C. This critical threshold can explain much of the variation in snowfall trends, including regional, temporal, seasonal and elevational contrasts. Other factors such as synoptic forcing and changes in precipitation regimes are the subject of further research.

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Appendix A T_{max} , T_{min} , and T_w

T_{max} is the temperature value at which the occurrence probabilities of rain and sleet are equal to each other, and T_{min} is that for the equal occurrence probabilities of snow and sleet. T_{max} and T_{min} were calculated by following:

$$T_{min} = \begin{cases} T_0 - \Delta S * \ln \left[e^{\left(\frac{\Delta T}{\Delta S}\right)} - 2 * e^{\left(-\frac{\Delta T}{\Delta S}\right)} \right], & \frac{\Delta T}{\Delta S} > \ln 2 \\ T_0, & \frac{\Delta T}{\Delta S} \leq \ln 2 \end{cases} \quad (A.1)$$

And,

$$T_{max} = \begin{cases} 2 * T_0 - T_{min}, & \frac{\Delta T}{\Delta S} > \ln 2 \\ T_0, & \frac{\Delta T}{\Delta S} \leq \ln 2 \end{cases} \quad (A.2)$$

where ΔT and ΔS are the temperature difference and temperature scale, respectively. $\ln 2$ is equivalent to $RH = 78\%$. If $RH > 78\%$, the T_{max} and T_{min} have different values; otherwise, T_{max} and T_{min} are equal. Therefore, the method is a dynamic method. The two parameters were details described in the literature of *Ding et al.* [2014]. Their dependences on RH and is given by:

$$\Delta T = 0.215 - 0.099 * RH + 1.018 * RH^2 \quad (A.3)$$

And,

$$\Delta S = 2.374 - 1.634 * RH \quad (A.4)$$

where RH is daily relative humidity (ranges is from 0 to 1).

And T_0 depends on both elevation and RH is given by:

$$T_0 = -5.87 - 0.104 * Z + 0.0885 * Z^2 + 16.06 * RH + 9.624 * RH^2 \quad (A.5)$$

Where Z is elevation (km) of observation stations.

Daily wet-bulb temperature (T_w) contains air temperature, humidity, and pressure information, and as can be calculated by:

$$T_w = T_a - \frac{e_{sat}(T_a) * (1 - RH)}{0.0006336 * P_s + e'_{sat}(T_a)} \quad (A.6)$$

Where T_a is daily air temperature ($^{\circ}C$), P_s is daily air pressure (h pa), $e_{sat}(T_a)$ is the saturated vapor pressure (h pa) at T_a and is one of the key variables of control evaporation process is given by:

$$e_{sat}(T_a) = 6.1078 * e^{\left(\frac{17.27 * T_a}{T_a + 237.3}\right)} \quad (\text{A.7})$$

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Table 1 Detailed description of each sub-region in Figure 3 b

Sub-regions	Explanation
a+b+c+d	Air temperature and snowfall changes are insignificant
e	Significant temperature increase and snowfall decrease
f	Significant temperature increase and snowfall increase
g	Significant temperature decrease and snowfall decrease
h	Significant temperature decrease and snowfall increase
i	Significant air temperature increase, snowfall insignificant
j	Significant air temperature decrease, snowfall insignificant
k	Air temperature insignificant, significant snowfall decrease
l	Air temperature insignificant, significant snowfall increase

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Table 2 Statistical parameters describing the distribution of temperature, precipitation, rain, and snowfall over different sub-periods

Variables	Periods	μ	$\mu \pm 1.96\delta$
Air temperature (°C)	1961-1990	3.1556	[-4.9095 11.2208]
	1971-2000	3.3642	[-4.6687 11.3971]
	1981-2010	3.7476	[-4.2221 11.7174]
	1991-2014	4.0379	[-3.8792 11.9549]
Precipitation (mm)	1961-1990	456.8575	[-41.8149 955.5298]
	1971-2000	463.6866	[-48.0666 975.4397]
	1981-2010	471.0736	[-36.6172 978.7643]
	1991-2014	473.8116	[-22.3651 969.9884]
Rain (mm)	1961-1990	399.6065	[-82.7192 881.9323]
	1971-2000	405.3381	[-93.6833 904.3594]
	1981-2010	412.8771	[-81.1378 906.8921]
	1991-2014	417.0333	[-64.9775 889.0441]
Snowfall (mm)	1961-1990	5.5120	[0.4669 10.5572]
	1971-2000	5.5281	[0.4542 10.6062]
	1981-2010	5.8645	[0.3659 11.3632]
	1991-2014	5.8947	[0.1504 11.6390]

μ is mean value, δ is standard deviation

Table 3 Frequency of different patterns of temperature and snowfall changes for different periods based on all stations. See Table 1 for definition of columns.

Periods	a+b+c+d	e	f	g	h	i	j	k	l
1961-1990	33	1	8	0	2	25	0	1	1
1971-2000	26	1	2	0	0	37	1	1	3
1981-2010	3	8	0	0	0	60	0	0	0
1991-2014	12	5	1	0	0	53	0	0	0

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Table 4 Frequency of different patterns of temperature and snowfall at all stations for each season during 1961-2014. See Table 1 for definition of columns.

