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Review Article

Elevation dependent warming over the Tibetan Plateau: Patterns, mechanisms and perspectives

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ABSTRACT

The Tibetan Plateau (TP) is also known as the “Third Pole”. Elevation dependent warming (EDW), the phenomenon that warming rate changes systematically with elevation, is of high significance for realistically estimating warming rates and their impacts over the TP. This review summarizes studies of characteristics and mechanisms behind EDW over the TP based on multiple observed datasets and model simulations. Spatial expression of EDW and explanatory mechanisms are still largely unknown because of the lack of suitable data over the TP. The focus is on the roles played by known mechanisms such as snow/ice-albedo feedback, cloud feedback, atmospheric water vapor feedback, aerosol feedback, and changes in land use, ozone and vegetation. At present, there is limited consensus on the main mechanisms controlling EDW. Finally, new perspectives and unresolved issues are outlined, including quantification of EDW in climate model simulations, explanation of the long-term EDW reconstructed from proxies, interaction between the Asian summer monsoon and EDW, importance of EDW for future environmental changes and water resources, and current gaps in understanding EDW over extremely high elevations. Further progress requires a more comprehensive ground observation network, greater use of remote sensing data, and high-resolution climate modeling with better representation of both atmospheric and cryospheric processes.

1. Introduction

The Tibetan Plateau (TP hereafter, Fig. 1), mostly lying above 4000 m, is the highest and largest plateau in the world, often called “Third Pole” and “the roof of the world” (Kang et al., 2010; Kang et al., 2019; Pithan, 2010; Qiu, 2008, 2016; Yang et al., 2019; Yao et al., 2019). Due to its high terrain, the TP is important for biodiversity conservation, carbon balance and other environmental issues (Ghatak et al., 2014; Kang et al., 2019; Sun et al., 2018; Yang et al., 2010; Yang et al., 2019; Yao et al., 2019). The TP contains the largest fresh water resource stored in cryosphere (snow, glacier and permafrost) outside the polar regions, and is often referred to as the “Asian water tower”. It is the birthplace of the great rivers of Asia, such as the Yangtze, the Yellow River, the Ganges and Indus Rivers, and is also the source of many inland rivers, providing fresh water for more than 1.4 billion people over the region and South/East Asia as a whole (Duan and Xiao, 2015; Immerzeel and Bierkens, 2012; Immerzeel et al., 2010; Kang et al., 2010; Qiu, 2008; Yao et al., 2012b). Understanding climate changes over the TP also has consequences beyond the immediate environment of the plateau itself. The uplift of the TP enhanced the Asian summer monsoon by enlarging thermal differences between land and sea, and future warming may influence the monsoonal system (Kang et al., 2010; Zhang et al., 2015). Further, the TP provides important ecosystem services, with numerous benefits for people including water

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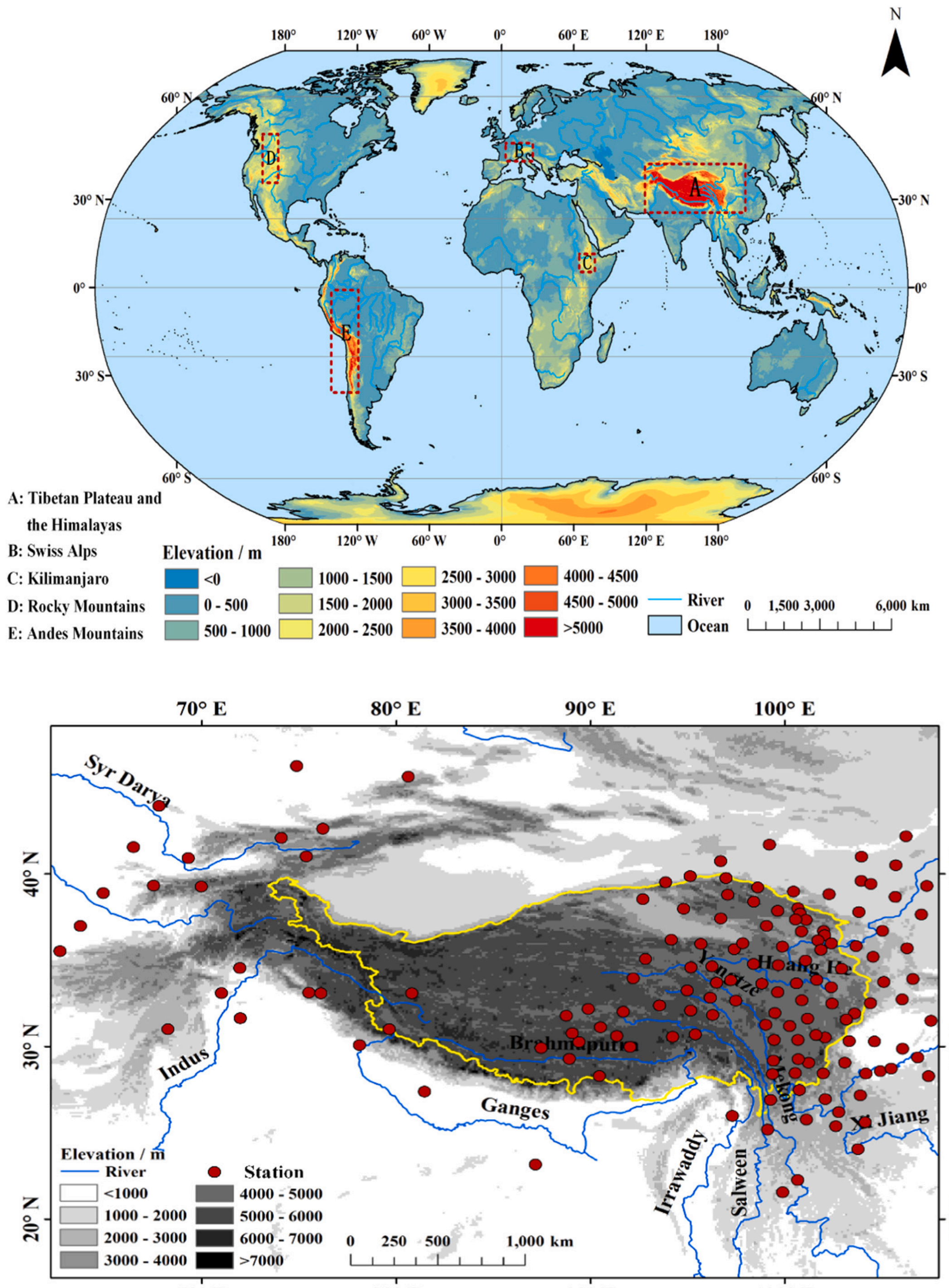

Fig. 1. The distributions of global land surface elevation (top panel). The dashed boxes indicate the five mountain regions considered in this review, which are the Swiss Alps, the Rocky Mountains, the Andes Mountains, the Kilimanjaro, and the Tibetan Plateau and the Himalayas, respectively. The distribution of 165 stations over the Tibetan Plateau and its surrounding areas (bottom panel), which is based on fig. 1 in Liu and Hou (1998). The yellow solid line is the contour of the Tibetan Plateau. The stations in/outside China are provided by the National Meteorological Information Center, China Meteorological Administration/the global historical climatology network (GHCN), respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

resources, tourism, varied ecosystems and their contributions to human well-being (Chen et al., 2015b; Chen et al., 2017; Grizzetti et al., 2016; Qin et al., 2002; Yao et al., 2017a, 2017b, 2017c; Zhang et al., 2015). Broader environmental responses to recent rapid warming, including glacier melt, permafrost degradation, desertification in some areas and flooding in others, changing ecological systems and increased natural disasters such as landslides and glacial lake outburst floods, are known to be occurring to a high degree of confidence over the TP (Kang et al., 2010; Kuang and Jiao, 2016; Thakuri et al., 2016; You et al., 2016; You et al., 2013; You et al., 2017; You et al., 2020). The TP is recognized as a potentially vulnerable area in terms of its ecological environment and future development needs to understand environmental issues such as climate change (Barnett et al., 2005; Gao et al., 2019; Immerzeel and Bierkens, 2012; Immerzeel et al., 2010; Yao et al., 2017b, 2017c).

In recent decades, climate change over the TP and its surrounding areas has become a source of debate, which makes it being one of the hotspots in the global field of geosciences (Chen et al., 2015b; Chen et al., 2017; Ding and Zhang, 2008; Duan et al., 2016; Liu and Hou, 1998; Liu et al., 1998; Liu et al., 2008; Minder et al., 2016; Qin et al., 2002; Wu et al., 2004; Wu et al., 2013; Xu et al., 2015; Yao et al., 2017b, 2017c). Since the turn of the 20th/21st century, a strong warming trend over the TP is evident, synchronizing with the global warming trend (Cuo et al., 2013; Yao et al., 2012a; Yao et al., 2015; Yao et al., 2012b). Significant warming of the TP has been shown through analysis of ground observation stations (Duan and Wu, 2006; Duan and Xiao, 2015; Liu and Hou, 1998; Liu and Chen, 2000; You et al., 2008; You et al., 2013; You et al., 2017), remotely sensed satellite data (Cai et al., 2017; Pepin et al., 2019; Qin et al., 2009), paleoclimate proxy indicators (Kang et al., 2007; Thompson et al., 2018; Yao et al., 2012b) and climate models (Chen et al., 2003; Chen et al., 2015a; Yan et al., 2016; You et al., 2016; You et al., 2020). The warming over the TP has accelerated in the second half of the 20th century (the early 1950s), somewhat earlier than that of the Northern Hemisphere as a whole (the mid-1970s) (Kang et al., 2010; Liu and Hou, 1998; Liu and Chen, 2000). During 1955–1996, the warming rate of the TP was highest in winter and autumn (+0.32 and + 0.17 °C/decade, respectively) (Liu and Chen, 2000). Seasonal variations in warming rate are common, and the warming rate in winter being twice that of the annual average is consistent with findings in other regions (Chen et al., 2003; Kang et al., 2010; Krishnan et al., 2019). A later analysis showed that the annual warming rate ranges from +0.16 to +0.36 °C/decade since the 1950s but +0.50 to +0.67 °C/decade from the 1980s (Kuang and Jiao, 2016). The warming rate has also been spatially heterogeneous across space and time periods. The northern TP experienced more warming than the southern TP in all seasons from 1982 to 1998, while the pattern was reversed in the period from 1998 to 2015 (Liu et al., 2019).

Elevation dependent warming (EDW) refers to the phenomenon that the warming rate changes systematically with elevation. The warming rate does not necessarily increase monotonically with elevation, although it sometimes does (Diaz and Bradley, 1997; IPCC, 2019; Pepin and Coauthors, 2015a; Rangwala and Miller, 2012; Rangwala et al., 2010). Table 1 lists previous studies on EDW including information on time period, elevation range and the number of stations considered in various mountain regions, including examples from the Swiss Alps, Rocky Mountains, Andes Mountains, the Kilimanjaro, and the TP and Himalayas (Fig. 1). EDW could also have a significant impact on the conservation of the cryosphere at high elevations and associated runoff, ecosystem stability and agricultural production. Further, EDW could disproportionately restrict species that are dependent on high elevation habitats for survival. EDW has consequences for changes in the fractional distribution of solid and liquid precipitation (e.g. the elevation of the rain vs snow line) as well as melting rates of glacial deposits, ultimately affecting streamflow and water resources downstream (Bolch et al., 2012; Dimri et al., 2018; Gao et al., 2018; Immerzeel and Bierkens, 2012; Pepin and Coauthors, 2015b). EDW has therefore become an important issue for high mountain areas (Table 1), and studies at global and regional scales show that there are significant spatial differences in its manifestation (Gao et al., 2018; Kotlarski et al., 2012; Liu et al., 2008; Pepin and Coauthors, 2015a; Rangwala and Miller, 2012). Taking the results as a whole, it is clear that the majority of these studies suggest an EDW in both observations and climate models, but with strong seasonality in some regions (Table 1). Existing studies on EDW over the TP have focused on certain aspects of this complex issue (Table 1), due to limitations in surface observations, the relatively recent development of the surface meteorological network over the TP, and uncertainties/limitations associated with methods of quantification.

We review recent studies on EDW over the TP in Section 2. Physical mechanisms are discussed in Section 3. Section 4 summarizes new perspectives and unanswered questions. This review has important theoretical and practical significance, aiming to providing a historical reference for understanding EDW and its consequences over the TP.

