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Contrasting long-term alpine and subalpine precipitation trends in a mid-latitude North American mountain system, Colorado Front Range, USA

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Background: Long-term climate trends in mountain systems often vary strongly with elevation.
Aims: To evaluate elevation dependence in long-term precipitation trends in subalpine forest and alpine tundra zones of a mid-continental, mid-latitude North American mountain system and to relate such dependence to atmospheric circulation patterns.
Methods: We contrasted 59-year (1952–2010) precipitation records of two high-elevation climate stations on Niwot Ridge, Colorado Front Range, Rocky Mountains, USA. The sites, one in forest (3022 m a.s.l.) and the other in alpine tundra (3739 m), are closely located (within 7 km horizontally, ca. 700 m vertically), but differ with respect to proximity to the mountain-system crest (the Continental Divide).
Results: The sites exhibited significant differences in annual and seasonal precipitation trends, which depended strongly on their elevation and distance from the Continental Divide. Annual precipitation increased by 60 mm (+6%) per decade at the alpine site, with no significant change at the subalpine site. Seasonally, trends at the alpine site were dominated by increases in winter, which we suggest resulted from an increase in orographically generated precipitation over the Divide, driven by upper-air (700 hPa) north-westerly flow. Such a change was not evident at the subalpine site, which is less affected by orographic precipitation on north-westerly flow.
Conclusions: Elevation dependence in precipitation trends appears to have arisen from a change in upper-air flow from predominantly south-westerly to north-westerly. Dependence of precipitation trends on topographic position and season has complex implications for the ecology and hydrology of Niwot Ridge and adjacent watersheds, involving interactions among physical processes (e.g. snowpack dynamics) and biotic responses (e.g. in phenologies and ecosystem productivity).

Keywords: alpine tundra; elevation-dependent climate change; high mountain regions; Niwot Ridge LTER; precipitation orographic ratio; precipitation seasonal cycle; snow water equivalent; subalpine forest; synoptic circulation

Introduction

Climate change in high mountain regions

The ecology and hydrology of high mountain regions are sensitive to regional climate change, as are human societies relying on resources provided by these systems (Baron et al. 2000; Inouye et al. 2000; Barry 2002; Williams et al. 2002; Millar et al. 2004; Huber et al. 2005; Beniston 2006; Löffler et al. 2011; Gurung et al. 2012). Changes in mountain climates often substantially differ from those in adjacent lowlands and broader regional signals, as seen in both observations and model simulations (Beniston et al. 1997; Giorgi et al. 1997; Fyfe and Flato 1999; Urrutia and Vuille 2009; Rangwala and Miller 2012; Mountain Research Initiative EDW Working Group 2015). Elevation dependence in long-term (multidecadal to centennial) trends has been documented extensively for surface air temperature in the instrumental record (e.g. Diaz and Bradley 1997; Pepin and Lundquist 2008) and less so for other variables, such as precipitation, wind speed and specific humidity (Vuille et al. 2003; Mote 2004; Knowles et al. 2006; Shenbin et al. 2006; Rangwala et al. 2009; McVicar et al. 2010).

Topographic dependence in precipitation trends

Elevation dependence in precipitation trends derives from atmospheric dynamics that control the spatial distribution of precipitation in mountain systems. These dynamics arise from a diverse array of cloud to synoptic-scale processes in complex interaction with topography (Egger and Hoinka 1992; Roe 2005; Barry 2012). In this interaction, orographic precipitation generation develops from processes including mechanical lifting of stable flow over a barrier and convective instability forced by mechanical or thermal lifting (the latter driven by solar heating of high terrain) (Roe 2005). Such mesoscale vertical motion is also forced by synoptic-scale flow. In the mid-latitudes, key synoptic dynamics are (1) lifting associated with warm and cold fronts within mid-latitude cyclonic storms and (2) upper-

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air (e.g. 500 hPa pressure level) flow associated with the mid-latitude jet stream which dynamically supports/suppresses lifting on the east side of upper-air troughs/ridges (Barry and Chorley 2010, 247–249). The result of such complexity is that different precipitation-generating mechanisms tend to produce characteristic precipitation–elevation profiles (e.g. Lundquist et al. 2010).

The location, amount and form (e.g. snow, rain) of orographic precipitation depend, by definition, on fixed topographic and variable atmospheric factors. Key topographic features include orientation and slope of mountain flanks relative to moisture-bearing atmospheric flow, mountain crest elevation relative to depth of these flows and topographic complexity (Barry 1994; Smith et al. 2003; Daly et al. 2008). Major atmospheric factors include air mass layer depth, direction and rate of air flow, moisture source strength, and vertical profiles of temperature and water vapour (determining column stability and condensation and freezing levels) (Barry 1994; Pandey et al. 1999; Roe 2005). Time-averaged (e.g. monthly mean) elevation profiles of orographic precipitation vary strongly with latitude, continentality and time of year, reflecting corresponding geographic and seasonal differences in the type and frequency of air masses and of synoptic circulation patterns which determine the movement and subsequent modification of air masses over a region (Barry 1990). In mid-latitude winter, precipitation–elevation profiles characteristically increase monotonically on the windward side (or with an optimum before the mountain crest) and show some degree of spillover onto the lee (e.g. Colton 1976; Barry 1990, 1992; Wratt et al. 2000).

Precipitation trends arise from interannual and multidecadal changes in synoptic circulation patterns and hemispheric circulation regimes. These changes alter time-averaged precipitation–elevation profiles by shifting, for example, the frequency, intensity, vertical structure and source region of mid-latitude cyclonic storms (Beniston 1997; Giorgi et al. 1997; Losleben et al. 2000b; Dettinger et al. 2004; Roe 2005; Lundquist and Cayan 2007). As a result, within a mountain system, locations that differ in terms of elevation, ridgeline versus valley topographic position and windward versus leeside position have been shown to exhibit different precipitation long-term trends (Changnon et al. 1993; Mote 2003, 2004; Rangwala et al. 2009).

Re-evaluating high-elevation historical climate change in the Colorado Front Range

The subalpine and alpine climates of Niwot Ridge, a high-elevation ridge (3000–4000 m a.s.l.) in the Front Range, Colorado Rocky Mountains, USA, have been shown to have strong elevational differences in precipitation and surface air temperature trends. These differences include switching trend sign with elevation, both regionally (with decoupling extending down to the adjacent Central Great Plains, 1300–1600 m a.s.l.) and, within the landscape, between alpine and subalpine zones (Williams et al. 1996; Pepin 2000; Greenland and Losleben 2001; McGuire et al. 2012). Such regional and within-landscape decoupling is in both long-term trends and interannual variability (Williams et al. 1996; Greenland and Kittel 2002). For example, on Niwot Ridge from the mid-twentieth century to the early twenty-first century, annual average maximum temperature (T_{max}) increased in the subalpine by 0.4°C decade^{-1}, but did not change in the alpine; at the same time, annual average minimum temperature (T_{min}) did not change at either location (1952–2010 trends, Kittel et al. 2011; McGuire et al. 2012). Elevational differences in long-term T_{max} trends were also season-dependent, with the strongest warming occurring in the subalpine in spring and summer and cooling occurring in the alpine in early winter (October–December) (Kittel et al. 2011; McGuire et al. 2012). The contrast between high mountain versus adjacent lowland trends has also been reported for elsewhere along the Front Range (Brown et al. 1992).

Here, we contrast long-term (1952–2010) trends in precipitation for an alpine site and a subalpine site on Niwot Ridge. These sites are closely located (within 7 km horizontally, ca. 700 m vertically), but are distinct in their position on the landscape. Key physical differences between the two sites are (1) distance from the mountain system crest, that is, the Continental Divide (ca. 3 vs. 10 km east, respectively) and (2) alpine tundra versus subalpine forest vegetation. With a recent upgrade of station records and the benefit of a longer observation period than that used in previous studies, we reanalysed Niwot Ridge precipitation data to assess whether significant long-term trends (1) occur in the extended record, (2) differ over the elevation span (700 m) from the subalpine to alpine, (3) vary seasonally and (4) are linked to landscape- and continental-scale atmospheric processes.