Periods	Seasons	a+b+c+d	e	f	g	h	i	j	k	l
1961-2014	DJFM	1	2	19	0	1	48	0	0	0
	AM	11	1	12	0	0	44	1	1	1
	JJAS	4	16	0	0	0	50	0	0	0
	ON	15	11	1	0	0	44	0	0	0
1961-1990	DJFM	11	1	18	0	1	31	0	1	8
	AM	39	2	3	1	0	1	0	0	25
	JJAS	52	0	0	0	1	9	2	0	7
	ON	57	0	0	0	1	3	4	0	6
1971-2000	DJFM	46	0	2	0	1	10	0	0	12
	AM	45	1	2	0	0	2	1	1	19
	JJAS	27	1	4	0	0	37	0	1	1
	ON	43	1	0	0	0	22	0	1	4
1981-2010	DJFM	9	1	22	0	0	38	0	0	1
	AM	2	1	7	0	0	60	0	0	1
	JJAS	10	3	0	0	0	58	0	0	0
	ON	11	5	0	0	0	55	0	0	0
1991-2014	DJFM	17	3	0	0	0	50	0	1	0
	AM	52	4	0	0	0	7	0	8	0
	JJAS	11	13	0	0	0	47	0	0	0
	ON	42	6	0	0	0	15	0	8	0

Table 5 Regionwide snowfall trend magnitudes (mm/yr) based on the Mann-Kendal trend test for different periods in the Tibetan Plateau, region A, region B, and region C. Results correspond to the gradients of the best fit lines in Figure 11.

	1961-1990	1971-2000	1981-2010	1991-2014
Tibetan Plateau	0.40 ^{**}	0.23	-0.26 [*]	-0.27
Region A	0.25	0.22	-0.07	-0.27
Region B	0.38	0.17	-0.37 ^{**}	-0.33
Region C	0.58 ^{**}	-0.07	0.18	0.48

Note: A Mann-Kendal test was used to detect the trends of snowfall (mm/a). “*” Significant at $p < 0.1$; “**” Significant at $p < 0.05$.

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Table 6 Seasonal snowfall trend magnitudes (mm/yr) for the Tibetan Plateau (1961-2014), region A, region B, and region C. Results correspond to the gradients of the best fit lines on Figure 12.

	Tibetan Plateau	Region A	Region B	Region C
Annual	0.07	0.08	-0.007	0.30**
DJFM	0.08***	0.09**	0.09**	0.07***
AM	0.05	0.10**	0.03	0.05**
JJAS	-0.04***	-0.003***	-0.04***	-0.03
ON	-0.03***	0	-0.04***	-0.05

Note: A Mann-Kendal test was used to detect the trends of snowfall (mm/a). “*” Significant at $p < 0.1$; “**” Significant at $p < 0.05$ and “***” significant at $p < 0.01$.

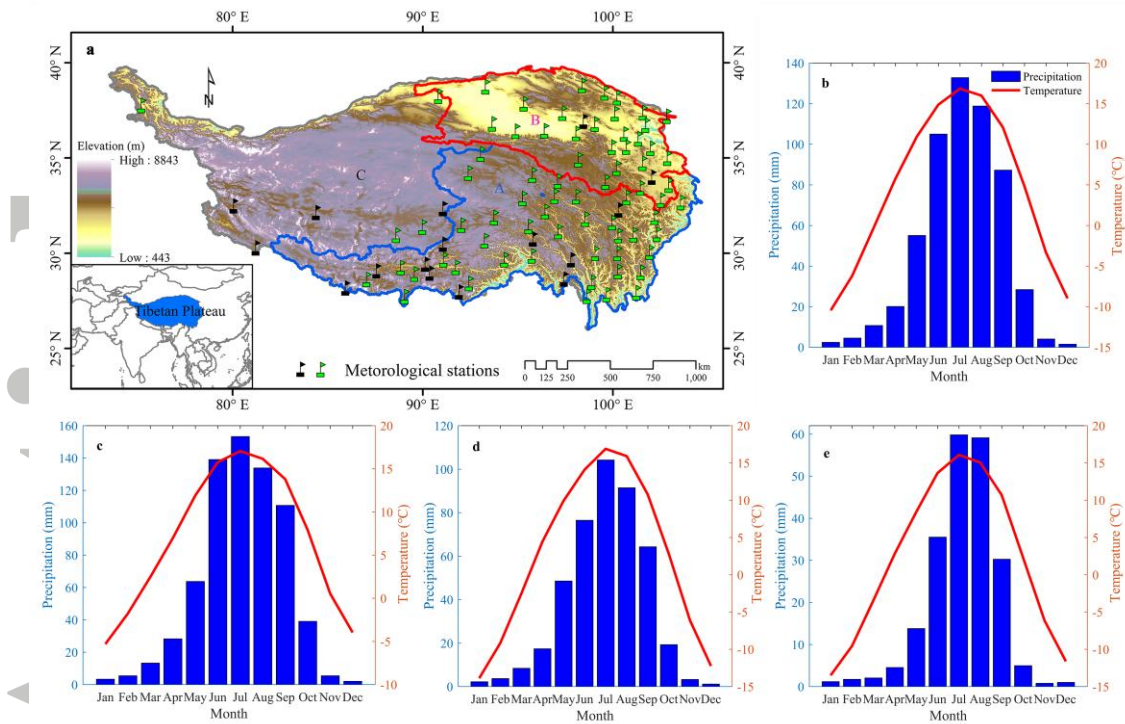


Figure 1 Study area. The Tibetan Plateau was divided into three sub-regions: Region A includes the basins of the Brahmaputra, Salween, Mekong, and Yangtze; Region B includes the Yellow river basin, Hexi corridor, and most of Qaidam basin; Region C includes parts of Qaidam basin and other smaller basins in north-west China. The flag symbols represent meteorological stations. Panels b, c, d, and e show mean monthly temperature and precipitation based on all stations, region A, region B, and region C respectively.

