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Climate change in mountains: a review of elevation-dependent warming and its possible causes

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Abstract Available observations suggest that some mountain regions are experiencing seasonal warming rates that are greater than the global land average. There is also evidence from observational and modeling studies for an elevation-dependent climate response within some mountain regions. Our understanding of climate change in mountains, however, remains challenging owing to inadequacies in observations and models. In fact, it is still uncertain whether mountainous regions generally are warming at a different rate than the rest of the global land surface, or whether elevation-based sensitivities in warming rates are prevalent within mountains. We review studies of four high mountain regions – the Swiss Alps, the Colorado Rocky Mountains, the Tibetan Plateau/Himalayas, and the Tropical Andes – to examine questions related to the sensitivity of climate change to surface elevation. We explore processes that could lead to enhanced warming within mountain regions and possible mechanisms that can produce altitudinal gradients in warming rates on different time scales. A conclusive understanding of these responses will continue to elude us in the absence of a more comprehensive network of climate monitoring in mountains.

1 Introduction

During the last century, global surface air temperature has increased by 0.75°C according to the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC; Trenberth et al. 2007). Between 1975 and 2010, land temperatures have been increasing at a rate of 0.30°C/decade, which is more than double the rate (0.12°C/decade) of ocean warming. It has been proposed that mountainous regions may be more sensitive to

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global scale climate change than other land surface at the same latitude (e.g., Messerli and Ives 1997; Beniston et al. 1997). Several studies have suggested that mountain regions have warmed at a greater rate than their low elevation counterparts often with greater increases in daily minimum temperatures than daily maximum temperatures (e.g. Diaz and Bradley 1997; Beniston et al. 1997; Rangwala et al. 2009; Liu et al. 2009; Qin et al. 2009; Pederson et al. 2010). Most climate models find enhanced warming in mountains and do so more consistently than found in observations (Pepin and Lundquist 2008).

Much of the world's supply of surface water has its source in mountains, which makes it critical to understand how climate will change in these regions. The Tibetan Plateau, with more than 36,000 glaciers (Liu et al. 2010) that drain into the main rivers of Asia (e.g. Yangtze, Yellow, Mekong, Brahmaputra, Ganges, Indus), directly and indirectly supplies water to the most populous region of the world with more than two billion people. Some of the Tibetan glaciers are melting rapidly, with surface losses in some places of 0.77 km^2 per year between 1999 and 2003 (Kehrwald et al. 2008). Beniston (2003) states that worldwide 30 to 50% of existing mountain glacier mass could disappear by 2100.

A continuous warming trend at high altitudes during this century could significantly modify the hydrologic cycles in these mountains (e.g., Nijssen et al. 2001). Increased warming will cause decreases in winter and spring snowpack leading to changes in the pattern of seasonal streamflow – generally a reduction in summer flows (Dettinger and Cayan 1995; Arnell 2003; Saunders et al. 2008). Small shifts in precipitation regimes in the mountains could cause widespread disruptions of freshwater availability (e.g., Beniston et al. 1997). Furthermore, increased evaporation from warming in elevated regions may cause severe drying in summer months (e.g., Beniston 2003). Increased warming and changes in precipitation are likely to have significant consequences for humans and ecosystems within the mountains as well as those downstream (Dettinger and Cayan 1995; Nijssen et al. 2001; Arnell 2003; Beniston 2003). These impacts include reduction in reservoir storage, increases in the intensity of wildfires, drought and pest induced plant mortality, and temperature induced changes in aquatic life (e.g., Overpeck and Udall 2010).

Within several mountain regions, studies have found an elevation dependency in surface warming where in several cases greater warming rates were reported at higher altitudes (e.g. Beniston and Rebetez 1996; Diaz and Bradley 1997; Liu and Chen 2000; Diaz and Eischeid 2007; Liu et al. 2009; Rangwala et al. 2009), however, there are other studies which found greater warming rates at lower elevations (e.g. Vuille and Bradley 2000; Pepin and Losleben 2002; Lu et al. 2010). Beniston et al. (1997) suggests that a tendency for greater warming rates at higher altitudes may be more apparent in the tropics. Furthermore, there are also studies where no elevation dependency in warming rates was found (e.g., Vuille et al. 2003; Pepin and Lundquist 2008; You et al. 2010). In fact, there are even studies within the same mountain region which obtain different results regarding elevation-dependent warming rates. For example, Liu and Chen (2000), Liu et al. (2009), Qin et al. (2009) and Rangwala et al. (2009) found increased warming rates at higher altitudes in the Tibetan Plateau in the latter half of the 20th century, while You et al. (2010) found no elevation-based trends and the study by Lu et al. (2010) suggests greater warming trends at lower elevations. Because some studies have found increases in warming rates with elevation in mountain regions and others have found decreases, references to elevation-dependent responses in this review refer to either case.

There are two fundamental questions regarding climate change in high elevation regions. One is whether the rate of warming within mountain regions is changing at a different rate than rest of the global land surface, and the second is whether the rates of change differ depending on elevation within a specific mountain region. In this paper, we review studies that have examined these questions using both observations and climate model output from selected high mountain regions. Table 1 lists these studies including the information on time period, elevation range and number of stations considered in these studies. We summarize the results from these studies and then examine the role of relevant feedbacks and mechanisms that can produce elevation-sensitive responses. The next section explores our current understanding of whether mountain regions are warming faster than other land areas. Section 3 examines the question of elevation-dependent warming within a mountain region, and summarizes findings of observed and modeled climate change in mountains with a focus on four mountain regions: The Swiss Alps, The Colorado Rocky Mountains, The Tibetan Plateau/Himalayas and The Tropical Andes (see Fig. 1). In section 4, we discuss potential mechanisms that might enhance temperature change in mountains. We discuss some key insights that emerge from these studies in the final section.

Table 1 List of studies that were reviewed to assess elevation sensitive warming within mountain regions. Also included are the information on time period, elevation range and number of stations considered in these studies