Materials and methods

Climate stations

Instrumental records of daily and monthly precipitation were evaluated for the two high-elevation sites on Niwot Ridge: ‘C1’ in the subalpine at 3022 m a.s.l. and ‘D1’ in the alpine at 3739 m a.s.l. (Figure 1, Table 1). These stations were established in the early 1950s as part of a climatic and ecological transect originating in the lower foothills of the Colorado Front Range (Marr 1961, 1967; Barry 1973). C1 is located in a closed-canopy subalpine conifer forest (mean height ca. 11 m, leaf area index ca. 4) (Monson et al. 2002). The site is on locally level terrain on the south-eastern flank of the Ridge, 9.7 km east of the Divide (Figure 1). In contrast, D1 is in low-stature alpine tundra (<5 cm tall, leaf area index < 0.5) (Bowman and Seastedt 2001) and is sited on a narrow, exposed ridge on the westernmost part of Niwot Ridge, 2.6 km east of and ca. 200 m lower in elevation than the Divide (Figure 1, Table 1). Both sites experience strong westerly winds from the Divide in autumn through spring (Figure 2) (Erickson

Table 1. Geographic attributes of Niwot Ridge long-term climate stations, Colorado Front Range.

		Location			Elevation difference (m) down from		Distance (km) east from	
Station	Ecological zone	Latitude (ºN)	Longitude (ºW)	Elevation (m)	Continental Divide*	D1	Continental Divide	D1
D1	Alpine tundra	40.06	105.62	3739	200	–	2.6	–
C1	Subalpine forest	40.04	105.55	3022	900	717	9.7	6.7

Note: *Based on a generalised Divide elevation of 3950 ± 100 m.

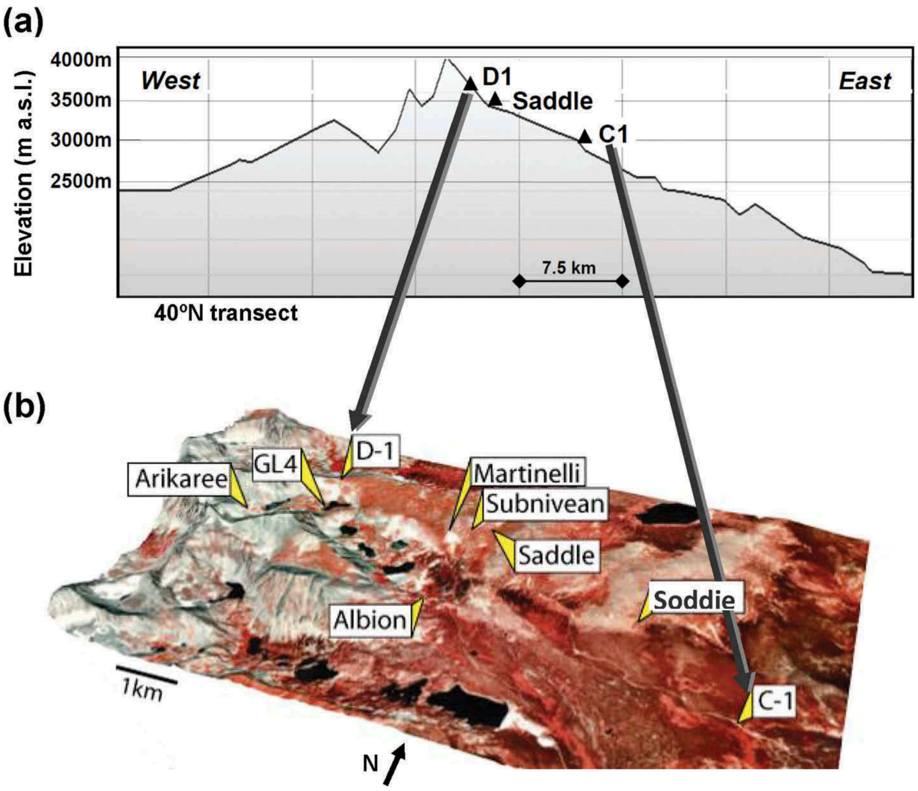

Figure 1. Location of the subalpine forest (C1) and alpine tundra (D1) climate stations at Niwot Ridge, in context of (a) a west-east topographic profile of the Colorado Front Range at ca. 40° N latitude (base profile from Google Earth) and (b) local geomorphology and vegetation cover. In (b), vegetation cover is shown in a colour infra-red image draped over 3D topography (red and brown indicate denser photosynthetically active vegetation cover; grey indicates minimal vegetation/bare ground and black indicates waterbodies) (Williams et al. 1999). In (a) and (b), other labelled locations include research sites for the Niwot Ridge Long Term Ecological Research Program and other ecological, hydrologic and geomorphological studies (Bowman and Seastedt 2001).

et al. 2005; Blanken et al. 2009). The role of climate variability and climate change on the ecology, hydrology, and biophysical and biogeochemical fluxes has been stressed in synthetic studies across the Niwot Ridge landscape (e.g. Bowman and Seastedt 2001; Seastedt et al. 2004; Williams et al. 2009; Schmidt et al. 2015; Suding et al. 2015; Knowles et al. 2015b).

Station instrumentation and histories. Climate data have been collected at the C1 (subalpine forest) and D1 (alpine tundra) sites almost continuously from 1952 to the present. At C1, precipitation was recorded using an unshielded US Weather Bureau standard totalising gauge (with observations manually recorded on an approximately weekly basis) through 1964. The gauge was located in an open area with sparse tree cover adjacent to the forest proper. Starting in late 1961, a Belfort Universal weighing-bucket gauge (with chart recorder) has been operated in a nearby 8-m diameter clearing in the forest, with the forest providing natural shielding for the gauge. At D1, an unshielded Weather Bureau totalising gauge was operated through 1969. Since 1965, observations have been recorded using a Belford weighing-bucket gauge with an Alter-type shield encircled by a Wyoming-snow fence. We evaluated the station records for inhomogeneities arising from these instrument and siting changes and, where possible, adjusted the records as follows.

To account for the switch in the D1 precipitation record from the unshielded Weather Bureau gauge to the shielded Belfort gauge, we developed monthly adjustment factors based on 5 years of overlapping