2. EDW characteristics over the TP

2.1. Annual and seasonal patterns of EDW in observations

Different conclusions have emerged concerning patterns of EDW over the TP based on in situ observations, which often vary by season and the number of stations (Table 2). One of the earliest analyses (Liu and Hou, 1998) based on 165 stations over the TP and its surroundings for 1961–1990 (Fig. 1 bottom panel), revealed that the annual temperature trends were 0.00, 0.11, 0.12, 0.19 and 0.25 °C/decade for areas < 500, 500–1500, 1500–2500, 2500–3500 and > 3500 m, respectively (Liu and Hou, 1998), providing clear evidence for EDW (Fig. 2). This broad amplification of warming was confirmed in some later studies. During 1955–1996, based on 97 stations above 2000 m, the warming rates of temperature over the TP increased by 0.16 °C/decade for the annual mean, which exceeded the mean rates for the Northern Hemisphere and the same latitudinal zone in the same period (Liu and Chen, 2000). Furthermore, climate warming revealed by 178 stations during 1961–1990 over the TP showed clear EDW, especially above 2000 m, although the number of stations at higher elevations was relatively small (Liu; Chen 2000) (Fig. 3a). EDW based on the annual mean temperature is confirmed through recent analysis of 144 stations during 1961–2010 (Wei and Fang, 2013) (Fig. 3b and e), 122 stations during 1961–2012 (Pepin et al., 2015a) (Fig. 3c), 150 stations with elevations over 2000 m over the TP (Fig. 3d), 150 stations with elevations over 2000 m over the TP (Fig. 3d) and 139 stations during 1961–2012 over the TP (Yan and Liu, 2014) (Fig. 3f). Observed EDW is also pronounced for annual minimum temperature (Liu et al., 2009; Yan and Liu, 2014) (Fig. 4a and b) and maximum temperature (Fig. 4c), although it is weaker for maximum temperature (Fig. 4c).

On a seasonal scale, the studies of EDW over the TP are limited. Based on 144 stations with a 250 m elevation interval during 1961–2010 (Wei and Fang, 2013), the warming trends from mean temperature increase with elevation for each season, suggesting EDW over the TP is remarkably high (Fig. 5a). Average linear trends of seasonal mean temperature, minimum temperature and maximum temperature based on 139 stations during 1961–2012 (Yan and Liu, 2014) also offer ample evidences of EDW, with more robust for minimum temperature, followed by mean temperature, and is weakest for maximum temperature (Fig. 5b, c and e). EDW is most notable in winter, followed by spring and autumn, and is weakest in summer (Fig. 5b, c and e), while the rate of increase for maximum temperature in spring exhibits no clear EDW over the TP (Yan and Liu, 2014). A further analysis based on 116 weather stations during 1961–2006 (Liu et al., 2009) confirmed the presence of EDW in trends of monthly mean minimum temperature in each season, with more pronounced EDW in winter and spring (Fig. 5d).

Clear observed EDW (Figs. 3, 4, 5) over the TP is not evident in more regional studies, reporting negative EDW (reduced warming at the highest elevations) is also evident over some regions of the TP based on the observations (Cuo et al., 2013; Du et al., 2019; Guo and Wang, 2012; Lu et al., 2010). Using 65 meteorological stations during 1951–2014 (Du et al., 2019), the TP was divided into three latitude zones: < 30°N, 30°N-35°N and > 35°N and negative EDW occurred in all three latitude zones when considered separately. In another analysis based on 72 stations with elevations ranging from 2111 m to 4700 m during 1961–2007, the most significant warming (mean annual air temperatures) occurred at relatively low elevations over the northern TP, indicating negative EDW in this sub-region (Guo and Wang, 2012). This finding has been supported by a more recent study based on 81 stations during 1957–2009 over the northern TP (Cuo et al., 2013). By using observations from 140 stations during 1960–2005, it is also found that warming in the high altitude area of the TP is less sensitive to the

Table 1
Recent observational studies on elevation dependent warming across global and regional mountain areas, including information on time period, elevation range and the number of stations, updated from a previous review (Rangwala and Miller, 2012).

Regions	Area	Time period	Elevation range (m)	Number of stations	Reference
Global scale	Global	20th century	1055–3310	126	Diaz and Bradley (1997)
	Global	1948–2002	500–4700	1084	Pepin and Seidel (2005), Pepin and Lundquist (2008)
	Global	1961–2010	Not available	2367	Fan et al. (2015)
European mountains	Swiss Alps	1979–1993	271–3572	88	Beniston and Rebetez (1996)
	Swiss Alps	1901–1999	271–3572	19	Jungo and Beniston (2001)
	Italian Alps	1975–2010	84–2125	28	Tudoroiu et al. (2016)
	Switzerland	1959–2008	203–3580	2 km gridded data	Ceppi et al. (2012)
	Switzerland	1981–2017	203–3580	93	Rottler et al. (2019)
North American mountains	Rocky Mountain Front Range	1953–2008 and 1989–2008	1611–3749	5	McGuire et al. (2012)
	Colorado Rockies	1987–2006	1250–4000	4 km gridded data	Diaz and Eischeid (2007)
	San Juan mountains in Colorado Rockies	1895–2005	1763–3536	58	Rangwala and Miller (2010)
	Front Range of the Rocky mountain	1952–1998	1059–3749	3	Pepin and Losleben (2002)
	Cariboo mountain	1950–2010	330–3520	Gridded data of five arc minutes	Sharma and Déry (2016)
South American mountains	Tropical Andes (1°N–23°S)	1939–1998	0–5000	268	Vuille and Bradley (2000)
	Tropical Andes	1950–1994	0–5000	277	Vuille et al. (2003)
	Andes (1.3°N–62.5°S)	1950–2010	0–4800	626	Vuille et al. (2015)
African mountains	Kilimanjaro	2004–2015	990–5803	22	Pepin et al. (2016)
Asian mountains	Tibetan Plateau	1961–1990	0–5000	165	Liu and Hou (1998)
	Tibetan Plateau	1955–1996	200–4801	197	Liu and Chen (2000)
	Tibetan Plateau	1961–2000	1591–4670	63	Chen et al. (2006)
	Tibetan Plateau	1961–2003	2000–4500	66	Liu et al. (2006)
	Tibetan Plateau	1961–2006	0–5000	116	Liu et al. (2009)
	Tibetan Plateau	1960–2005	1000–5000	140	Lu et al. (2010)
	Tibetan Plateau	2000–2006	2000–5000	71	Qin et al. (2009)
	Tibetan Plateau	1961–2000	1000–5000	43	Rangwala et al. (2009)
	Tibetan Plateau	1961–2005	2000–4700	71	You et al. (2016), You et al. (2008)
	Tibetan Plateau	1961–2004	2000–4700	71	You et al. (2010)
	Tibetan Plateau	1961–2010	Not available	144	Wei and Fang (2013)
	Tibetan Plateau	1951–2014	2750–4900	65	Du et al. (2019)
	Northern Tibetan Plateau	1961–2007	2111–4700	72	Guo and Wang (2012)
	Tibetan Plateau	1961–2013	2000–4700	71	You et al. (2017)
	Northwestern Himalayas	1901–1989	1200–3800	10	Bhutiyani et al. (2007)
	Indian Himalayas	1901–2007	Not available	121	Kothawale et al. (2010)
	Nepal Himalayas	1971–1994	72–3705	49	Shrestha et al. (1999)
	Southern central Himalaya	1976–2015	72–2566	58	Thakuri et al. (2019)
	South slopes of central Himalaya	1994–2013	2660–5600	7	Salerno et al. (2015)
	South slopes of central Himalaya	2007–2011	2660–5600	5	Yang et al. (2018)

globe than the low-altitude neighboring area both in onset time and magnitude (Lu et al., 2010). Local environmental changes such as urbanization and land cover changes were suggested to be at least partly responsible. This highlights clear regional differences in EDW across the TP. Although enhanced warming at higher elevations may occur for the plateau as a whole over long time periods, it may not be valid for some sub-regions within it and/or over shorter timescales.

Although there is a fairly strong consensus on the rapid warming of the TP and its surrounding areas, no significant EDW and/or the largest warming rates at an elevation are observed in some cases (Guo et al., 2019; Salerno et al., 2015; You et al., 2008; You et al., 2010). There is no significant EDW based on indices of temperature extremes from 71 stations during 1961–2005 (You et al., 2008). The lack of a simplistic elevation increase of warming rates also agrees with results from 7 stations located between 2660 and 5600 m over the southern slopes of Mt. Everest, which showed no significant EDW in minimum temperature and maximum temperature trends (Salerno et al., 2015). Other studies revealed the largest warming rates at an intermediate elevation near 4500 m (Guo et al., 2019) and in association with the 0 °C isotherm (You et al., 2008; You et al., 2010).

To summarize, although there are many studies that demonstrate enhanced warming at higher elevations, others have shown opposing patterns or no significant EDW (Table 2). The scarcity of observed data over the higher elevations of the TP makes accurate mapping of the distribution of the EDW challenging.

2.2. EDW from multiple reanalysis products

Lack of ground-based observations across much of the plateau has motivated the use of multiple reanalyses to investigate climate change over the TP. The extent to which different reanalyses have been used to study EDW varies and there are relatively few comparative analyses of multiple reanalysis products. In order to test their ability to reproduce surface observations, six reanalysis products (i.e., MERRA, NCEP1, CFSR, ERA-40, ERA-Interim, and GLDAS) were evaluated using 63 weather stations over the TP during 1992–2001 and using data from 9 field campaigns during 2002–2004. ERA-Interim showed the best overall performance in both daily and monthly air temperatures, while MERRA also had a high correlation with observations during 1992–2001 (Wang and Zeng, 2012). Another validation study using three reanalysis products (NCEP2, ERA-Interim, and JRA-25) in comparison with satellite and radiosonde observations over the TP during 1979–2010 showed ERA-Interim and JRA-25 to be more reliable than NCEP2 (Zou et al., 2014). ERA-Interim data indicate dramatic warming over the TP during 1979–2010, with warming rates of 0.33 °C/decade in annual mean temperature, 0.22 °C/decade in summer and 0.47 °C/decade in winter, and EDW is also more pronounced in ERA-Interim than in other reanalysis products (Zou et al., 2014). In contrast, during

Table 2
Summary of published results on elevation dependent warming (EDW) over the Tibetan Plateau and its surroundings. EDW is investigated based on the mean temperature (Tmean), maximum temperature (Tmax) and minimum temperature (Tmin) using ground-based observations and other datasets such as reanalysis products, remote sensing data and climate model simulations.

EDW phenomenon	Ground-based observations			Other datasets		
	T_{min}	T_{max}	T_{mean}	T_{min}	T_{max}	T_{mean}
Positive EDW	➢ Annual (Liu et al., 2009; Pepin et al., 2015b; Yan and Liu, 2014) ➢ All seasons (Liu et al., 2009; Yan and Liu, 2014)	➢ Annual (Shrestha et al., 1999; Thakuri et al., 2019; Yan and Liu, 2014) ➢ All seasons (Yan and Liu, 2014)	➢ Annual (Liu and Hou, 1998; Liu and Chen, 2000; Pepin et al., 2015a; Wei and Fang, 2013; You et al., 2017) ➢ All seasons (Liu and Hou, 1998; Wei and Fang, 2013; Yan and Liu, 2014) ➢ Autumn (Pepin et al., 2015a) ➢ Winter (Pepin et al., 2015b)	➢ Annual (Liu et al., 2008; Liu et al., 2009) ➢ All seasons (Liu et al., 2009) ➢ Winter (Rangwala et al., 2016) ➢ Autumn (Palazzi et al., 2019)	➢ Annual (Liu et al., 2008) ➢ Autumn (Palazzi et al., 2019)	➢ Annual (Dimri et al., 2018; Liu; Chen 2000; Liu et al., 2008; Yan et al., 2016) ➢ All seasons (Dimri et al., 2018; Yan et al., 2016) ➢ Winter (Chen et al., 2003; Gerlitz et al., 2014; Rangwala et al., 2010) ➢ Spring (Gerlitz et al., 2014; Rangwala et al., 2010)
Negative EDW	➢ Annual (Cuo et al., 2013)	➢ Annual (Cuo et al., 2013)	➢ Annual (Cuo et al., 2013; Du et al., 2019; Guo and Wang, 2012; Lu et al., 2010)	not available	not available	➢ Annual (Gerlitz et al., 2014)
No significant EDW	➢ Annual (Salerno et al., 2015)	➢ Annual (Salerno et al., 2015)	➢ Annual (Du et al., 2019; Salerno et al., 2015; You et al., 2008; You et al., 2010)	Not available	Not available	➢ Annual (You et al., 2010)
No significant EDW but largest warming rates at an intermediate elevation	Not available	Not available	➢ Annual (Guo et al., 2019; You et al., 2008; You et al., 2010)	➢ Annual (Palazzi et al., 2017)	➢ Annual (Palazzi et al., 2017)	➢ Annual (Gao et al., 2018; Guo et al., 2019; Pepin et al., 2019; Qin et al., 2009)

the post-monsoon season, negative EDW is found in the elevation-bias corrected ERA-Interim over the TP (Gerlitz et al., 2014).