Study	Region	Time Period	Elevation Range (m)	No. of Stations
Diaz and Bradley 1997	Global	20th Century	1055–3310	126
Pepin and Lundquist 2008	Global	1948–2002	500–4700	1084
Pepin and Seidel 2005	Global	1948–2002	500–4700	1084
Seidel and Free 2003	Global	1960s–2000	2–3649	52; incl. radiosonde data
Beniston and Rebetez 1996	Swiss Alps	1979–1993	271–3572	88
Beniston et al. 1994	Swiss Alps	1901–1992	276–2500	4
Ceppi et al. 2010	Swiss Alps	1959–2008	200–3500	2 km gridded data
Jungo and Beniston 2001	Swiss Alps	1901–1999	271–3572	19
Clow 2010	Colorado Rockies	1986–2007	2560–3536	70
Diaz and Eischeid 2007	Colorado Rockies	1987–2006	1250–4000	4 km gridded data
Rangwala and Miller 2010	Southern Colorado Rockies	1895–2005	1763–3536	58
Pepin and Losleben 2002	Colorado Front Range	1952–1998	1059–3749	3
Chen et al. 2006b	Tibetan Plateau	1961–2000	1591–4670	63
Liu and Chen 2000	Tibetan Plateau	1955–1996	200–4801	197
Liu et al. 2006	Tibetan Plateau	1961–2003	2000–4500	66
Liu et al. 2009	Tibetan Plateau	1961–2006	0–5000	116
Lu et al. 2010	Tibetan Plateau	1960–2005	1000–5000	140
Qin et al. 2009	Tibetan Plateau	2000–2006	2000–5000	71; incl. satellite data
Rangwala et al. 2009	Tibetan Plateau	1961–2000	1000–5000	43
You et al. 2008	Tibetan Plateau	1961–2005	2100–4700	71
You et al. 2010	Tibetan Plateau	1951–2004	2100–4700	71
Bhutiyani et al. 2007	Indian Himalayas	1901–1989	1200–3800	10
Kothawale et al. 2010	Indian Himalayas	1901–2007	not available	12
Shrestha et al. 1999	Nepal Himalayas	1971–1994	72–3705	49
Vuille and Bradley 2000	Tropical Andes	1939–1998	0–5000	268
Vuille et al. 2003	Tropical Andes	1950–1994	0–5000	277

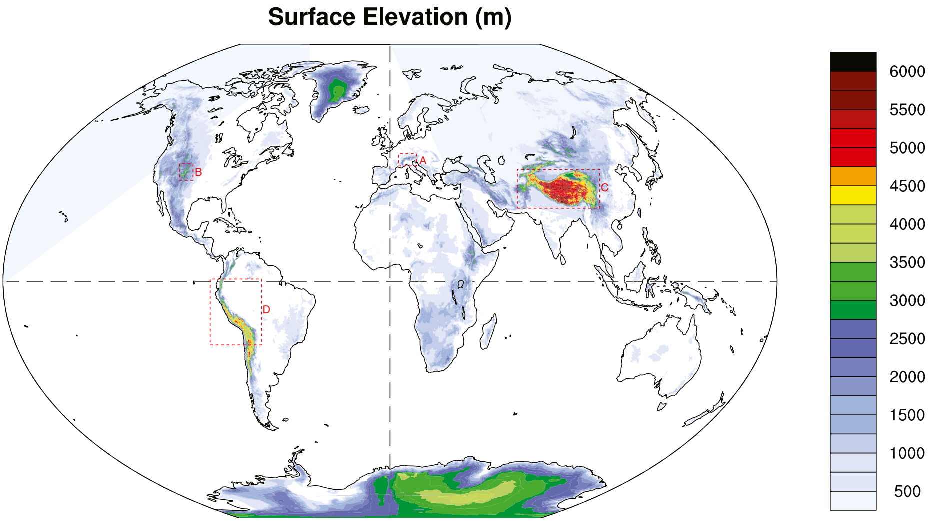

Fig. 1 Global land surface elevation (m). The dashed boxes and letters indicate the four mountain regions considered in this review: (A) the Swiss Alps, (B) the Colorado Rockies, (C) the Tibetan Plateau and the Himalayas, and (D) the Tropical Andes

2 Are mountains warming faster than other land areas globally?

This question has not been adequately addressed in the existing literature primarily because of the current state of climate monitoring and the relative sparsity of observations in mountainous regions. Several studies have suggested that mountain regions may be warming at higher rates globally (e.g., Messerli and Ives 1997; Beniston et al. 1997; Diaz and Bradley 1997). However, it also is suggested that long-term trends in mean annual temperatures in mountains are comparable to lowlands, although there are significant differences on seasonal and diurnal scales (e.g., Barry 2001). Figure 2 compares seasonal and annual temperature trends in the late 20th century between the hemispheric land and selected mountain regions covered in this study. For the Southern Hemisphere, we only examine the annual trends because of the lack of significant seasonality in the Tropical Andes. For the Northern Hemisphere, there are substantial differences in warming rates on a seasonal basis such that some mountain regions showing much greater warming trends in some seasons. In the Swiss Alps, warming rates are large in all seasons except autumn. For the Colorado Rockies and Tibetan Plateau, warming trends are largest in summer and winter, respectively. However, annual warming rates in the mountains are comparable to the associated hemispheric land averages.

Most studies have analyzed temperature trends in a specific mountain region and compared them to global or hemispheric trends (e.g., Beniston et al. 1997; Liu and Chen 2000; Ceppi et al. 2010). They often find that a mountain region is warming faster than the global or hemispheric average. However, these analyses cannot be used to answer the primary question posed in the title of this section because (a) the spatial scales being compared are vastly different with a much greater spatial variability expected across the global and hemispheric scales and, (b) the observation stations in mountain regions do not adequately sample the region, both geographically and topographically.

Seidel and Free (2003) performed a station-based comparison of temperature trends (1960s–2000) at a high and a representative low elevation site for 26 station pairs across

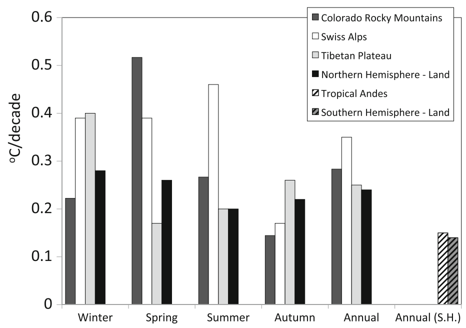

Fig. 2 Comparison of daily average temperature trends between the Northern Hemisphere (Land) and three high elevation regions (Colorado Rockies, Swiss Alps and Tibetan Plateau), and between the Southern Hemisphere (Land) and the Tropical Andes during the latter half of the 20th century. For the Southern Hemisphere, only the annual trends are compared because of the lack of significant seasonality in the Tropical Andes. Time period used to estimate trend magnitudes is slightly different based on the source study for each region. North-central Colorado Rocky Mountains: 1957–2006 (Ray et al. 2008); Swiss Alps: 1959–2008 (Ceppi et al. 2010); Central & eastern Tibetan Plateau: 1961–2004 (You et al. 2010); Tropical Andes: 1959–2006 (Vuille et al., 2008); Northern and Southern Hemisphere – Land: 1959–2008 (www.ncdc.noaa.gov). Seasons are defined as winter (DJF), spring (MAM), summer (JJA) and autumn (SON)

the globe. They found that in the tropics the mountain stations had higher warming rates relative to their low elevation counterparts, however the results were mixed for the extra-tropics. They state that the high elevation stations selected in their studies do not sufficiently represent the associated mountain region because they are often at relatively low elevation, rarely on mountain peaks and usually in topographic hollows.