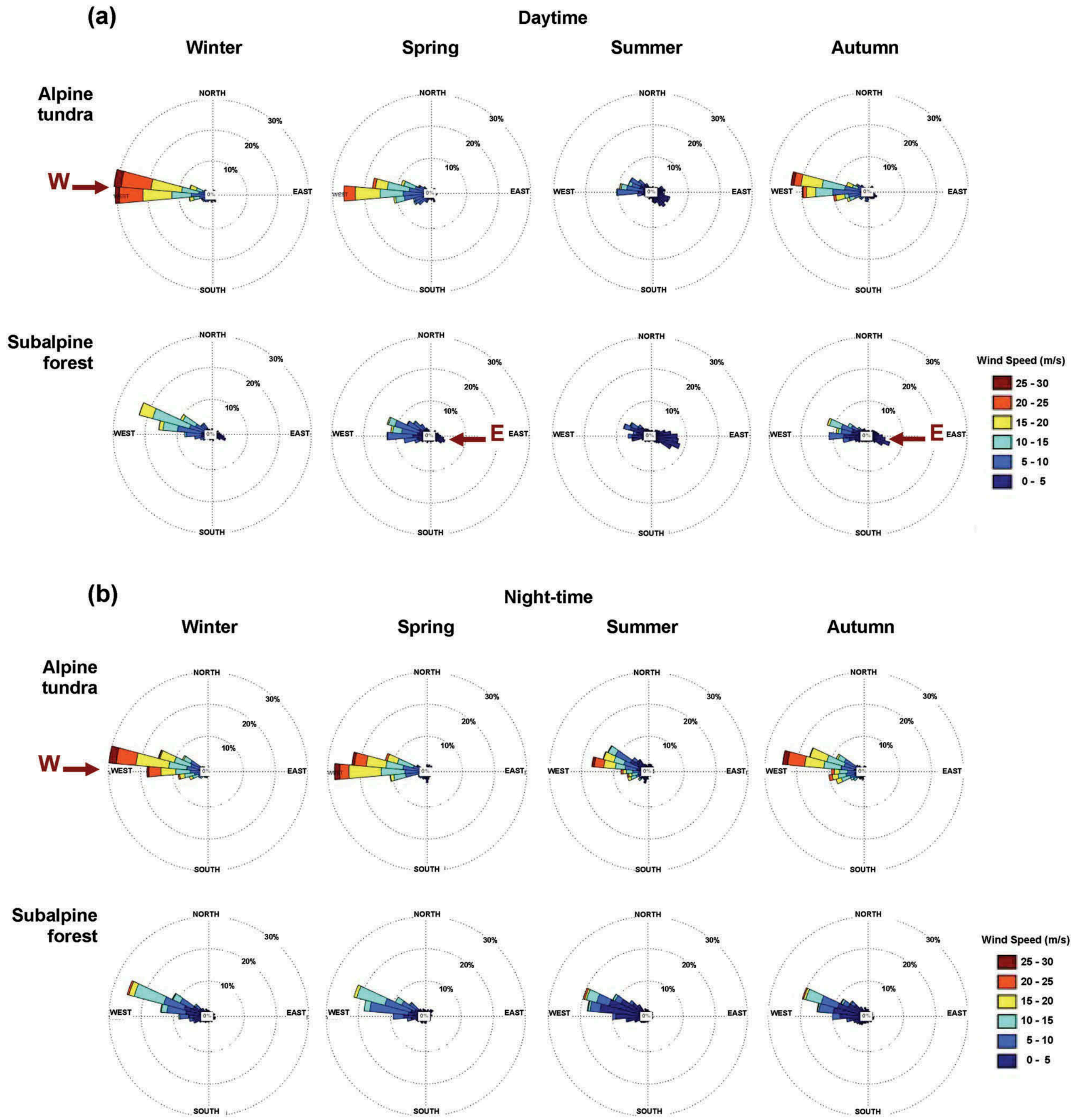

Figure 2. Wind roses showing greater intensity and frequency in the alpine zone of westerly and west-northwesterly near-surface winds (arrows labelled 'W') for winter, spring and autumn than in the subalpine zone and, in (a), the occurrence in the subalpine zone of daytime upslope easterly winds in spring and autumn (arrow labelled 'E' in [a]) versus in the alpine zone. Wind roses are for (a) daytime (08:00–16:00 local time) and (b) night-time (20:00–04:00) frequency of near-surface (3 m height) winds by direction and speed (m s^{-1}) by season for alpine and subalpine Niwot Ridge sites in Blanken et al. (2009) over a year (June 2007–July 2008). The alpine site is at an elevation of 3480 m a.s.l. and ca. 2.5 km east of the alpine station (D1); the forest site is at 3050 m a.s.l. and located near the subalpine station (C1). Seasons are winter (December, January, February), spring (March, April, May), summer (June, July, August) and autumn (September, October, November). The bars extend out from the centre towards the direction from which winds come; north is at the top of the roses. The frequency scale (concentric circles) is in 10% steps (0–30%); the wind speed scale is in intervals of 5 m s^{-1} (dark blue, 0–5 m s^{-1}; dark red, 25–30 m s^{-1}). Winter and summer data are replotted from Blanken et al. (2009).

observations, following the regression method of Yang et al. (1999). These factors were strongest for winter months, reflecting significant snow undercatch by the unshielded gauge, consistent with catch biases seen elsewhere (Yang et al. 1998). The D1 Weather Bureau gauge data were modified by these factors to bring the early record (1952–1964) into line with the more recent Belfort record.

We similarly assessed the impact of changing the location and instrumentation of the C1 precipitation station in the 1960s and found that no seasonal adjustment factors were required (Appendix A). As part of this assessment, we also evaluated the record for a step change just prior to and after the station change transition period. In the cold season (October–April), when precipitation observations in the more open site would have been subject to snow undercatch, we found no significant shift in precipitation means ($P = 0.89$, t-test for means, water years 1957–1961 vs. 1965–1969). Nonetheless, because of the nature of these station changes, the possibility of an uncorrected inhomogeneity in the subalpine precipitation record in the early 1960s should be kept in mind when interpreting trend results.

Data processing

We reprocessed the long-term precipitation station data set used in previous analyses (Barry 1973; Greenland 1987, 1989; Williams et al. 1996, 2002). Processing included quality checks, station change corrections, missing-data infilling, parsing to daily resolution as needed and archiving. Major improvements to the data set were the following:

- Recovery and incorporation of early missing written and analogue chart precipitation observations.
- Reassessment of station record inhomogeneities, resulting in revised adjustments of records where possible.
- Revised statistical infilling of missing observations based on daily regressions with records from nearby stations (Appendix B), resulting in 2% and 6% of daily values infilled in the C1 and D1 records, respectively.
- Application of the same daily statistical approach to parse multiday-accumulated precipitation totals to daily values.

These processes resulted in complete, gauge-corrected, daily and monthly precipitation records for the C1 and D1 sites from 1952 through 2010. These data are made available online by the Niwot Ridge Long Term Ecological Research Program (NWT LTER) (http://culter.colorado.edu/NWT/).

Trend analysis

Our analysis of trends combined linear-trend calculation and non-parametric significance testing (as implemented in Suppiah and Hennessy 1998; Ray et al. 2008). Trend values were determined as the slope from least-squares linear regression analysis. Statistical significance was evaluated by using the Mann-Kendall test for trends, a non-parametric test based on Kendall's tau rank correlation (Press et al. 1992; ITT Visual Information Solutions 2006; and Wessa 2012). Trends were evaluated for the full record (1952–2010) unless stated otherwise. Water year (WY) results include data from the start of winter in a given year continuing into the next calendar year, for example, October 1982–April 1983 values are reported for WY1983. As temporal autocorrelation can lead to over- or understatement of Mann-Kendall trend significance levels, we evaluated precipitation time series for autocorrelation at lag – 1 year (r_{-1}) (Harcum et al. 1992; Yue et al. 2002); we indicate in the figures the few instances where autocorrelation was strong enough to possibly influence results one way or the other.

The main effects of seasonality and site differences on trends were evaluated by using two-factor (months × sites) analysis of variance (ANOVA) without replication. We also evaluated elevation dependence in trends using the ratio of alpine (D1) to subalpine (C1) precipitation, referred to as the 'orographic ratio' (Dettinger et al. 2004). Serial correlations between selected variables and between the two sites over the record were tested using Pearson's product-moment correlation coefficient and lag-lead cross-correlation analysis (Wessa 2011). For comparison of subalpine precipitation and snowpack trends, we analysed snow water equivalent (SWE) observations from a SNOTEL snow course, also located at C1 ('Niwot', Site 663/Station 05J42S, http://wcc.sc.egov.usda.gov/nwcc/site?sitenum=663; elevation 3021 m a.s.l.). For a precipitation–snowpack comparison for the alpine zone, we used snow depth measurements from a long-term snow-fence experiment from 'Saddle' site, another Niwot Ridge alpine site (Figure 1) (Litaor et al. 2008).

We specified seasons based on calendar months, for example, with the 'cold season' defined as October–April and the 'warm season' as May–September. These seasonal designations have a climatological basis. During the cold season, record-average monthly mean temperature minima were less than −2°C at the subalpine (C1) site and less than −5°C at the alpine (D1) site (Greenland 1989). In addition, the cold season spans the snowpack accumulation period at the Niwot SNOTEL site, which starts on average in October and peaks in April. The warm season includes the plant growing season (in terms of frost-free and snow-free periods) and shoulder months (e.g. May and September). May is generally when snowmelt initiates in the alpine and is most rapid in the subalpine (Greenland 1989; Williams et al. 2006; Molotch et al. 2008; Caine 2010). September has on average the first run of 3 days of frost (daily $T_{min} < 0$°C) and the first day with a hard frost of $T_{min} < -3$°C at both sites (Kittel et al., unpublished data). From here on, we generally refer to the C1 and D1 climate stations by their environments, subalpine and alpine, respectively.

Results

Climatological means

Over the 59-year (1952–2010) record, average (±standard deviation) annual precipitation at the D1 alpine station was

1090 ± 230 mm year^{-1}, whereas that at the C1 subalpine station was 670 ± 130 mm year^{-1}. Seasonal elevational differences were greatest in the cold season, with the alpine station receiving twice as much precipitation as the subalpine station from November through April (680 vs. 340 mm 6-month totals, respectively; D1:C1 orographic ratio = 2.0). In contrast, differences between the alpine and subalpine sites were least in the warm season, especially from June to August (180 vs. 170 mm 3-month totals; orographic ratio = 1.06). A greater orographic ratio in winter versus summer is consistent with that generally found for the Colorado Rocky Mountains (Doesken et al. 2003).

Long-term trends and seasonal shifts

Annual total precipitation at the alpine station increased over the 1952–2010 record by 60 mm year^{-1} decade^{-1} ($P \ll 0.001$), resulting in an increase of 230 mm year^{-1} or 25% from the first to last decade of the record (Figure 3(a)). In contrast, the annual precipitation trend at the subalpine station was not significant, with a weak negative tendency of −11 mm year^{-1} decade^{-1} ($P = 0.21$, Figure 3(a)); first-to-last decade change = −55 mm year^{-1} or −8%).