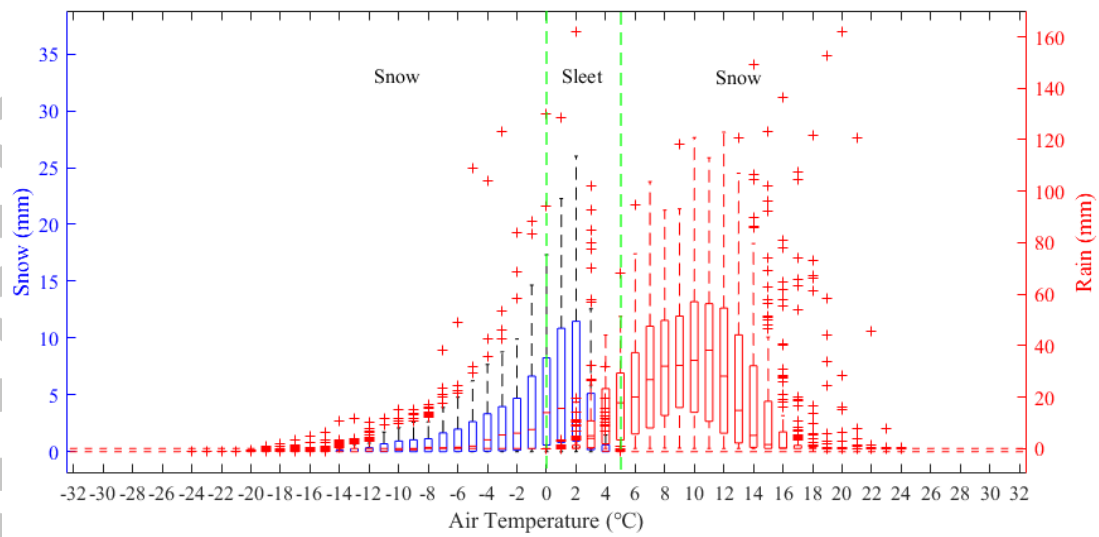


Figure 2. Box and whisker plot of precipitation type (rain vs snow) for different air temperature bands, based on all stations over the Tibetan plateau (1961-2014).

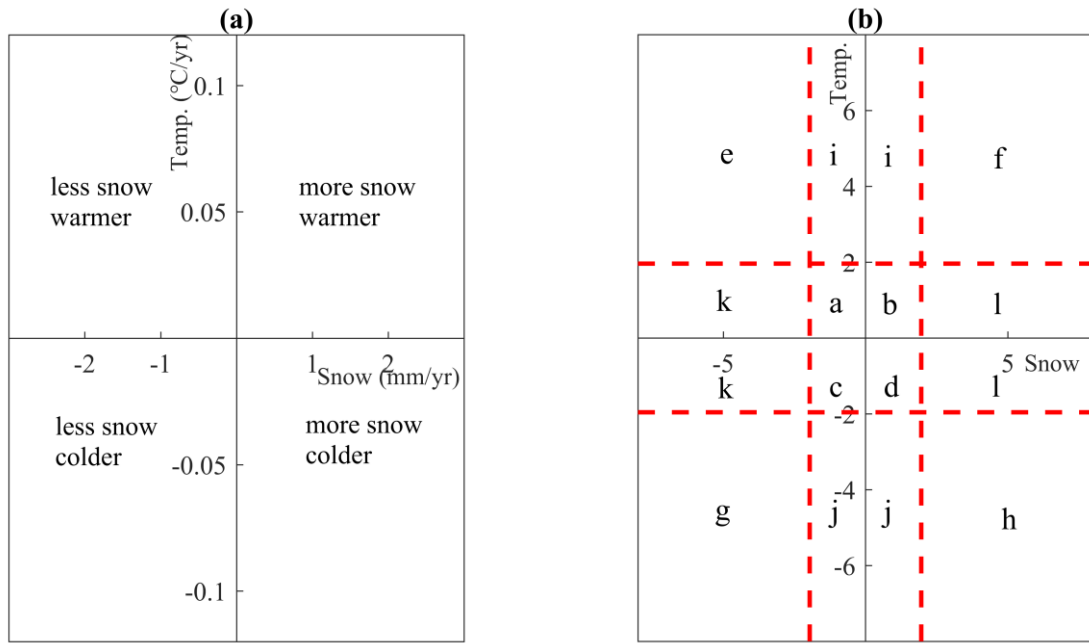


Figure 3 Cross-Quadrant diagram plotting a) air temperature trend magnitude (y axis) vs snow trend magnitude (x axis) and b) temperature trend Z value (y axis) against snowfall trend Z value (x axis). Dotted lines represent the boundaries of statistically significant changes. Areas a-d are insignificant for both variables. These definitions are also used in Tables 3 and 4.