One way to address this question using available observations might be to perform a regional-pair-based comparison, akin to the station-pair-based comparison by Seidel and Free (2003). For such a comparison at the global scale, a lower elevation reference region or regions must be appropriately identified relative to a specific mountain region such that the variability associated with geographical scales, latitudinality and continentality are comparable. In addition to examining this question for mean temperature trends, it may be preferable to assess trends in minimum and maximum daily temperatures separately because they can change at different rates and for different reasons.

3 Is there elevation-dependent warming within mountain regions?

3.1 Global

There are very few analyses that examine elevation-dependent warming trends in mountains at the global scale. Most studies focus on a particular region. In this section we summarize some of these global studies. Diaz and Bradley (1997) used observations from more than 100 sites between 30 and 70°N latitude and found that mean temperature warming rates were enhanced at many higher elevation sites between 1951 and 1989. Furthermore, they found that most of the increase was associated with increases in minimum temperatures, and that the trends in maximum temperatures were small. Beniston et al. (1997) indicate that enhanced warming rates at higher elevations may be more apparent in the tropics. Pepin and Seidel (2005) found that surface temperatures at the majority of high elevation stations across the globe are increasing faster than the free air temperature at the same elevation between 1948 and 2002. They also found less discrepancy between surface and free-air temperatures at mountain summits relative to mountain valleys. Overall, however, they found no elevation dependence of warming rates.

One of the more comprehensive global studies is that of Pepin and Lundquist (2008) who analyzed more than 1000 high elevation stations. They found no elevation dependency in temperature trends and concluded that this was, in part, because mountain summits and freely draining slopes show less variance in temperature trends than do incised valleys where local effects are more important. However, most of the stations analyzed in their study are below 2000 m. Their analysis suggested that the warming rates are strongest near the annual 0°C isotherm (Fig. 3), which implies a role of the snow/ice albedo feedback mechanism causing enhanced warming at these elevations.

Several modeling studies have addressed elevation-dependent warming. Diaz and Bradley (1997) suggest that models tend to simulate elevation-dependent warming in the mountains. Bradley et al. (2004) analyzed seven global climate model simulations and found large increases in free air temperature along the whole extent of the American Cordillera (Alaska to southern Chile) when atmospheric CO_2 levels were doubled. These free air temperature changes increased with elevation. They argued that such increases in the free air temperature could augment the warming rates at higher surface elevations. In one of the more comprehensive global studies, Nogues-Bravo et al. (2007) used five climate models at 0.5° resolution and four IPCC scenarios to examine projected climate change in mountains later this century. They found future rates of warming to be two to three times higher than recorded in the 20th century and larger temperature increases in higher northern latitude mountains than in temperate and tropical zones. However, they did not specifically address elevation-dependent warming. Some modeling studies have suggested that the increasing influence of snow/ice albedo feedback mechanisms, primarily during spring and summer, could be responsible for increased warming rates at high elevations (e.g., Giorgi et al. 1997). Although there are plausible reasons why elevation-dependent climate responses might arise, mountain systems are inherently difficult to understand owing to their complex topography that is still difficult to represent accurately in models.

3.2 The Swiss Alps

Historically, the Swiss Alps may be the best observed among high mountain regions (e.g., Beniston et al. 1997). However, even here the climate at higher elevations (above 3000 m) is

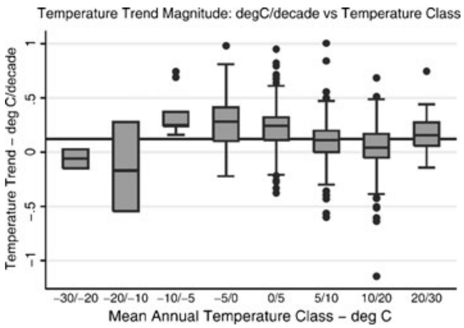

Fig. 3 Mean annual temperature trend magnitudes, between 1948–2002, for different mean annual temperature bands from 1084 mountain stations described in Pepin and Lundquist 2008 (Fig. 2c in Pepin and Lundquist 2008). Trend magnitudes are generally higher for regions associated with -5 – 0°C and 0 – 5°C temperature bands

not adequately observed. The 20th century warming in the Alps has been greater than the global and hemispheric (land and ocean combined) average (Beniston et al. 1994; Beniston et al. 1997; Ceppi et al. 2010). There is, however, a diurnal asymmetry in the warming because minimum temperatures have risen at a higher rate than maximum temperatures. In fact, the daytime maximum temperatures in the Swiss Alps experienced a cooling trend in the 1980s (Beniston et al. 1994).

Beniston and Rebetez (1996) reported altitudinal gradients in nighttime warming anomalies during winter. Using observations from 88 stations in the Swiss Alps, they found greater increases in the wintertime minimum temperatures at higher elevations during the late 1980s and early 1990s, a period that experienced warm, wet winters in connection with positive North Atlantic Oscillation (NAO) indices. However for an anomalously cold period between 1983 and 1987, when some of the years had negative NAO indices, they found a switch in the sign of the altitudinal gradient in the wintertime minimum temperature anomalies. They concluded that temperature variability generally increased with altitude owing to a damping out of temperature anomalies at lower elevations by local climatic effects. For the 1961–1999 period, Jungo and Beniston (2001) found greater increases in the minimum temperature in winter at high altitudes in the Swiss Alps, however, they found greater increases in the maximum temperature in summer at low altitudes on the north side of the Alps.

Giorgi et al. (1997) used a regional climate model (50 km spatial resolution) forced by doubled atmospheric CO_2 levels to simulate seasonal climatic changes in the Swiss Alps. Their analysis suggested larger increases in average surface air temperature at higher elevations during all seasons, although the elevation dependence was most pronounced in winter and spring. The surface elevation range in their model, however, was only between 100 and 2000 m. Both spring and winter experienced large decreases in snow depth at higher elevations. However, they found a large altitudinal gradient in surface absorption of solar radiation only for spring; a very weak gradient was simulated during winter. Their analysis suggests a strong snow albedo feedback mechanism to explain the greater springtime warming at higher elevations but no clear explanation for the wintertime warming simulated in their study.

Ceppi et al. (2010) used a 2 km×2 km gridded mean temperature data set, interpolated from 91 homogenized stations from the Swiss Alps, to examine trends between 1959 and 2008. Their analysis found high warming rates during summer (0.46°C/decade) and winter (0.40°C/decade). Using a statistical model, they also concluded that circulation changes did not significantly affect these trends except during winter when they could explain 50% of the variance. Seasonally, their analysis did not reveal altitudinal gradients in warming rates, although they did find larger temperature anomalies below 500 m in all seasons except spring. In spring, they also found larger trends at mid to high elevations in association with the 0°C isotherm.

