Long-term change in precipitation and its variability and consequences in a montane environment

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Abstract

The climate at high altitudes is changing more rapidly than in many other areas, with a variety of consequences for montane flora and fauna. We analyze long-term weather records from Crested Butte (2703m, 1910- 2024) and Gothic (2921m, 1974 – 2024), Colorado, USA, in the West Elk Mountains. In Crested Butte, mean monthly precipitation in the growing season (May – September) can vary by over 2-fold and peak weekly precipitation occurs in late July with the beginning of the summer monsoon. Mean rainfall per storm has been declining significantly since 1910. Many of the variables exhibit a curvilinear relationship with year, with high points at the beginning and end of the instrumental record period. Some of these curves have significant breakpoints, ranging from 1973 (for mean rain per rainy day) to 1985 (number of rainy days/summer, and number of consecutive days without rain). Some measurements are showing increasing variability over time (e.g., coefficient of variation of rainfall amount per rainy day). In Gothic, which typically gets more precipitation than Crested Butte, June is also the driest month, and December the wettest. Warming temperatures are leading to rainfall vs. snow earlier and later in the year, and mean precipitation per rainy day is declining significantly. Although it has snowed every month of the year in Gothic, total winter snowfall has been declining significantly since a peak in 1995. Mean snowfall per snowy day and per storm are both declining. Mean snow density is increasing, and is correlated with rising temperatures. Dates of last snow cover are trending earlier. There is evidence of a 12-14yr cycle of precipitation. Variability of some but not all of the measures we examined is changing. A reconstruction of historic snowfall back to the 1930s (using stream gage data) suggests that it peaked in the 1970s and is now declining. These climate changes have the potential to impact many aspects of the ecology of montane plants and animals, and we give some examples.

I. Introduction

Many aspects of the ecology of high-altitude environments are controlled by climate. Productivity, phenology, distribution, and species diversity are examples of such variables. Our understanding of the ecological roles of snowpack, snowmelt dates, and variation in precipitation is growing, as is our appreciation of how the changing climate is affecting these climate variables, and how they in turn are affecting the plants and animals characteristic of high elevations. For example, a long-term study of flowering by a perennial herb showed how climate influences flowering phenology and phenotypic selection on phenology, with consequences for fitness via natural selection (Ehrlén and Valdés 2022). (Ehrlén and Valdés 2022), and other studies at our field site have also shown significant effects of climate on flowering phenology, flower abundance, and butterfly and bumble bee abundance (Inouye and McGuire 1991, Inouye et al. 2003, Lambert et al. 2010, Boggs and Inouye 2012, Ogilvie et al. 2017).

Long-term data are essential to our ability to document and interpret changes in climate variables, particularly given the variability of these parameters, which requires long data records to discern significant trends. For example, a study of decadal and multidecadal variability in North American precipitation found evidence for a decadal component of autumn precipitation over the central Rocky Mountains on a 12–14-yr time scale (Ault and St. George 2010), and a study of tree rings in the southern Rockies documented multidecadal (>30–70 yr) patterns of drought variation (Gray et al. 2003). Decades of data are required to elucidate such patterns and to be able to draw conclusions about them with confidence, and century-long records make possible an evolutionary as well as ecological perspective on how the environment is shaping the biota and its interactions. In some circumstances, evolution in response to climate may proceed rapidly enough to detect it with shorter datasets (Franks et al. 2007, Franks et al. 2014, Hamann et al. 2018), but the long lifespans of many high-altitude plant species may make for slow evolutionary responses.

Much of the phenology at high altitudes is determined by the date when the snowpack disappears, allowing the ground to warm up and the growing season to begin (Inouye and Wielgolaski 2024). Key phenological events such as initiation of growth and date of maximum leaf length, and morphological traits such as leaf number and flower number, are related to snowmelt patterns (Walker et al. 1995). The flowering of wildflowers throughout the summer is linked to date of snowmelt (e.g., Inouye et al. 2002, Pardee et al. 2019, Inouye 2020), as is abundance of some species (Inouye and McGuire 1991). Thus, the environmental variables of winter precipitation and spring temperatures (which influence snowmelt dates) are of particular interest, but summer precipitation can also influence flowering (Inouye et al. 2003), sometimes with time lags of up to four years (Prather et al. 2023). In turn, flowering phenology and abundance influence populations of pollinators (Ogilvie et al. 2017, Pardee et al. 2019) and probably seed predators and their parasitoids, and these same climate variables can affect many other aspects of montane ecology (Prather et al. 2023).

Forecasts of the effects of climate change include the likelihood of increasing precipitation and increased variability (van der Wiel and Bintanja 2021, Thackeray et al. 2022), and a continuation of the declining trend in global snow cover (Mudryk et al. 2020). In this context we examine data on precipitation from weather stations at the Rocky Mountain Biological Laboratory and the nearby town of Crested Butte, Colorado, to look for evidence of changes that could be consequences of local, regional, and global climate change. This high-altitude site in southwest Colorado's Rocky Mountains is in an area with particularly rapid climate warming, at the edge of the largest 2°C hot spot in the lower 48 states (Eilperin 2020). It is also unusual because of a relatively large number of long-term ecological studies conducted there that can provide insights into responses by the flora and fauna to the climate (e.g., Prather et al. 2023).

II. Methods

We accessed climate data from NOAA's National Centers for Environmental Information, via the portal Search | Climate Data Online (CDO) | National Climatic Data Center (NCDC) (noaa.gov). Station USC00051959 in Crested Butte, CO, is at 2,702.7m above sea level, 38.8738°N, -106.9772°W, with a data record beginning 1 June 1909 and 98% data coverage up to present (1919 data and some other months are missing). This location is the closest weather station to the Rocky Mountain Biological Laboratory (RMBL) with such a long data record. RMBL is a high-elevation research station in Gothic, at 2,900m and 9.4km north of the Crested Butte weather station. Data downloaded 11 August 2022 are available as a supplementary CSV file (see the note in Supplementary Material about a bug in the download process if you attempt to reconstruct this process). Additional data up to July 2025 were similarly accessed.

Daily summaries for May – September (the growing season) were downloaded as CSV files. Summary statistics for summer precipitation were calculated for five-month blocks of annual data (May – September), including total rain, number and percentage of rainy days, maximum daily rainfall, maximum number of consecutive days with and without rain, mean quantity of rain per rainy day, and the amount of rain per storm (consecutive rainy days). For 81 of the first 111 years there were no missing data, and for an additional 19 years no more than three observations were missing, but in a few cases an entire month's data are missing. If only a few days of data were missing we still calculated the list of statistics shown above. 1930 rain/snow data were not used because snow water equivalent (SWE) data were missing for some measurements (and that year was an outlier); 1925 and 1927 were also outliers. We used SigmaPlot (version 15; Systat Software) to generate graphs and basic statistics.

1 May was used as the beginning of the dataset for summer precipitation, and 30 September as the end (this interval encompasses almost all flowering at this site), so estimates of the number of consecutive rainy (or dry) days, and total rainfall per storm, are slight underestimates because those beginning and end dates don't always correspond with the start or stop of storms (or dry spells). In a few cases, data from the Aspen, Colorado weather station (about 38 km NNE of Crested Butte, 28 km from Gothic) were used to inform decisions made about whether a day with missing data was likely to have had precipitation in Crested Butte.

A reviewer of an earlier version of this paper expressed concern about the reliability of instrumental weather records prior to about 1950. There is no doubt that recorded precipitation may differ significantly from actual precipitation. Grosman and Legates (1994) point out that "Gauge measurements tend to be underestimates of the true precipitation, largely because of wind-induced turbulence at the gauge orifice and wetting losses on the internal walls of the gauge. These are not trivial as monthly estimates of this bias often vary from 5% to 40%. Biases are larger in winter than in summer and increase to the north in the United States due largely to the deleterious effect of the wind on snowfall.". Metadata about equipment used at the Crested Butte weather station are only available back to 1980 (https://www.ncdc.noaa.gov/cdo-web/datasets/GHCND/stations/GHCND:USC00051959/detail). The correlation between Crested Butte and Aspen July precipitation from 1935-1979 is $r^2 = 0.1071$, $p =.03$; the Aspen station (USC00050370) is about 287m lower than the one in Crested Butte. Given that the NOAA data have undergone rigorous quality-control measures, and that they are the only ones available, we have used them for this analysis.

Snowfall, snowpack, snowmelt, and other snow-related data have been collected daily by barr at a weather station adjacent to RMBL from November 1974 to present, and rainfall measurements were initiated in 2000. This measurement site is at 38.9628°N -106.9929°W, 2921m. Snowfall is measured with a ruler from a snow board that is cleared twice each day at 7 a.m. and then usually between 4 and 4:30 p.m.; 24-hour snowfall records are from sunset to sunset. Snowpack depth is measured against a snow pole marked to show current snowpack. Snow water content (SWE) is determined with a standard bucket and scale as issued by the CO Avalanche Center and USDA Forest Service. Density of newly fallen snow was calculated each day beginning in November 1975 by dividing the day's snowfall (inches) by water content (inches); water content of snowfalls on 7, 17, 19, and 20 October 1990 was estimated by using the correlation for water content in November and December 1990 snowfalls between Crested Butte weather station and barr's data from Gothic. Data used for this project are available (from 2016 to present) at http://www.gothicwx.org/, and detailed data are available upon request from DWI or bb, or through spreadsheets archived for this paper (see Data Access below).

Stream gage data are reported for peak runoff and date of peak from the USGS gage on the East River at Almont ("USGS 09112500 EAST RIVER AT ALMONT, CO"; Lat. 38°39'52", Long. -106°50'51" NAD27, 2,440m ASL), from the National Water Information System nwis.waterdata.usgs.gov Web site. We used the correlations between instantaneous peak runoff in the East River at Almont and snowpack and snowmelt data from Gothic to hindcast snowpack and snowmelt dates in Gothic from 1935 to 1974 (i.e., prior to barr's observations).

To look for evidence of a cyclical pattern in fall and winter precipitation at RMBL, we used the observed and calculated (as in the previous paragraph) dates of snowmelt in Gothic. We calculated 8-year running correlations between year and date of snowmelt, and plotted the resulting regression lines.

For some analyses that indicated a significant curvilinear relationship we calculated the breakpoint where trends reversed, using a breakpoint analysis in R version 4.2.1 with the package "segmented" (Muggeo 2008, Muggeo 2017). For each variable we entered the estimated breakpoint year from visual observation. We were then able to 1) see if a statistical breakpoint occurred, 2) the year in which it occurred, and 3) the slope of the regressions before and after the breakpoint.

Variability of weather variables could be measured in multiple ways, but the standard deviation and coefficient of variation are most commonly employed. If there is no significant change in mean of the variable under consideration then standard deviation may be the best index to quantify variability, but if there is an ongoing change in mean value, then coefficient of variation may be more appropriate. We report both indices, with the choice being influenced by whether means appear to be changing.

III. Results

A. Summary (May – September) precipitation

1. Crested Butte

Mean monthly precipitation varies by month, with June the driest month and January the wettest (Table 1). The summer pattern of daily rainfall, 1 May – 30 September, 1910 – 2024, lumped into weekly means, shows a peak in late July and early August (Fig. 1; see Supplementary Figure 1 for totals for 3-day periods grouped by 20-year intervals). Total summer precipitation measured in Crested Butte has varied greatly from 1910 – 2024, ranging from 10.16 cm in 2011 to 69.7 cm in 1927 (an outlier), with a mean of 22.01 cm (s.d. = 8.41), but the slight trend toward lower totals is not significant (p = .22; Supplementary Figure 2). Broken down by month, June shows the strongest declining trend over time (but p = .155; Supplementary Figure 3). A curvilinear relationship is found for the mean amount of rain per day on rainy days over the same time period (Fig. 2; $r^2 = .278$ p = .01); some measurements about 90 years ago are more than double what they have been in recent decades (mean = 5.14 mm, maximum = 11.3 mm), although there is a recent trend for increasing amounts. Breakpoint analysis indicated a break in 1985 (p = .003).

Table 1. Mean monthly precipitation (cm) in Crested Butte, 1910 – 2024 (June)

	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Mean	6.82	5.97	5.86	4.55	3.67	3.21	5.10	5.28	5.23	3.92	4.33	5.77
Max.	28.17	20.65	20.32	13.21	12.22	17.91	13.97	14.38	37.26	12.42	10.64	32.00
Min.	0.61	0.86	0.41	0.10	0.00	0.00	0.00	0.46	0.00	0.00	0.08	0.03
N	112	111	114	111	113	112	111	110	112	113	113	112

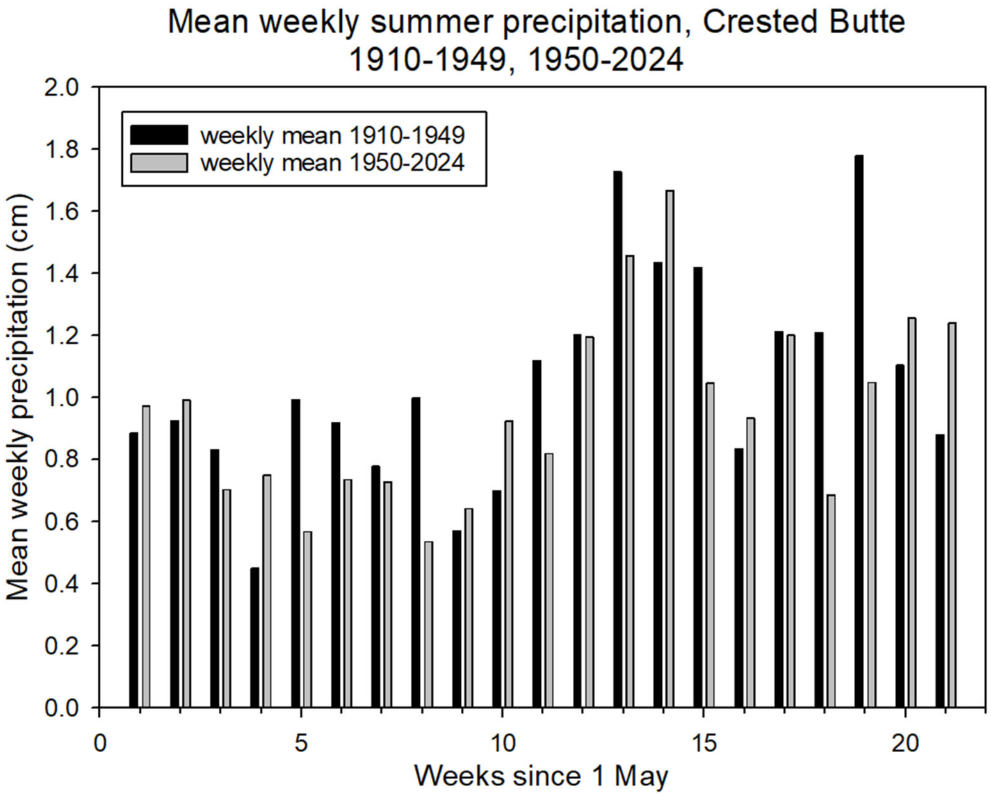

Figure 1. The summer pattern of rainfall in Crested Butte, 1 May – 26 September, 1910 – June 2024, lumped into 1-week totals. The data are split into two time periods, 1910-1949, and 1950 – June 2024.