A comparison between trends based on in-situ temperature at 71 homogenized surface stations above 2000 m over the eastern and central TP and trends at the 56 equivalent grid points from surface NCEP1 and ERA-40 reanalyses showed that ERA-40 was the best at capturing the spatial pattern of warming trends, especially in winter. However, no systematic EDW over the TP was shown in either ERA-40 or NCEP1 (You et al., 2008; You et al., 2010). Fig. 6 shows the spatial trends and patterns of EDW from gridded-observations and multiple reanalysis products (NCEP1, NCEP2, ERA-Interim, MERRA and JRA-55) during 1979–2018 over the TP. Each reanalysis of surface temperature (NCEP1, NCEP2, ERA-Interim, MERRA and JRA-55) can produce warming patterns over the TP, and capture the surface warming in categorized elevation bands to some extent (Fig. 6 top panels). Both NCEP1 and NCEP2 exhibit greater warming at higher elevations, suggesting they can reproduce EDW consistent with the gridded-observations over the TP. ERA-Interim, MERRA and JRA-55 however fail to capture EDW over the TP (Fig. 6 bottom panels). Overall, application of reanalysis products over the TP requires attention to their capability to reproduce realistic EDW rates.

2.3. *EDW from MODIS remote sensing data*

Satellite data has the advantages of extensive regional coverage and relatively high spatial resolution, and thus has been frequently used to detect EDW across different regions of the TP. MODIS (Moderate Resolution Imaging Spectroradiometer) is particularly useful providing data through Terra and Aqua Platform. MODIS Land Surface Temperature (LST) at 1 km resolution is the most common data source, and validation studies indicate it reasonably describe warming rates consistent with ground-based observations over the TP (Dimri et al., 2018; Guo et al., 2019; Qin et al., 2009; Wan et al., 2002; Zhang et al., 2018a; Zhang et al., 2018b). The earliest analysis (Qin et al., 2006) was restricted to 2000–2006 indicating strongest warming rates between 3000 m and 4800 m (Fig. 7a). However, the length of record at the time was too short for reliable warming trend analysis (Fig. 7a). Above this level, warming rates became quite stable with a slight decline near the highest elevations. In a more recent study using LST data (corrected to more closely represent air temperature at 2 m) during 2002–2017, maximum warming was recorded around 4500–5500 m in the NyenchenTanglha range, consistent with snowline retreat (Pepin et al., 2019). No strong EDW pattern was found in the Qilian Mountains, and there was a stabilization of warming at very high elevations in the Himalaya, including cooling above 6000 m during 2002–2017 (Pepin et al., 2019)**.** Based on recent comprehensive satellite-based datasets during 2001–2015, negative EDW in annual mean surface temperature warming occurred above 4500 m for the TP as a whole, but between 2000 m and 4500 m warming increased with elevation (Guo et al., 2019) (Fig. 7b). However, the amount of data at extremely high elevations (> 6000 m) is limited and often contaminated by clouds (particularly in the southern and eastern regions of the TP), which increases the uncertainty associated with any stabilization of warming at the highest elevations.

2.4. *EDW from climate model outputs*

EDW over the TP has also been examined in global and regional climate model simulations, and there are differences in EDW profiles across different climate models. EDW over the TP is projected to continue under future climate change scenarios (Liu and Hou, 1998; Liu et al., 1998; Liu et al., 2008; Liu et al., 2009). It is noted that annual surface mean temperature differences between a doubled CO_2 concentration experiment and the 1 × CO_2 experiment (control run) from the Geophysical Fluid Dynamics Laboratory (GFDL, 2.25° × 3.75°), the Canadian Climate Center (CCC, 3.75° × 3.75°) and the UK

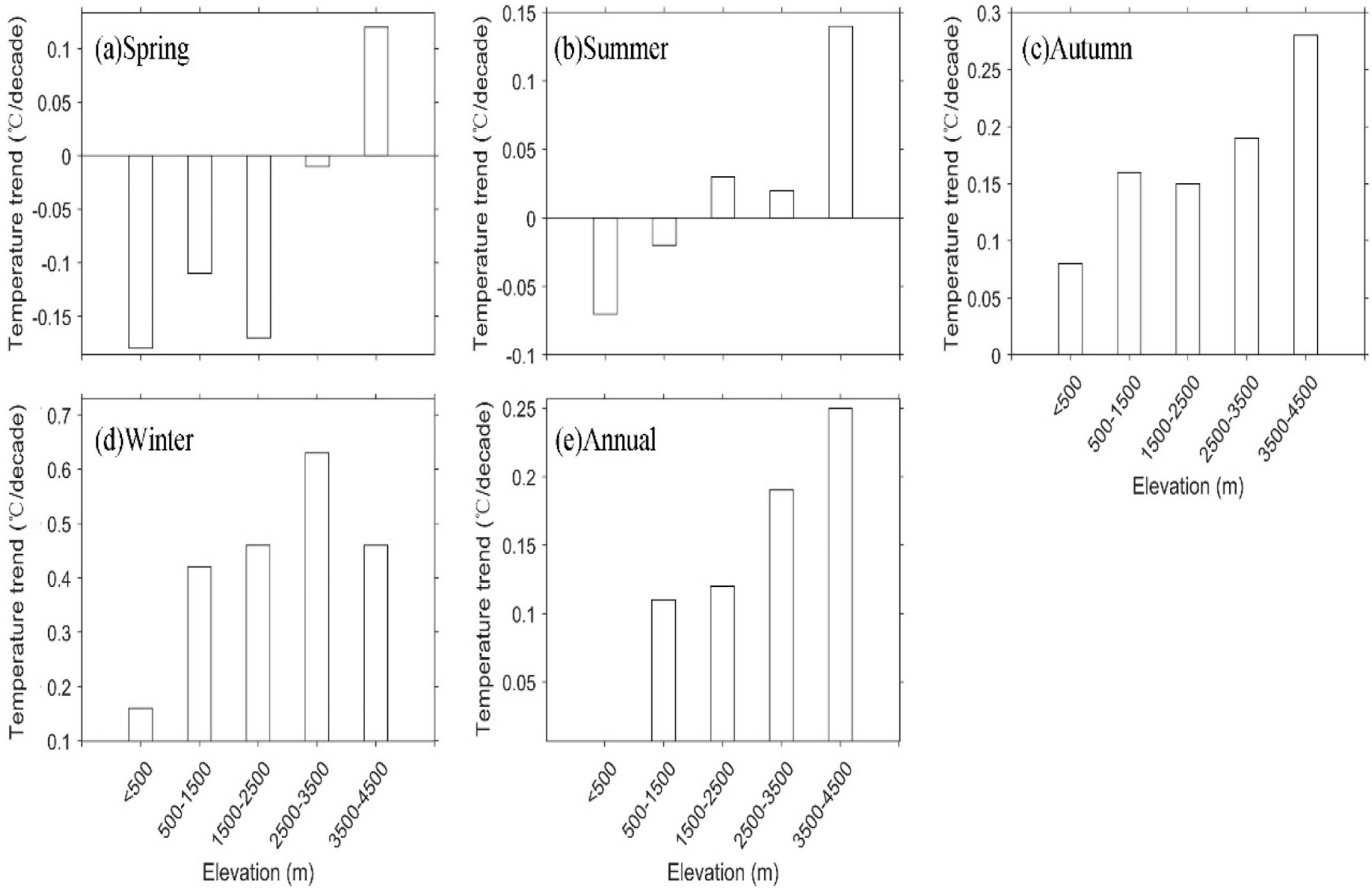

Fig. 2. Rates of annual and seasonal surface mean temperature increase over the Tibetan Plateau and its surrounding areas for different elevation bands from 1961 to 1990 (°C/decade) adopted from table 1 in Liu and Hou (1998). The numbers of stations for elevation bands < 500, 500–1500, 1500–2500, 2500–3500 and > 3500 m are 34, 37, 26, 38 and 30, respectively.

Meteorological Office (UKMO, 2.5° × 3.75°) climate models over the TP fails to reproduce EDW (Fig. 8), suggesting that climate models with relatively coarse spatial resolutions are not suitable for investigating EDW over the TP (Liu and Chen, 2000). Differently, EDW was confirmed in high resolution numerical simulations of mountain climate using doubled CO_2 concentration (Chen et al., 2003; Liu et al., 1998). Using high-resolution (T85, 140 km horizontal resolution) climate model output from the 1% per year CO_2 increasing experiment with the fully coupled Community Climate System Model (CCSM 3), developed by the National Center for Atmospheric Research (NCAR), averaged for various elevation ranges over the TP (Liu et al., 2008), EDW in annual and winter surface mean temperatures were found to be significant (Fig. 9). Based on the regional climate model of the NCAR Community Climate System Model (CCSM3) during three 20-year mean periods (1980–1999, 2030–2049 and 2080–2099) for the IPCC mid-range emission (A1B) scenario and using an analysis of elevation zones of 500 m interval (Liu et al., 2009), presence of EDW of the mean minimum temperatures over the TP was occurred on an annual and seasonal basis (Fig. 10).

Using regional climate model (Weather Research and Forecasting WRF) simulations over the TP for RCP 4.5 and RCP 8.5 global emission scenarios, no EDW is projected above 5000 m in both the near-term (2016–2035) and long-term (2080–2100) (Gao et al., 2018). Maximum warming is projected around 5000 m, casting doubt on whether future enhanced warming will extend up to high elevations (Gao et al., 2018). Similarly, an experiment using the MIROC5/WRF simulations shows that the warming rates first increase and then decrease with increasing elevation for two future periods (2006–2050 and 2006–2099) under the RCP6.0 scenario (Fig. 11), suggesting that the elevations with the largest warming rates are 4600–4800 m and 5000–5200 m for the periods 2006–2050 and 2006–2099, respectively (Guo et al., 2016).

Investigating EDW based on the gridded-observations, 21 global climate models participating in the Coupled Model Intercomparison Project phase 5 (CMIP5) and their multi-model ensemble mean over the TP (historical run 1979–2005) (Fig. 12), revealed that most CMIP5 models underestimate the observed warming rate (Fig. 12). Further, and the multi-model ensemble mean exhibited a cold bias relative to the ground-based observations (Fig. 12 top panels). Moreover, only CanESM2, CESM1-BGC, MIROC-ESM, MPI-ESM_MR models reproduced the observed EDW over the TP (Fig. 12 bottom panels). This suggests that most CMIP5 models have deficiencies in reproducing relevant processes over complex terrain (Lau and Kim, 2018; Lau et al., 2010; Su et al., 2013; You et al., 2019; You et al., 2020). Using a global climate model (GISS-AOM -Goddard Institute for Space Studies-Atmosphere Ocean Model, NASA), simulated EDW over the TP was found between 1950 and 2100 based on both SRES A1B and control model experiments (Rangwala et al., 2010). 24 out of 27 CMIP5 produced significant amplification of EDW over the TP/Himalayas (Rangwala et al., 2016). Subsequent analyses have confirmed that the clear simulated EDW over the TP/Himalayan region in the 20th century will likely strengthen by the end of the 21th century under a high-emission scenario (Palazzi et al., 2017). Strong enhancement of warming at higher elevations (up to 5500 m) was seen for both maximum and minimum temperatures. However warming was most pronounced in regions with temperatures below freezing, suggesting that snow and phase change processes were major factors contributing to the enhanced warming (Palazzi et al., 2017; Palazzi et al., 2019). Using CMIP5 models under RCP4.5 and RCP8.5 scenarios (You et al., 2019), it was found that EDW is present