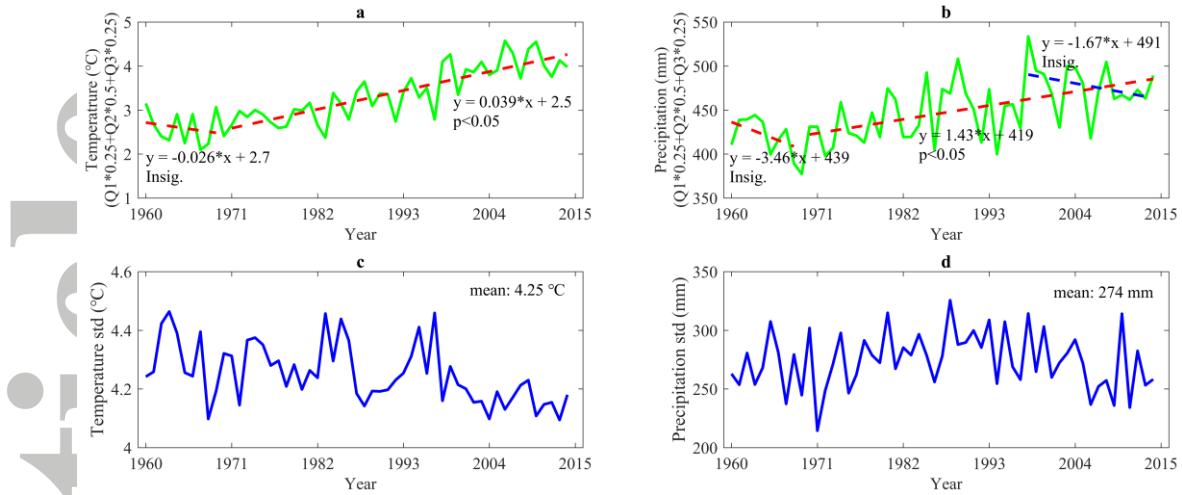


Figure 4. Region-wide air temperature (a) and precipitation (b) over the Tibetan Plateau during 1960-2014. Panels c and d show trends in the standard variances of temperature and precipitation between all stations (the spatial variation of temperature and precipitation across the plateau).

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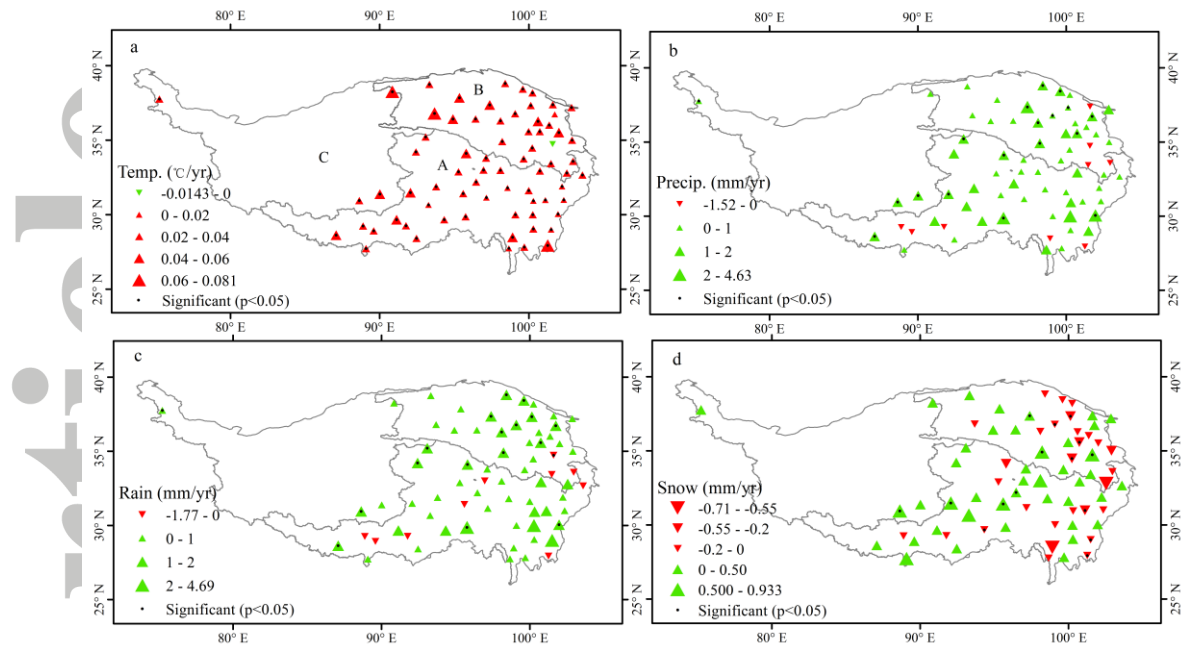


Figure 5 Trends of temperature (a), precipitation (b), rain (c), and snow (d) in the Tibetan Plateau during 1960-2014. Black points show significant trends ($p < 0.05$).

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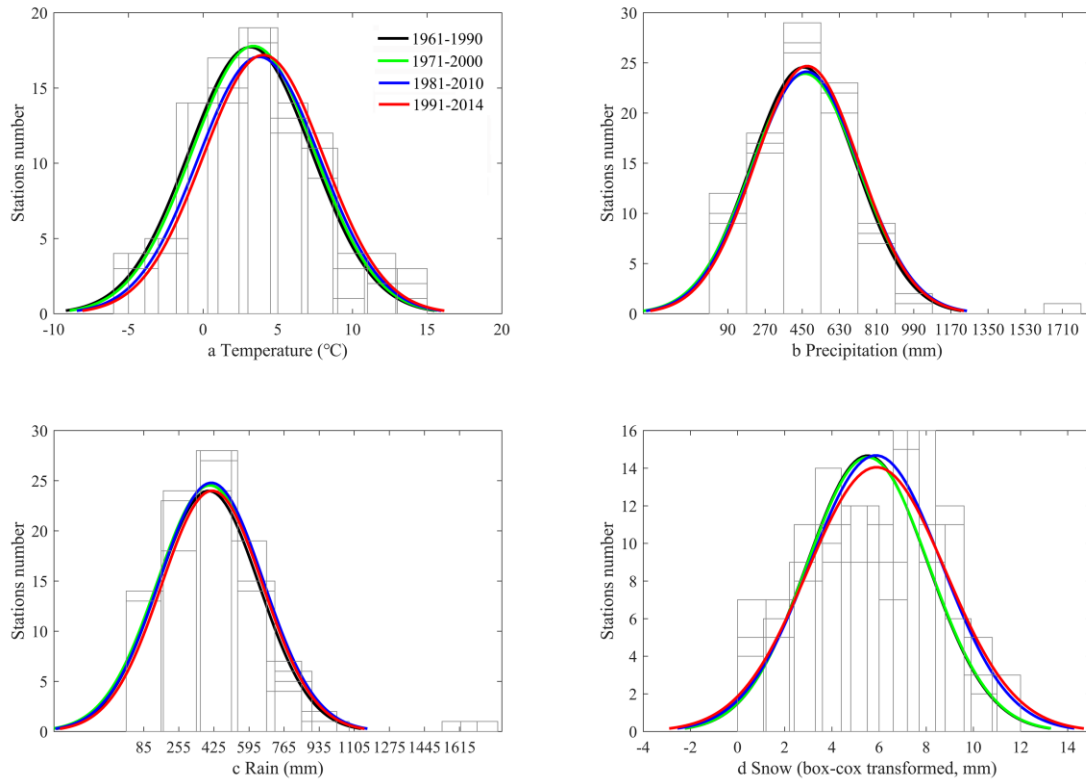