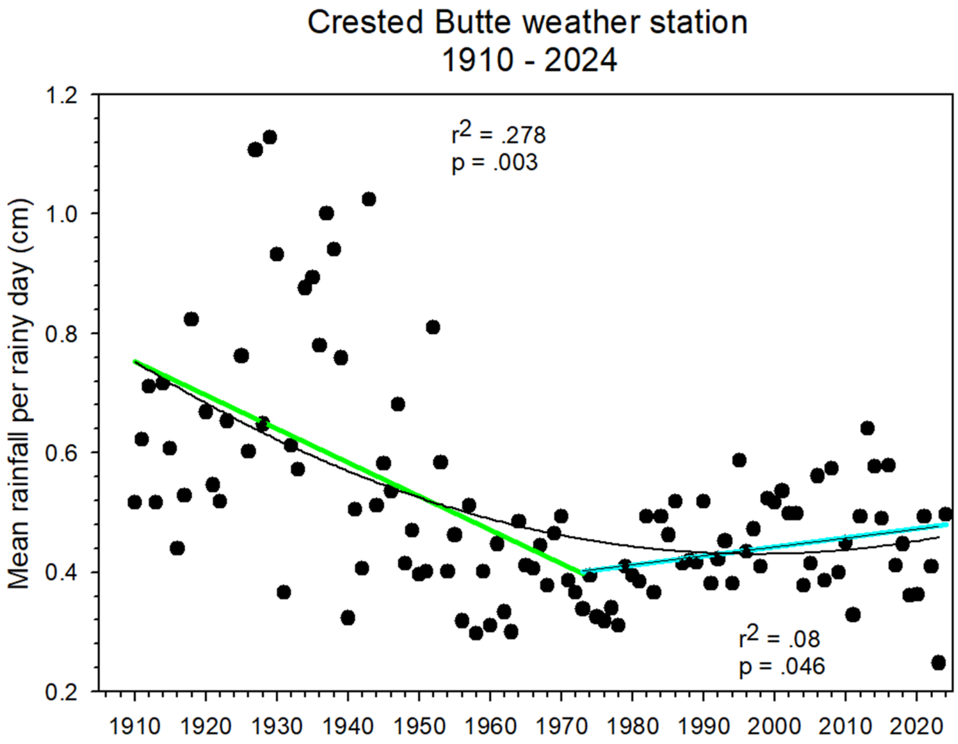

Figure 2. The mean amount of rain/day on days with rain has declined some since the first few decades of the record, but may be increasing again. The calculated breakpoint is 1973. A quadratic fit is shown for the whole data set ($r^2 = .278$) Linear regression lines are shown for data from 1910 – 1973 and 1973 – 2024 ($p = .046$).

There is no significant trend in the maximum rainfall per day (Supplementary Figure 4), or per storm (Supplementary Figure 5); note that the value for maximum rainfall per storm (consecutive days of rainfall) in 1927 (31.83 cm) is an outlier (next highest value is 9.4 cm in 1961). The mean amount of rain dropped by storms is declining (Supplementary Figure 6; $r^2 = .078$, $p = .003$). Precipitation per storm event is unevenly distributed across the summer, with a peak in late July to early August (Supplementary Figure 7 shows data for 2000 - 2024).

The pattern of summer rainfall has changed over time. For example, the number of rainy days trended higher until the 1980s and has since begun to decline (Fig. 3; $r^2 =0.231$, $p<.0006$), with a breakpoint calculated as 1985 and a significant linear decline since then. The mean number of consecutive rainy days per year (overall mean = 2.18, maximum = 3.6 in 1989) is increasing ($r^2 = 0.057$, $p = .01$; Supplementary Figure 8). The maximum number of consecutive rainy days per year is not changing significantly ($p = .135$), however there is a remarkable string of high values from 1967-1973, with up to 29 consecutive rainy days (in 1968; Supplementary Figure 9), and there may be an overall increasing trend. The percent of rainy summer days per year is changing, having increased until about 1980 but with a declining trend since then ($r^2 = .241$, $p = .0005$, for a curvilinear fit; Supplementary Figure 10).

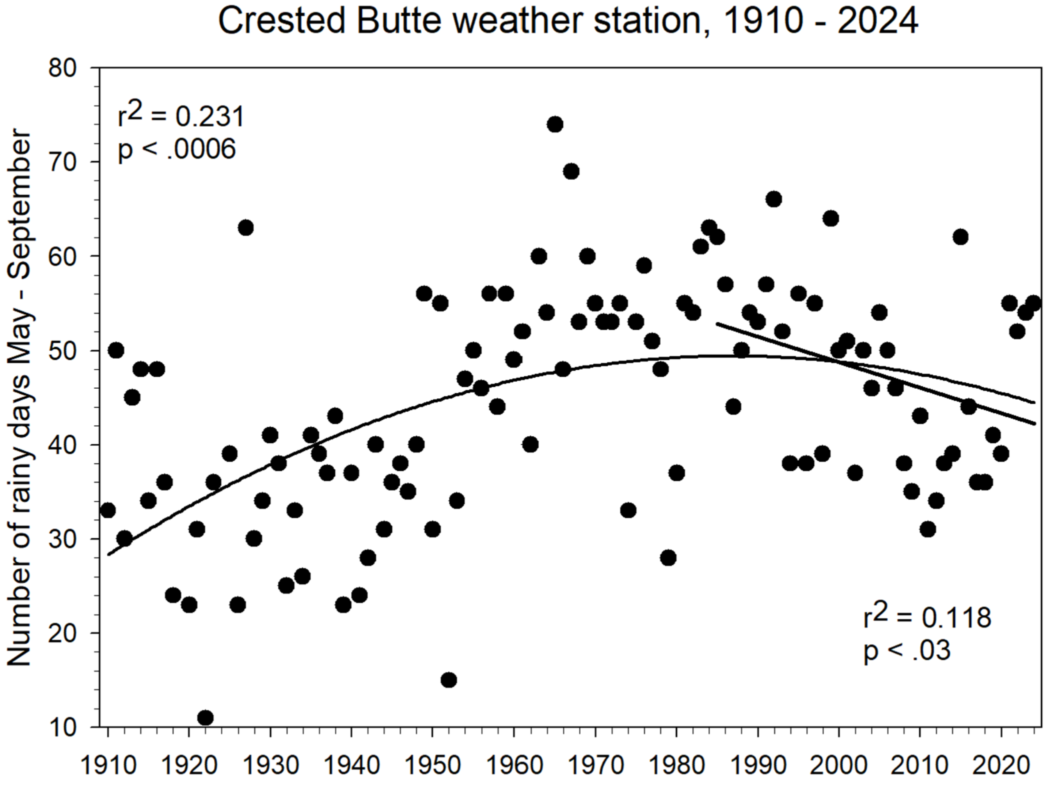

Figure 3. The number of rainy days during the growing season increased from the beginning of the study period until the 1980s, and has subsequently declined. A quadratic fit ($r^2 = .231$) is shown for the whole dataset and a linear fit ($r^2 = .118$) since the breakpoint calculated as 1985.

The maximum number of consecutive days without rainfall is changing (mean 21.8 days, maximum 71 days (in 1924, when some data are missing; Supplementary Figure 11; $r^2 = 0.155$, $p < .002$; there is a significant breakpoint in 1985, $p = .001$). Similarly, there is a significant curvilinear trend for an increasing (since about 1970) total number of days without rain ($r^2 = 0.201$, $p = .0004$; Supplementary Figure 12) and the mean number of consecutive days per summer with no rain ($r^2 = 0.183$, $p = .001$; Supplementary Figure 13). There is no significant relationship between summer (May – September) and winter (December – March) precipitation ($r^2 = .01$, $p = .24$) measured from 1912 – 2024.

Variability

The means of some of our measures of precipitation are changing over time (e.g., rainfall per rainy day (Fig. 2), and mean rainfall per storm (Supplementary Figure 6)), and because some models of climate change show increasing variation in precipitation over time, we calculated standard deviation (SD) and coefficient of variation (CV) for our measures of precipitation to look for evidence of changing variability. Variability of some but not all of the measures we examined is changing. The CV of rainfall per rainy day during the summer is increasing over time ($p < .0001$; Supplementary Figure 14), although the SD of rain per rainy day is declining ($p = .057$; Supplementary Figure 15). The standard deviation of mean rain per day (all days) is declining ($p = .058$; Supplementary Fig. 16); variability for daily summer precipitation is trending lower, but not significantly for individual months (e.g., for June, $p = .06$; Supplementary Figure 17). There is no significant trend in variation in mean rainfall amount per day (Supplementary Figure 18) or per storm (Supplementary Figure 19), but the standard deviation of the number of consecutive days without rain has declined significantly, and the trend for an increasing standard deviation of the number of consecutive days with rain is not quite significant ($p = .07$; Supplementary Figure 20).

2. Gothic

The first date of rain in the spring (i.e, preceded by snow) is occurring earlier but the date of the last rain in the fall (i.e., followed by snow) has not changed significantly (Fig. 4), and the number of rainy days from May – September is trending lower, but not significantly (Supplementary Figure 21), with range of 39 – 89 days/yr. Neither the mean or maximum number of consecutive days with or without rain from May – August have changed significantly, although mean consecutive days with rain is trending shorter and days without rain is trending longer (Supplementary Figure 22); length of dry spells, as measured by maximum number of consecutive days without rain, is trending longer and days with rain is trending shorter although neither trend is significant (Supplementary Figure 23). The amount of precipitation (rain and

SWE) per day is trending lower (but not significantly) from May – September (Supplementary Figure 24), as is the mean amount of rain (including SWE) per rainy day (Supplementary Figure 25).

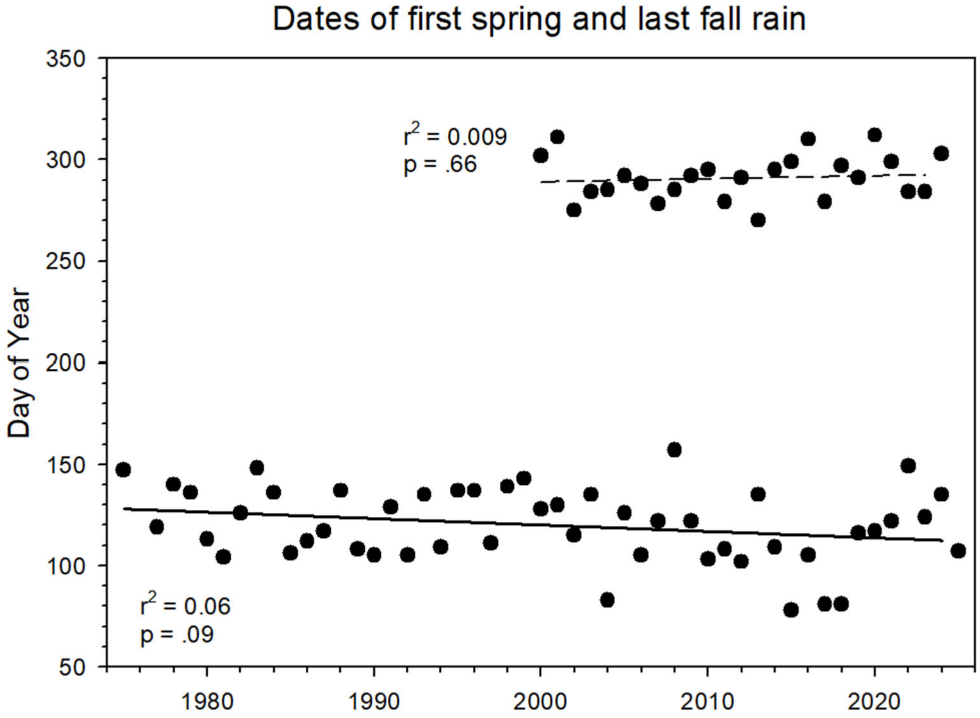

Figure 4. The date of the first rain in the spring in Gothic is occurring earlier, and the date of the last fall rain is trending later (but not significantly).

The maximum amount of precipitation per day from May – September is trending lower (Supplementary Figure 26), as is the maximum amount per storm (i.e., consecutive days of precipitation) (Supplementary Figure 27), but the changes are not significant. June is the driest month and December the wettest, with the wettest three consecutive months being December – February (Table 2, Supplementary Figure 28). Monthly precipitation can vary greatly among years (see Supplementary Figure 29 for July and August).

Table 2. Mean monthly precipitation (cm water) in Gothic, 2000 – 2024 (June). All months except July and August include some water from snow (SWE).

	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Mean	9.51	9.68	9.01	7.84	6.62	2.95	6.05	6.03	6.87	6.28	7.10	10.90
Max.	28.55	17.17	21.18	17.10	15.85	6.45	14.0	10.41	16.36	13.23	15.88	28.12
Min.	2.34	2.82	2.18	2.90	0.41	0.25	2.08	1.83	1.32	1.22	2.34	2.03
N	24	24	24	24	24	24	23	23	24	24	24	24

Monthly precipitation in Gothic from April through September is mostly snow in April and May (e.g., Supplementary Figure 30, for April, with no significant trends over time), and primarily rain from June – September. There are no significant trends since 2000, although the steepest (negative) slope is for September (Supplementary Figure 31; p = 0.15).

Variability of some precipitation measures from Gothic is changing. For example, the Standard Deviation of the number of consecutive days each year without precipitation is trending higher for the water-year, although the SD of consecutive days with precipitation is trending lower (Supplementary Figure 32). For the summer (May – September) the trend for increasing variation in the number of consecutive days without precipitation is significant (p = .05), but not for consecutive days with precipitation (Supplementary Figure 33).

3. Comparison of Crested Butte and Gothic summer precipitation

Monthly summer (May – September) precipitation means of both rain and SWE from 2000 – 2024 (N = 120; 1 missing summer in Crested Butte) are higher in Gothic than in Crested Butte for 95 of 120 months. However, there is no significant change in the ratio of the two sites over the study period (for May or July; Supplementary Figure 34). The maximum rain per day for each month in this same period is usually higher in Gothic (84 of 125 monthly comparisons), but again there is no significant trend. On a water-year basis (September – August), Gothic has had more precipitation than Crested Butte in all years from 2000 – 2023 except 2017 (Supplementary Figure 35).