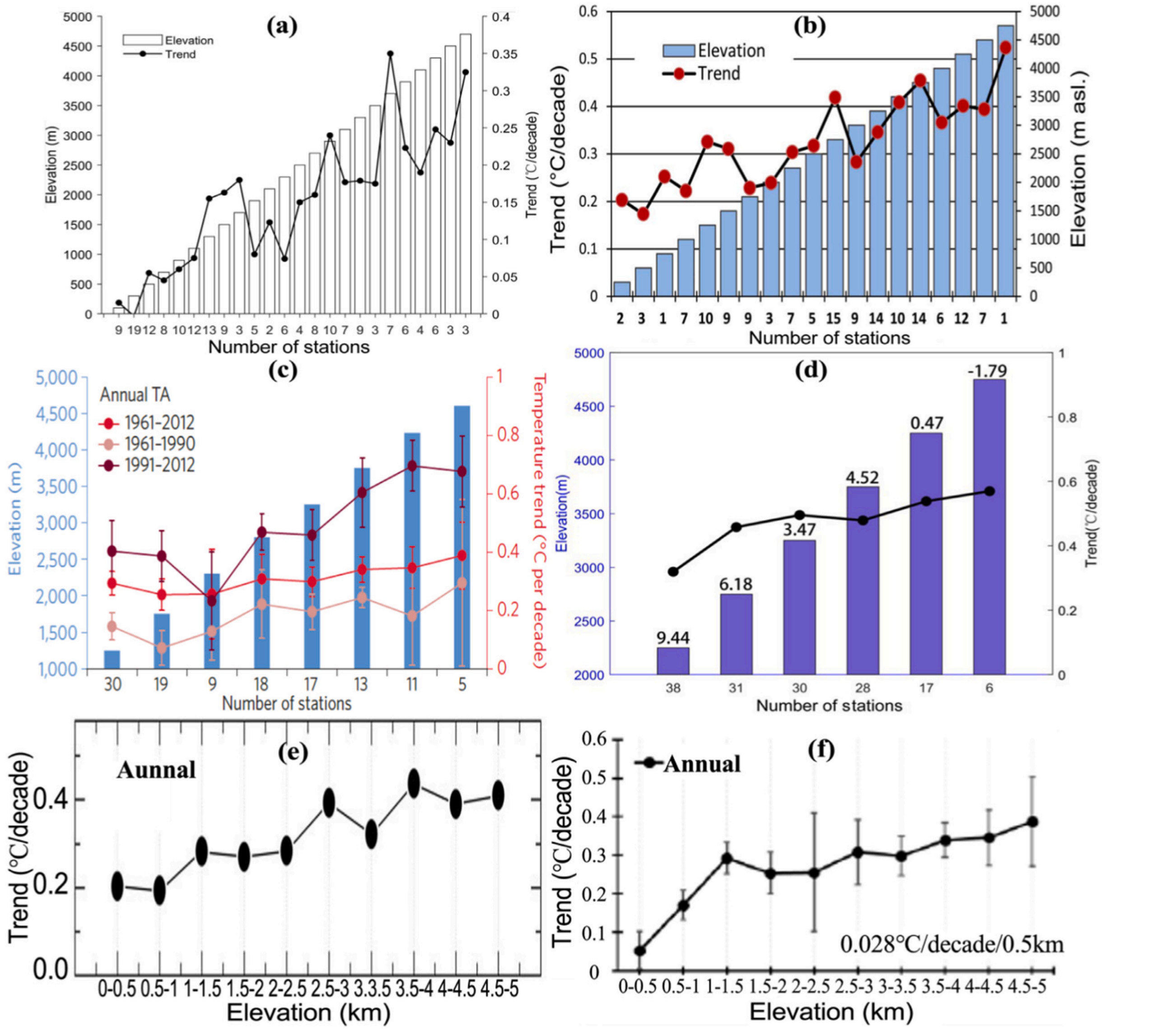

Fig. 3. Observed elevation dependent warming based on the annual mean temperature over the Tibetan Plateau from different studies. Panel A is based on 178 stations with a 200 m elevation interval during 1961–1990 adopted from fig. 9 in Liu and Chen (2000). Panels B and E are based on 144 stations with a 250 m elevation interval during 1961–2010 adopted from figs. 19 and 17 in Wei and Fang (2013). Panel C is based on 122 stations with a 500 m elevation interval during 1961–2012 adopted from fig. 1 in Pepin et al. (2015a). Panel D is based on 150 stations with elevation over 2000 m over the Tibetan Plateau with a 500 m elevation interval during 1979–2018 (this study), provided by the National Meteorological Information Center, China Meteorological Administration. Panel D above/below the columns are the annual surface mean temperature/number of stations in categorized elevation bands, respectively. Panel F is based on 139 stations during 1961–2012 over the Tibetan Plateau adopted from fig. 6 in Yan and Liu (2014).

under global warming of 1.5 °C and 2 °C, but with some significant differences, particularly in the elevation range between 3500 m and 4000 m where snow/ice albedo feedback is expected to play an important role. Warming is amplified preferentially in this elevation band under global warming of 2 °C in comparison with 1.5 °C, showing that the former scenario would have relatively serious consequences for the TP environment (You et al., 2019).

3. Physical mechanisms controlling EDW over the TP

There are a number of physical mechanisms that can produce enhanced warming rates in certain elevation bands over the TP. Table 3 describes the specific responses and the underlying physical mechanisms along with more detailed seasonal and diurnal responses based on changes in relevant climate drivers over the TP. More details on the underlying these physical mechanisms are discussed in the following sub-sections.

3.1. Snow/ice-albedo feedback

Snow/ice-albedo feedback is thought to be one of the most important feedbacks for explaining elevational profiles of warming over the TP (Che et al., 2019; Ghatak et al., 2014; Kang et al., 2010; Liu and Chen, 2000; Liu et al., 2009; Pepin and Coauthors, 2015a; Pepin et al., 2019; Rangwala et al., 2013; Rangwala et al., 2016; You et al., 2016; You et al., 2019). Snow cover reduces the solar radiation absorbed at the surface which decreases surface temperature. However, if this is replaced by water or bare rock, the albedo decreases, leading to more warming. Seasonal snow cover varies with altitude over the TP, and the maximum heating rate is to be expected where the snowline is retreating upward, mainly in areas near the 0 °C isotherm (Pepin and Coauthors, 2015b; Rangwala and Miller, 2012; Rangwala et al., 2013).

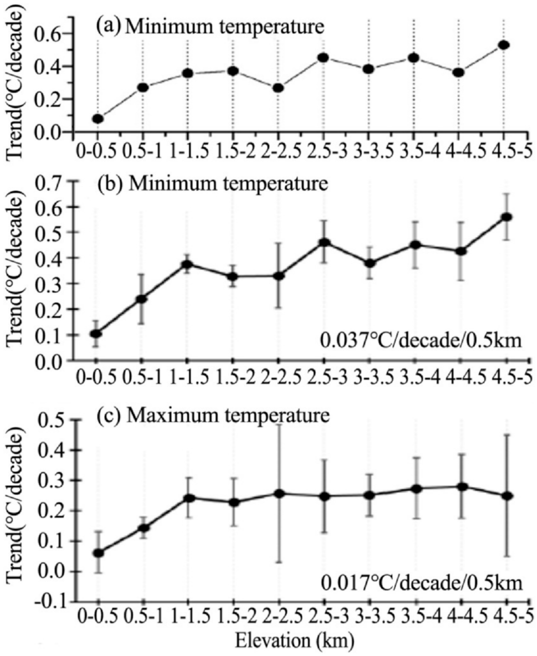

Fig. 4. Observed elevation dependent warming based on the annual mean minimum temperature (a and b) and maximum temperature (c) over the Tibetan Plateau from different studies. Panel A is based on 116 stations with a 500 m elevation interval during 1961–2006 adopted from fig. 3 in Liu et al. (2009). Panels B and C are based on 139 stations during 1961–2012 over the Tibetan Plateau adopted from fig. 6 in Yan and Liu (2014).

Since snow/ice-albedo feedback modulates shortwave radiation absorbed by the surface, it is expected to have a more dominating effect on maximum temperatures (Liu et al., 2009; Rangwala et al., 2010). Over the TP, the snowline typically lies around 5000-6000 m in summer but varies significantly (and typically at much lower elevations) in winter. Many studies have suggested warming rates to peak around 4500–5500 m and then decline with elevations (Gao et al., 2018; Guo et al., 2016; Pepin et al., 2019; Qin et al., 2009).

It is evident that the regions over the TP with thicker annual mean snow depth have experienced stronger decreasing trends of snow cover and snow depth during 1980–2018, and coincide with areas of larger positive radiative forcing due to snow loss (Che et al., 2019) (Fig. 13). Although this appears to support the important role of snow/ice-albedo feedback as an amplifying factor, there was little evidence for exact correspondence of enhanced warming with declining snow and ice cover. Model-based studies also show that snow/ice-albedo feedback has an increasingly important role on EDW in spring and summer over the TP, but the spatial and temporal variability of EDW rates are still poorly quantified (Chen et al., 2003; Giorgi et al., 1997; Liu et al., 2009). It should be noted that current climate models generally overestimate the feedback of spring snow cover change on EDW (Guo et al., 2018), and exhibit deficiencies in reproducing snow/ice albedo feedback (Lau et al., 2010). Although positive snow/ice-albedo feedback may account for much of the EDW over the TP, accurate attribution of the EDW to this effect is still challenging due to limited observations in high elevations, particularly above the current snowline (You et al., 2016; You et al., 2013; You et al., 2017).

3.2. Cloud feedback

Both total cloud amount and cloud characteristics affect the surface energy budget through their effects on shortwave and longwave radiation, and can influence EDW (Duan and Wu, 2006; Hua et al., 2018). Over the TP, particularly over the northern part, low-level cloud amount at night has increased, leading to enhanced nocturnal surface warming, while both total and low-level daytime cloud amounts have decreased, resulting in more solar absorption and increased warming particularly over the southern TP (Duan and Wu, 2006; Duan and Xiao, 2015; Duan et al., 2016). Thus several studies have argued that changes in cloud have contributed to enhanced warming over the TP during 1998–2013 (Duan and Xiao, 2015). Studies using 116 meteorological stations and high resolution climate models, found that EDW over the TP was partly caused by cloud effects on radiation balance (Liu et al., 2006; Liu et al., 2008; Liu et al., 2009). This is consistent with a study in Nepal, where reduced cloudiness and consequent increase of solar radiation was found to account for enhanced warming of maximum temperatures at higher elevations (Thakuri et al., 2019). Based on 5 stations deployed at 2660–5600 m in the south of the central Himalaya during 2007–2011, higher elevations were found to have more clouds in the afternoon, and hence, lower solar radiation and higher downward longwave radiation in the monsoon season, influencing EDW during the South Asian summer monsoon period (Yang et al., 2018).

Future greenhouse gas emissions can drive EDW over the TP through local cloud feedbacks (Bradley et al., 2004; Chen et al., 2003; Liu et al., 2009). For example, a 130-year transient simulation of the NCAR Climate System Model shows that when CO_2 is doubled, strong ground warming occurs over the TP (Chen et al., 2003). In a later study (Yan et al., 2016), four CCSM3 experiments using quadrupled CO_2 levels imply that EDW over the TP could be attributed to local cloud feedback. Using results of an annual 1% CO_2 increase numerical experiment, the annual mean total cloud cover increases over the TP at lower elevations due to increased vigor of the hydrological cycle and atmospheric humidity, but decreases at higher elevations (Fig. 14a). This will result in an increasing trends of atmospheric downward longwave radiation but mostly at lower elevation (Fig. 14b) and increases in surface-absorbed solar radiation but most strongly at higher elevation (Fig. 14c). Thus longwave/shortwave changes are gradually depressed/strengthened with elevation respectively, under the influence of the changes in cloud amount. Additionally, decreases in annual mean snow depth (Fig. 14d) become more prominent as elevation increases, which in turn leads to the reduced surface albedo and further increased absorption of solar radiation with increasing elevation, contributing to EDW over the TP (Fig. 14e) especially in winter and spring (Liu et al., 2009). Other studies have shown similar patterns of cloud changes. In the eastern TP, when CO_2 is doubled, cloud amount also increases/decreases at lower/higher elevations in winter. As a consequence, shortwave solar radiation absorbed at the surface increases at higher elevations but the downward longwave flux reaching the surface decreases. Since the former effect is dominant, the net result of anthropogenic greenhouse gas emission is enhanced surface warming at higher elevations, contributing to EDW over the TP (Chen et al., 2003; Duan and Wu, 2006).

Cloud height may also play a critical role. In a study based on cloud area fraction from the Earth's Radiant Energy System (CERES) products during 2000–2015 (Hua et al., 2018), an increase in high clouds coupled with a decrease in middle-level cloud has resulted in a positive net cloud radiative forcing and thus enhanced warming over much of the TP, especially in the cold season. Cloud cover can also influence the rate of atmospheric warming through latent heat release. If the condensation level rises, a warming slowdown/intensification should occur below/above the new condensation height, which could influence EDW over the TP (Pepin and Coauthors, 2015a; Rangwala and Miller, 2012). However, so far this effect has not been investigated in detail using ground-based observations.