3.3 The Colorado Rocky Mountains

Analysis of long-term trends in annual mean temperatures from the Colorado Rocky Mountains indicates large warming trends (0.5–1°C/decade) during the last three decades, but particularly since the mid-1990s (Diaz and Eischeid 2007; Ray et al. 2008; Saunders et al. 2008; Clow 2010; Rangwala and Miller 2010). The recent warming trend in the Colorado Rocky Mountain region appears to be among the largest in the contiguous United States (e.g., Saunders et al. 2008). This warming is found in all seasons although the largest increases in temperatures are observed mostly in winter and summer (Clow 2010).

Williams et al. (1996) and Pepin and Losleben (2002) reported decreases in mean annual temperatures at high elevation stations (Niwot Ridge) between 1952 and the mid-1980s, and increases since then. This cooling trend has been attributed to short-term climate oscillations. Rangwala and Miller (2010) found decreases in the maximum temperature during the same time period in the San Juan Mountains in southwestern Colorado; however, they also found a slight increasing trend in the minimum temperature during that period. They attributed the decreases in the maximum temperature, in part, to increases in the anthropogenic aerosol loading in the region. Between 1994 and 2005, they found similarly high annual warming rates (~1°C/decade) in the minimum and maximum temperatures. However, between 2006 and 2009, Rangwala and Miller (2011) found decreasing trends in the maximum temperature while no such trends were found in the minimum temperature. They suggest that the maximum temperature could be affected more by the interannual variability in precipitation, particularly in the cold season, as compared to the minimum temperature.

Long-term observations in the Rockies are available from NOAA's National Weather Service (NWS) cooperative stations. Since the 1980s, temperature observations are also available from the Natural Resources Conservation Service's (NRCS) Snow Telemetry (SNOTEL) stations that are located between 3000–3500 m elevations and are about 800 m higher relative to the NWS stations. Clow (2010) points out that the observed temperature increases at the SNOTEL sites in the Colorado Rocky Mountains between 1979 and 2006 (~1°C/decade) are substantially greater that those for the entire state of Colorado (0.4°C/decade) which are based on Ray et al. (2008) who only considered the NWS stations. Although Rangwala and Miller (2010) did not find any significant difference in annual warming rates between the SNOTEL and NWS stations within the San Juan Mountains, they did find greater warming rates at NWS sites in winter and at SNOTEL sites in summer.

Diaz and Eischeid (2007) used PRISM (Parameter-elevation Regressions on Independent Slopes Model) data to evaluate warming rates at different elevations (1000–4000 m) in the Colorado Rocky Mountains between 1979 and 2006. PRISM is a 4 km grid-based mapping of observations that includes NWS and SNOTEL sites (Daly et al. 2008). They found enhanced warming rates in both the minimum and maximum temperatures at elevations above 2000 m and also found an elevation-dependent gradient in the warming rates of both variables between 2000 and 4000 m with higher warming rates at higher elevations. However, caution must be exercised in interpreting these results because observations above 3000 m are limited, and since there are almost none above 3500 m, it is difficult to assess how well the PRISM data represent temperatures at these higher elevations. Gutmann et al. (2011) used a high resolution (2 km) model driven by historical boundary conditions to obtain a precipitation climatology for the Colorado Rocky Mountains and found significant differences, relative to PRISM, at higher elevations and away from an observation site.

3.4 The Tibetan Plateau and the Himalayas

The last decade has seen a rapidly growing interest in understanding climate change on the Tibetan Plateau and in the Himalayas, in part, because climatic changes there are likely to have significant impacts on water resources for a large percentage of the population in eastern and southeastern Asia. The Tibetan Plateau is occasionally referred to as the third pole because of the large expanse of snow and ice found there. Weather observations are available primarily for the central and eastern parts of the Plateau, and much less information is available for a large portion of the western half, which is relatively higher in elevation and may be more important in a hydrologic sense. The average elevation for the region is more than 4000 m.

The Tibetan Plateau may be among the most sensitive regions to the ongoing global climate change. Evidence suggests that the warming during the latter half of the 20th century started earlier (early 1950s) than the Northern Hemisphere trend (mid-1970s), and Niu et al. (2004) found that the region experienced a rapid temperature increase in the mid-1980s. For the 40-year period from 1955 to 1996, Liu and Chen (2000) found that the Tibetan Plateau experienced large warming rates in winter (0.32°C/decade) and autumn (0.17°C/decade). Even larger warming rates were reported for the central and eastern Plateau when more recent years were included (Duan and Wu 2006; Liu et al. 2006; Wang et al. 2008; Rangwala et al. 2009). The increases are thought to be associated with enhanced warming in spring and summer in the latter part of the 1990s.

There is significant seasonal variability in warming rates on the Plateau. Liu and Chen (2000) found that the warming rate in winter was about twice as large as the annual mean warming rate, a result which is consistent with other studies (Chen et al. 2006b; Liu et al. 2006; You et al. 2007; Rangwala et al. 2009). Autumn has the next highest warming rate, while summer and spring have only recently (since the 1990s) experienced significant increases in warming (Liu et al. 2006; Rangwala et al. 2009). Liu et al. (2006) also examined trends in minimum and maximum temperatures for the 42-year period from 1961 to 2003. They found that minimum temperatures were increasing faster than maximum temperatures in all months, but more so in winter and spring. The absolute increases in the minimum temperature were also highest in winter and spring. Temperature changes in the Tibetan Plateau are similar to those in the northern high latitudes where the warming rate during the last 50 years has also been greater in winter and spring, and smaller in summer (Serreze et al. 2000).

Although there is evidence of elevation-dependent warming within the Tibetan Plateau, the sparsity of data at the higher elevations limits our ability to interpret these results conclusively. Using an extensive selection of weather stations within the Tibetan Plateau, Liu and Chen (2000) found a differential increase in the rate of surface warming dependent primarily on the elevation of the observing station for the 1960–1990 period. Rangwala et al. (2010) found comparable trends of increases in warming rates with elevation on the Plateau from a climate model for the same historical period and during the 21st century (Fig. 4).

A recent study by You et al. (2010) did not find any significant elevation dependence in warming rates of mean temperature between 1961 and 2005. However, for the same period and considering mostly the same stations, Liu et al. (2009) found that the warming rates for

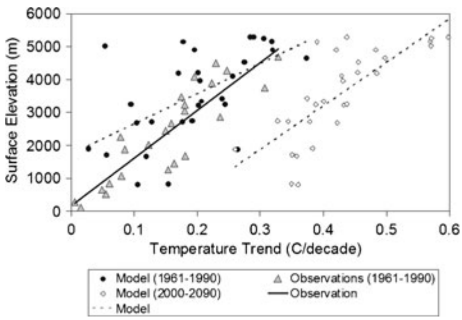

Fig. 4 Observed (1961–1990; from Liu and Chen 2000) and modeled (1961–1990 and 2000–2090) trends in surface temperature (C/decade) in the Tibetan Plateau as related to the elevation of the observing station and the model grid, respectively (Fig. 2b in Rangwala et al. 2010). Model results are based on 29 grid cells using updated 3°×4° version of Russell et al. (1995) model

daily minimum temperature were greater at higher elevations and more pronounced in winter and spring. You et al. (2008) examined correlation of different indices of temperature extremes with elevation but did not find any significant correlation. Possible reasons for the differences between these results could be that they were examining different variables and were using different stations.