B. **Winter precipitation in Gothic: Snowfall, snow density, snowpack, and snowmelt date**

1. **Snowfall**

Snow has fallen at least a few times every month of the year in Gothic (b. barr, unpublished), but using 1 September as beginning of the conventional water year, measurable snowfall has been as early as 5 September (2004) and as late as 5 July (1993) within water years; only trace amounts of snow have fallen in August. Winter snowfall measured in Gothic has been declining since 1995 (Supplemental Figure 36), although there is no significant breakpoint. From 1976-2024 the annual snowfall average is 1,021 cm, with a range of 474 (1976-77) to 1,641 cm

(1994-95). Because of the close relationship between snowfall in Gothic and runoff in the East River (measured at a USGS stream gage downstream in Almont), we can estimate the historic snowfall in Gothic back to 1935. Figure 5 shows a curvilinear fit to the data, which suggest a peak in this longer time series around 1970.

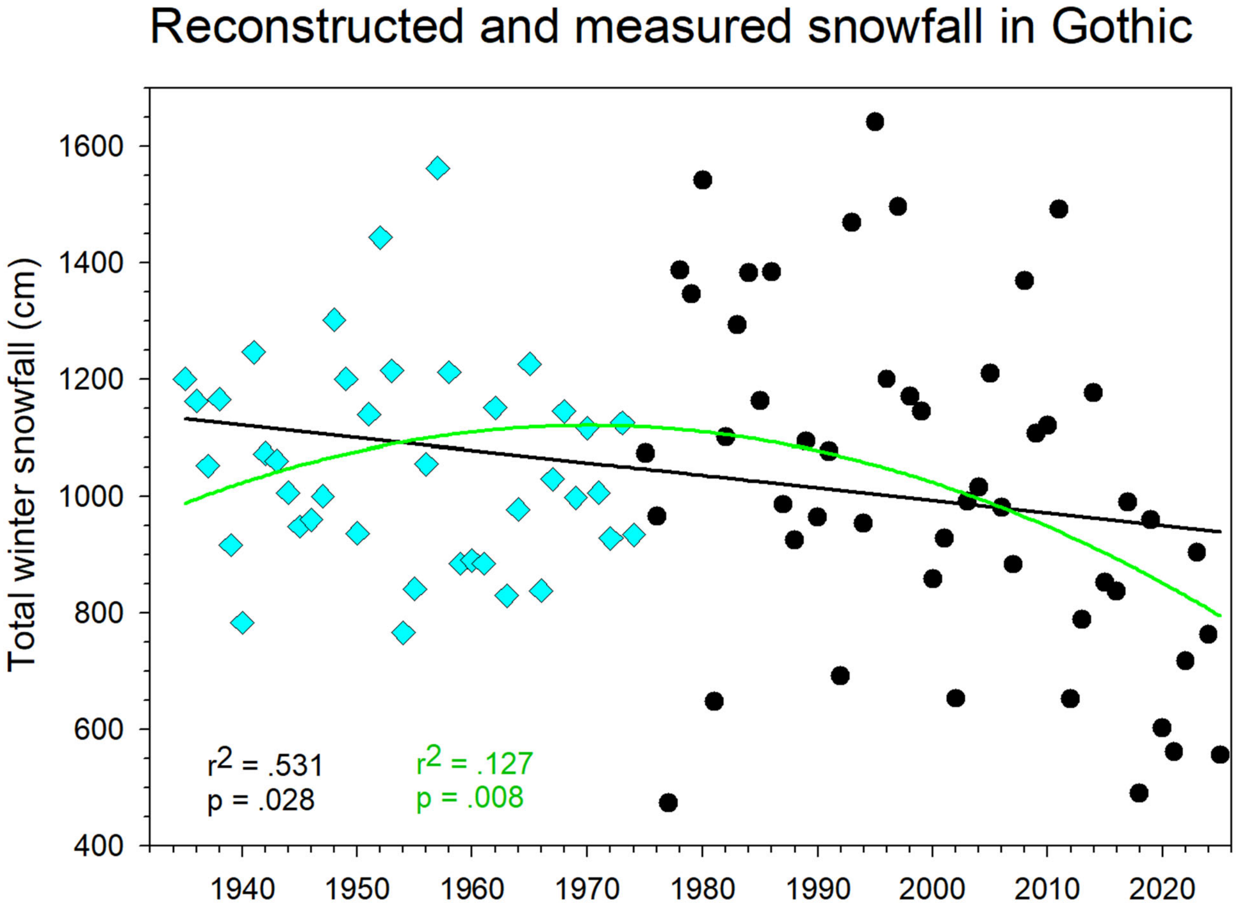

Figure 5. Winter snowfall in Gothic, measured since 1975 and estimated (cyan diamonds) from 1935-1974 (based on the relationship between snowfall in Gothic and peak runoff in the East River at Almont).

The average daily snowfall for snowy days in Gothic has declined significantly during the study period (Figure 6; $r^2 = 0.212$, $p < .0007$). There is no significant trend in the maximum daily snowfall (p = .185), but the mean amount of snow dropped per storm (consecutive snowy days) has declined significantly (Supplemental Figure 36; $r^2 = 0.169$, p = .003). The mean number of snowstorms per year (38.1, S.D. = 3.64) has remained relatively stable, but variable (range 32 – 48). The mean number of consecutive days with snow falling (2.79, S.D. = 0.399, range 1.9 – 3.7) and consecutive days without snow falling (5.9, S.D.= 0.75, range 2.5 – 5.9) are not changing.

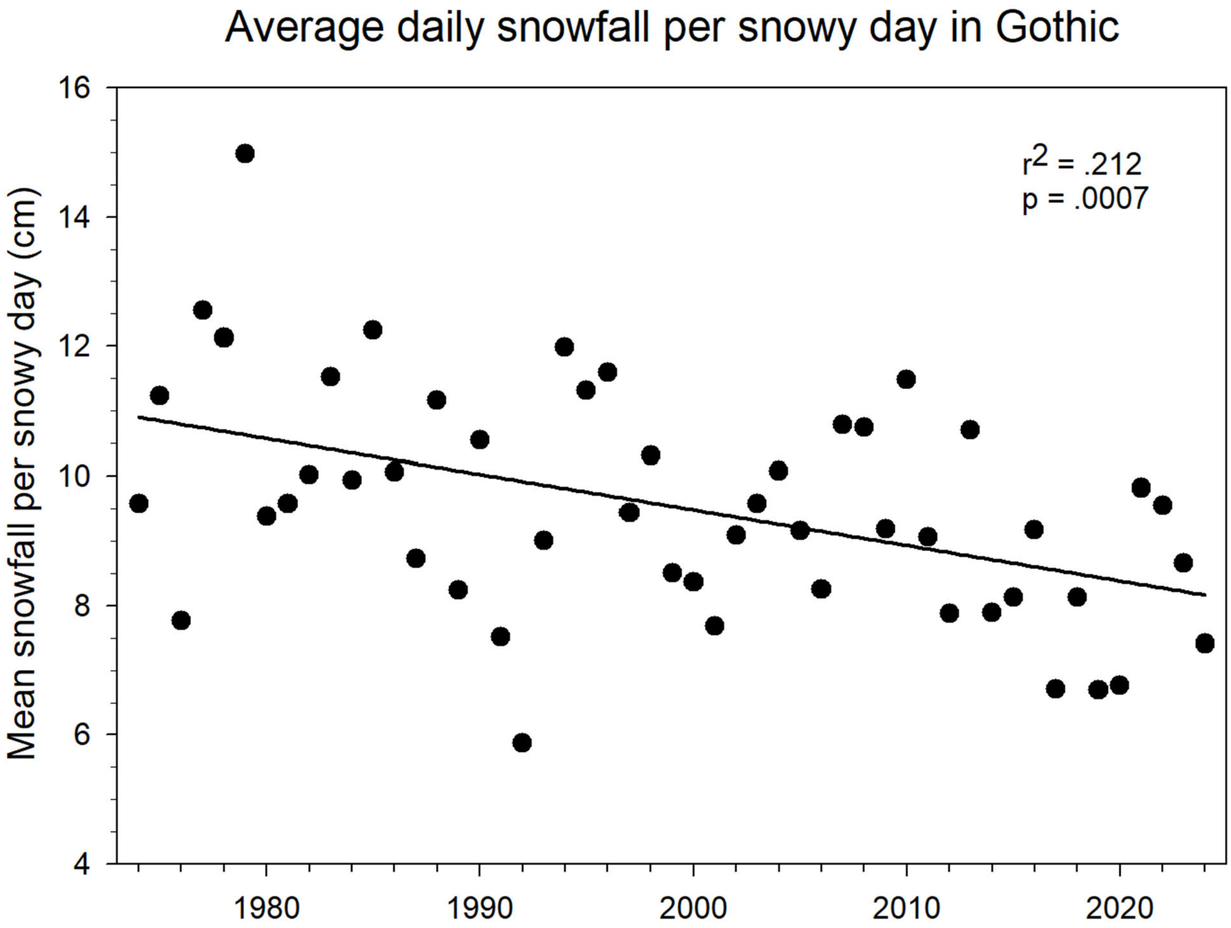

Figure 6. The average daily snowfall per snowy day in Gothic is declining; $r^2 = 0.212$, $p < .0007$.

The temporal distribution of snowfall (plotted as 3-day totals;) shows an approximately normal distribution with a peak in late January. Breaking this into two time periods, 1974-1999 and 2000-2021, shows that snowfall totals have declined in the second period, although the temporal distribution has not (Supplementary Figure 37). The top three snowiest 3-day periods have occurred in early January and mid-February. Mean daily and monthly snowfall measurements since November 1974 are shown in Table 3.

Table 3. Daily snowfall statistics by month from Gothic (cm of snow), November 1974 – June 2025.

	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Mean	5.10	5.76	4.92	4.05	2.27	0.27	0.003	0	0.41	1.99	4.15	4.71
Mean Max.	34.57	36.37	33.06	25.24	21.65	5.31	.06	0	6.38	21.56	31.69	33.82
Mean Min.	0	0	0	0	0	0	0	0	0	0	0	0
Monthly mean total	158.02	163.12	153.67	121.41	70.45	8.29	.08	0	12.18	61.66	124.42	146.06

Monthly snowfall statistics from Gothic (cm of snow), November 1974 – June 2025.

	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Max.	441	383	342	252	275	51	3	0	116	125	328	465
Min.	35	49	37	31	7	0	0	0	0	0	36	18

2. Snowfall density in Gothic

In addition to the changing amount of snowfall, the density of the snowfall (snow water content/snow depth, measured as kg/m^3) is changing, increasing over time (Table 4) (Figure 7 shows data for December and January, and Supplemental Figure 38 shows November - May). Minimum and mean daily density are lowest in the middle of the winter, and the maximum is highest in the spring (Supplemental Figure 39 shows a representative winter, 1992-93). Given that temperature can affect snow density, it’s not surprising that there is a significant correlation between air temperature on the day of snowfall (Heating Degree Days (calculated with base of 65°F)) and the density of the snow that falls ($r^2 = .123$; Supplementary Figure 40 shows data from winter 1994-95). The correlation between density and maximum daily temperature of the day of snowfall is also significant ($r^2 = .094$, $p = .0003$, Supplementary Figure 41).

Although both total winter snowfall and average daily snowfall on snowy days are declining, because snowfall density is increasing, SWE has not declined significantly ($p = .355$ for a linear model of data from 1975 – 2021).

Table 4. Snowfall density (kg / m^3) by month, split into two time periods.

Years	November	December	January	February	March	April	May
1975-1997	630	626	596	628	597	613	714
1998-2024	659	656	664	677	667	713	770

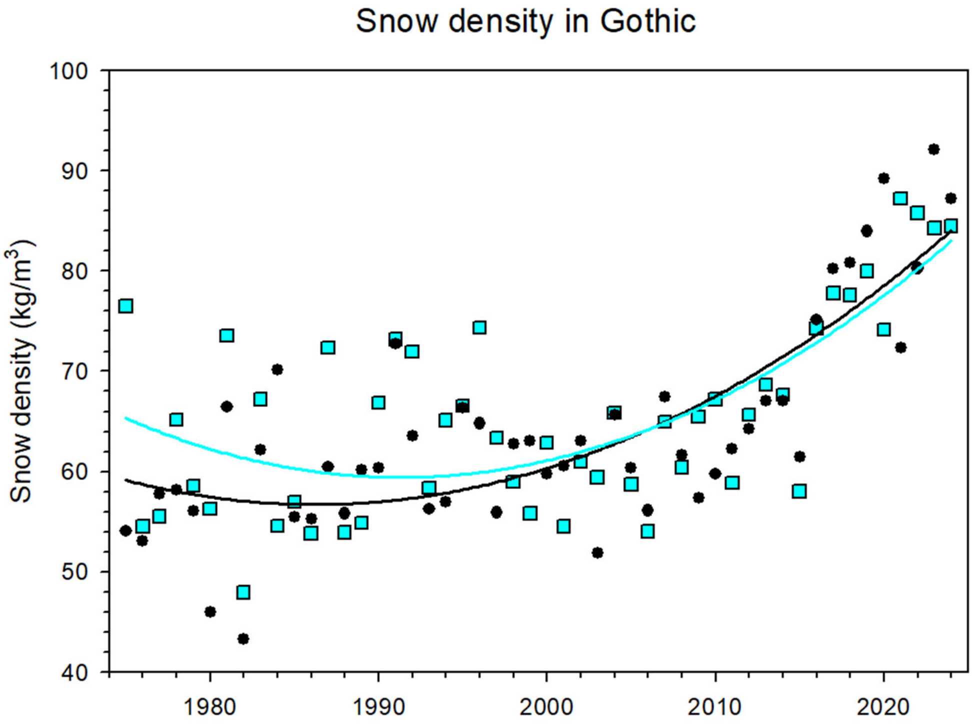

Figure 7. Mean December ($r^2 = 0.439$, p = .001) and January ($r^2 = .334$, p = .02) snow density measured in Gothic. Density has been increasing markedly in the past decade.

Variation in winter snowfall density is changing at least in some months, e.g., coefficient of variation of winter snowfall density has changed significantly (using a quadratic fit) for Jan – April, but not for December (Supplementary Figures 42a, b), with higher values at the beginning and end of the study period. Mean variation is highest for May (mean CV = 0.278), which also has the highest density (Supplementary Figure 38), followed by February, January, December, March, and April, with the lowest (mean CV = 0.220, p = .274).