3.3. Atmospheric water vapor feedback

Atmospheric water vapor content also influences radiative fluxes, contributing to positive EDW over the TP. The most critical relationships are the sensitivity of downward longwave radiation to specific humidity, and the relationship between temperature and outgoing

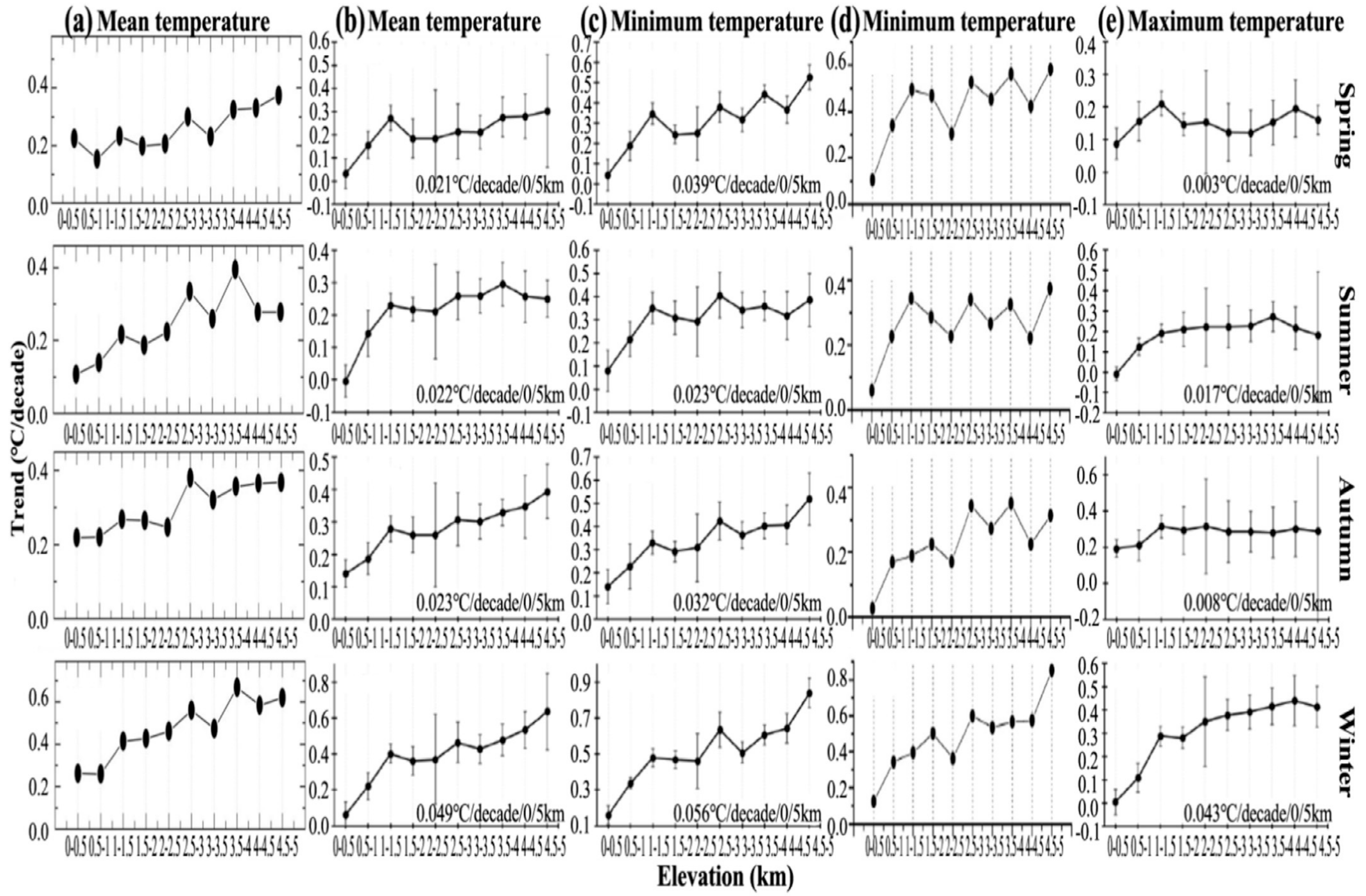

Fig. 5. Observed elevation dependent warming based on the seasonal mean temperature (a and b), minimum temperature (c and d) and maximum temperature (e) over the Tibetan Plateau from different studies. Panel A is based on 144 stations with a 250 m elevation interval during 1961–2010 adopted from fig. 17 in Wei and Fang (2013). Panels B, C and E are based on 139 stations during 1961–2012 over the Tibetan Plateau adopted from fig. 6 in Yan and Liu (2014). The values denote the gradients of the best-fit lines fitted to warming rate against elevation (line not shown). Error bars represent the 95% confidence interval of the mean trends in each elevation band. Panel D is based on 116 stations with a 500 m elevation interval during 1961–2006 adopted from fig. 3 in Liu et al. (2009).

longwave radiation (Pepin and Coauthors, 2015b; Rangwala and Miller, 2012). Over the TP, seasonal increases in downward longwave radiation (which cause warming) result from increases in specific humidity but the relationship between the two is non-linear. Thus the sensitivity of longwave radiation to increased moisture content is enhanced at low humid levels (Ruckstuhl et al., 2007). This occurs mostly during cold seasons and at higher elevations (Rangwala et al., 2009). Thus in winter the increase of surface specific humidity is a key process leading to enhanced warming with elevation over the TP (Rangwala et al., 2009; Rangwala et al., 2010). This mechanism is also important for explaining positive EDW in spring and autumn, but the contribution is relatively weak (Rangwala et al., 2009; Rangwala et al., 2010).

A second mechanism which favors enhanced warming with increasing height relates to the relationship between temperature and outgoing longwave radiation (Pepin and Coauthors, 2015a). The functional shape of the Stefan–Boltzmann law indicates that the low temperature of the environment amplifies the effect of the energy balance variation on surface temperature, which contributes to enhanced temperature variability at high altitudes (Ohmura, 2012). While some studies have investigated this potential driver, the underlying processes and potential significance of atmospheric water vapor are not yet fully understood.

3.4. Aerosol feedback

Aerosols, such as the Asian brown cloud and deposition of black carbon on mountain glaciers and snowpack, may have made an important contribution to EDW over the TP (Kang et al., 2019; Lau et al., 2010; Ramanathan et al., 2007; Ramanathan et al., 2005; Ramaswamy et al., 2006; Xu et al., 2009). The Asian brown cloud, which is mostly the result of biomass fossil fuel burning, could enhance lower atmospheric solar heating regionally by about 50% (Ramanathan et al., 2007; Ramanathan et al., 2005; Ramaswamy et al., 2006). It is unclear, however, how this may influence patterns of warming over the TP, but one possible mechanism is through its influence on atmospheric circulation (Kang et al., 2019).

Most atmospheric aerosol pollutants (such as brown clouds associated with black carbon in Asia) accumulate at relatively low altitudes (less than 3000 m), which may reduce the shortwave radiation flux to lower altitudes, but there is little or no effect on higher mountain regions for much of the year, thus creating a warming EDW gradient (Kang et al., 2019; Xu et al., 2009). On the other hand, in spring, measurement of desert dust and locally emitted black carbon have been observed up to an elevation of 5000 m (Ramanathan and Carmichael, 2008; Ramanathan et al., 2007; Ramanathan et al., 2005) and during the active period of the convective monsoon, aerosols have been transported from the surface up to altitudes of 8–12 km (the upper troposphere) (Lawrence, 2011). The transportation and deposition of black carbon and other anthropogenic aerosols from lowland South Asia into the TP, forces snow and glaciers to melt more rapidly and reduces surface albedo, thereby enhancing EDW in some regions (Kang et al., 2019; Xu et al., 2009). However, there is almost no systematic

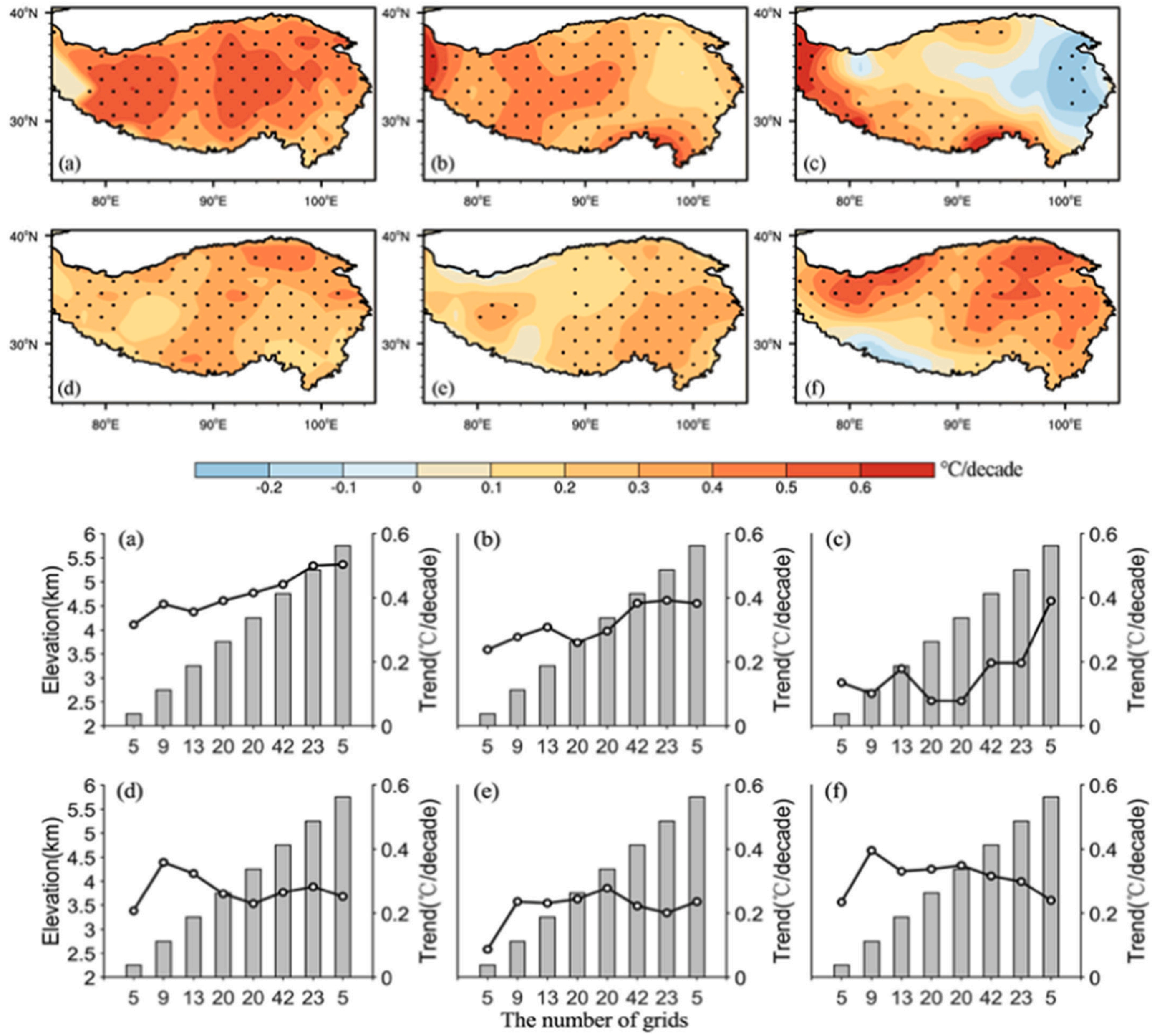

Fig. 6. Spatial trends (top panels) and patterns of elevation dependent warming (bottom panels) from the gridded-observations (a), NCEP1 (b), NCEP2 (c), ERA-Interim (d), MERRA (e) and JRA-55 (f) during 1979–2018 over the Tibetan Plateau (this study). Values below vertical bars in the bottom two rows are the number of grid points in categorized elevation bands. The reanalysis data used include: National Centers for Environmental Prediction (NCEP)-National Center for Atmospheric Research (NCAR) Reanalysis Project (NCEP1); NCEP-Department of Energy (DOE) Reanalysis Project (NCEP2) (http://www.esrl.noaa.gov); European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Reanalysis (ERA-Interim) (http://www.ecmwf.int); Japan Meteorological Agency (JMA) 55 year Reanalysis Project (JRA-55) (http://jra.kishou.go.jp); and the National Aeronautics and Space Administration (NASA) Modern-Era Retrospective Analysis for Research and Applications (MERRA) (http://disc.sci.gsfc.nasa.gov).

and temporally consistent observation of black carbon deposition in high altitude areas and the influence of aerosol feedback may only be analyzed and assessed using climate models.

3.5. Land use changes

Land use changes are regarded as an important factor for EDW over the TP (Cui and Graf, 2009; Cui et al., 2006; Frauenfeld et al., 2005; Zhang, 2007). Significant land use changes over the TP include permafrost and grassland degradation, urbanization, deforestation and desertification, and may account for spatial differences in EDW in the region (Cui and Graf, 2009; Cui et al., 2006). Most general atmospheric circulation models support the notion that human induced land use changes over the TP have had a significant impact on local to regional scale climate (Cui and Graf, 2009; Cui et al., 2006). For example, the production of livestock or meat over the TP has increased by nearly 300% since 1978, which means that there should be an equivalent increase in the consumption of plant biomass (Du et al., 2004). There is a positive feedback in which degradation of grassland by overgrazing will increase sensible heat flux and decrease latent heat flux, thereby promoting further climate warming over the TP (Du et al., 2004). This effect will potentially influence patterns of EDW over the region.