Figure 6 Spatial distributions of temperature (a), snowfall (b), rain (c) and precipitation (d) in different sub-periods ($n=71$). Since snowfall is not normally distributed it has been transformed using the box-cox method.

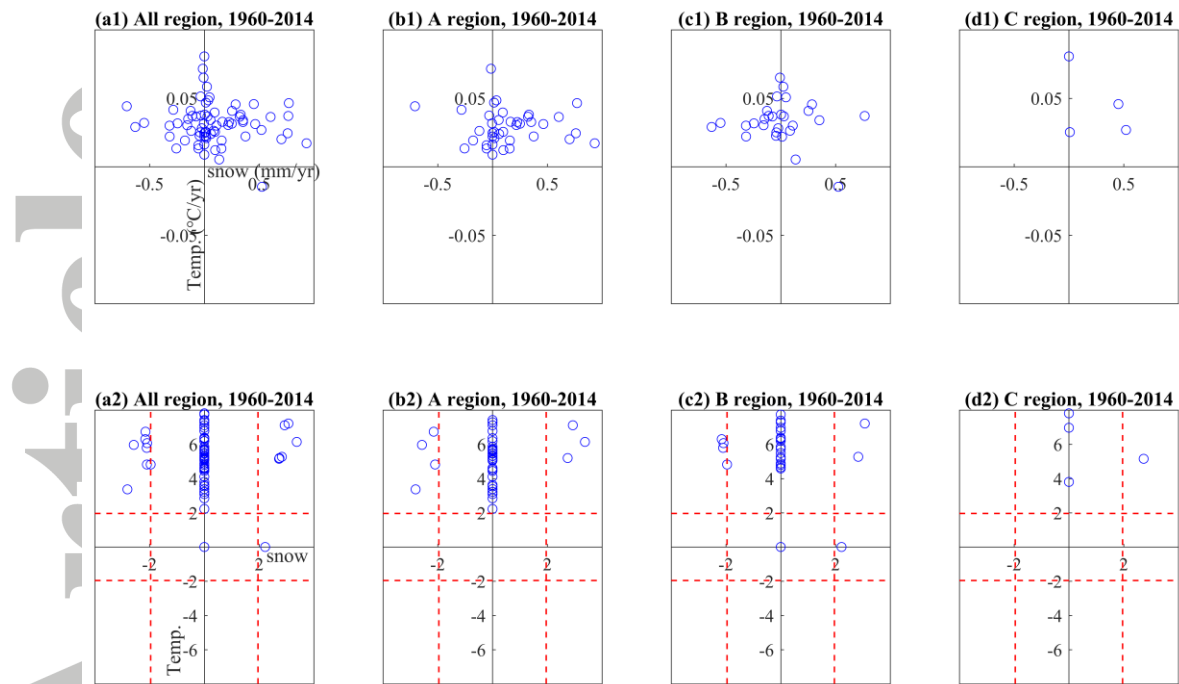


Figure 7 Changes of temperature versus snowfall in the Tibetan Plateau during 1960-2014. The top graphs plot trend magnitudes for temperature ($^{\circ}\text{C}/\text{yr}$) on y axis vs snow (mm/yr) on x axis. The bottom graphs plot Z values for temperature (y axis) vs snow (x-axis) (by M-K trend test). The red dashed lines demarcate significant trends. Regions A, B, and C are shown separately and are identified on Figure 1a.