3. Snowpack

The accumulated snowfall creates a snowpack, and its character is changing. For example, the date when the permanent snowpack starts has ranged from 15 October (1980) to 24 November (1989), but is trending later (see curvilinear fit with a low in the mid-1990s, p = .12; Supplementary Figure 43). The date when the permanent snowpack reaches 10 cm, or approximately what is needed to keep the ground below it from freezing, has been variable, as early as 17 October and as late as 7 December, but is trending later (Supplementary Figure 44), while the last date of 10cm snowpack in the spring is trending significantly earlier (about two weeks earlier now than when the study began; Supplementary Figure 44). The number of days the ground is covered by the permanent winter snowpack is also highly variable but is trending lower (Supplementary Figure 45; p = .13), matching the pattern for number of days with at least 10cm of snow (approximately what it takes to prevent the ground from freezing; Supplementary Figure 46; p = .06).

4. Timing of snowmelt

Snowmelt date (date of the last snow at barr's permanent measurement site) has been trending significantly earlier dates during our study (range 23 April to 19 June; Supplementary Figure 47. It has advanced about 11 days during our study period. We have also estimated the date of snowmelt from 1935 to 1974, by using the correlation between observed snowmelt date in Gothic and the date of peak runoff in the East River downstream at Almont (Figure 8). The trend for this longer time period (p = .074) is also toward earlier dates.

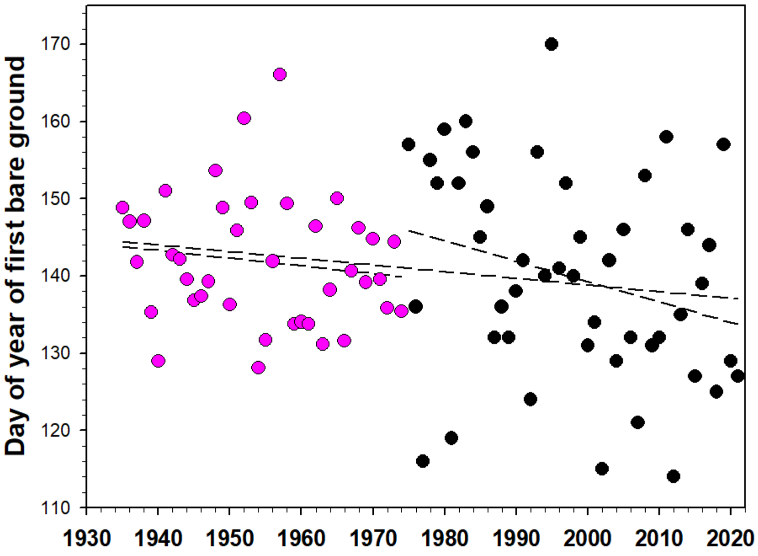

Figure 8. The change in day of year when snow has melted at the snow measurement site in Gothic, plotted against year. Dates from 1935 – 1974 (purple points) are estimated by using the relationship between observed snowmelt date in Gothic and date of peak runoff at the stream gage on the East River at Almont. Regression lines are shown for the estimated (r^2 = .006, p = .38), measured (r^2 = .052, p = .068), and combined (r^2 = .066, p = .0001) datasets.

5. Variability of snowfall and snow density

All metrics of snowfall and snow density in Gothic are characterized by variation, on time scales ranging from daily (or shorter) to decadal (or longer), as illustrated in figures above. For example, total winter snowfall has ranged from a low of 474 cm (1976-77) to a high of 1641 cm (1994-95). Snow density changes within a season (Supplementary Figure 39), but is also changing over the years (Supplementary Figure 38), in concert with warming temperatures, measured either as Heating Degree Days or maximum daily temperature (Supplementary Figures 40, 41).

6. Interactions among temperature, precipitation, and snowmelt date

Ratio of rain to snow

Another aspect of precipitation that is changing is the amount of annual precipitation that is falling as rain or as snow. That ratio of rain to snow is changing, with a tipping point around 1970 in Crested Butte (Fig. 9).

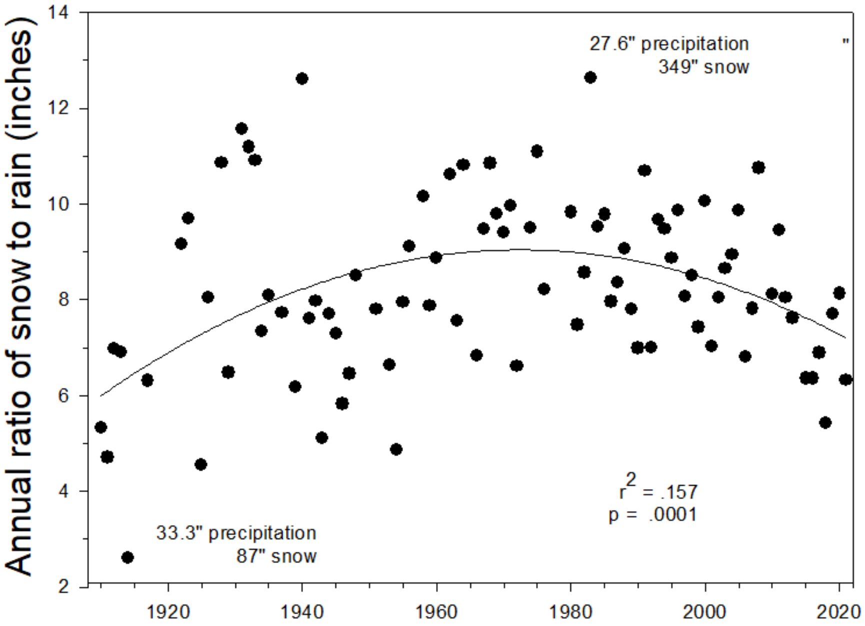

Figure 9. The changing ratio of snow (inches per year of snowfall) to rain (annual inches) measured at the Crested Butte weather station. A tipping point was reached in the 1970s.

The number of days between snowmelt date and the last hard frost (temperature below -3.9°C (25°F) in the spring is variable but has changed significantly (p = .028 since 1975 (Supplementary Figure 48).

7. Soil moisture

The typical annual pattern of soil moisture includes a peak in the spring associated with snowmelt, when the soil is saturated, followed by drying out during the summer that is modified by summer precipitation. Data for a representative year (September 2019 to October 2020) are shown in Supplementary Figure 49.

8. Cycles

The result of plotting the regression of snowmelt date in Gothic against year, in 8-yr running intervals, shows a cyclical pattern (Fig. 10) of 12-14 years.

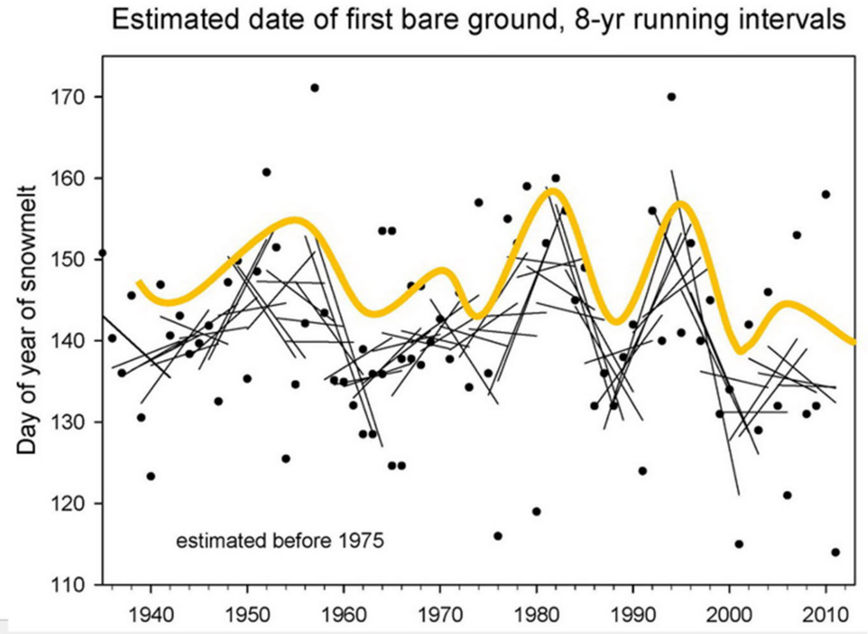

Figure 10. The solid line, redrawn from Ault and St. George (2010), supporting their conclusion that there is a 12-14 yr cycle in autumn precipitation in the central Rocky Mountains, is closely matched by the results of 8-year running regressions of data from Gothic on day of year when snowmelt occurred.

Discussion

Precipitation is an important component of climate that creates the mesic meadow habitat in our montane study area. To understand how responses to changes in recent climate could affect ecological and evolutionary trajectories in montane biota, quantitative data on trends in temperature and precipitation, and their variation, are required. It's clear that in many parts of the world there has been significant warming, and not far from the geographical focus of this paper at the Rocky Mountain Biological Laboratory there is already the largest 2°C hot spot in the Lower 48 (Eilperin 2020).

Data recorded in tree rings and glaciers document that there have been significant changes in this area's climate in the past. For example, tree-ring chronologies record late 20th century snowpack reductions that "are almost unprecedented in magnitude across the northern Rocky Mountains" (Pederson et al. 2011a); these authors attribute the snowpack declines (and their synchrony across Colorado, Columbia, and Missouri River drainages) to positive reinforcement of anthropogenic warming by decadal variability. Going further back into time, Woodhouse et al. (2010) found that “during the medieval period, ~AD 900–1300, the Northern Hemisphere experienced temperatures warmer than all but the most recent decades.” This was a time of extensive and persistent drought in western North America, such that the mid-12th century could be used as an analog of severe droughts that may occur in this area in the future (Woodhouse et al. 2010). The drought in the upper Colorado River basin in the mid-1100s was characterized by a decrease of more than 15% in mean annual flow (averaged over 25 years) over about six decades (Meko et al. 2007).

In more recent history, and more locally, a tree-ring-based reconstruction of 1 April snow water equivalent (SWE) generated for the Gunnison River basin (which includes our study site) for 1569-1999 explained 63% of the variance in the instrumental record (Woodhouse 2003). The twentieth-century part of the record was similar to the longer-term record, although years of extreme SWE did not occur evenly throughout the record, and few extreme years occurred in the 20th century. Woodhouse attributed winter drought to the influence of multiple circulation patterns, including both the El Niño–Southern Oscillation (ENSO) and Pacific-North American (PNA) pattern, but pointed out that the Gunnison River basin “is on the edge of the area with a strong relationship to the PNA and is generally in a transitional zone with respect to regional ENSO influences.” (Woodhouse 2003). The patterns observed in Woodhouse’s dendrochronlogical studies are also reflected in Crested Butte precipitation (Supplementary Fig. 50) (Griffin et al. 2013) (personal communication, C. Woodhouse). A study of climate changes in the northern Rockies based primarily on snow telemetry network (SNOTEL) records since 1969 reported that changes in spring minimum temperatures corresponded with atmospheric circulation changes and surface-albedo feedbacks in March and April (Pederson et al. 2011b).

Long-term droughts have occurred in the Western US in the past, but not since at least the late 1500s (Woodhouse and Overpeck 1998, Stahle et al. 2020). Persistence of the current megadrought is probably caused by a combination of decadal-scale oscillations in Eastern Pacific sea surface temperatures, consequent shifts in precipitation patterns, and enhanced regional terrestrial evaporation (Williams et al. 2020). Szejner et al. (2020) found that the probability of extreme drought has increased up to 70% in the Southwest during the past two decades. Ault et al. (2016) concluded that “the risk of multidecadal megadrought in the American Southwest will increase at least twofold from 5 to 15% over the last millennium to between 20 and 50% this century.” Historically, meteorological drought in the western United States has been driven primarily by precipitation deficits, but recent analyses show that, “since around 2000, rising surface temperature and the resulting high evaporative demand have contributed more to drought severity (62%) and coverage (66%)” over the western US than precipitation deficit, and attributed the increase in evaporative demand during droughts primarily to anthropogenic warming (Cook et al. 2014, Zhuang et al. 2024).

Variation in and effects of summer precipitation

Climate change may result in increased precipitation variability and consequently more frequent extreme events (Zhang et al. 2021), in the case of rainfall at least in part due to a warming-induced shift from snow to rain (Ombadi et al. 2023). Zhang et al. (2021) found trends of reduced annual precipitation across most of the western USA, with increasing interannual variability of precipitation. Extreme-duration drought became more common, with increases in both the mean and longest dry interval between precipitation events and greater interannual variability in these dry intervals. Although some models predict that climate change will result in increased precipitation and increased variability, data from the Crested Butte weather station show a trend toward a lower number of rainy days from May - September (Supplementary Fig. 10), a trend toward an increasing amount of rainfall per rainy day (Fig. 2), a decreasing trend in the mean rainfall per storm (Supplementary Fig. 5), and an increase in the number of consecutive days with no rain (Supplementary Fig. 13). We also found increasing variability in daily precipitation (p = .0001; Supplementary Fig. 14). Rainfall from atmospheric rivers may increase with temperature (Payne et al. 2020), and this might bring additional precipitation to our study area, perhaps influencing variables such as the precipitation amount per day and per storm.

Gothic, 849m higher than Crested Butte, typically receives more precipitation (Supplementary Figure 34), although in the month of June that trend is reversed. From 2000 – 2024, the daily maximum precipitation per summer month was greater in Gothic in 82 of 122 months. The general trend of increasing precipitation with altitude is also conspicuous along the gradient from Gunnison (2348 m, with a NOAA weather station) to Schofield Pass (3263 m; the top of the East River valley, with a nearby SNOTEL site). Rainstorms can be quite localized in this area, and it would be useful to have a network of precipitation gauges in the area to document this effect. Summer monsoons, however, tend to be more regional in effect so the differences across altitude might be less from such events.

Precipitation can influence plants and animals in the montane ecosystem. Plant survivorship, flowering, nectar production, and seed set can all be affected, and mushroom production by many fungi is highly dependent on summer precipitation. Pollinators that rely on flower abundance and production of nectar and pollen can be influenced by summer precipitation both directly (e.g., flooding of subterranean nests, curtailment of foraging) and indirectly (e.g., via changes in floral resource abundance and phenology (Ogilvie et al. 2017)). Changing precipitation dynamics can drive shifts in ecological interactions such as herbivory. For example, a regime shift in the dynamics of a moth species in California was documented during a 34-year study and attributed to contemporaneous changes in regional precipitation (Pepi et al. 2021). In our study area, variability in summer precipitation generates mass flowering by a common wildflower (*Frasera speciosa*; Gentianaceae), but with a 4-year lag time due to multi-year preformation of leaves and inflorescences (Taylor and Inouye 1985, Inouye 1986).