3.6. Ozone change

A reduction in the total stratospheric ozone over the TP in

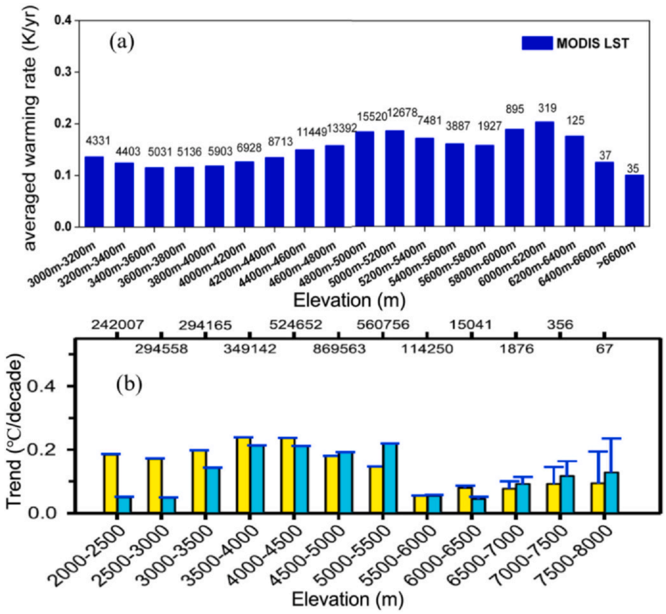

Fig. 7. Elevation dependent warming based on Moderate Resolution Imaging Spectroradiometer (MODIS) Land Surface Temperature (LST) data. Panel A is based on MODIS LST during 2000–2006 adopted from fig. 4 in Qin et al. (2009). Panel B is based on annual mean satellite-based 2 m air temperature (yellow bar) and MODIS LST (cyan bar) trends during 2001–2015 adopted from fig. 9 in Guo et al. (2019). The numbers at the top of bars are the number of satellite pixels in each elevation range. Blue error bars in Panel B are based on 95% confidence intervals around the mean. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

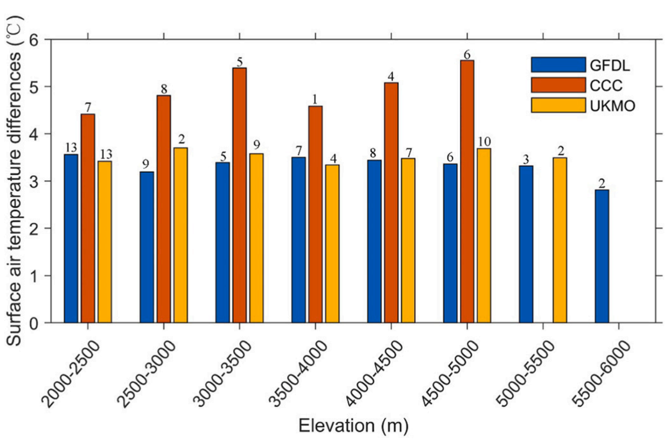

Fig. 8. Annual surface mean temperature differences (°C) between the doubling CO_2 concentration experiment and the 1 × CO_2 experiment from the Geophysical Fluid Dynamics Laboratory (GFDL, 2.25° × 3.75°), the Canadian Climate Center (CCC, 3.75° × 3.75°) and the UK Meteorological Office (UKMO, 2.5° × 3.75°) climate models averaged for various elevation ranges over the Tibetan plateau (22°-46°N , 70°-110°E). Numbers above vertical bars are the number of grid points in categorized elevation bands. This is based on table 3 in Liu and Chen (2000).

comparison with adjacent regions may be an important reason enhancing EDW in recent decades (Zhang and Zhou, 2009; Zhou and Zhang, 2005). Strong reduction in ozone has resulted in increased ultraviolet radiation reaching the Earth's surface and greater warming in the lower troposphere (Zhang and Zhou, 2009; Zhou and Zhang, 2005). This may to some extent explain EDW over the TP. Since the mid-1980s, summer air temperature has been strongly correlated with column ozone in the region, indicating that the most significant warming over the northern TP may be related to radiative and dynamical heating resulting from pronounced stratospheric ozone depletion (Guo and Wang, 2012).

3.7. *Vegetation greenness*

Changes in vegetation are another component of environmental change over the TP which may have an influence on EDW (Shen et al., 2015; Tian et al., 2014). Based on three normalized difference vegetation index (NDVI) datasets during 2000–2012, there are evidences of overall vegetation greening over the TP but negatively correlated with elevation (Liu et al., 2019). This finding was consistent with a study in the Yarlung Zangbo River basin with a wide range of elevations from 147 m to over 7000 m using NDVI data during 1999–2013 (Li et al., 2015). Although the NDVI on average increased by 8.83% from 1999 to 2013, increases showed negative elevation dependency (Li et al., 2015). The reason given was that the effect of increasing dryness at higher elevations has contributed to enhanced warming, which eventually prevented rapid greening there (Liu et al., 2019). This suggests that the strong vegetation greening at lower elevations may decrease surface albedo driven by dense vegetation in these areas (Tian et al., 2014). Forest or grassland tends to have lower albedo than dry scrubland. However, any extra absorption of radiation from lower albedo can be efficiently dissipated by evaporative cooling over the TP (Shen et al., 2015). Increased vegetation activity at lower elevations over the TP will therefore tend to attenuate daytime warming by enhancing evapotranspiration (Shen et al., 2015). Thus, a daytime cooling effect from increased vegetation activity at low elevations is thought to partly encourage positive EDW over the TP.

However, another study examining the high mountains of the southeastern TP in Sichuan showed an opposite greening rate and elevation behavior (Tao et al., 2018). Based on NDVI during 1999-–2013, it was shown that the vegetation greening rate driven by temperature shows positive elevation dependency (Tao et al., 2018). At higher elevations, vegetation greening is more likely to occur in otherwise snow-covered areas, and in such a case the warming effect due to reduction of albedo may override any evaporative cooling effect, although there have been no detailed studies of this topic. Clearly there are contradictory results in different regions and for different time periods. More work on the interactions between elevation-profiles and vegetation greening is required to understand the role of this potential driver.

4. New perspectives on EDW over the TP

4.1. *Can climate models realistically simulate profiles of EDW?*

EDW has strong significance for climate change in mountain regions, and the question has been raised as to whether models which simulate observed patterns of EDW more successfully, may be more suitable for future mountain predictions (Pepin and Lundquist, 2008; Pepin and Coauthors, 2015a; Pepin et al., 2019; You et al., 2019). Nearly all climate models suggest that enhanced warming with elevation is a feature of future mountain warming during the first half of the 21st century, but the amount of enhancement depends on season and the region under consideration (Kotlarski et al., 2015; Kotlarski et al., 2012). Some early studies (Diaz and Bradley, 1997; Giorgi et al., 1997) suggested that climate models simulated enhanced warming at high elevations and later this was confirmed by a study of seven global climate models which showed that all of them simulated EDW in free-air temperatures when atmospheric CO_2 levels were doubled (Bradley et al., 2004). Regional climate model simulations over the TP show a mean warming of 0.6°–0.9 °C for the period 2015–2050, with pronounced EDW (Zhu et al., 2013). Many model studies have also shown stronger warming rates in northern high latitudes and future studies should investigate whether those models with strong Arctic amplification also simulate strong EDW (Bulygina et al., 2007). Most climate models also simulate EDW associated with elevation dependent

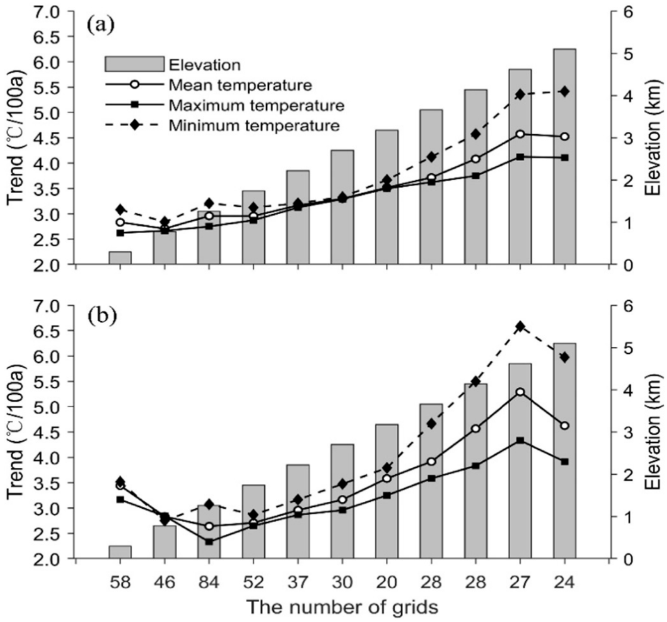

Fig. 9. Trends in annual (a) and winter (b) surface mean temperature using climate model outputs of a 1% per year CO_2 increasing experiment with the high resolution (T85, 140 km horizontal resolution) version of the fully coupled Community Climate System Model (CCSM 3) developed by the National Center for Atmospheric Research (NCAR). Mean trends are averaged for 500 m wide elevation bands over the Tibetan Plateau and surroundings (26°–42°N , 70°–105°E). This is based on fig. 3 in Liu et al. (2008).

increases in atmospheric water vapor and elevation dependent decreases in surface albedo (Rangwala et al., 2016). Over the TP, the ability of climate models to simulate the existence of EDW has been shown to be dependent on model resolution (Palazzi et al., 2019). Furthermore, climate model deficiencies in simulating cloud properties (Zhou and Li, 2002), snow/ice albedo feedback (Lau et al., 2010) and the aerosol effects on snow/ice albedo (Lau and Kim, 2018) may induce insufficient warming over the TP particularly at higher elevations, which will of course influence the simulated EDW phenomenon. As a result there are obvious regional differences in EDW simulations between models and much more work is needed to understand the drivers of such variation.

4.2. Can EDW explain the rapid warming periods reconstructed from proxies?

EDW is widely cited to explain the rapid rise in reconstructed temperature from proxies (such as ice cores and tree rings) over the TP (Gou et al., 2007; Kang et al., 2010; Kang et al., 2007; Thompson et al., 2018; Tian et al., 2006; Yang et al., 2009). For example, an ice core record retrieved from a flat part of the Guoqu Glacier (5720 m) on Mt. Geladaindong has been used to reconstruct a history of summer air temperature over the last 70 years (Kang et al., 2007). Results show a warming rate of 0.5 °C/decade since the 1970s and 1.1 °C/decade since the 1990s respectively, much higher than over the central TP and the Northern Hemisphere as a whole. A 41.6 m ice core drilled at 7010 m on a rather gentle slope of the Muztagata glacier in the eastern Pamirs provided a record over 150 years and a warming magnitude of 2 °C–2.4 °C/decade from the 1960s to 1990s (Tian et al., 2006), again much higher than the Northern hemisphere mean. Records of recent climate change from ice cores drilled on the Guliya ice cap in the western Kunlun Mountains also demonstrate that this region has become warmer and moister since at least the middle of the 19th century, and shows the highest rate of warming since the end of the Little Ice Age (Thompson et al., 2018).

Analyses of tree rings from juniper in the Xiqing Mountains (northeastern TP), shows a rate of increase for winter (Nov-April) minimum temperature of approximately +0.5 °C/decade from 1940 to 1990, again much higher than the rate of increase in mean temperature over the TP as a whole (Gou et al., 2007). The rapid warming rates reconstructed from ice cores and tree rings at higher elevations over the TP are possibly explained by enhanced warming at high elevations.

4.3. Would EDW influence the Asian summer monsoon?

The pattern of EDW over the TP will influence the distribution of the surface heat source and may therefore have far reaching effects on the Asian monsoon system (Duan and Wu, 2005; Duan et al., 2012; Ge et al., 2019; You et al., 2020). As a huge elevated heat surface projecting into the atmosphere, the TP transfers heat from the surface to the air in the form of sensible and latent heat transfer (Wu et al., 2004; Wu et al., 2015; Zhu et al., 2019). The sensible heat flux trend (1980–2015) over the TP displayed a negative elevation dependency above 2000 m, but a positive elevation dependency at low elevation stations (Zhu et al., 2019). Any elevational pattern of warming over the TP in combination with spatial differences could change the role of the TP as a heat source and thus influence the Asian monsoon and rainfall patterns through interaction with Rossby wave trains (Wang et al., 2008). Numerical experiments with general circulation models have shown that wider atmospheric heating induced by rising TP temperatures can enhance East Asian subtropical frontal rainfall through two distinct Rossby wave trains. Isentropic uplift to the east of the plateau deforms the western Pacific Subtropical High and enhances moisture convergence toward the East Asian subtropical front (Wang et al., 2008). Previous studies have suggested that warming of the plateau drives the land-sea thermal contrast and thus the strength of the Asian summer monsoon (Kang et al., 2019; Wang et al., 2008; Wu et al., 2015). Changes in the features of EDW would lead to changes in the distribution and magnitude of overall TP heating, particularly if EDW changes over time. For example, due to rapid warming in some lower elevation valley over the TP, EDW based on annual and monthly mean temperatures during 1991–2010 has been weakened compared with that during 1961–2010 (Wei and Fang, 2013). It is not clear how such variations in the magnitude of EDW could influence the heating of the plateau and its consequences in the monsoonal context. Due to discrepancies in current explanations concerning the mechanisms involved, it remains to be better understood how EDW may related to overall TP warming and the Asian monsoon.