Monthly precipitation trends were season- and site-dependent ($P < 0.05$ and $P \ll 0.001$, respectively, two-factor ANOVA; Figure 3(b)). Seasonal dependence resulted in an increase over the record in the ratio of cold season (October–April) to warm season (May–September) precipitation at both alpine (trend: $P < 0.05$; first-to-last decade shift in ratio from 2.1 to 2.4) and subalpine stations (trend: $P < 0.05$; shift from 1.1 to 1.4). This is consistent with an overall shift in precipitation from summer to winter months observed for the Colorado Rockies (for 1949–2004, Knowles et al. 2006). At the alpine station, this shift resulted from increases in precipitation throughout most of the cold season (trend: +52 mm season^{-1} decade^{-1}, $P \ll 0.001$), with no change in late spring through early autumn (warm season trend, $P > 0.40$) (Figures 3(b) and 4(a)). At the subalpine station, on the other hand, the warm-to-cold season shift was largely due to moderate decreases in warm season precipitation (−16 mm season^{-1} decade^{-1}, $P < 0.05$), while the cold season trend was not significant ($P > 0.70$; Figure 3(b)).

At the subalpine station, the early winter peak in precipitation shifted from November to October (trend in difference in precipitation between months, $P < 0.05$), with October precipitation increasing by 59% from the first to last decade of the record (Figure 4(b)). The peak in spring precipitation also shifted to earlier, from May to April ($P = 0.05$; Figure 4(b)). At the alpine station, the timing of precipitation within the cold season was also altered. A strong increase in early winter (October–December) precipitation (Figures 3(b) and 4(a)) shifted the distribution of precipitation from predominantly late-winter/spring (March–May) to being more equally partitioned between these periods (early to late winter precipitation ratio trend, $P < 0.05$; first-to-last decade ratio shift from 0.6 to 0.9). On a percentage basis, the increase in early winter precipitation

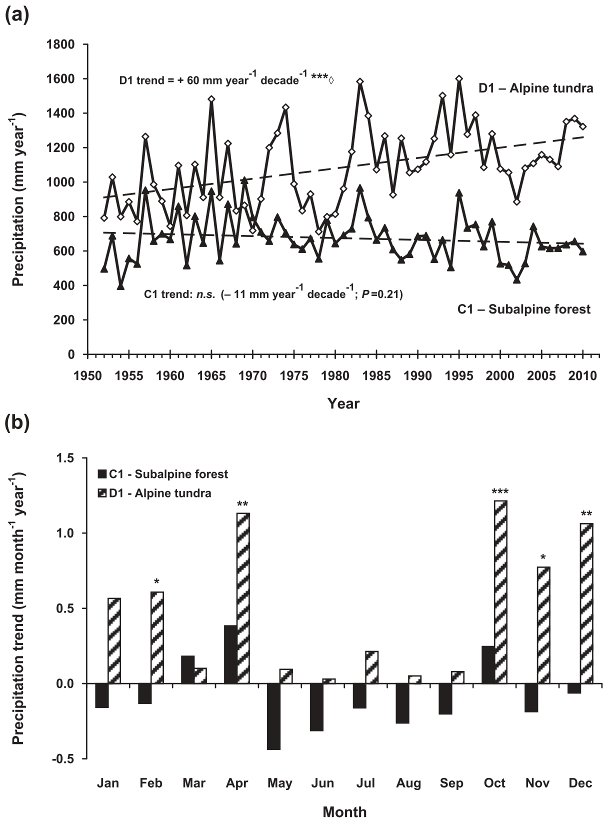

Figure 3. Precipitation (a) annual time series and (b) monthly trends for the C1 subalpine forest and D1 alpine tundra stations. In (a), the subalpine series is in closed triangles and alpine in open diamonds. Also in (a), dashed lines are trendlines, annotated with trend values and Mann-Kendall test significance levels (***$P \leq 0.001$; **$P \leq 0.01$; *$P \leq 0.05$; n.s., $P > 0.05$). An appended diamond (◊) indicates when significance is likely to have been overestimated due to high positive detrended serial correlation, specifically when the 1-year lag correlation coefficient $r_{-1} > +0.2$ (Harcum et al. 1992).

came primarily at the very start of the cold season, with a more than doubling of October precipitation from the first to last decade of the record (126% increase) and November and December precipitation increasing by 32% and 78%, respectively (Figure 4(a)).

Along with these increases in early winter, precipitation at the alpine site also increased strongly in early spring (27% in April; Figures 3(b) and 4(a)). Concurrent with little change in mean surface air temperatures (T_{mean}) in the cold season through early summer (alpine October–April and May–June T_{mean} trends, $P > 0.20$) (Kittel et al. 2011; Leopold et al. 2016), the increase in precipitation for winter overall and in early spring may have led to deeper winter alpine snowpack and its persistence longer into summer. Deeper snowpack resulting from higher winter precipitation is supported by a positive correlation of October–May alpine (D1) station precipitation with winter snow depth at the Saddle alpine site (Figure 1) (correlation of precipitation with winter-average snow depth: $r = +0.57$, $P = 0.01$, and with maximum snow depth: $r = +0.52$, $P < 0.05$, for record overlap of 1982–2003).

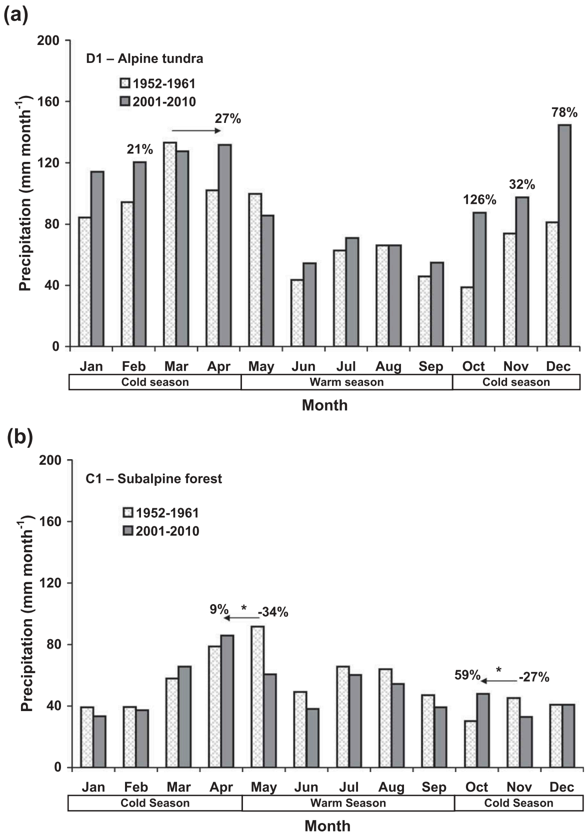

Figure 4. Comparison of the annual cycle of monthly precipitation for the first versus last decade of the record at the (a) D1 alpine and (b) C1 subalpine stations. In (a), there was an observed overall upward shift in the alpine station's seasonal pattern ($P < 0.05$, two-tailed paired t-test). Per cent change is given above the bars for months, in (a), whose long-term trend was significant (Figure 3(b)) and, in (b), where there was a significant shift in a seasonal precipitation maximum (arrows, with significance level as in Figure 3).

Snowpack changes at the subalpine station

Focusing on that part of the cold season with substantial snowpack accumulation, subalpine (C1) station precipitation for October through March lacked a long-term trend ($P > 0.90$). In the more recent record (WY1978–2010), however, October–March precipitation decreased (−29 mm season^{-1} decade^{-1}, $P < 0.05$; Figure 5). This was concurrent with a negative tendency in SWE on 1 April at the Niwot SNOTEL site at C1 (−19 mm or −6% decade^{-1}, $P = 0.09$; Figure 5). The decline in SWE is consistent with a regional analysis by Clow (2010) of Colorado Front Range SNOTEL sites (which are primarily in the subalpine zone and include the Niwot SNOTEL site). He found for the period WY1978–2007 that SWE on 1 April decreased for the Front Range by −12 to −27 mm decade^{-1} ($\alpha = 0.05$).

For this same period (WY1978–2007), we tested for a trend at C1 in the ratio of SWE on 1 April to antecedent (October–March) precipitation. A lack of a trend in the ratio would indicate that SWE changes tracked changes in precipitation inputs to the snowpack. On the other hand, a negative trend in the ratio would indicate that snowpack losses exceeded inputs. We found no trend in the SWE to precipitation ratio at C1 ($P > 0.25$; Figure 5), which is also consistent with Front Range regional results in Clow (2010). This suggests that the decline in winter SWE at C1 was due primarily to changes in precipitation inputs rather than from an increase in snowpack losses. This is in spite of increasing mean winter surface air temperatures over the period (subalpine WY1978–2007 October–March T_{mean} trend = +0.8°C decade^{-1}, $P < 0.01$) (Kittel et al. 2011). With this temperature trend, mean winter surface air temperature, nonetheless, remained below 0°C over this period. While the occurrence of daily temperature-driven melt events may have increased with this warming, their effect may have been compensated for by decreases in other processes that result in the loss of mid-winter SWE, and so possibly resulted in the lack of a trend in the SWE to precipitation ratio. Possible factors reducing losses from melting, sublimation or erosion include (1) weaker surface winds, (2) reduced solar and infrared radiative heating of the snowpack surface (e.g. through increased daytime or decreased night-time cloud cover) and (3) decreased ground heating of the snowpack due to lower stored heat in the soil.