Studies of the changing climate in the U.S. Southwest, including Colorado, have focused on regional changes in precipitation and temperature. This area has “experienced a strong hydroclimate trend from the 1980s to the 2010s, from cool and wet to warm and dry conditions.” (Lehner et al. 2018). They attributed the warming primarily to greenhouse gas forcing, and the drying to internal climate variability, which is large in this region. McKinnon et al. (2021) found that specific humidity on dry days during July – September has decreased over the past seven decades in Southwest USA (including a site in Colorado), and attribute this to the anomalously low evapotranspiration that results from low summer soil moisture. Changing precipitation regimes can increase biodiversity loss; for example, long droughts and warming temperatures can stress plants, causing increased mortality and impoverished communities (Breshears et al. 2005, Mueller et al. 2005, Reynaert et al. 2021).

Seedling survival can be strongly influenced by rainfall patterns. If seedlings don't put down roots quickly enough to reach soil with sufficient moisture, they will die. Although soils in our area are saturated by snowmelt in the spring (Supplementary Fig. 46), we have had periods after snowmelt as long as 71 days with no precipitation, resulting in significant soil desiccation and seedling mortality (Inouye, unpublished observations). Thus, the trend over the past four decades for a decreasing number of rainy days (Fig. 3) and increasingly long periods without rain (Supplementary Fig. 11) could bode poorly for future seedling survival and probably result in declining seed production by species that ripen seeds in late summer. There also appears to be (Supplementary Fig. 3) a trend for decreasing amount of rainfall per rainy day.

Nectar production is dependent on soil moisture (Powers et al. 2022), and increasing periods of drought could result in lower nectar production and therefore fewer resources for pollinators, and potentially reduced visitation and seed set. For example, drought was found to affect visitation and seed set in a subalpine plant species in the central Chilean Andes; nectar standing crop was reduced, although its concentration was increased (Arroyo et al. 2020). The reduced seed set was predicted to have the potential to result in a decline of the plant population if drought conditions continue. In our study site drought was found to reduce both nectar production and pollen production in a common wildflower, *Ipomopsis aggregata* (Waser and Price 2016), and this effect on floral resources has also been documented elsewhere (Carroll et al. 2001, Alqudah et al. 2011, Phillips et al. 2018), reducing male fitness (Powers et al. 2022)

Drought can also affect other floral traits, including flower size and abundance. Kuppler et al. found that decreasing soil moisture content correlated with decreasing size of all measured morphological floral traits except stamen length and nectar tube width, resulting in decreased visitation by bumble bees (Kuppler et al. 2021). They did not measure pollen transfer, but it seems likely that decreased visitation could result in lower seed set. Induced experimental drought has also been shown to decrease flower size and nectar volumes (but not concentration) in *Epilobium angustifolium*, a common plant species in our study area (Carroll et al. 2001). Water stress can also affect flower phenology in addition to decreasing flower size and flower number (by 22-45%) (Descamps et al. 2020).

Floral volatile emissions can also be affected by drought while negatively affecting floral resources for pollinators, seed set, and seed mass (Rering et al. 2020). Powers et al. (2025), in an experimental study manipulating snowmelt and precipitation at the Rocky Mountain Biological Laboratory, found that within years, floral emissions of *Ipomopsis aggregata* (Polemoniaceae) did not respond to precipitation treatments but shifted with snowmelt treatment depending on the year, while across the three-year study, emissions correlated with both precipitation and snowmelt date; these effects were driven by changes in soil moisture.

The negative effects of drought can also extend to decreased protein quality of pollen, and cascade to decrease survival and productivity in both honey bees (*Apis mellifera*) and bumble bees (*Bombus impatiens*) (Wilson Rankin et al. 2020). Experimental studies of the addition of water or reduction of natural precipitation are useful tools for understanding the consequences of altered water regimes, and have been employed at our study site and elsewhere (Sloat et al. 2015, Gallagher and Campbell 2017, Gorton et al. 2019, Recart and Campbell 2021, Faust and Iler 2022).

Variation in and effects of winter precipitation

Significant changes have occurred in snowpack in many mountain areas. For example, snow cover in the Alps has declined 5.6% per decade over the past 50 years, and duration of snowpack cover is now 36 days shorter than the long-term mean (Carrer et al. 2023). It has also declined in the western United States in recent decades (Mote et al. 2018), as well as across the Northern Hemisphere (Mudryk et al. 2020). Between the 1980s and 2000s there was a 10-20% loss in the annual maximum amount of water in the western U.S. regional snowpack, with a loss of up to 60% projected by 2047 (Fyfe et al. 2017). From 1982 to 2016 annual maximum snow water equivalent (SWE) decreased by 41% on average over much of the western USA, and snow season shortened by 34 days on average for some parts of it, primarily from earlier ending of the season (Zeng et al. 2019). In addition to this trend, simulations and models of mountain snowpack suggest that interannual variability of maximum SWE and its timing are likely to decline in areas where more of annual precipitation falls as rain instead of snow (Marshall et al. 2019), although variability of SWE_{max} timing could increase in some area. These models also predict that average frequency of consecutive snow drought years will increase from 6.6% to 42.2% of years. However, another modeling study of cold-season storms in the western USA suggests that under a high-emissions scenario for the future, precipitation volume from the top 20% of winter storms may increase by up to 40% by mid-century (Chen et al. 2023).

Snow is a good insulator, and a loss of snowpack could have significant impacts on species sensitive to freezing (Kearney 2020). In our study area ptarmigan burrow into the snow that provides both insulation and camouflage, and snow provides insulation for hibernating mammals and many insects overwintering underground as eggs, larvae, or adults, as well as plants (many of them have pre-formed buds not far underground); typically there is enough snow on the ground to protect the ground from freezing all winter. As snowfall declines (Fig. 5) and becomes denser (Fig. 7), the insulation from cold air temperatures declines, potentially stressing both plants and animals. This effect may be most pronounced at mid-elevations (Roberts et al. 2021). Loss of snowpack can also have a significant economic effect, because of the value of the water it contains; “snow cover also provides social and cultural services for human beings, such as aesthetics and entertainment, scientific research, environmental education, religion and culture, etc.” (Gao et al. 2024).

Snow hardness, which can be influenced by rain-on-snow events as well as snow density, can have a negative effect on small mammals in the arctic (Poirier et al. 2021), and it seems likely this can also happen in our study area. The subnivium, however, may be relatively unaffected as long as there is snow cover for insulation (Thompson et al. 2021). Loss of this insulation can also have an impact on ecohydrological interactions such as evapotranspiration

(Pirk et al. 2023) and the protection that snowcover provides for abrasion from blowing snow and ice. Increasing density of the snow can also have implications for avalanche activity. The winter of 2019-20 had record avalanche activity in the mountains near our study area (Art Mears, personal communication), and had the highest monthly snow densities recorded to that point for January, February, and May, and the second-highest for November, December, and March. Denser snow facilitates the propagation of fracture lines when avalanches are triggered, and there were unusually long facture lines that winter. Avalanches can have significant effects by knocking down trees, altering the vegetative community (Patten and Knight 1994).

An approximate 18% loss of SWE in the Rocky Mountains is forecast for 2025-49, and >40% SWE loss forecast for 2075-99, with about 85% of the Upper Colorado River basin forecast to have low to no snow by 2100 (Siirila-Woodburn et al. 2021). For a variety of reasons, this scale of decline could have a major impact on natural and anthropogenic landscapes in the mountains, affecting agriculture, recreation, and plant and animal populations. One potential mitigating factor could be the effect of winter cloud seeding, which can make it snow up to 12% more in a single storm, and 8% more over a winter season (Andrew Rickert, CO Weather Modification Program Manager; https://www.denverpost.com/2022/03/26/colorado-cloud-seeding-snow-drought/).

1. Timing of snowfall– temporal changes and ecological consequences

There is significant variation in the timing of snowfall in Gothic, and the only month without measurable snowfall since 1974 is August. In 1969 a 3-day snowstorm beginning 24 June left up to 50cm of snow in Gothic (Ehrlich et al. 1972), and 11.4 cm in Crested Butte (NOAA weather station data). We estimate that the first date of bare ground in Gothic was 19 May that year, so that many plants were already up and flowering, and most animal populations active, by the date of the storm. The storm caused extensive damage to herbaceous plants, as well as some insect and small mammal populations, and caused the local extinction of at least one butterfly (Ehrlich et al. 1972). The low temperature in Crested Butte on 27 June, the day after the snow ended, was -3.9°C. Marmot reproduction was poor that summer, and populations of small mammals were also depressed (Ehrlich et al. 1972).

There is a relationship between the length of time between snowfall events and the stability of the winter snowpack. Exposure to the sun, sublimation, air temperatures, and precipitation amount and temporal pattern can result in either more or fewer avalanches, which can cause landscape-scale changes by pushing over trees. For example, a link has been established between ENSO cycles and avalanches (McClung 2013), and with weather in the Alps (Castebrunet et al. 2012). The changing climate in our study area may promote more frequent avalanche activity, but this effect could be moderated if snowfall continues to decline.

2. Density – changes over time and ecological consequences

Density of snow is affected by multiple variables, including air temperature, humidity, and whether the crystals have fallen through high concentrations of supercooled water, which causes them to become rimed (Judson and Doesken 2000). Density influences avalanche occurrence and snow safety, and the insulation value of snow. Density changes over time in the snowpack, as a function of initial density, prevailing weather conditions, pressure from snow above, temperature in the snowpack, and time since deposition, as snow crystals metamorphose into more rounded shapes (Judson and Doesken 2000). Density can change during a storm cycle, and whether the initial snow is denser or lighter than the following snow can influence avalanche probability (and the quality of powder for skiing) (https://www.eoas.ubc.ca/courses/atsc113/snow/met_concepts/07-met_concepts/07c-up-down-snow/).

Changes in the density of snow, as are ongoing in Gothic, can have significant implications for animals that rely on snow as a substrate for walking, for insulation, or experience it as a barrier to access to plants or subterranean habitat (Hansen et al. 2013, Descamps et al. 2017, Poirier et al. 2021). Changes in rain-on-snow events have implications for flood risk (Musselman et al. 2018).

Thermal conductivity of snow increases with snow density, with lighter snow serving as a more effective insulator for plants and animals in or below the snowpack. Density has a complicated relationship with temperature, but in some cases there can be a linear relationship with increasing air temperature resulting in denser snow, as seen in Supplementary Figure 41 (showing density increasing with air temperature). Supplementary Figure 39 shows that the minimum density typically occurs during the coldest months of the winter, and the highest in May. The temporal changes we have observed in snow density are likely a consequence of changes in temperature over the study period, both within and among years. Snow density can also impact soil temperature under the snow, and as the insulating capacity of snowpack declines with increasing density, both soil temperature and nutrient cycling in the soil can be impacted (Kelsey et al. 2021).

Density can also change across a landscape, with significant differences occurring over a few km (Judson and Doesken 2000). The marked differences observed in snowmelt dates across the landscape in the East River valley (Ian Breckheimer, pers. comm.; Inouye and Barr, personal observations) are suggestive of the spatial variation in snowpack density in our study area, which could result from variation in slope and aspect. For example, south-facing slopes receive more sun, which will transform the snowpack more quickly than north-facing slopes, making the snow denser.

Some previous measurements of snow density from Gothic are available, such as a record from the U.S. Forest Service Westwide Network (Judson 1970) of 46 cm of 44 kg m^3 snow , as are comparisons of snow density in Gothic with Red Mountain Pass (r = 0.68; 129 km away) and Dry Lake (r = 0.36, 175 km away) (Fig. 6, Judson and Doesken 2000). Our measurements are comparable, but avoid some sources of variation (those incurred as snowpack metamorphoses in response to temperature and pressure) by restricting calculations to newly fallen (within the past 24 hours) snow rather than snow that has settled to some degree.

3. Snowpack

Average snowfall fraction of annual precipitation has decreased steadily in the Northern Hemisphere since 1950; this decline has a significant effect on streamflow seasonality (e.g., when 50% of total streamflow has passed a geographical location) and the strength of the seasonal pattern of streamflow (typically with high flows in spring and low flows in autumn in our study site) (Han et al. 2024). As climate change results in smaller snowpacks, these characteristics of streamflow will change, with consequences ranging from effects on aquatic organisms (e.g., lower flows, higher temperatures) to changes in availability of water for recreation and agriculture. (Suni et al. 2023)

The date when the permanent winter snowpack begins in Gothic is trending toward later dates over the past few decades (Supplementary Figure 43), and the number of days of permanent snowcover is declining (Supplementary Fig.). The date at which 10 cm of snow cover is achieved is occurring later (Fig.), and the date when snowpack drops below 10 cm is occurring earlier (Supplementary Figure 32), resulting in a significant drop in the number of days with 10 cm or more (Supplementary Figure 45). The loss of insulating snow cover exposes the soil and the plants and animals it shelters in the winter to colder temperatures, which could result in increased mortality. For example, the very low and late snowpack in the winter of 1976-77 in our study area may be responsible for the unusually low abundance of bumble bees the following summer (Inouye, personal observation); overwintering queens could have been exposed to unusual freezing temperatures in the absence of the insulating snowpack. This might also explain the extremely low abundance of flowers of *Delphinium nuttallianum* in 1977 (Inouye and McGuire 1991).

Changes in snowpack depth can influence bird distributions (Bosco et al. 2022) and those of big-game animals that migrate altitudinally to avoid deep snowpacks, and whose migratory phenology is linked to snowfall and snowmelt dates (Bischof et al. 2012, Merkle et al. 2016, Middleton et al. 2018, Laforge et al. 2021). Another variable that is increasingly affecting snowpack in the mountain west is wildfires (Kampf et al. 2022). Experimental manipulations of snowpack and snowmelt date are useful tools for investing the consequences of earlier, or later, snowmelt (Gezon et al. 2016, Sherwood et al. 2017, Iler et al. 2019, Kudo and Cooper 2019, Hamann et al. 2021, Jerome et al. 2021, Rixen et al. 2022).

4. Timing of snowmelt – temporal changes and consequences

The growing season for montane and alpine plants doesn't begin for most species until the snow is gone, although two species in our area (*Claytonia lanceolata* and *Erythronium grandiflorum*) will grow up through the last few cm of snow. Activity of many animal species is also tied to this environmental event, as warming soils signal species overwintering underground, such as bumble bees and many other species of solitary bees, that the summer season is beginning. Some hibernating species may emerge before the snow melts, such as marmots (*Marmota flaviventris*), although their appearance is also linked to the warming temperatures and melting snow (Inouye et al. 2000). The flowering phenology of almost all plant species in these environments is also closely tied to snowmelt date (Inouye and Wielgolaski 2013), as is the abundance of some species of flowers, sometimes mediated by effects of late spring frosts that can have a significant effect in years with early snowmelt (Inouye 2000, Inouye et al. 2002, Inouye 2008). These frost effects can also have a lagged effect on subsequent abundance of insects (Boggs and Inouye 2012). Earlier snowmelt (and lower summer precipitation) can result in decreased corolla length, style length, corolla width sepal width, and nectar production in *Ipomopsis aggregata* (Polemoniaceae); earlier snowmelt also resulted in shorter inflorescences (Powers et al. 2022). Velocity of snowmelt across landscapes also predicts green wave velocity (O'Leary et al. 2020).