One way to augment the sparsity of observations in high elevation regions is to use reanalyses. You et al. (2010) compared temperature trends in the Tibetan Plateau obtained using two different reanalyses with 71 homogenized surface stations. They found that one reanalysis (ERA-40) showed a general warming trend consistent with the surface-based trend, especially in winter. However, the trend did not appear in the other reanalysis (NCEP). They did not find any elevation dependency in temperature trends and concluded that different reanalyses can lead to different trends because of such factors as topographic differences between data sets and other reanalysis model differences.

Another way to increase the number of observations is to incorporate satellite data into the analysis. Qin et al. (2009) used MODIS monthly-averaged land surface temperatures to examine elevation-dependent warming in a wider portion of the Tibetan Plateau. They were able to supplement the surface observations in the western Tibetan Plateau and obtain temperature trends for regions above 5000 m. They first showed that the temperature trends from MODIS were consistent with surface-based trends where surface measurements were available. They then extended their results to the broader Tibetan Plateau and did find elevation-dependent warming with the warming rate increasing with altitude between 3000 m and 5000 m. However, they also found that above 5000 m, there was no additional enhancement of the warming rate. They suggest that a reason for the mixed results regarding elevation-dependent warming found in previous studies (i.e., some studies found it and others didn't) may have occurred because the surface-based stations were not sufficiently representative of the spatial heterogeneity of the region.

The Yunnan Plateau in southwestern China is a high elevation region located southeast of the Tibetan Plateau. The mean elevation of the Yunnan Plateau is above 2000 m with mountain peaks as high as 3700 m. Fan et al. (2010) analyzed records from 119 meteorological stations over the Yunnan Plateau and found that regional temperature there has been increasing at a rate of 0.3°C/decade for the last four decades. They also found that the warming trends are enhanced during winter, minimum temperatures are increasing faster than maximum temperatures, and increases are greater in the higher elevation regions. These results are consistent with most of the studies for the nearby Tibetan Plateau.

Observations from the Indian and Nepal Himalayas yield a different climate change narrative than found on the Plateau. Shrestha et al. (1999) and Bhutiyani et al. (2007) found that the maximum temperatures have been increasing at a greater rate than minimum temperatures in recent decades. These studies suggest that monsoonal circulation has a role in producing these trends. Shrestha et al. (1999) also found elevation dependence in the rate at which maximum temperatures were increasing in the Nepal Himalayas, with higher rates at higher elevations. They report that the warming was greater during the monsoon (Jun–Sep) and post-monsoon (Oct–Nov) months.

Kothawale et al. (2010) find that, between 1971 and 2007, the western Himalayan region of India has warmed at a higher rate (0.46°C/decade) than rest of India (0.20°C/decade). Similar to other studies from the Indian and Nepal Himalayas, they find that maximum temperatures are increasing at a faster rate (0.53°C/decade) than minimum temperatures (0.37°C/decade). Seasonally, winter is experiencing the highest warming rates with increases in maximum temperatures (0.82°C/decade) almost twice as large as increases in minimum temperatures (0.47°C/decade). Possible reductions in cloud cover, snow cover or precipitation during winter could cause such large increases in the maximum temperatures. However, there is some evidence that winter precipitation has a small upward (Archer and Fowler 2004) or insignificant (Bhutiyani et al. 2010) trend in this region during the latter half of the 20th century. Although, Bhutiyani et al. (2010) suggest that in recent decades a large proportion of winter precipitation has been falling as rain instead of snow on the windward side of the northwestern Himalayas. This implies a possible reduction in snow cover in winter that could cause increases in maximum temperature through the snow-ice albedo feedback mechanism.

3.5 The tropical Andes

The Andes Mountains in South America differ from the Alps, Rocky Mountains, and Himalayas in a number of ways, but one in particular is that they are aligned along the west coast of the continent. The southern Andes are in the westerly wind belt and directly affected by their proximity to the ocean, but the tropical Andes derive much of their moisture from the east. Since we found very few studies of the central and southern Andes, the focus here is on the tropical Andes, where many of the studies have been concerned with the melting of tropical glaciers. Ames (1998) reports that between 1932 and 1994, the ten monitored tropical glaciers of Peru have been retreating. Vuille and Bradley (2000) used temperature data from 268 stations ranging from 0 to 5000 m above sea level, to investigate the inconsistency between the observed glacier retreat and slight cooling trend in the lower tropical troposphere after 1979 as reported by Gaffen et al. (2000). They found that annual mean temperatures in the tropical Andes were increasing at a rate of 0.33°C/decade between 1975 and 2000, and that the warming rate, although positive, was reduced at higher elevations. However, they also found that warming rates did increase with altitude between 1000 m and 2500 m on the eastern slopes of the mountains. The warming rate is smaller when the record is extended further into the past; it is 0.11°C/decade for 1939–1998 (Vuille and Bradley 2000) and 0.15°C/decade for 1950–1994 (Vuille et al. 2003). Vuille et al. (2008) extended this record to include observations from 279 stations between 1939 and 2006, and reported a warming trend of 0.1°C/decade, which is similar to the trend found in NCEP-derived upper (500mb) air temperature increases in the last 50–60 years (Bradley et al. 2009).

At the larger scale, sea surface temperatures affect the circulation in the region. Vuille and Bradley (2000) indicated that during their 60-year record (1939–1998), all of the major warm anomalies in the tropical Andes coincided with El Nino events and all of the major cold anomalies coincided with La Nina events. Diaz and Graham (1996) found that the freezing level height (FLH) has been rising between 1958 and 1990 and noted that this rise was related to sea surface temperatures in the eastern tropical Pacific. A more recent study by Bradley et al. (2009) using upper air temperatures from the NCEP/NCAR reanalysis data set also found that FLH has been rising and that the interannual variability of FLH is controlled by the phase of ENSO variability.

There have also been modeling studies of the tropical Andes. Vuille et al. (2003) used an atmospheric GCM to better understand the reason for the rapid glacier retreat between 1950 and 1998. Their analysis indicated that temperature increases were more responsible for the retreat than changes in the hydrologic cycle. Their model simulation generally reproduced the spatial pattern of the observed warming which is larger on the western slopes. They attributed the increasing temperatures to warmer sea surface temperatures in the equatorial Pacific and changes in clouds and atmospheric water vapor. Urrutia and Vuille (2009) used a regional model and projected that there would be significant future warming in the tropical Andes and that the warming would be enhanced at higher elevations.