Elevation dependence in trends

Differences in precipitation trends between the alpine and subalpine stations resulted in significant long-term increases in precipitation orographic ratios for much of the year (Figure 6(b)); trend in annual mean orographic ratio: +0.12 decade^{-1}, $P \ll 0.001$). These increases arose in different seasons for different reasons. Orographic ratios increased throughout the cold season (Figure 6(a) and 6 (b)) due to large increases in precipitation at the alpine site (Figure 3(b)) (first-to-last decade cold-season ratio shifted from 1.9 to 2.5). On the other hand, increases in warm season ratios (Figure 6(b)) arose from decreasing precipitation at the subalpine site (Figure 3(b)).

Discussion

Relationship between cold-season elevation-dependent precipitation trends and upper-air circulation patterns

From the pattern of an increasing elevational contrast in cold-season precipitation arising from substantial increases at the alpine station and little change at the subalpine station, we may be able to infer precipitation-generation mechanisms responsible for these trends. A key consideration is that the alpine station is east of, but in close proximity to the high crestline of the Continental Divide, whereas the subalpine station, also to the east, is lower in elevation and more removed from the Divide (Figure 1, Table 1). A mechanism common to Niwot Ridge winters that favours precipitation largely restricted to the Divide is one which develops from westerly air flow over the Divide crest under stable atmospheric conditions

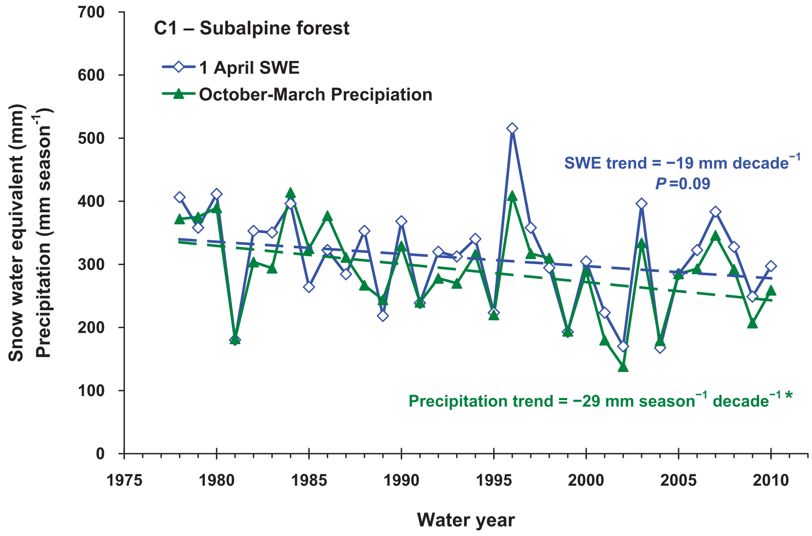

Figure 5. Comparison of WY1978–2010 time series of subalpine (C1) October–March precipitation (green, closed triangles) and Niwot SNOTEL snow water equivalent (SWE) on 1 April (blue, open diamonds). Dashed lines are trendlines (in corresponding colours), with significance levels as in Figure 3. The Niwot SNOTEL site is located in a forest clearing near the C1 precipitation station.

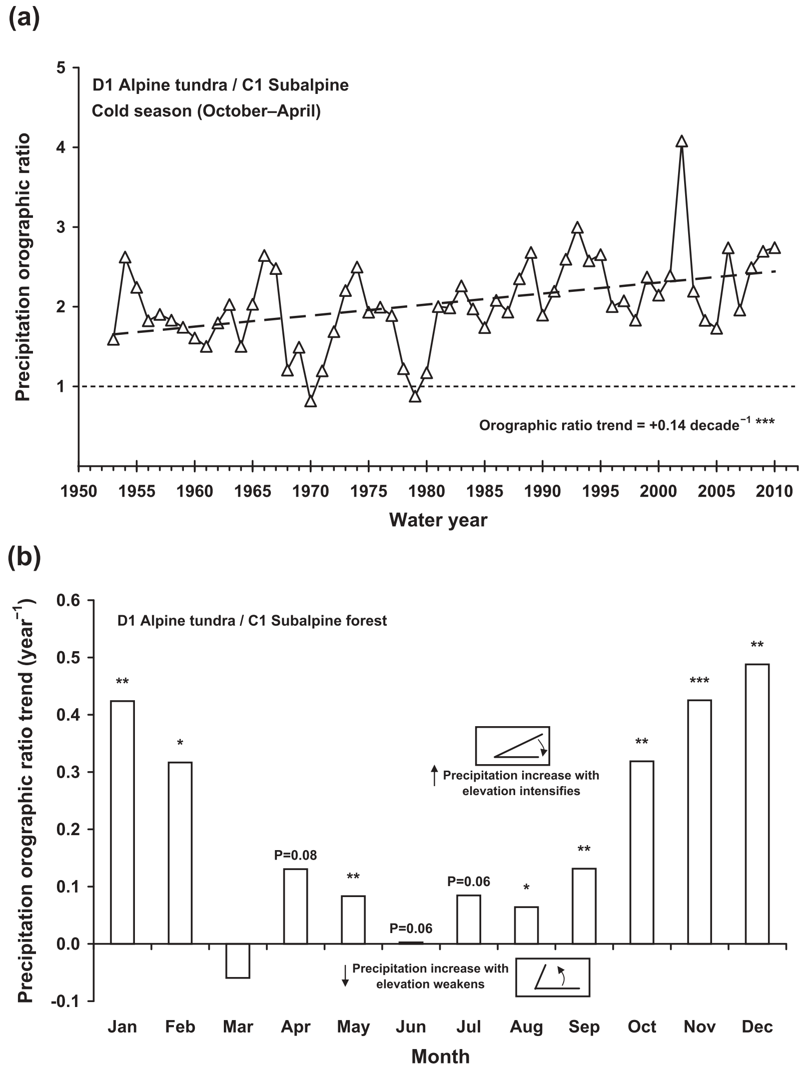

Figure 6. Orographic ratio of precipitation (D1 alpine to C1 subalpine) (a) cold season (October–April) time series and (b) monthly trends. Trend significance levels are as in Figure 3. In (a), heavy dashed line is the trendline and light dashed line is where the ratio = 1. Inset cartoons in (b) are precipitation–elevation profiles (precipitation on the *x*-axis and elevation on the *y*-axis) and contrast the direction of change in profiles when orographic ratio trends are positive (elevation gradient sharpens, top inset) or negative (gradient weakens, bottom inset).

(Losleben et al. 2000b). With sufficient moisture, such flow generates orographic precipitation along the windward (western) side of the Divide, with substantial spillover and leeward rotor entrapment of precipitation on the east side, including in the alpine zone of Niwot Ridge (Greenland 1989). These dynamics extend less frequently farther to the east, such that the subalpine zone, more distant from the Divide (Figure 1), receives substantially less precipitation from this mechanism than the alpine zone (Barry 1973). Such westerly flow predominates the wintertime surface wind regime in both zones, but with greater intensity and frequency in the alpine (Losleben et al. 2000a; Blanken et al. 2009) (Figure 2).

The dynamics of precipitation generation along the Divide is tied to wintertime upper-air (e.g. 700 hPa) flow regimes, where north-westerly flow is observed to favour stronger alpine (D1) to subalpine (C1) differences in daily precipitation amount and frequency, compared to weaker, or reversed, differences with south-westerly flow (Barry 1972, 1973) (Figure 7). This arises because north-westerly flow often precedes (is on the leading eastern edge of) a strong, blocking upper-air ridge, whose axis is centred over the Intermountain West (Changnon et al. 1993; Losleben et al. 2000b). East of the ridge axis, stable, subsiding air regionally suppresses synoptic-scale precipitation generation (such as that associated with mid-latitude cyclonic storms). While this limits Niwot Ridge precipitation from cyclonic storms, the strong north-westerly flow aloft generates orographically forced precipitation along the Divide, restricted more to the alpine zone rather than to the subalpine zone.