Given the significance of snowmelt timing, the fact that it has advanced significantly during the period of our study documents the importance of declining winter snowfall, warmer spring temperatures, and the increasing frequency of dust-on-snow events that accelerate snowmelt in the spring (Steltzer et al. 2009, Painter et al. 2018), for the phenology of many species and the abundance of some. An experimental study simulating future flow regimes in Sierra Nevada mountain streams in California found that early snowmelt "destabilized stream epilithic biofilm metabolism and altered key ecosystem functions such as insect production and emergence, via shifts in community composition, structure, and phenology" (Leathers et al. 2024). Pulses of the dominant insect group (Chironomidae) emerging from streams almost doubled in magnitude, benefiting a generalist riparian predator, Brewer's Blackbirds (*Euphagus cyanocephalus*). This study illustrates how climate change effects in mountain streams can alter the dynamics of stream food webs via fine-scale changes in phenology, with consequent phenological mismatches (Leathers et al. 2024).

5. Variability of snowfall and snow density

One prediction of some models of climate change is that the variability of climate measures will increase, and this is borne out in some analyses. Over the last 50 years greater interannual variability has been recorded in dry periods in the western USA (Zhang et al. 2021). On a global scale variability in precipitation can have an impact on terrestrial carbon capture (Knapp et al. 2002, Ritter et al. 2020), with increases in productivity with higher interannual rainfall variability in some ecosystems and decreases in others. Large-scale climate variability has also been shown to influence both plant phenology and herbivore demography (Post and Stenseth 1999). A research need for our study system is investigation of how it is being affected by climate variability with regard to carbon sequestration, and how plant-herbivore and plant-pollinator interactions are being affected.

6. Interactions among temperature, precipitation, and snowmelt date

There is evidence globally of shifts in the ratio of precipitation falling as rain vs. snow (Knowles et al. 2006, Feng and Hu 2007, Wang et al. 2016, Lynn et al. 2020), with more precipitation falling as rain and less as snow as temperature warm. Crested Butte is no exception, and while the ratio was increasing toward more snow and less rain for the first decades of the instrumental record, about 50 years ago the ratio switched to an increasing amount of rain compared to snow (Fig. 32). This change has significant implications for water management in the western U.S.A., as snow has historically served to create a reservoir of water that is slowly released during the summer as it melts. This new pattern will require changes to agriculture and recreation that relies on water in rivers, as snowmelt dates continue to get earlier. Earlier snowmelt dates can also have significant consequences for plant demography. Campbell (2019) concluded that earlier snowmelt will lead to negative population growth of a common hummingbird-pollinated wildflower (*Ipomopsis aggregata*) in montane Colorado, driven especially by changes in seedling establishment and seed production. Earlier snowmelt can also expose plants to frost damage (Inouye 2008).

7. Soil moisture

In our study area the typical pattern for soil moisture is that maximum levels are reached at the time of snowmelt. The subsequent decline, which typically continues for the remainder of the growing season, is reversed only by significant monsoon rainfall. Soil moisture then stabilizes throughout the winter (Supplementary Fig. 46). It's not uncommon for soil moisture to drop to levels that create water stress for plants in our study area. In 1924 there were 71 consecutive days with no rain, and while this may be an outlier (Supplementary Fig. 11), in 2013 there was a stretch of 45 days with no rain. In Gothic if there are intervals as long as a month during the summer with no precipitation, some plants will stop flowering and their leaves may dry out (Inouye, personal observation).

8. Runoff and stream flows

The declining winter snowpack and warmer temperatures will mean that in the future spring runoff quantity and timing will be less predictable from snowfall amounts (Wieder et al. 2022), making management of water resources more challenging. Variability in precipitation and temperature will be reflected in the sto chasticity of runoff. Altered ecohydrological function will affect soil moisture, water levels in kettle ponds, lakes, and streams in our study area, and the aquatic organisms associated with them.

9. Cycles

Cayan et al. reported evidence of decadal-scale variations of precipitation over western North America, which explained 20-50% of the variance of annual precipitation (Cayan et al. 1998). They attributed the variations to decadal atmospheric circulation and sea surface temperature anomaly patterns, as well as short-period circulation patterns such as the Pacific - North American pattern. More recently, Ault and St. George identified decadal variability in autumn precipitation that contributed 25-30% of total variance in the central Rocky Mountains (Ault and St. George 2010). The pattern they found was not explained by major climate modes such as ENSO or the Pacific Decadal Oscillation, so they suggested it was not a product of a simple linear translation of any single climate forcing. The pattern they described is closely matched by our calculations of regressions between snowmelt and year (Fig. 10). We don't know whether this pattern has persisted in the subsequent two decades, but it would be interesting to look for evidence of this pattern in variables such as abundance and phenology of species in our study areas.

Conclusions

There is no doubt that the climate in our study area has changed during the period of instrumental records from Crested Butte and Gothic. It is likely that these changes are ongoing and will persist, given the continuing trends in production of greenhouse gases and the effects they are having on temperature and precipitation, globally, regionally, and locally; a recent projection is that the ongoing megadrought in the southwestern US will continue for decades (Todd et al. 2025). The changes we have observed in phenology of a variety of species in our study area (Prather et al. 2023) can be attributed to changes in climate, leading to the conclusion that many aspects of montane ecosystems in the southwestern Rocky Mountains will change in the future. This highlights the importance of collecting long-term ecological data, for climate, and plant and animal populations, if we want to understand the changes that are in progress.

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Author Contributions: barr collected the Gothic weather data, Prather did breakpoint analyses and assisted with writing, and Inouye conceived the study, did most analyses, and writing.

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Supplementary Figures

Figure SF1. Mean May – September 3-day precipitation totals (both Snow Water Equivalent and rainfall, cm), 1910 – 2023, lumped in 20-year blocks. Data from the Crested Butte weather station.

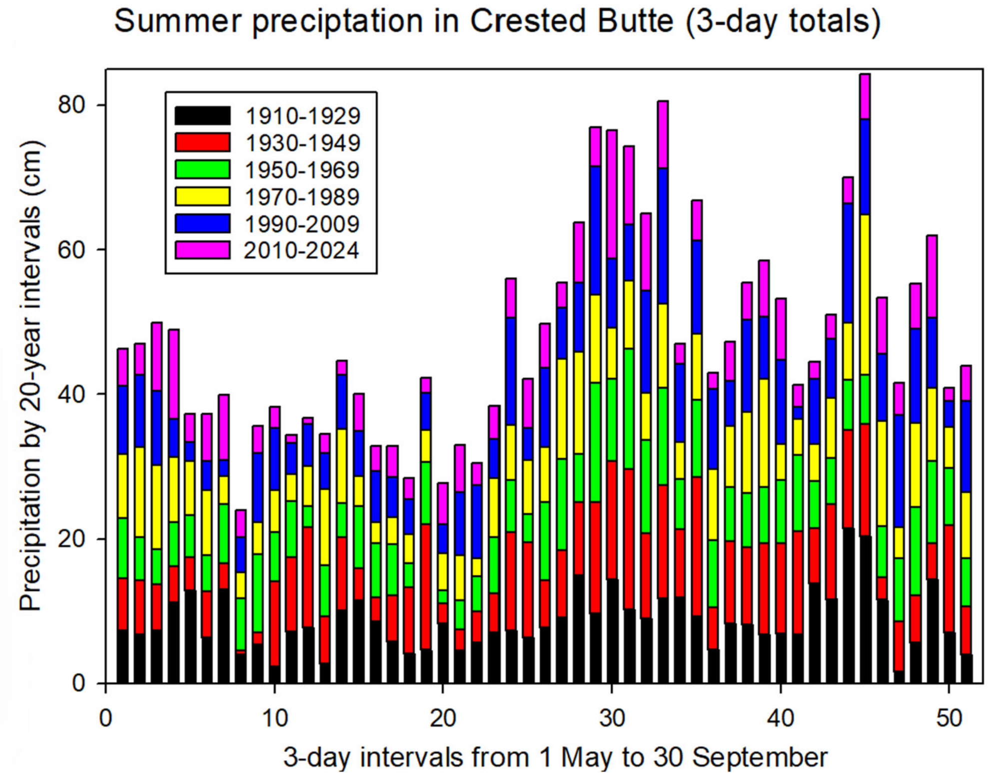


Figure SF2. Total summer (May – September) precipitation in Crested Butte by year, 1910 – 2024. 1919 data are not available, and 1927 appears to be an outlier. There is no significant trend. Data from the Crested Butte weather station.

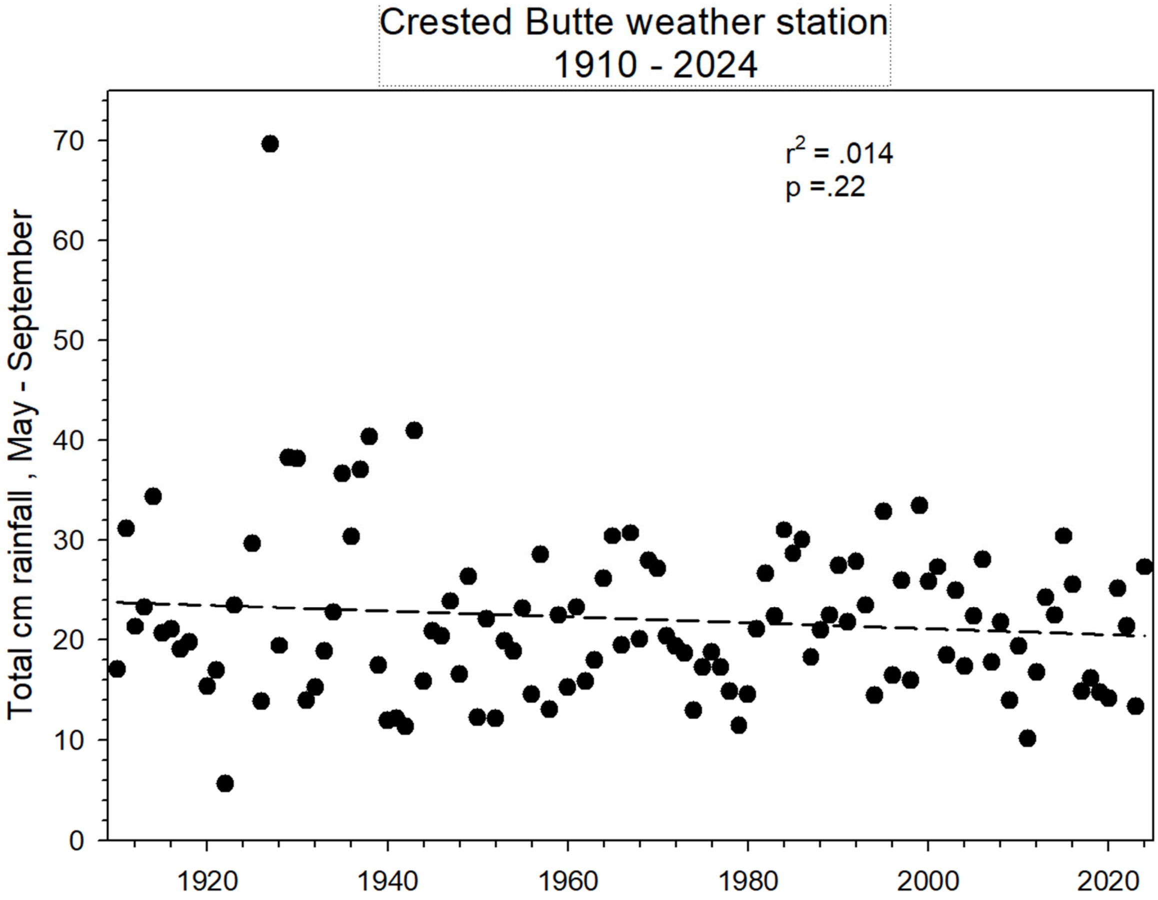


Figure SF3. Mean daily precipitation (cm) in June in Crested Butte, 1910 – 1924. There is no significant trend. Data from the Crested Butte weather station.

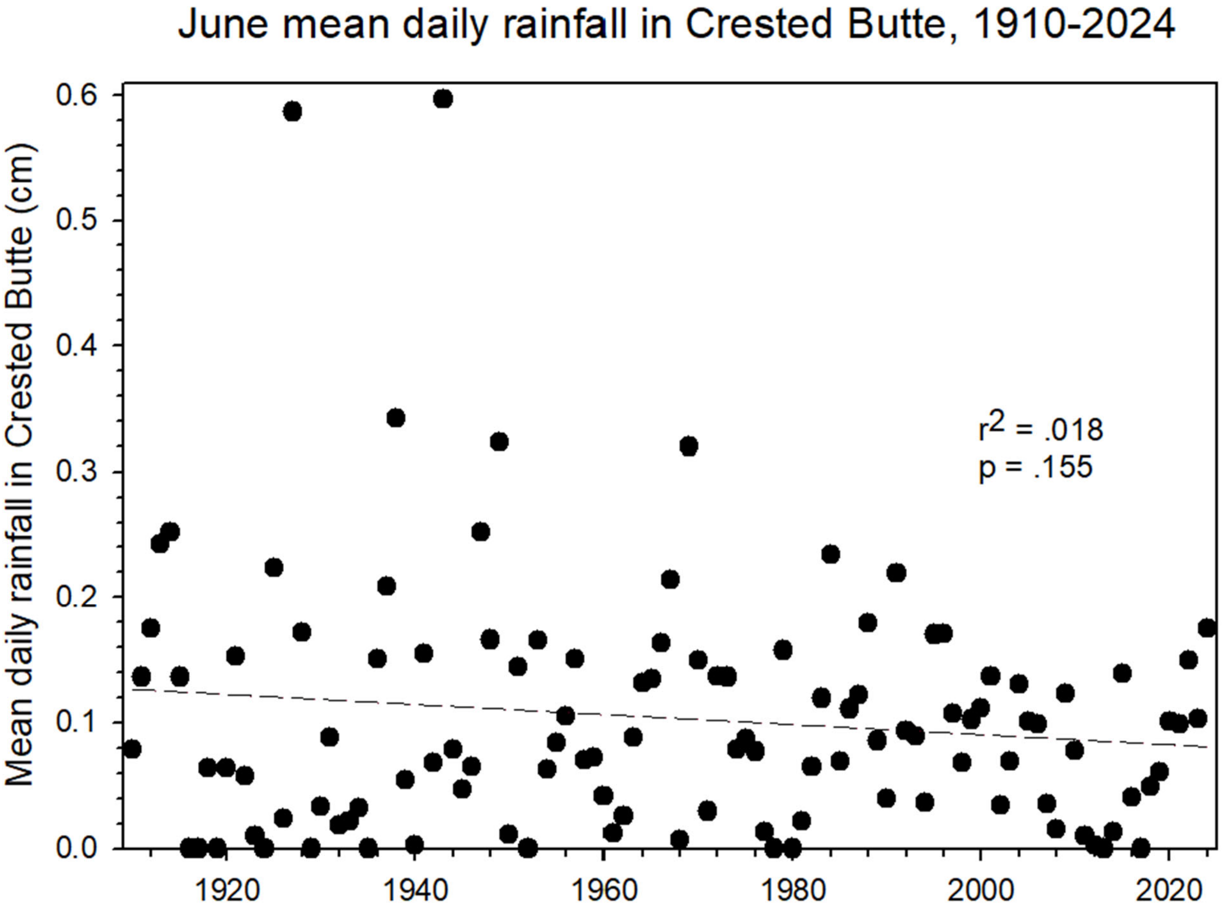


Figure SF4. Maximum daily precipitation (mm) each summer in Crested Butte, 1910 – 2023. There is no significant trend. Data from the Crested Butte weather station.