4 Mechanisms responsible for possible temperature enhancements

There are a number of mechanisms that can produce enhanced warming rates in certain elevation bands, and they often have a strong seasonal dependence. These mechanisms arise from either elevation based differential changes in climate drivers, such as snow/ice cover, clouds, water vapor, aerosols, and soil moisture, or differential sensitivities of surface warming to changes in these drivers at different elevations. Table 2 describes the specific responses and the physical mechanisms based on changes in these climate drivers, and we discuss some of these mechanisms in more detail.

4.1 Snow/ice albedo feedbacks

This is certainly one of the strongest feedbacks in the climate system, and it has a rapid response time. In response to a positive temperature anomaly, more snow melts thus decreasing the local albedo, which in turn leads to increased absorption of solar radiation and enhancement of the initial positive temperature anomaly. Since this feedback modulates the surface absorption of incoming solar radiation, it primarily affects changes in the maximum temperature. Increases in the minimum temperature are also possible if decreases in snow cover are accompanied by increases in soil moisture and surface humidity which can facilitate a greater diurnal retention of the daytime solar energy in the land surface and amplifies the longwave heating of the land surface at night (Rangwala et al. 2012).

This feedback should be strongest at elevations associated with the snow line or the 0°C isotherm, and it should be more important at lower elevations earlier in the cold season and more important at higher elevations later in the year. For the Tibetan Plateau, Rikiishi and Nakasato (2006) found that the length of the snow cover season has been declining at all elevations between 1966 and 2001, with the highest rate of decline in the 4000–6000 m range. However, they found that the largest decrease in the mean snow-covered area occurred at the lowest elevation (0–500 m). Modeling studies have suggested that the increasing influence of snow/ice albedo feedback mechanisms, primarily during spring and summer, are important in causing elevational gradients in warming rates (e.g., Giorgi et al. 1997; Chen et al. 2003; Liu et al. 2009; Rangwala et al. 2010). Global analysis of observed temperature trends in mountain regions by Pepin and Lundquist (2008) suggest that the warming rates are strongest near the annual 0°C isotherm (Fig. 3), which could imply a role of the snow/ice albedo feedback mechanism causing enhanced warming at these elevations.

4.2 Cloud cover

Clouds are still arguably the most uncertain component of global climate model simulations of greenhouse gas scenarios because of their impact on both shortwave and longwave radiation. Changes in cloud cover and cloud optical depth strongly modulate both insolation and longwave radiation at the surface. Daytime decreases in cloud cover and optical depth will enhance the maximum temperature; however such a trend at night may lower the minimum temperature. Conversely, nighttime increases in cloud cover will cause increases in the minimum temperature. Although it’s likely that changes in clouds are partly responsible for some of the temperature trends found in section 3, observational datasets of clouds are generally insufficient to resolve their local impact on climate. Assessing changes in clouds and quantifying cloud feedbacks will remain challenging in the near term.

For the Swiss Alps, Beniston and Rebetez (1996) found that stations located in valleys experienced lower increases in nighttime temperature anomalies during winter because of

Table 2 Description of mechanisms that can produce an elevation sensitive temperature response at the land surface which will be dependent on elevation based changes in the climate drivers. Superscripts refer to the following citations: [1]Albrecht 1989; [2]Dai et al. 1999; [3]Durre et al. 2000; [4]Hansen et al. 1997; [5]Pepin and Lundquist 2008; [6]Rangwala et al. 2009; [7]Twomey 1974

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[5]	≻ 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}; strongest effect when the cloud base is low[2]
Increases in **Cloud Cover** (Nighttime)	≻ Increases downwelling longwave radiation	All seasons but greater effects in winter	≻ Increases T_{min}
Increases in **Specific Humidity (*q*)**	≻ Increases downwelling longwave ≻ Downwelling longwave has high sensitivity to changes in q when q is less than 5 g/kg[6]	Primarily winter; smaller effects are possible in autumn and spring	≻ Increases T_{min}
Increases in **Aerosols: non-absorbing** e.g. sulfates	≻ Decreases surface insolation ≻ Increases cloud albedo[7] and cloud lifetime[1]	Dependent on seasonal emissions	≻ Decreases T_{max} ≻ Small increases in T_{min} when cloud lifetime is enhanced ≻ Effect is somewhat localized to near the emission source
Increases in **Aerosols: absorbing** e.g. black carbon, dust	≻ Decreases surface insolation but increases mid-tropospheric heating ≻ Decreases albedo of clouds ≻ Decreases albedo of snow on ground ≻ Decreases cloud cover[4]	Dependent on seasonal emissions and insolation	≻ Increases T_{min} ≻ Increases T_{max} when cloud cover is reduced ≻ Effect is somewhat localized to near the emission source
Increases in **Soil Moisture**	≻ Increases latent heat fluxes and decreases sensible heat fluxes during the day	Snowmelt effects are strongest in spring and winter; rainfall effects are strongest in summer	≻ Decreases diurnal temperature range[2] ≻ Strong T_{max} - soil moisture link in summer[3]

strong and persistent temperature inversions whereas stations located in regions (e.g. plateaus) exposed to stratus clouds, which trap outgoing infrared radiation, had much warmer nighttime temperature anomalies. Overall, they found that there was an altitudinal dependence of temperature anomalies that exhibited a linear trend except at lower elevations where changes in fog and stratus clouds affected the results. This is consistent with Ceppi et al. (2010) who found pronounced increasing temperature trends in autumn at low altitudes (below 800 m) in the Swiss Alps, which they ascribed to decreases in the frequency of fog duration that lead to increases in the incoming solar radiation and the daytime heating of the land surface.

For the Tibetan Plateau, Duan and Wu (2006) found that low level nocturnal cloud cover was increasing over the central and eastern parts of the Plateau between 1961 and 2003. They suggest that these increases explain part of the increases in minimum temperatures over the Plateau in the latter half of the 20th century. During the same time period, they found that daytime and total cloud cover were decreasing which could increase the absorption of solar radiation and enhance daytime warming.

High-resolution climate models are likely to provide one of the best ways to investigate cloud feedbacks in high-elevation regions. Liu et al. (2009) examined 116 weather stations and high-resolution climate model output under a greenhouse-warming scenario and found that the increases in monthly minimum temperatures were greater at higher elevations in the Tibetan Plateau. They suggested that this elevation dependency was, in part, caused by cloud-radiation effects. Another modeling study by Vuille et al. (2003) found that changes in cloud cover were responsible for some of the temperature changes in the Tropical Andes.

4.3 Water vapor modulation of longwave heating

The globally positive temperature perturbation produced by increasing levels of anthropogenic greenhouse gases (e.g., carbon dioxide) leads to more water vapor in the atmosphere that in turn absorbs and emits longwave radiation, thus enhancing the surface warming. Increases in surface specific humidity have been suggested to be partly responsible for a rapid increase in surface warming across central Europe (Philipona et al. 2005) and the Tibetan Plateau (Rangwala et al. 2009, 2010) in the late 20th century. These studies suggest that the increases in specific humidity cause significant increases in longwave downwelling radiation (LDR) producing a surface warming.