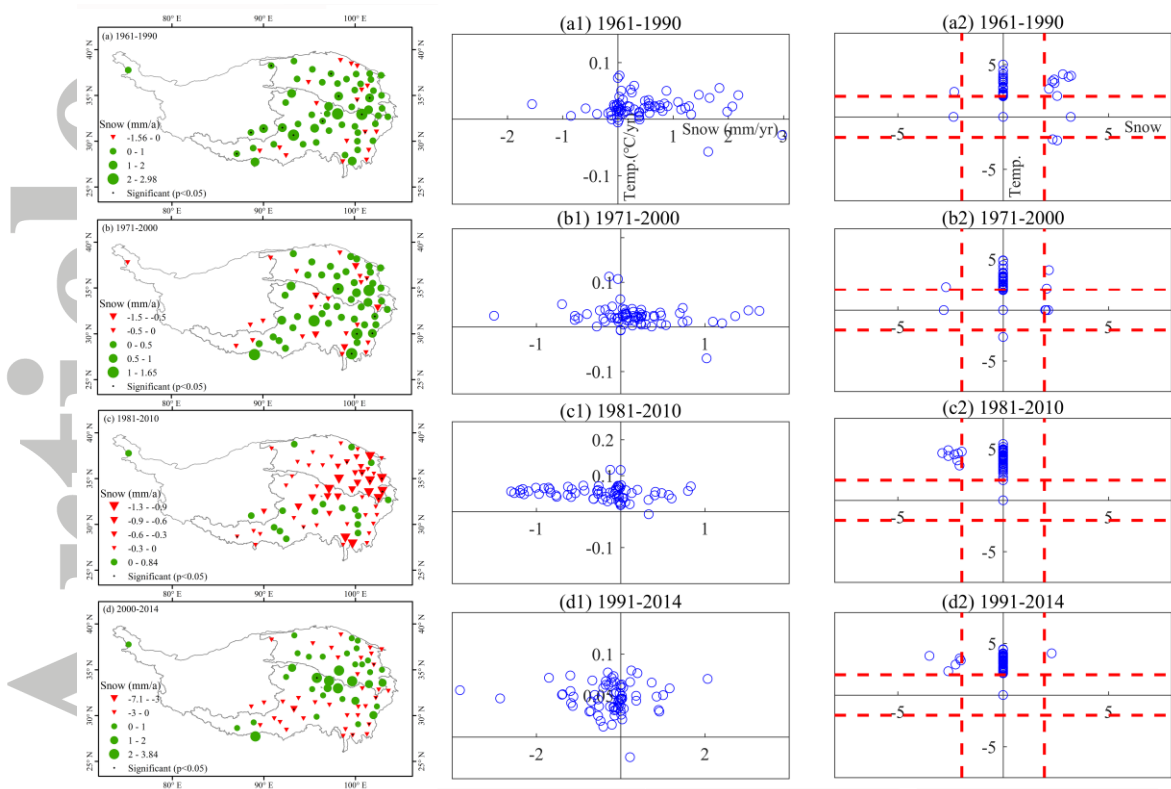


Figure 8 Snowfall trends across the plateau for sub-periods a) 1961-1990, b) 1971-2000, c) 1981-2010 and d) 1991-2014. The left panel shows the spatial variation of trend magnitudes. The centre and right panels show the cross-quadrant diagram (similar to Figure 7) for each sub-period.

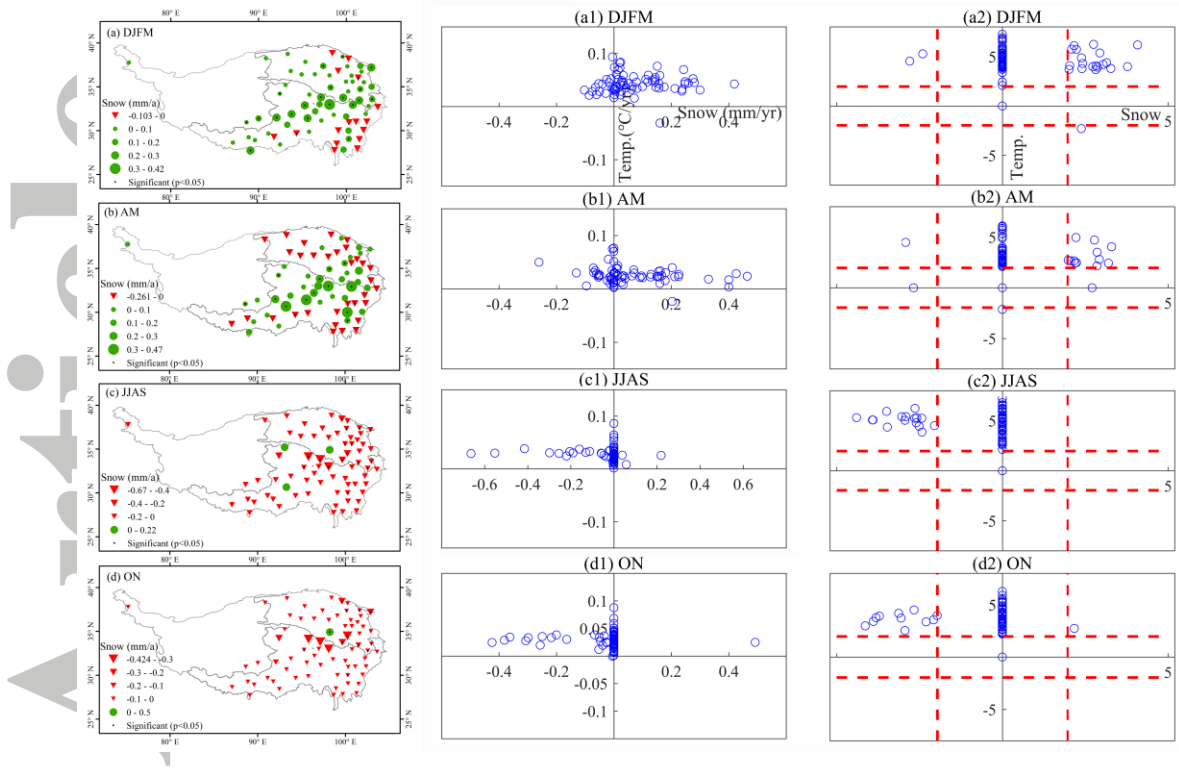


Figure 9 Similar analyses can be performed for different seasons. We define winter as DJFM (usually extensive snow cover), spring as AM (snow-melt), summer as JJAS (relatively little snow) and fall as ON (snow build-up). Figure 9 shows similar snowfall trend magnitudes (left panel) and cross-quadrant diagrams (middle and right panel) for each season (1960-2014).