4.4. What implications of EDW for water resources changes?

EDW has significant implications for water resources and environmental changes over the "Asian water tower" (Immerzeel and Bierkens, 2012; Immerzeel et al., 2010; Pithan, 2010; Qin et al., 2006; Qin et al., 2018; Zhang et al., 2020). Studies of permafrost and seasonally frozen ground across the plateau have shown total permafrost area to decrease at 9.2 km^2/decade, and the area-mean active layer thickness to increase by 0.15 m/decade during 1981–2000 (Guo and Wang, 2013). The area-mean maximum freezing depth of seasonally frozen ground decreased by 0.34 m/decade (Guo and Wang, 2013). Other studies have shown that the thickness of the active layer has increased by 0.15 to 0.50 m on average and ground temperature at a depth of 6 m has risen by about 0.1 to 0.3 °C during 1996–2001 (Cheng and Wu, 2007). Also, most Himalayan glaciers are losing mass at alarming rates similar to glaciers elsewhere around the world (Bolch et al., 2012). The projected loss of glacial area over the TP is estimated to be around 43% by 2070 and 75% by the end of the 21st century (Latif et al., 2019). Under RCP 2.6 to RCP 8.5 projections, 36 ± 7 to 64 ± 5% of glacier ice in this region would disappear by 2100, with glaciers on the eastern and southern TP experiencing the most loss (50–75%) (Kraaijenbrink et al., 2017). Part of the reason for the relatively wide range of uncertainty in the above

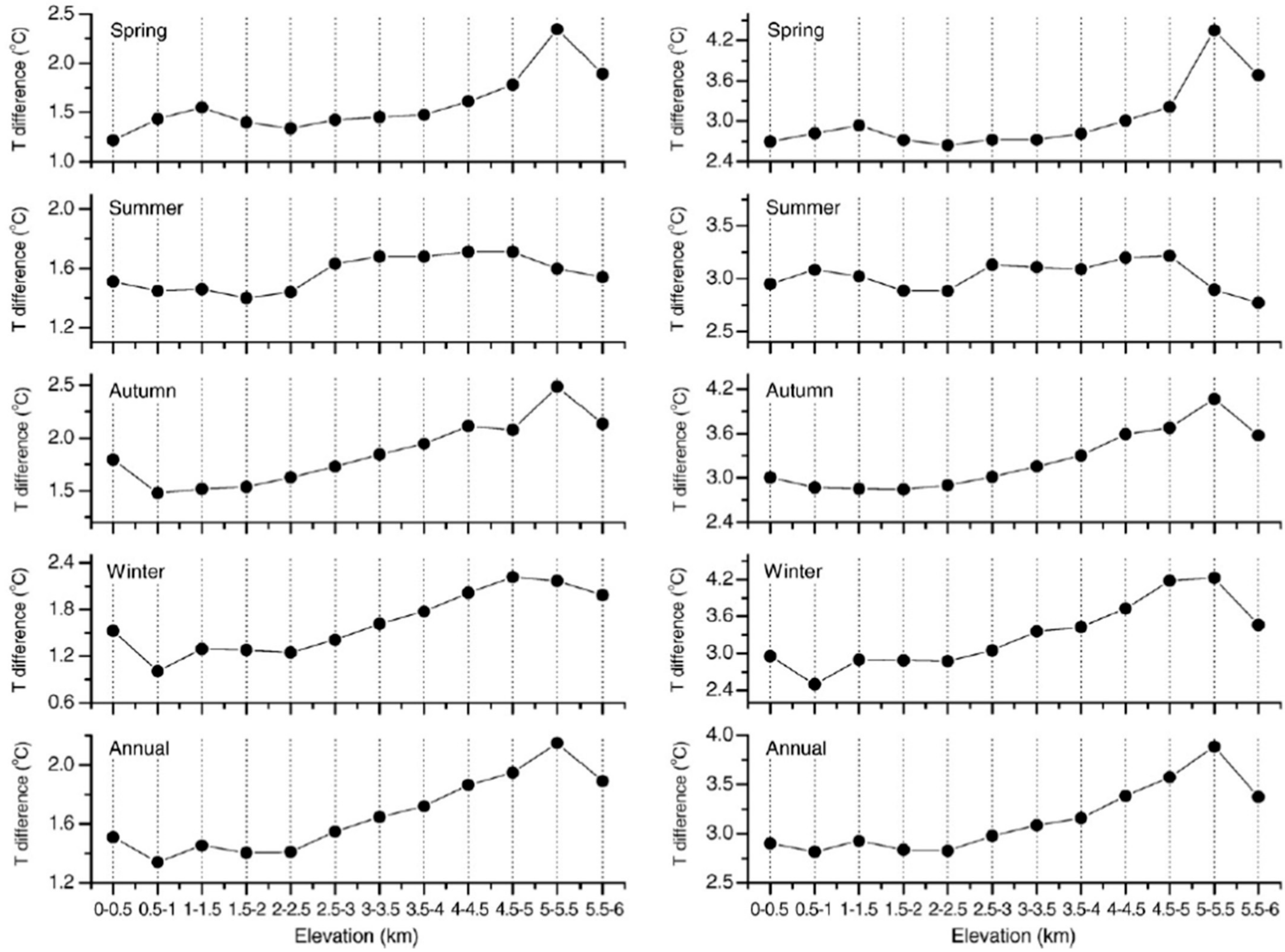

Fig. 10. Changes of spring, summer, autumn, winter and annual mean minimum temperature projected with downscaled CAM3/CLM3 model under the A1B emission scenario during 2030–2049 (left panel) and 2080–2099 (right panel) with respect to the reference period 1980–1999 in 12 elevation zones over the Tibetan Plateau. This is adopted from fig. 6 in Liu et al. (2009).

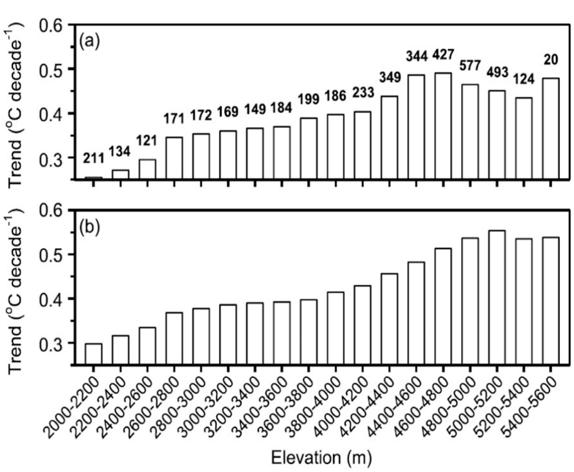

Fig. 11. Changes in temperature trends with increasing elevation over the Tibetan Plateau based on (a) MIROC5/WRF from 2006 to 2050 and (b) MIROC5/WRF from 2006 to 2099 under the RCP6.0 scenario. The trends are the means values over individual elevation bins. The numbers above bars are the number of simulation grids in each elevation bin. This is adopted from fig. 4 in Guo et al. (2016).

projections is the uncertainty around EDW, making it a critical issue for evaluating changes in water resources in the future (Yang et al., 2019; Yao et al., 2012a; Yao et al., 2019).

4.5. Is there a way to overcome limitations in our observation of EDW?

Knowledge and understanding of EDW over the TP based on the surface station network is inadequate, because most stations are concentrated in the relatively accessible southern and eastern parts of the plateau below 4000 m (Kang et al., 2010; Kang et al., 2019; Yao et al., 2019; You et al., 2019). Many stations are in incised valley locations and/or in urban areas, so their representation of the range of topographic situations and land cover is questionable. Thus, the uncertainty in EDW due to the gap in station observations at higher elevations is a serious concern (Kang et al., 2010; Pepin et al., 2019; Yan et al., 2016).

Lack of in situ weather station data above 5500 m and sparse stations over the western TP are particular limiting factors for understanding of EDW based on ground-based observations. Gridded simulations from climate models and reanalyses are often too coarse especially at high elevations in a region with complex topography. Thus, a full assessment of EDW requires other datasets, such as remotely-sensed observations, regional reanalyses, and downscaled global/regional climate models (Kang et al., 2010; Kang et al., 2019; Yao et al., 2019; You et al., 2019).

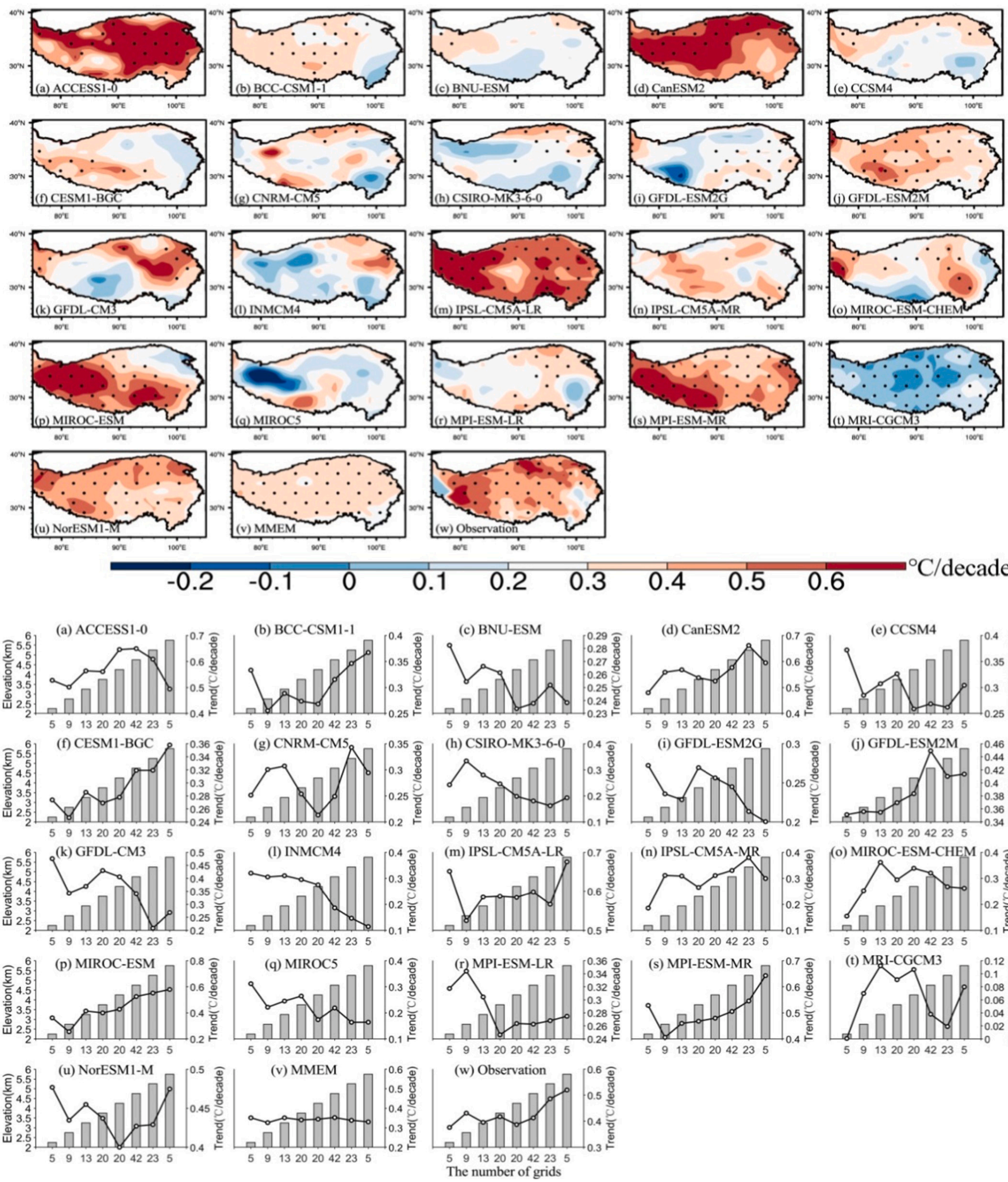

Fig. 12. Spatial trends (top panels) and patterns of elevation dependent warming (bottom panels) from 21 Coupled Model Intercomparison Project Phase 5 (CMIP5) models, their multi-model ensemble mean, and the gridded-observations for the historical run (during 1979–2005), respectively. Values below the column bars (bottom panels) are the number of grid points in categorized elevation bands. Surface (2 m) mean temperatures for models participating in CMIP5 are obtained from https://pcmdi.llnl.gov. All model simulations are bi-linearly interpolated to a common 1.5° grid before analysis (Wu et al., 2019; You et al., 2019).

5. Summary

This review aims to synthesize the state of knowledge on EDW over the TP. We have examined recent assessments of EDW using multiple datasets and aevaluated the relative importance of various physical mechanisms in explaining patterns of EDW (Table 3). Further, the review contributes to outlining new perspectives on EDW and highlighting current research gaps related to EDW over the TP.