Although increases in LDR associated with increasing specific humidity occur globally, the sensitivity is non-linear and is particularly large when the initial humidity is low as found at high elevations during the cold season. Ruckstuhl et al. (2007) examined the seasonal relationship between observed surface specific humidity (q) and LDR at four different surface elevations in the Swiss Alps. They found that LDR has large sensitivities to q, particularly when q is below 5 g/kg. Such low values of q occur during cold seasons and more widely at higher elevations (Fig. 5). Based on these findings, Rangwala et al. (2010) examined a global climate model simulation for the Tibetan Plateau and found a LDR-q relationship similar to that of Ruckstuhl et al. (2007). They found that modeled LDR-q relationships obtained for the Tibetan Plateau had greater sensitivities at higher elevations, particularly during the cold season. Ruckstuhl et al. (2007) found similar relationships between LDR and q for “clear sky” and “all sky” conditions, although the sensitivity of LDR to q was found to be slightly enhanced for “clear sky” conditions.

The mid-latitude boundary layer at high altitudes is expected to be under-saturated in longwave absorption in the water vapor absorption lines. Therefore, an increase in near surface water vapor during winter when the specific humidity is lowest will cause a large increase in LDR at the surface (Rangwala et al. 2009 and 2010). Such humidity modulation of increases in LDR will primarily cause increases in the minimum temperature. This process might be, in part, responsible for higher warming rates found during winter in the upland areas, globally, in both the observations (e.g., Liu and Chen 2000; Jungo and

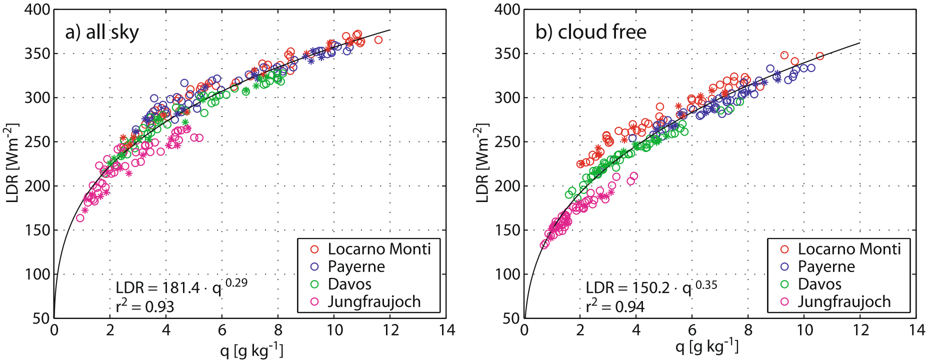

Fig. 5 Relationship between downwelling longwave radiation (LDR) and surface specific humidity (q), for (a) all sky and (b) cloud free conditions, obtained from observations from the Swiss Alps from four station observations at different surface elevation: Locarno-Monti (388 m), Payerne (498 m), Davos (1598 m) and Jungfraujoch (3584 m) (Fig. 5a,b in Ruckstuhl et al. 2007)

Beniston 2001; Holden and Rose 2011; Fan et al. 2010; Pederson et al. 2010) and climate models (e.g., Giorgi et al. 1997; Chen et al. 2003; Rangwala et al. 2010).

4.4 Aerosols

4.4.1 Absorbing aerosols: black carbon

Aerosols, black carbon (soot), and dust are additional contributors to warming. During the boreal spring, an atmospheric layer of dust from deserts and locally-emitted black carbon can be found up to 5 km high in the Indo-Gangetic Plain against the foothills of the Himalayas and Tibetan Plateau (Ramanathan and Carmichael 2008). Lau et al. (2010) used the NASA finite volume climate model to examine the potential impacts of this layer. They found that it absorbs solar radiation and warms the mid-troposphere, which in turn increases the rate of spring snowmelt and leads to enhanced warming of the atmosphere-land system. Their model results indicate that the increased heating occurs primarily because of changes in sensible and latent heat fluxes since the changes in shortwave and longwave radiation tend to offset each other. In a recent review paper by Ramanathan and Carmichael (2008), they suggest that black carbon in the Himalayan Mountains arising from anthropogenic activities might be responsible for half the total warming there during the last several decades. Because black carbon affects the radiation budget in two ways – it absorbs radiation in the troposphere and decreases the surface albedo when deposited on snow – it is very difficult to assess its effect on elevation-dependent warming. Depending on the elevation at which black carbon is deposited it could either contribute to enhanced or reduced warming with elevation during the melt season. Xu et al. (2009) suggest that atmospherically deposited black carbon can increase the absorption of visible radiation by 10–100% in the Tibetan Glaciers. Black carbon can also causes decreases in cloud cover and affect the radiation budget at the surface (Hansen et al. 1997).

4.4.2 Absorbing aerosols: dust

Similar to black carbon, dust absorbs radiation within the atmosphere and reduces surface albedo when deposited on snow. Painter et al. (2007) analyzed several snow events in the Colorado Rocky Mountains and found that strong wind events in spring transport large amounts of dust from arid regions to the west. This causes a significant reduction of surface albedo when snow melts during spring and increases the absorption of solar radiation and the rate of snowmelt. They suggest that this process could have a much greater impact on regional climate change than the longer-term snow-albedo feedback mechanism would without the dust deposition. For this particular region, the surface is particularly vulnerable to movement by wind because human activities and development perturb the soil, which would otherwise be more crusted and resistant to wind transport. Similarly to black carbon, the impact of dust on elevation-dependent warming will depend on the elevation at which it is deposited. During the melt season, we would expect enhanced warming at elevations where the dust is deposited.

4.4.3 Non-absorbing aerosols

Reflective aerosols (e.g., sulfate aerosols) can affect local heat budgets by reducing the incoming solar radiation. These aerosols also interact with clouds and increased concentration of these aerosols can increase cloud reflectivity (Twomey 1974) and lifetime (Albrecht 1989). To the extent that the spatial variability of atmospheric aerosol concentrations can be attributed to elevation, aerosols could either enhance or reduce heating rates in mountains. There is still very little evidence in the literature about the role of reflective aerosols in mountain regions. Rangwala et al. (2010) found some evidence that aerosol concentrations have been higher at lower elevations in the Tibetan Plateau. This would cause the rate of increase of daytime maximum temperatures to be lower at these lower elevations and consequently contribute to an elevation gradient in warming rates.