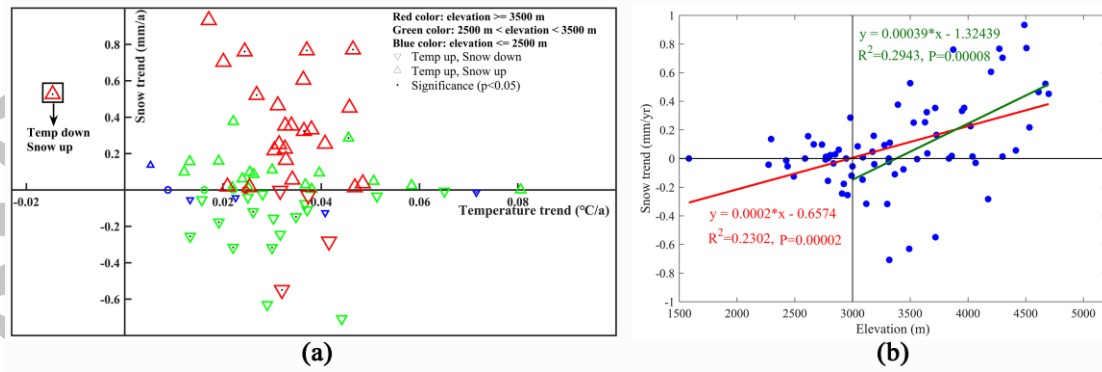


Figure 10 a) Temperature and snowfall trend magnitudes (1960-2014) classified according to station elevation band. Red symbols mean elevation is greater than 3500 m, green symbols between 2500 and 3500 m, and blue below 2500 m. Upward pointing triangles mean warming (temperatures) or more snow, and downward pointing triangles the opposite. Black dots represent significant trends ($p=0.05$), b) Relationship between snowfall trend magnitude (mm/yr) and elevation (m). Critical elevations where snow trends change sign are represented by where lines cross the central x axis (see text for explanation).

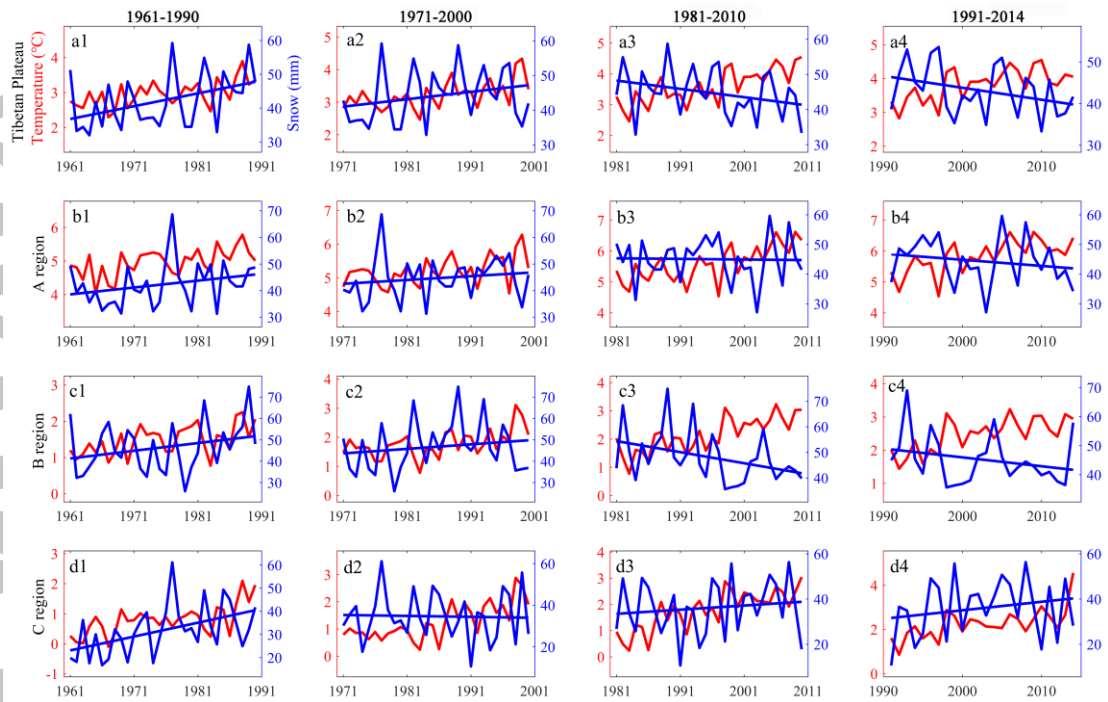


Figure 11 Trends in snowfall for different periods for different regions in the Tibetan Plateau during the past 55 years. From left to right: results for period 1961-1990, 1971-2000, 1981-2010, and 1991-2014. From top to bottom: results for the Tibetan Plateau, region A, region B, and region C. Each region's annual values of air temperature and snowfall are calculated using equation 4.