In contrast, south-westerly flow aloft favours a reduced or, in spring, a reversed alpine (D1) to subalpine (C1) difference in precipitation (Barry 1972, 1973) (Figure 7(a)). Such flow generally precedes an upper-air trough that dynamically supports mid-latitude cyclonic storms (Changnon et al. 1993). The wintertime alpine–subalpine contrast in precipitation generated by these storms, while still favouring the alpine zone, is not as strong as under north-westerly flow (e.g. in January, Figure 7(a)). In spring, south-westerly upper-air flow can alternatively result in a reversed orographic pattern (i.e. with more precipitation in the subalpine zone than in the alpine zone; for example, in April, Figure 7(a)). This arises when storms on a southern track, arriving from the south-western United States and passing to the south of or across southern Colorado, generate low- to mid-level upslope (easterly) flow on the eastern flank of the Colorado Front Range (e.g. Figure 2(a)): spring daytime subalpine windrose). Through orographic lifting, these easterly, upslope flows result in major snowfall events that favour lower elevations, including the subalpine zone, over the alpine zone, reversing the orographic ratio (Greenland 1989; Losleben et al. 2000b).

Analyses of the local wind regime in terms of the leeside rotor and location-dependent predominance of westerly versus easterly (upslope) surface winds at D1, C1 and other alpine and subalpine climate stations on and

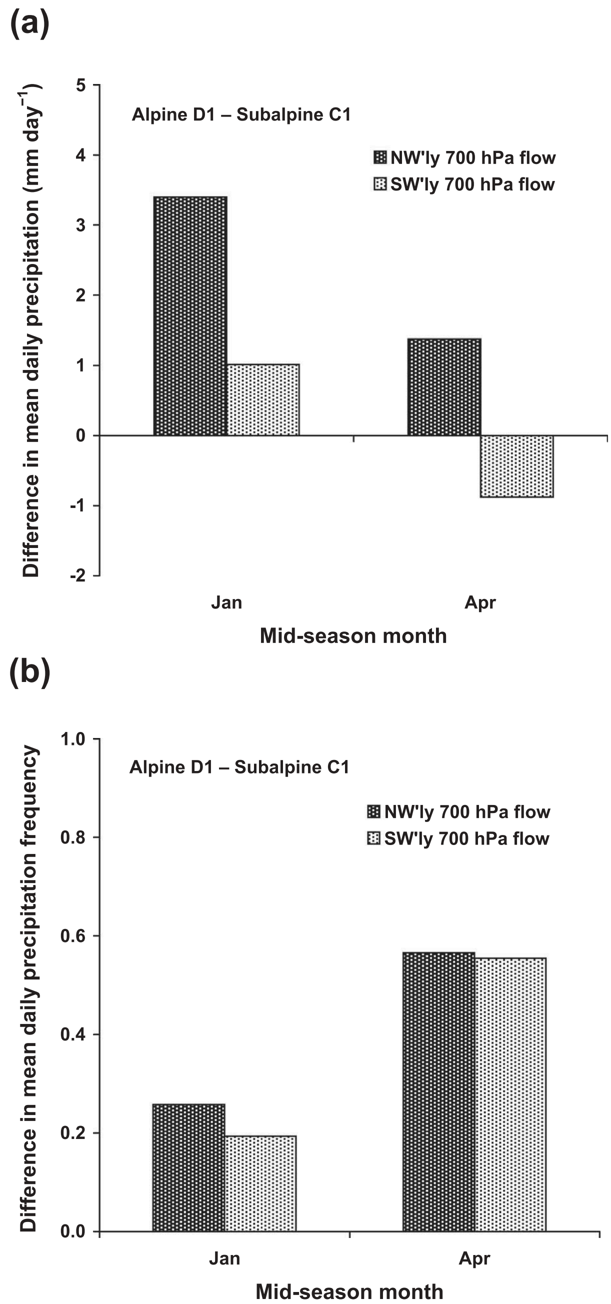

Figure 7. Difference in mean daily precipitation (a) amount and (b) frequency (fraction of days with precipitation) between D1 alpine and C1 subalpine stations under north-westerly (NW'ly) versus south-westerly (SW'ly) upper-air (700 hPa) flow regimes based on the analysis of synoptic patterns over western North America in relation to Niwot Ridge daily precipitation data for winter (January) and spring (April) over the period 1965–1970 by Barry (1972, 1973). We grouped Barry's synoptic types into those with NW'ly versus SW'ly flow over northern Colorado when specified in his text or maps; his results for October were not included as these were limited by substantially fewer daily observations and by a large portion of days with 'unclassified' synoptic patterns (Barry 1973). In (a), positive values indicate that precipitation lapse rate is strengthened (precipitation is favoured in the alpine over the subalpine) versus negative values, where the lapse rate is reversed with precipitation favoured in the subalpine (e.g. in April under SW'ly flow).

in the vicinity of Niwot Ridge (Losleben et al., unpublished data; Losleben et al. 2000a; Turnipseed et al. 2004; Erickson et al. 2005; Blanken et al. 2009) indicate that these atmospheric dynamics are characteristic of alpine and subalpine zones on the Ridge. Linkage of these dynamics and precipitation-generating mechanisms to upper-air circulation patterns suggest that these processes are regional and a key influence in distinguishing precipitation regimes at alpine and subalpine elevations along the Colorado Front Range east of the Divide (Barry 1973; Abbs and Pielke 1987; Baron and Denning 1993; Changnon et al. 1993; Cline 1997; Losleben et al. 2000b). Because of the linkage of elevation-dependent patterns in precipitation generation to upper-air dynamics, the strengthening of the cold-season orographic ratio observed over the 59-year Niwot Ridge record (Figure 6) suggests a shifting of upper-air circulation regimes from ones in which south-westerly flow plays a substantial role to ones more dominated by north-westerly flow.

Surface energy balance and precipitation trends

Elevation-dependent trends in precipitation as seen here are likely to co-occur with or lead to location-dependent responses of other surface climate variables, such as surface air temperature, relative humidity and surface winds. These potential linkages are through (1) associated changes in hemispheric and regional circulation patterns and (2) dependencies among the effects of cloud cover, precipitation (and precipitation form as snow vs. rain), snow cover and wind on the local surface energy balance (Barry 1990; Cohen and Rind 1991; Williams et al. 1996; Rangwala and Miller 2012). These interactions are reflected in a conceptual model for climate change in high mountain regions proposed by Barry (1990) and modified by Williams et al. (1996) that tied observed early record (e.g. through 1994) decreases in temperatures in the Niwot Ridge alpine to increases in snowfall through the surface energy balance.

Ecological implications

Increased winter snowpack in the alpine zone

Over the 59-year (1952–2010) record, precipitation substantially increased at the alpine site, with little overall change at the subalpine site. The difference in trends enhanced the already strong elevational contrast in precipitation between the two environments. Furthermore, the strongest trends at the alpine site were concentrated in the cold season (October–April), intensifying the cold-to-warm seasonal contrast in precipitation. This increase in winter and early spring precipitation, with little change in cold-season surface air temperatures over the record, may have led to deeper and longer-lasting snowpack in the alpine zone. Such snowpack changes would have both direct and interacting consequences for warm- and cold-season ecology and hydrology of Niwot Ridge and adjacent watersheds (Williams et al. 2016). Greater snowpack and delayed snowmelt would, for example, delay the onset of the plant growing season, but also lead to increased growing season soil moisture availability (Litaor et al. 2008). These changes, in turn, affect alpine plant productivity, species phenologies and demographics, community structure and plant–microbe interactions in ways that strongly depend on micro- and mesoscale geomorphology (e.g. due to wind-driven redistribution of snow) (Billings and Bliss 1959; Bowman 2000; Malanson et al. 2012; Gasarch and Seastedt 2016; Schmidt et al. 2015; Suding et al. 2016).

Alpine herbaceous plant responses to snowpack changes may have had nutritional consequences for survivorship of pika and marmot, both alpine-resident herbivores (Hill and Florant 1999; Morrison and Hik 2007; Erb et al. 2014; Bhattacharyya and Ray 2016). In addition, greater winter snowpack accumulation and delayed melt out affect the pika's talus habitat. For example, the insulating effect of deep snowpack during winter and increased intra-talus moisture and ice in summer (from later melt out) both serve to ameliorate talus habitat temperatures, reducing physiological stress, and so may have slowed their extirpation from alpine habitats under increased summer temperatures (June–August T_{mean} trend = + 0.3°C decade^{-1}, P < 0.01; Kittel et al. (2011)) (Beever et al. 2011; Wilkening et al. 2015).

Increases in snowpack may also have played a role in the observed increase in shrub cover (primarily *Salix* species) in alpine meadow communities (Spasojevic et al. 2013; Suding et al. 2016) and in treeline dynamics (Daly and Shankman 1985; Suding et al. 2016) on Niwot Ridge either by facilitating or suppressing woody species establishment and growth. The effects of increased snow cover on woody growth in tundra environments can be both positive (e.g. greater protection of above-ground tissues from frost damage and higher summer soil moisture levels) and negative (e.g. shortened growing season and increased exposure to snowpack plant pathogens) (Humphries et al. 2007; Wipf et al. 2009; Myers-Smith et al. 2011; Olofsson et al. 2011; Castanha et al. 2013; Moyes et al. 2013). Feedbacks of woody species expansion on the distribution, depth, melt rates, thermal conductivity and other physical properties of the snowpack may have, in turn, influenced alpine winter and summer microclimates, soil conditions and nutrient recycling (Seastedt and Adams 2001; Myers-Smith et al. 2011).