5 Discussion

In this review, we have explored available literature to address two important questions related to climate change in the mountain regions: (1) are mountain regions warming faster than low lying regions, and (2) is there an elevation-dependent climate response within mountain regions? From the available studies, it remains difficult to sufficiently assess whether mountains have warmed at a higher rate than the rest of the global land surface primarily because we lack adequate observations to resolve it conclusively. However, available observations suggest that some mountain regions may be experiencing higher warming rates on seasonal time scales.

To explore the impacts of elevation on rates of temperature change within mountain regions, we have synthesized several important studies from four high mountain regions across the globe. It should be noted that the different regions considered in our study are vastly different in area, which allows for considerable geographical variability. Three of these regions are in mid-latitudes and one is in the tropics. However, the discussion of physical mechanisms reviewed here is applicable globally.

Table 3 summarizes some key findings associated with these studies of elevation-dependent warming in mountain regions. A majority of these studies suggest an elevation-dependent climate response in both observations and climate models. Some studies also suggest strong seasonality to the elevation-dependent response, particularly for the minimum temperature increases during the cold season and for the maximum temperature during the warm season. Although the collective evidence from these studies does not allow us to make global generalizations, a large number of observational and modeling studies suggest that elevation-dependent climate responses do occur under specific spatial and temporal conditions.

Table 3 Compilation of the results from studies that investigated altitudinal gradient in warming rates. Superscripts refer to the following citations: [1]Bhutiyani et al. (2007), [2]Beniston and Rebetez (1996), [3]Ceppi et al. (2010), [4]Chen et al. (2003), [5]Diaz and Bradley (1997), [6]Diaz and Eischeid (2007), [7]Giorgi et al. (1997), [8]Liu and Chen (2000), [9]Liu et al. (2009), [10]Lu et al. (2010), [11]Pepin and Lundquist (2008), [12]Qin et al. (2009), [13]Rangwala et al. (2009), [14]Rangwala et al. (2010), [15]Seidel and Free (2003), [16]Shrestha et al. (1999), [17]Vuille et al. (2003), [18]Vuille and Bradley (2000), [19]You et al. (2010)

Altitudinal gradient in the warming rate	Observations			Models		
	T_{min}	T_{max}	T_{avg}	T_{min}	T_{max}	T_{avg}
Increases with elevation	Annual[2,5a] Winter[2,6,9]	Annual[2,16]	Annual[8,12,15b]	Annual[9] Winter[9]	–	Annual[14] Winter[3,4,7,14]
	Spring[13] Autumn[13]	Summer[13]	Seasonal[13]	Spring[9]		Spring[3,7,14]
Decreases with elevation	Winter[2]	Winter[13]	Annual[10,18] Winter[3] Autumn[3]	–	–	
No significant gradient	–	Annual[1]	Annual[3,19] Seasonal[17,19]	–	–	Annual[3]
No significant gradient but largest warming rates associated with 0°C isotherm			Annual[11] Spring[3]			Spring[3]

[a] No significant gradient but greater warming at higher elevations relative to regions between 0–500 m

[b] Greater warming at higher elevations in tropics and a mixed result for extratropics

As discussed in the previous section, there are plausible reasons why elevation-dependent climate responses might arise. These include elevation based differential changes in climate drivers, such as snow/ice cover, clouds, water vapor, aerosols, and soil moisture, or differential sensitivities of surface warming to changes in these drivers at different elevations. However, mountain systems are inherently difficult to understand owing to their complex topography, which leads to a high level of spatial and temporal variability in their climatic responses. Both observations and models are currently inadequate to provide us with a definitive understanding of the climate change signal in high elevation regions. There is a serious deficiency in weather observations along elevation gradients with generally poor observations in mid to high elevations, and only a few climate variables are usually observed. Satellite retrievals can help fill in some of the missing gaps in space and time as well as provide additional climate variables.

On the other hand, climate models can provide many more climate variables to explore feedbacks within the system, but they don't simulate realistic local-scale atmospheric processes until resolved at 1–6 km scale (Rasmussen et al. 2011). Simulations at these scales are computationally intensive and only recently have realizations been made at such scales. In the near future, these simulations should improve our understanding of elevation-based climate sensitivities in mountains. Nonetheless, we will still be challenged by the inadequacy in our observations to validate these simulations.

One reason why we cannot reconcile all of the studies summarized in this paper is that the analyses vary. For example, we cannot compare a study that analyzed averaged daily temperature with another that examined daily minimum temperature because these parameters respond differently to different climate drivers. For the same reason, we cannot compare annual trends from one study with seasonal trends in another. It also becomes more difficult to generalize when a climate variable is less sensitive to a specific climate driver, but instead is influenced by several climate drivers. Then there is a greater likelihood that the elevation-based signal will be lost in the noise. For example, it may be more likely to find elevation sensitivity in minimum temperatures in winter but not find such sensitivity in annual mean temperature for the same region. Furthermore, differences can arise from the location of observation stations within the complex topography of the mountain region, and on a larger scale from the location of the mountain region in relation to rest of the continent.

When investigating elevation-dependent sensitivities to climate change, it may be preferable to examine more sensitive variables such as the daily minimum and maximum temperatures for a particular season. Moreover, such variables are likely to be more relevant for studying the impacts of climate change on ecosystems. For example, changes in wintertime minimum temperatures affect the survival of the bark beetle and its impact on the conifer forest in the Rocky Mountains (e.g., Kurz et al. 2008), or the impact of increases in summertime maximum air temperature on stream temperatures can affect the survival of a particular species of fish (e.g., Merten et al. 2010).

Additional difficulties in reconciling all of the studies here arise because of the spatial and temporal differences in the studies and the influence of climate variability. Both of these differences introduce statistical uncertainties because we know that neither spatial nor temporal changes in temperature will be uniform globally. In some regions and during specific periods, temperatures will be increasing faster than the global average, while in other regions they will be increasing more slowly or even decreasing. There are similarities between some of the mechanisms associated with enhanced warming rates in mountains and enhanced warming rates in the Arctic region. Both have significant snow and ice at the surface, are very cold in winter, and have generally low concentrations of atmospheric water vapor in winter. The snow/ice-albedo feedback is important in the Arctic Ocean, and most of the mountain studies that address it find it to be important there, too. Chen et al. (2006a) showed that increasing water vapor in the Arctic could enhance winter warming rates because the sensitivity of downward longwave radiation to water vapor is much greater when the atmosphere is drier, as commonly found in winter. A similar effect can occur in high mountain regions in winter (Ruckstuhl et al. 2007; Rangwala et al. 2010).

Our review suggests that high elevation regions will remain sensitive to the projected warming during the 21st century and that we will need to improve our understanding of how different climate drivers influence changes in high mountain regions. An essential requirement for this is to increase climate monitoring of high elevations sites. This should also include monitoring of a greater number of climate parameters that help to better assess energy fluxes and moisture availability at the land surface. Greater use of remote sensing tools and high-resolution climate modeling will be required to augment the ground observations. Nevertheless, it will still require a comprehensive ground observation network to support and validate these products.

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