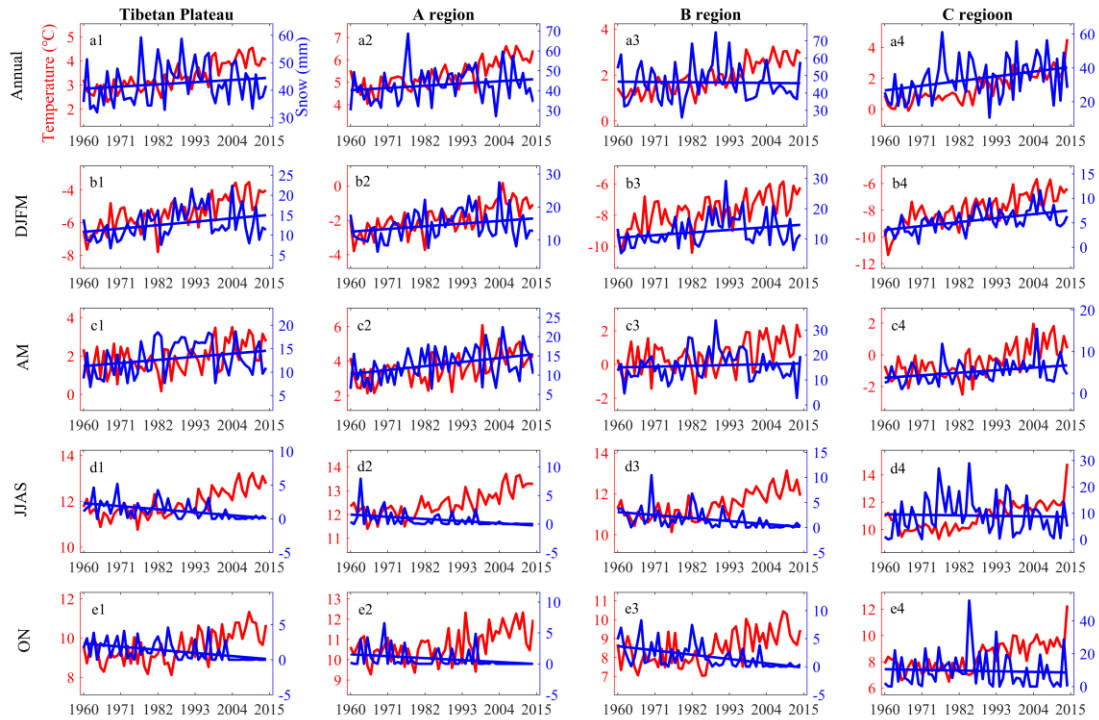


Figure 12 Seasonal snowfall changes in the Tibetan Plateau during the past 55 years. From top to bottom: results for annual, DJFM, AM, JJAS, and ON. From left to right: results for the Tibetan Plateau, region A, region B, and region C. Regional values of air temperature and snowfall are again calculated by equation 4.

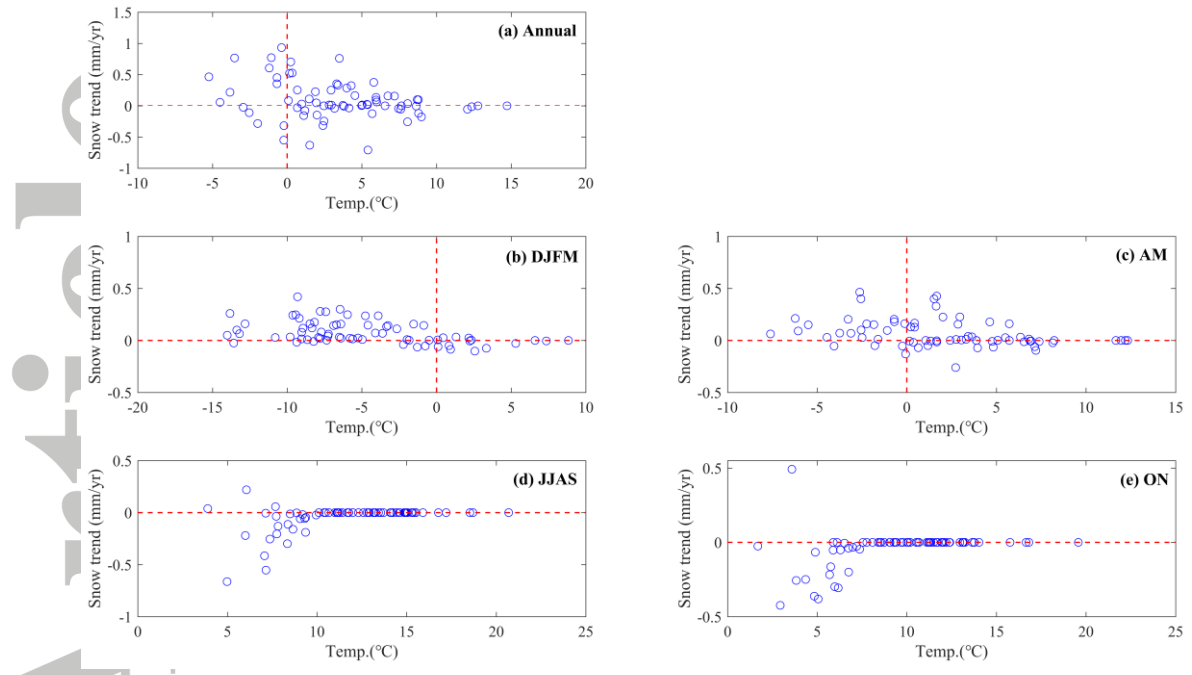


Figure 13 Relationship between mean annual or seasonal temperature and snow trend magnitude (as measured by the Mann-Kendall trend test).

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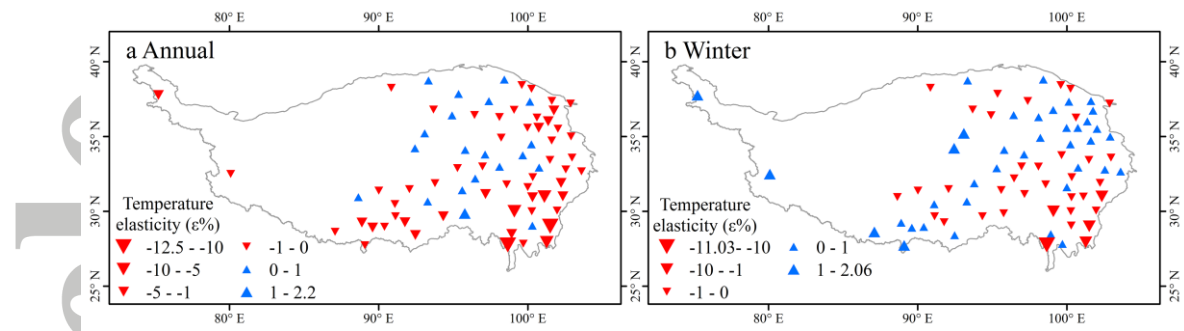


Figure 14 The contribution of temperature change to snowfall changes in the Tibetan Plateau over 1960-2014. For explanation of scale see text. Panels a and b are annual and wintertime respectively.

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