Increased snowpack melt from increased winter precipitation may have strongly influenced growing season microbial respiration. Higher levels of growing season soil moisture from snow melt has been seen to alter growing season carbon balance of alpine soils, with respiration on dry to mesic sites being most responsive to increased moisture levels and that on wet sites less so (already limited by saturated, anaerobic conditions) (Knowles et al. 2015a). In the cold season, increases in snow depth and cover elevate and stabilise winter and spring subnivean moisture and temperatures, which in turn increase tundra and talus microbial community biomass accumulation over the winter and turnover in spring (Ley et al. 2004; Lipson and Schmidt 2004; Schmidt and Lipson 2004). These microbial responses boost wintertime soil decomposition rates, the emission of nitrogen oxides and other trace gases through the snowpack, and subsequent release of dissolved organic and inorganic nitrogen and other nutrients into groundwater and downstream surface waters (Brooks et al. 1998; Williams et al. 2009; Liptzin et al. 2016).

In the Green Lakes Valley, an alpine watershed adjacent to Niwot Ridge, an increase in winter alpine precipitation decreased lake-ice thickness over the period 1980–2000, which was attributed to (1) the insulating effect of additional snow cover and (2) heating from increased winter groundwater inputs (from the previous winter's water basin inputs) (Caine 2002; cf. Leopold et al. 2016). The resulting earlier ice-off date and related surface and subsurface hydrologic changes in the watershed altered lake stratification, chemistry, and primary and secondary productivity during the subsequent longer ice-free season (Miller and McKnight 2016).

Winter wind regime change in the alpine zone

If the increase in winter precipitation at the alpine site arose from a higher frequency of north-westerly winds over the Divide as we suggest, the resulting change in wind speed and shift in wind direction may have altered the degree and location of wind scouring and redeposition of snow across the landscape (Erickson et al. 2005). Such micro- and mesoscale landscape distribution of snow-covered and snow-free areas affects the pattern (heterogeneity) and productivity of tundra plant communities and the size and shape of *krummholz* tree islands (Walker et al. 1993; Seastedt and Adams 2001; Gasarch and Seastedt 2016; Suding et al. 2016). In addition, white-tailed ptarmigan (*Lagopus leucura* Richardson), considered a year-round resident of the alpine zone, rely on winter snowdrifts for roosting and wind-blown areas for feeding and, in spring, breeding (Hoffman and Braun 1977; Braun 1980; Armstrong et al. 2001). Consequently, high snow cover limits their survivorship and population densities (Choate 1963). An increase in north-westerly winter winds, along with increased precipitation, may also have increased the rate of material exchange (water, particulates and nutrients) downwind to the subalpine forest, especially to the forest–alpine ecotone, with trees at the treeline acting as windbreaks and thus facilitating wind-drift deposition (Seastedt et al. 2004).

The possible change in upper-air circulation patterns from those favouring south-westerly flow to ones favouring north-westerly flow and accompanying changes in the frequency of synoptic weather patterns (e.g. of cyclonic storms) would likely to have changed source regions for air pollution and aeolian dust that arrives at the Front Range alpine and subalpine zones. Such source shifts would lead to changes in rates of dust and wet and dry nitrogen deposition, with consequences for, via altered biogeochemical dynamics, the structure and function of high-elevation terrestrial and aquatic ecosystems (Williams and Tonnessen 2000; Wolfe et al. 2001; Neff et al. 2008; Bowman et al. 2015). In addition, dust deposition on snow alters snowpack melt regime, shifting the timing of hydrologic processes (e.g. run-off) and ecosystem phenologies (Painter et al. 2007; Steltzer et al. 2009).

Snowpack change and decreased warm-season precipitation in the subalpine zone

At the subalpine site, there was no change in cold-season precipitation over the record. However, from the late 1970s, October–March precipitation decreased, with a similar decline in SWE on 1 April. This resulted in no change in the SWE to precipitation ratio, suggesting that subalpine winter snowpack was more strongly controlled by precipitation inputs than by increasing surface air temperatures. On the other hand, in spring (April–May), when temperatures are higher and snowmelt is most rapid, increases in surface air temperatures (1978–2010 April–May T_{mean} trend = +0.7°C decade^{-1}, P = 0.05) (Kittel et al. 2011) are likely to have resulted in an increased occurrence of early spring snowpack melt events (Burns et al. 2014), followed by an earlier and more rapid melt out of the snowpack. For the Colorado Rocky Mountains in general, both of these factors – decreasing maximum SWE (early April at the Niwot SNOTEL site) and increasing April and May surface air temperatures – contributed to earlier timing of when 50% of the snowpack had melted (Clow 2010). For the Front Range, dates of snowmelt onset and of when 50% of the snowpack had melted occurred earlier in the year by 2–4 days decade^{-1} (for the period WY1978–2007, Clow 2010). In addition, the shift of spring precipitation at the subalpine site from May to April (Figure 4(b)) reduced late spring inputs to the snowpack, also increasing the likelihood of an earlier melt out of the subalpine snowpack.

An increase in spring snowpack melt events likely would have favoured earlier development of snow moulds, significant in forest litter decomposition at a time when soils are saturated and anaerobic (Schmidt et al. 2008). In addition, late-winter/spring melt events are a precursor of the spring-to-summer turnover of soil microbial communities and winter-to-summer switching of forest net ecosystem-atmosphere CO_2 exchange from source to sink (Monson et al. 2005, 2006). An earlier snowmelt date also affects the phenology of subalpine herbaceous communities, in part by exposing emerging plants to frost (Inouye and McGuire 1991; Inouye 2008; Lambert et al. 2010). Changes in plant phenology, in turn, have consequences for the structure of these communities and for vertebrate and invertebrate herbivore and pollinator populations dependent on the timing of food resource availability (Inouye et al. 2000; Lambert et al. 2010; CaraDonna et al. 2014).

In the growing season, forest production is primarily reliant on deep soil water that has infiltrated from snowmelt and secondarily on shallower soil water arriving with growing season precipitation (Monson et al. 2002; Hu et al. 2010). Little precipitation change in winter, decreased summer precipitation, increased growing season surface air temperatures (e.g. June–August T_{mean} trend = +0.3°C decade^{-1}, $P < 0.01$; Kittel et al. 2011), and a longer growing season (initiated by earlier snowpack melt; Monson et al. 2005) are likely to have (1) increased growing-season evaporative demand and (2) more rapidly reduced plant available water over the growing season through evapotranspiration. These changes would have in turn increased the probability of growing season drought conditions and reduced forest productivity. If sustained and recurring, such drought has the potential for increased forest tree mortality either directly or through altered frequency and intensity of fires and insect outbreaks (Sherriff et al. 2001; Sibold and Veblen 2006; Bigler et al. 2007; Raffa et al. 2008).

Conclusions

Strong differences between long-term trends in precipitation at an alpine site and a subalpine site, both annually and seasonally, appear to have arisen from complex interactions among atmospheric processes operating at two spatial scales. These scales are (1) synoptic weather patterns across western North America, such as upper-air north-westerly flow and cyclonic storms supported by upper-air south-westerly flow, and (2) precipitation generation from the mesoscale interaction of synoptic circulation with local topography, such as from north-westerly flow over the crest of the Continental Divide. Changes in precipitation over the long term are likely to have resulted in different ecological and hydrologic responses in the alpine tundra and the subalpine forest because of not only system-specific dynamics but also divergence in their long-term precipitation trends.

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No potential conflict of interest was reported by the authors.