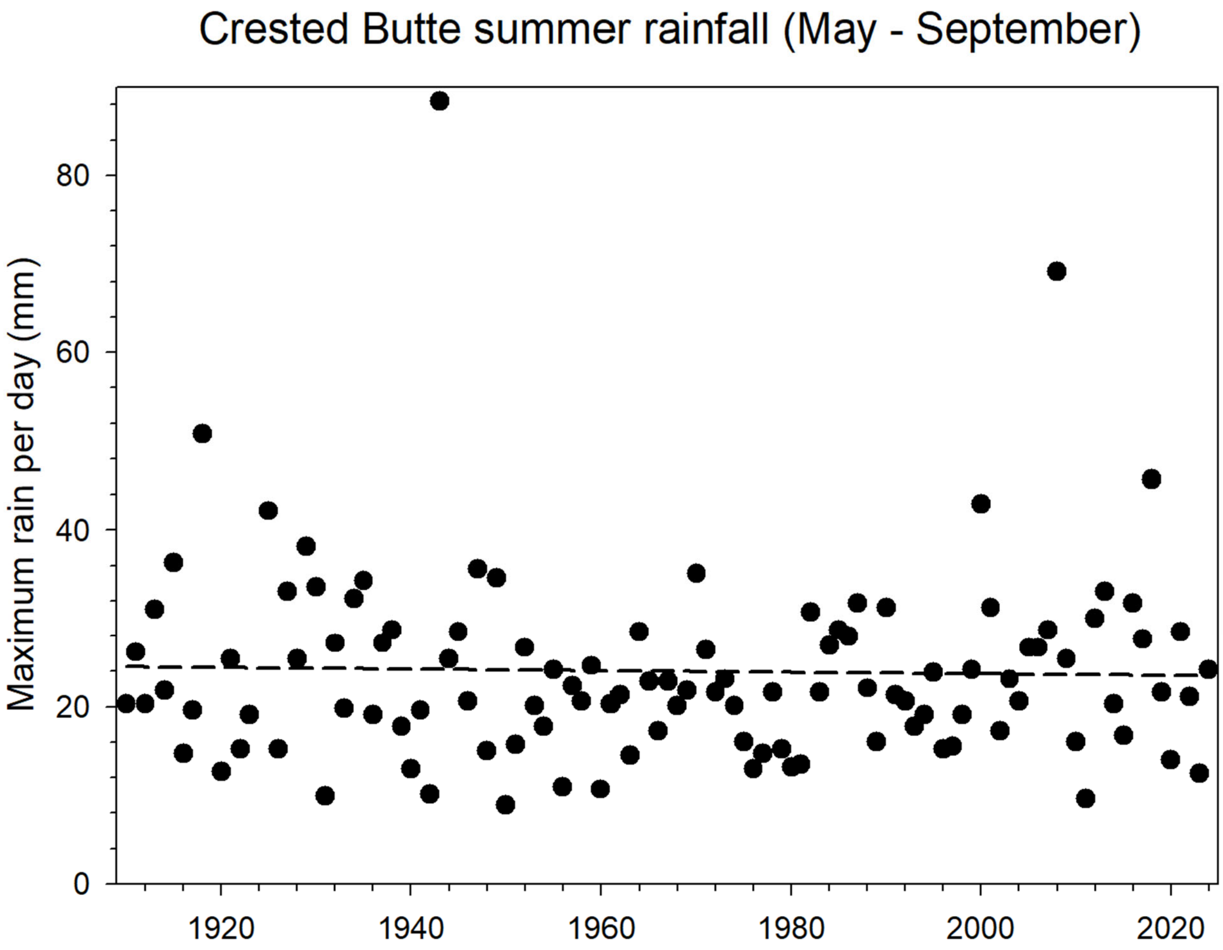


Figure SF5. Maximum precipitation per storm (consecutive days of precipitation, summer (May – September), 1910 – 2024 in Crested Butte. There is no significant trend. Data from the Crested Butte weather station.1927 appears to be an outlier.

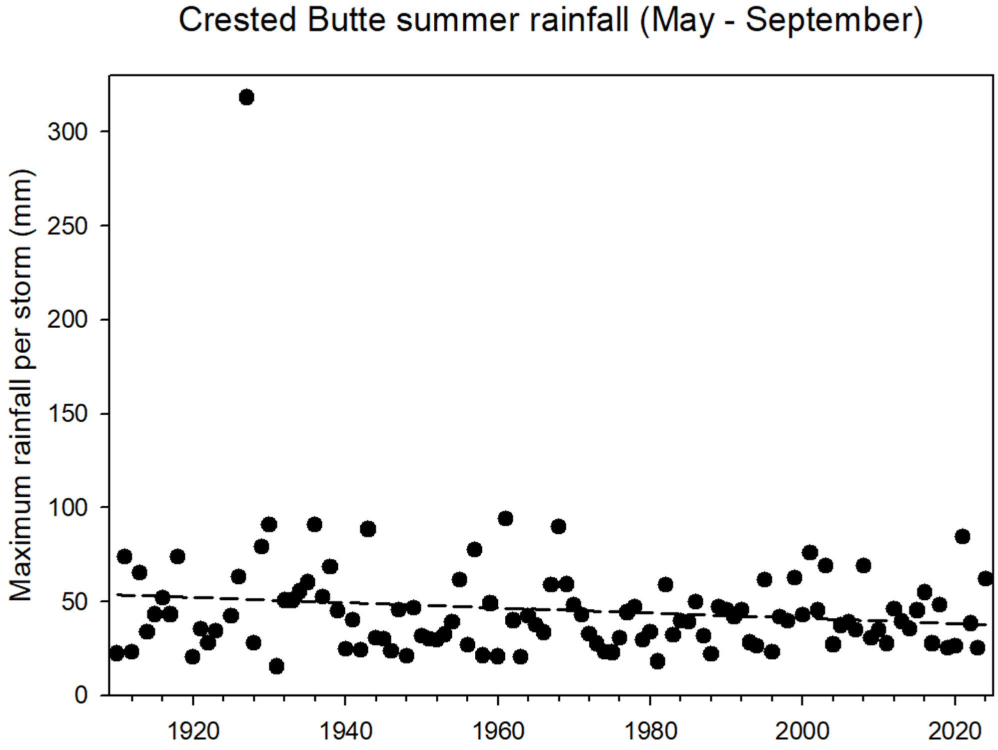


Figure SF6. Mean precipitation per storm (consecutive days of precipitation), summer (May – September), 1910 – 2024 in Crested Butte. There has been a significant (p = .003) decline. Data from the Crested Butte weather station.1927 appears to be an outlier; without it, r2 = .065, p = .007.

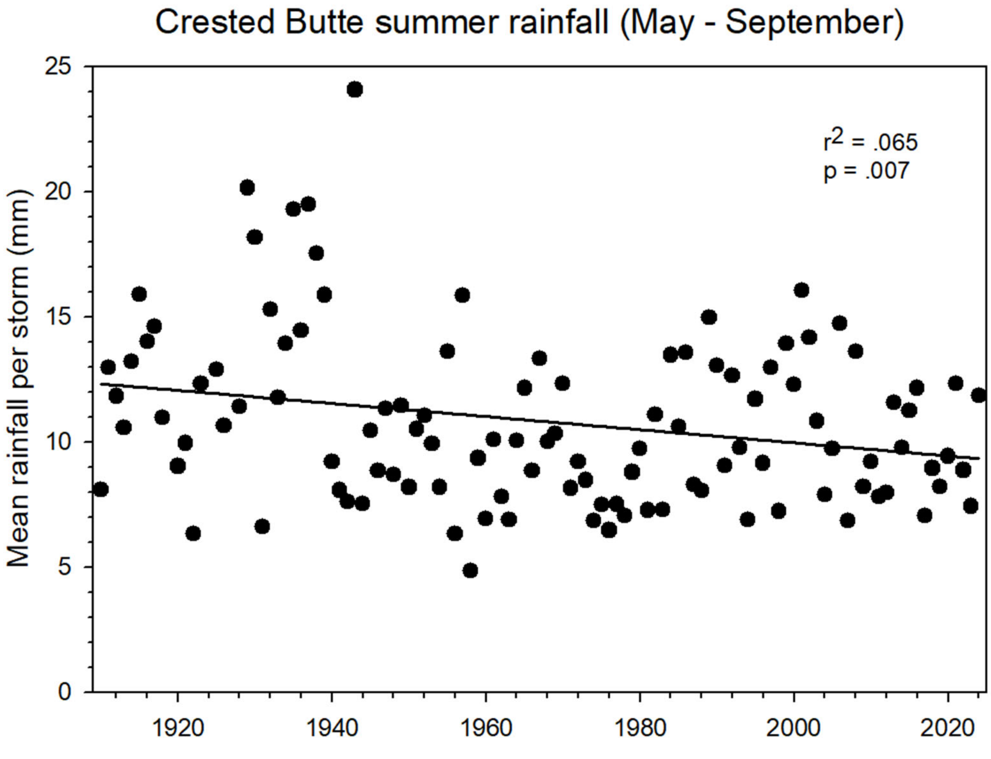


Figure SF7. Total precipitation (cm) per summer (May – September) storm in Crested Butte, 2000 – 2021. Data plotted by first day of the storm. Different colors represent different years. Data from the Crested Butte weather station.

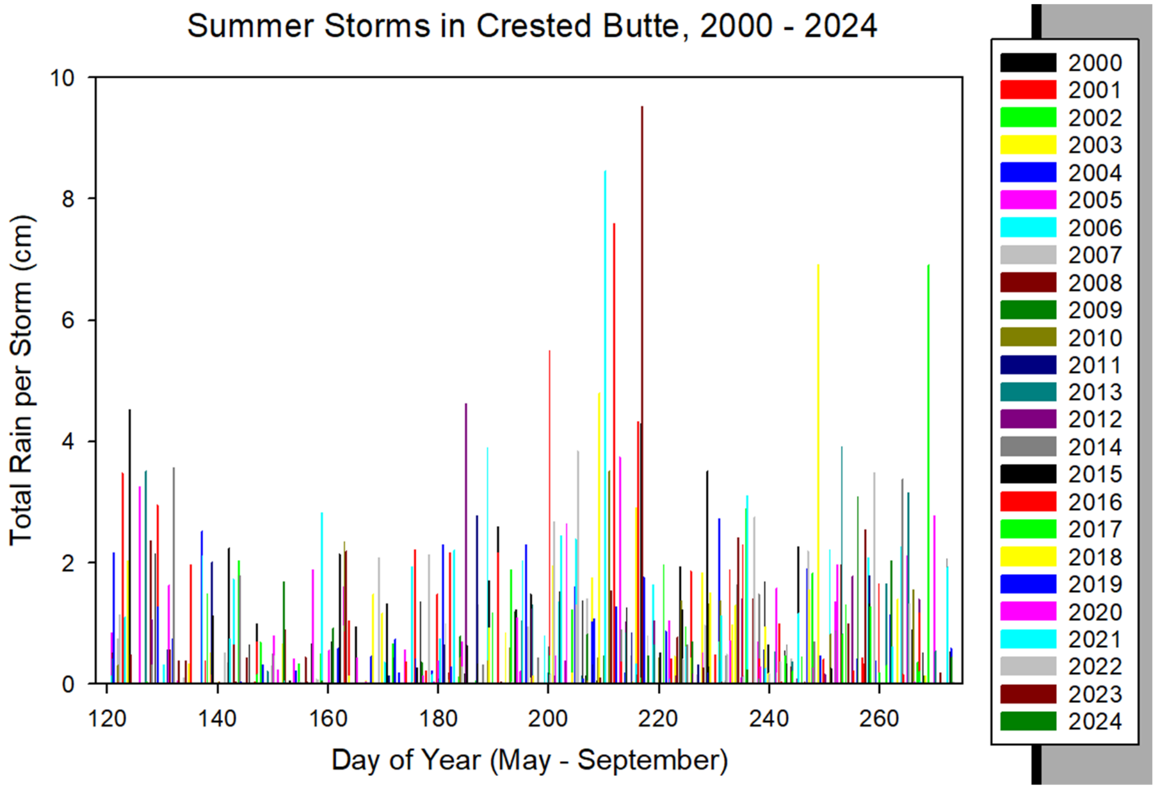


Figure SF8. Mean number of consecutive days with precipitation summer (1910 – 2023) in Crested Butte. There is a significant (p = .017) curvilinear fit, indicating a decline in the last few decades. Data from the Crested Butte weather station.