Table 3
Description of physical mechanisms that can produce an elevation-sensitive temperature response over the Tibetan Plateau. Temperature responses include mean temperature (Tmean), maximum temperature (Tmax) and minimum temperature (Tmin) adopted and updated from a previous review (Rangwala and Miller, 2012).

Climate driver	Mechanisms	Seasonal relevance	Temperature response
Decreases in **Snow/Ice Albedo**	➢ Increases surface absorption of insolation	Primarily spring; but also important in winter at lower elevations, summer at higher elevations, in association with the 0 °C isotherm (Pepin et al. 2015; Rangwala; Miller 2012)	➢ Increases T_{max}; suppressed effect if soil moisture also increases and causes daytime evaporative cooling
Increases in **Cloud Cover** (Daytime)	➢ Decreases surface insolation	All seasons but greater effects in summer	➢ Decreases T_{max}
Increases in **Cloud Cover** (Nighttime)	➢ Increases downwelling longwave radiation	All seasons but greater effects in winter	➢ Increases T_{min}
Increases in **Water Vapor**	➢ Increases downwelling longwave ➢ Downwelling longwave has high sensitivity to changes in surface specific humidity (Rangwala et al., 2009)	Primarily winter; smaller effects are possible in autumn and spring (Rangwala; Miller 2012)	➢ Increases T_{min}
Increases in **non-absorbing Aerosols**	➢ Decreases surface insolation ➢ Increases cloud albedo and lifetime Rangwala et al., 2009]	Dependent on seasonal emissions	➢ Decreases T_{max} ➢ Small increases in T_{min} when cloud lifetime is enhanced
Increases in **absorbing Aerosols**	➢ Decreases surface insolation but increases mid-tropospheric heating ➢ Decreases albedo of clouds and snow on ground	The total ozone amount declined in all seasons	➢ Increases T_{mean} ➢ Increases T_{max} when cloud cover is reduced
Land use changes	➢ Increases sensible heat fluxes and decreases latent heat fluxes during the day	Permafrost degradation is strongest in summer; Overgrazing is strongest in winter	➢ Increases T_{max}
Decreases in **total stratospheric ozone**	➢ Increases ultraviolet radiation	Permafrost degradation is strongest in summer; Overgrazing is strongest in winter	➢ Decreases T_{mean} in the upper troposphere to lower stratosphere ➢ Increases T_{mean} in the lower to middle troposphere
Increases in **vegetation**	➢ Increases surface albedo ➢ Increases evaporative cooling	Vegetation greening is strongest in growing season	➢ Decrease T_{max}

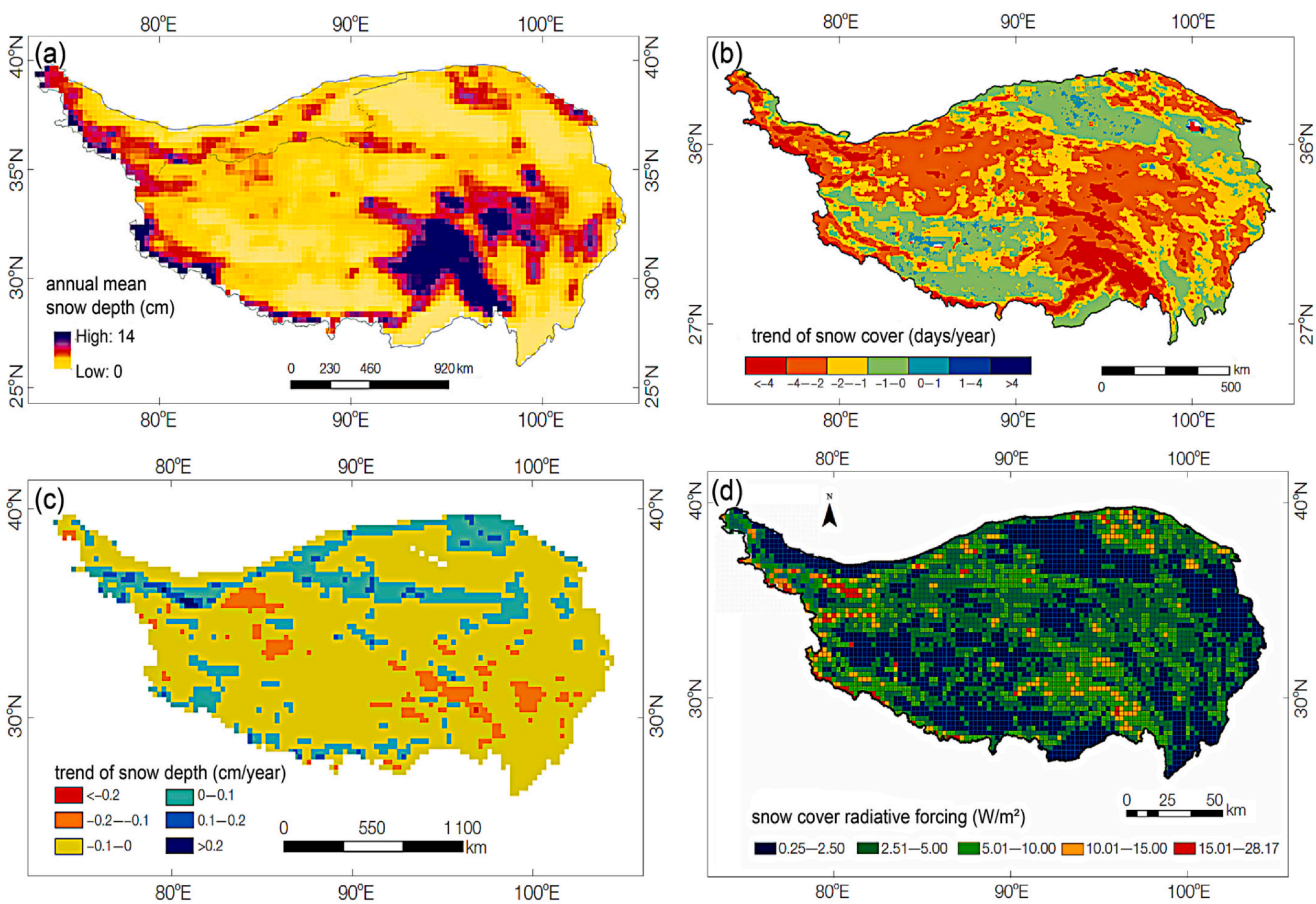

Fig. 13. The spatial patterns of a) annual mean snow depth (cm), b) trend of snow cover (days/year), c) trend of snow depth (cm/year) during 1980–2018, and d) snow cover radiative forcing (W/m^2) during 2001–2010 over the Tibetan Plateau based on figs. 2, 4, 6 and 7 in Che et al. (2019).

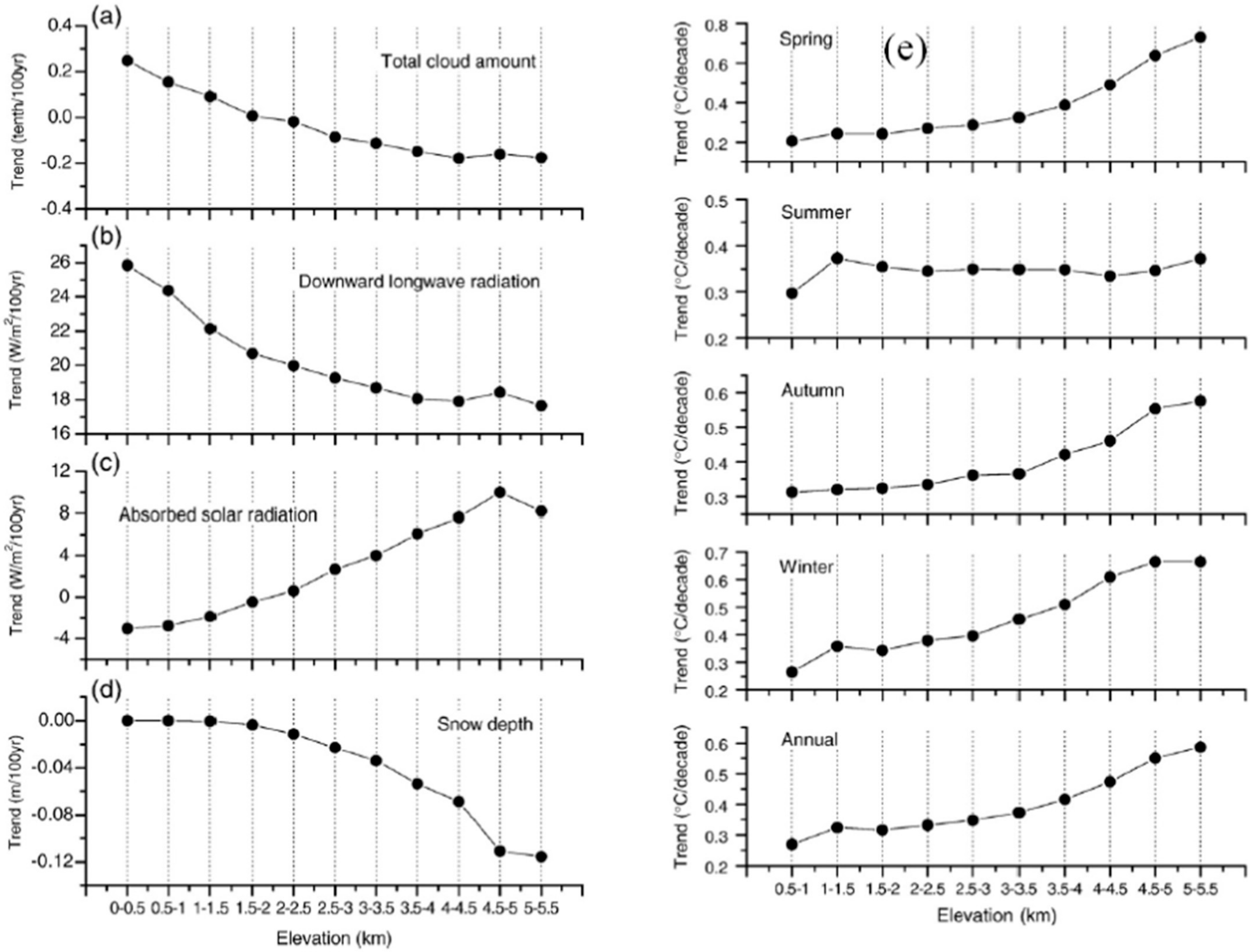

Fig. 14. Linear trends of a) annual-mean total cloud amount, b) atmospheric downward longwave radiation, c) surface-absorbed solar radiation, d) snow depth, and e) annual and seasonal minimum temperature in the annual 1% CO_2 increase experiment based on NCAR Community Climate System Model (CCSM3) for 100 simulation years over 500 m elevation bands over the Tibetan Plateau adopted from figs. 4 and 8 in Liu et al. (2009).

The main findings can be summarized as follows:

(1) The TP and its surrounding areas are experiencing rapid warming, but the patterns of EDW identified with multiple datasets do not always agree. Many physical mechanisms possibly controlling EDW have been proposed and studied, including snow/ice-albedo feedback, cloud feedback, atmospheric water vapor feedback, aerosol feedback, and changes in land use, ozone and vegetation. However, there is no consensus on the relative importance of each mechanism.

(2) Although great progress has been made in many studies of EDW, there are still limitations and caveats mainly due to limited ground-based observations across different elevations. The distribution of the existing plateau stations is uneven, and even basic information about regional patterns (such as warming characteristics, trend magnitudes) is lacking over the western TP.

(3) Most studies on EDW over the TP are based on a single data source. However, different sources of data have their own shortcomings. For example, reanalysis products are affected by inhomogeneities in assimilated data and climate models have deficiencies in simulating relevant atmospheric and cryospheric processes over complex terrain. Comparisons of multiple data sources also required reliable independent observations which are not always available.

(4) Whether climate models can realistically simulate the observed EDW can be used to assess their reliability of future EDW and the corresponding impacts on the regional climate and the Asian summer monsoon. The consequences of EDW at high elevations on water resource and variability are still under discussion, and there is not a clear consensus on the actual impacts of EDW on regional water availability.

This review has clearly demonstrated the need to improve the accuracy of local climatology through incorporating fine-scale topography into climatological assessments and model simulations. This will certainly help understanding the complex distribution of EDW over the TP and the potential impacts on the relevant processes. Current weather stations over the TP are nearly all located at lower elevations with very limited observations above 5500 m, resulting in critical gap in our understanding of temperature variability. Combining application of remote sensing and high-resolution modeling experiments, and ideally improving the ground-based observations, will be required to validate EDW and better understand recent and future high elevation climate.

Declaration of Competing Interest

The authors declared that they have no conflicts of interest to this work. We declare that we do not have any commercial or associative interest that represents a conflict of interest in connection with the work submitted.

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