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Appendix A. Record homogeneity assessment of 1960s Niwot Ridge C1 subalpine precipitation station gauge and siting change

We assessed the early 1960s change in the C1 precipitation station's location and instrumentation on record homogeneity using two approaches: (1) a comparison of overlapping records from the Weather Bureau gauge (in an open site near the site's hygrothermograph Stevenson screen) and the Belfort gauge (in the forest clearing) (this comparison was limited by only 2–3 years of overlap) and (2) an evaluation of a potential shift

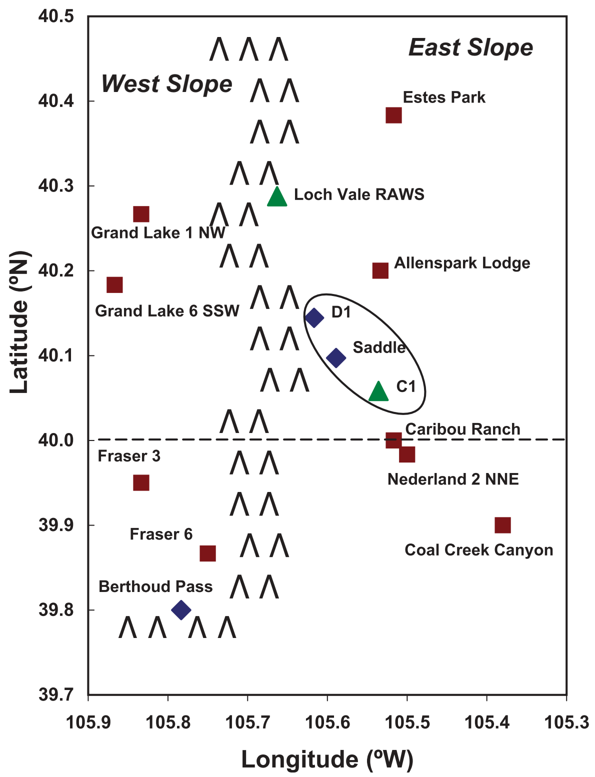

Figure A1. Location of Niwot Ridge subalpine (C1) and high alpine (D1) climate stations in context of northern Colorado Rockies stations used in record infilling. Shown are Niwot Ridge subalpine (C1) and alpine (D1 and Saddle) climate stations (within the oval), National Weather Service COOP stations and a Remote Automatic Weather Station (RAWS). All infilling source stations have daily minimum and maximum temperature and precipitation records (except the Loch Vale RAWS station which is temperature only). Station symbols indicate station vegetation zone and corresponding elevational range (per Peet 1988): blue ♦ alpine tundra (ca. 3400–4000 m a.s.l.), green ▲subalpine forest (ca. 2750–3400 m a.s.l.) and dark red ■ montane forest (ca. 2000–2750 m a.s.l.). 40° N latitude is shown by the horizontal dashed line and the general location of the Continental Divide by the double peaked 'Λ Λ' pattern ('West Slope' and 'East Slope' refer to locations west and east of the Divide, respectively). Longitude and latitude axes are scaled to approximately represent the same horizontal distances in *x* and *y*. The station 'Allenspark Lodge' is the current name for 'Allenspark 2NNW' used in an assessment of homogeneity in the C1 precipitation record (Appendix A).

at the time of these gauge and siting changes in the ratio of the C1 precipitation record to that of Allenspark 2NNW (COOP station 050183, Western Regional Climate Center [2009]), the station closest to Niwot Ridge that is east of the Divide and with a precipitation record spanning the change-over period ('Allenspark Lodge' in Figure A1). Allenspark 2NNW is relatively close to C1 (by 17 km to the north), similarly located east of the Continental Divide (by ca. 12 km vs. ca. 10 km for C1), and ca. 400 m lower in elevation than C1 (at 2591 m a.s.l.). Neither approach resulted in statistically significant seasonal adjustment factors that could be used to modify the subalpine record with any degree of reliability. However, these assessments suggest a winter undercatch bias by the open-sited gauge relative to that in the forest clearing potentially on the order of −4% to −9%; the two approaches did not give a consistent catch bias for summer. Although seasonal adjustment factors were not significant, we tested the sensitivity of trend results to adjusting the early record by factors derived from the first approach and found that monthly and annual trends were not altered in their statistical significance nor their magnitude (P = 0.19, paired t-test on monthly trends with and without adjustment). Nonetheless, as stated in the main text, the possibility of an uncorrected inhomogeneity in the subalpine precipitation record in the early 1960s should be kept in mind when interpreting trend results.

Appendix B. Methods for the statistical infilling of missing observations for Niwot Ridge C1 and D1 climate records

Infilling of missing observations in C1 and D1 climate station records was based on daily regressions with nearby stations. While we present, in the main text, an analysis of precipitation records, we include here infilling methods for minimum and maximum temperature (T_{min} and T_{max}, respectively), as well as for precipitation, as we used the same approach for the three variables. Upgrades to statistical infilling methods of missing precipitation and temperature observations were the following:

- An expanded set of infilling source stations, including four short-term Niwot Ridge instrumental records and 11 other mid- and high-elevation northern Colorado Rockies stations (up from three or fewer stations in earlier data sets) (Figure A1).
- Infilling of missing T_{min} and T_{max} observations through the modelling of daily mean temperature (T_{mean}) and diurnal temperature range (DTR). This allowed us (1) to replace analysing two well-correlated series (i.e. T_{min}, T_{max}) with an approach emphasising important differences between them, (2) to avoid incongruities arising from separately infilling T_{min} and T_{max} (such as infilled days with minima > maxima) and (3) to explicitly model DTR, an important variable in climate monitoring (Vose et al. 2005).
- Least-squares linear regression infilling models developed from daily records (upgraded from previous month-based non-statistical and regression approaches [Greenland 1987, 1989]), with logarithmic transformation of daily precipitation values.
- Two infilling regression approaches to model station-to-station daily covariation, distinguished by time frame: (1) same-season (within same year) and (2) long-term (multi-year) frames, discussed next.

Of these two regression approaches, the same-year/same-season method aimed to capture relationships between stations tied to synoptic dynamics under conditions of a year's hemispheric circulation regime (such as influenced by the Pacific Decadal Oscillation and El Niño–Southern Oscillation) around the time of a missing observation. The long-term approach, on the other hand, sought relationships persisting over the full period of record overlap for the day of the year of the missing observation. This dual regression model method was implemented because neither approach alone gave statistically significant infilling relationships for all occurrences of missing data. For a given missing observation, infilling models were computed using the two approaches applied across available source stations; final model selection was based on the regression with greatest percentage of variance explained (i.e. the magnitude of the coefficient of determination, R^2) drawn from models that were statistically significant (at significance level α = 0.05). The long-term approach was selected over the same-season method roughly 65% of the time for temperature and 90% for precipitation. Mann-Kendall trend significance results

Table A1. Comparison of Kendall's rank correlation significance values (P) for maximum temperature (T_{max}) and precipitation trend analyses run with and without in-filled data covering years 1952–2006 for Niwot Ridge C1 (subalpine) and D1 (alpine) sites.

	C1 Maximum Temperature		D1 Maximum Temperature		C1 Precipitation		D1 Precipitation	
Month	With infilling	Without infilling	With infilling	Without infilling	With infilling	Without infilling	With infilling	Without infilling
Jan	0.016 *	0.025 *	0.670	0.693	0.539	0.664	0.003 **	0.003 **
Feb	0.016 *	0.025 *	0.560	0.856	0.179	0.307	0.000 ***	0.000 ***
Mar	0.000 ***	0.000 ***	0.782	0.440	0.559	0.313	***0.001*** ***	***0.011*** *
Apr	0.001 **	0.003 **	0.525	0.782	0.534	0.569	***0.000*** ***	***0.003*** **
May	0.001 **	0.005 **	0.805	0.943	***0.071***	***0.034*** *	0.284	0.480
Jun	0.067	0.100	0.459	0.569	0.105	0.076	0.869	0.366
Jul	0.002 **	0.003 **	0.039 *	0.040 *	0.338	0.320	0.804	0.853
Aug	0.004 **	0.004 **	0.243	0.243	0.342	0.271	0.982	0.837
Sep	0.104	0.086	0.714	0.742	0.594	0.583	0.361	0.200
Oct	0.700	0.710	0.012 *	0.028 *	0.181	0.181	0.000 ***	0.000 ***
Nov	0.436	0.453	0.004 **	0.007 **	0.924	0.924	0.000 ***	0.000 ***
Dec	0.302	0.407	0.017 *	0.011 *	0.346	0.317	0.000 ***	0.000 ***
Annual	0.000 ***	0.000 ***	0.935	0.759	0.259	0.279	0.000 ***	0.000 ***

Notes: Values in bold italics indicate when a test's significance level changed (e.g. from *** to **).
***P < 0.001, **P < 0.01, *P < 0.05.

(Methods in main text) were robust in sensitivity tests omitting months with infilled missing values compared to results with infilled values (Table A1).

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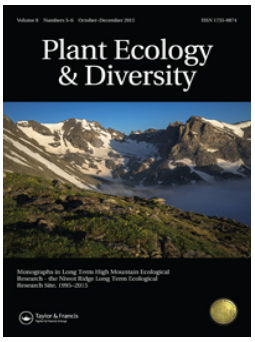

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Contrasting long-term alpine and subalpine precipitation trends in a mid-latitude North American mountain system, Colorado Front Range, USA

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