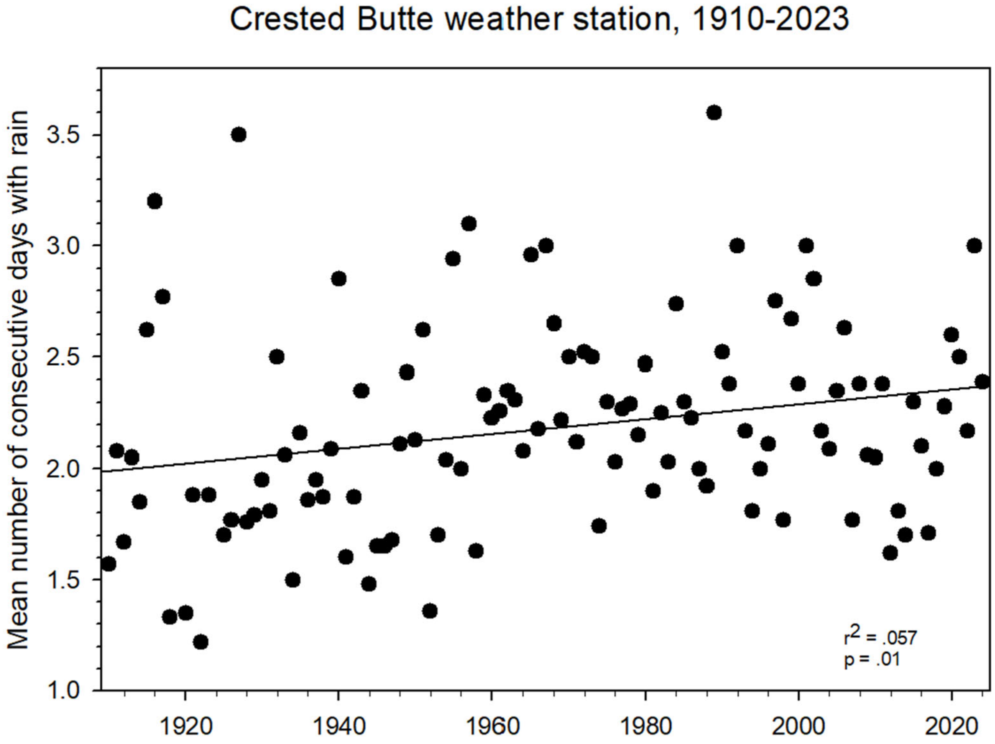


Figure SF9. The maximum number of consecutive days with precipitation during the summer (May – September), 1910 – 2023 in Crested Butte. The slight increasing trend is not significant (p = .135). Data from the Crested Butte weather station.

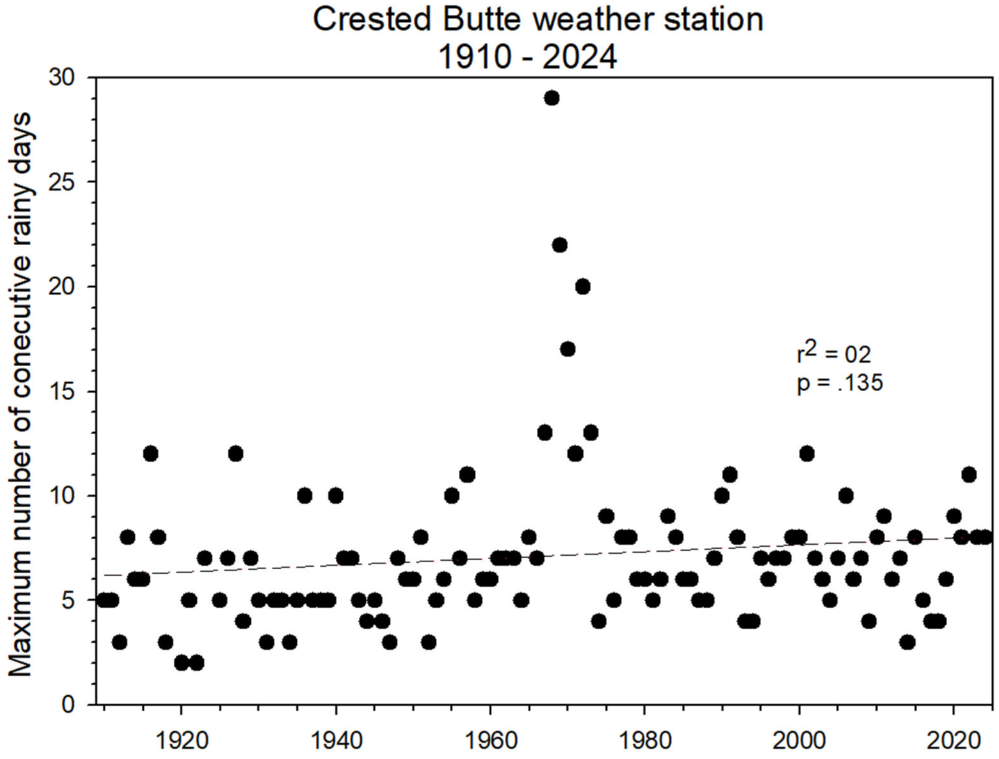


Figure SF10. The percentage of summer (May – September) days with precipitation in Crested Butte, 1910 – 2024. There is an overall significant (p = .0001) linear trend to more days with precipitation, but a quadratic fit suggesting a declining trend since the 1980s is also significant (p = .0003). Data from the Crested Butte weather station.

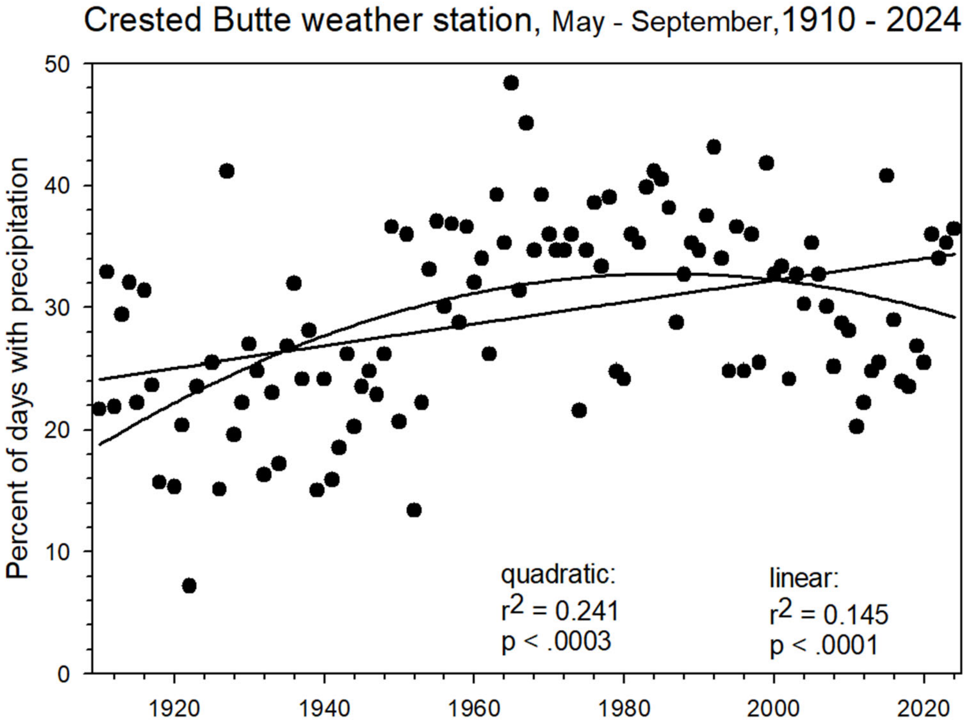


Figure SF11. The maximum number of consecutive summer (May – September) days per year with no precipitation in Crested Butte. There is a significant (p = .002) curvilinear fit, indicating an increasing trend over the past four decades. Data from the Crested Butte weather station, 1910 – 1924.

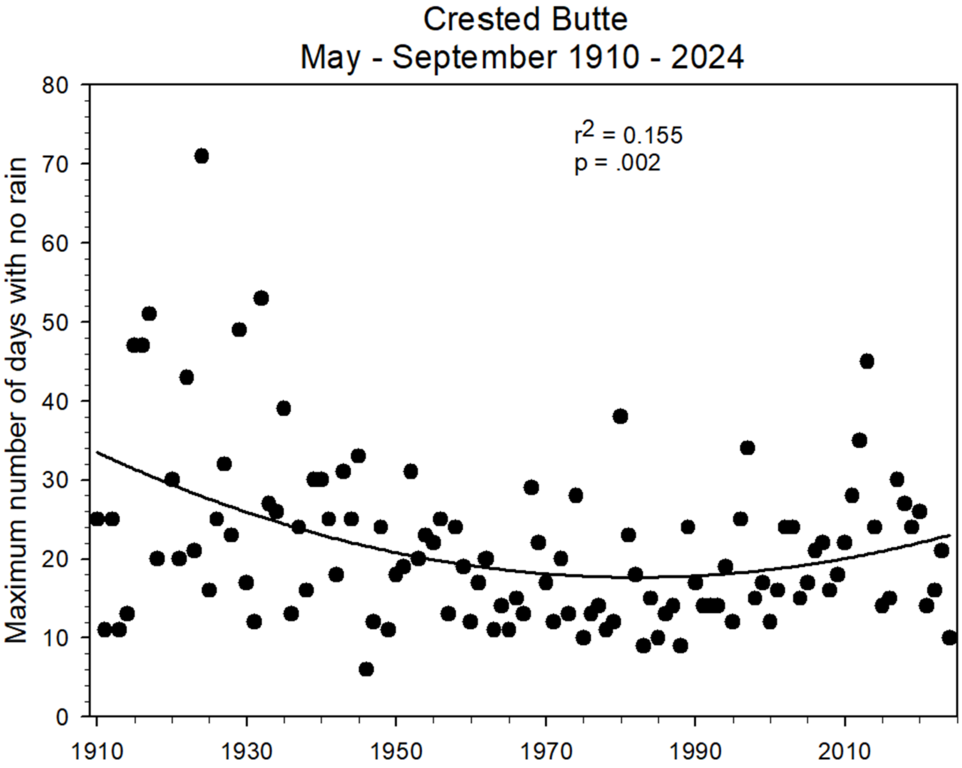


Figure SF12. The total number of summer (May – September) days per year with no precipitation in Crested Butte. There is a significant (p = .0004) curvilinear fit, indicating an increasing trend over the past four decades. Data from the Crested Butte weather station, 1910 – 1924.

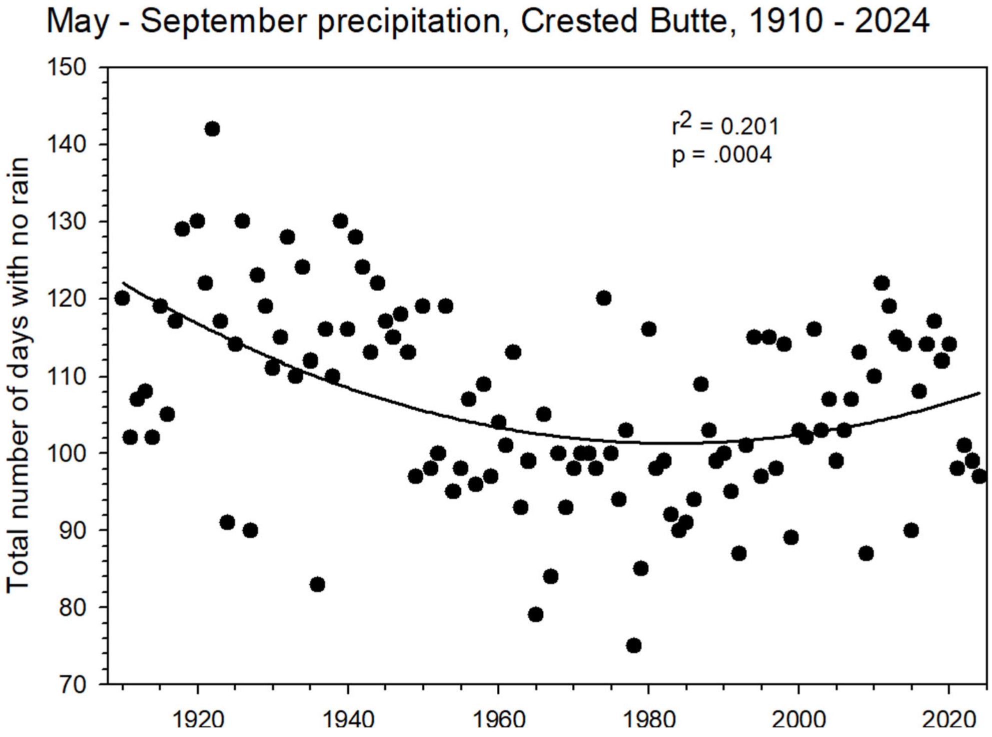


Figure SF13. The mean number of consecutive summer (May – September) days per year with no precipitation in Crested Butte. There is a significant (p = .001) curvilinear fit, indicating an increasing trend over the past four decades. Data from the Crested Butte weather station, 1910 – 1924.

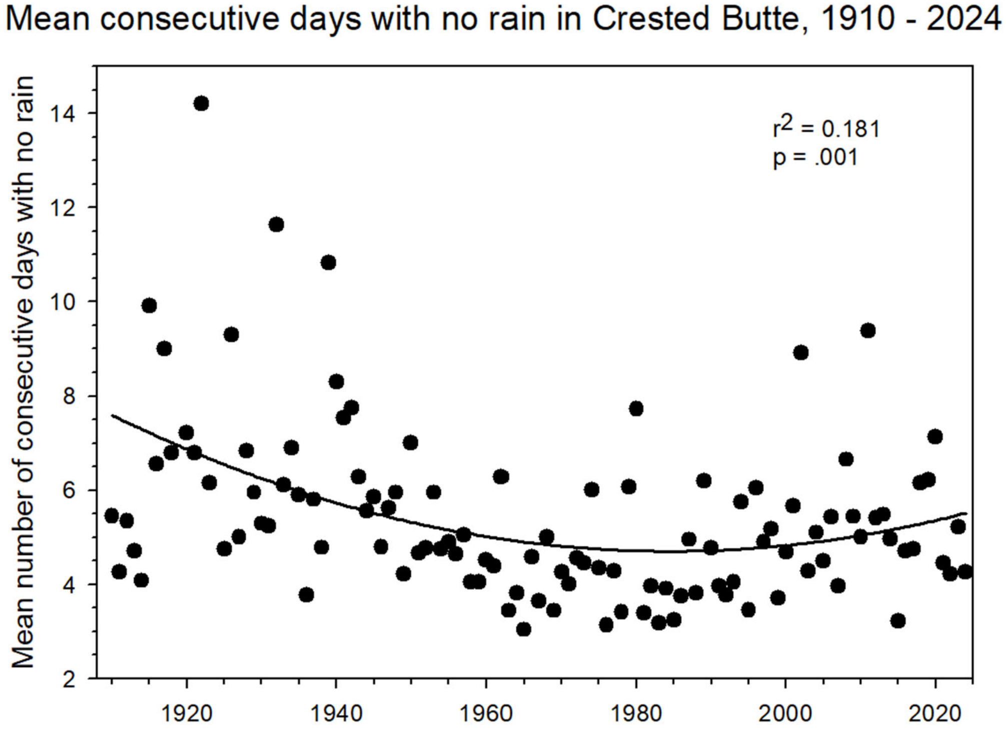


Figure SF14. Variation in amount of rain per rainy day in Crested Butte, as measured by the coefficient of variation, which is increasing significantly (p = .0001). Data from the Crested Butte weather station, 1910 – 1924.

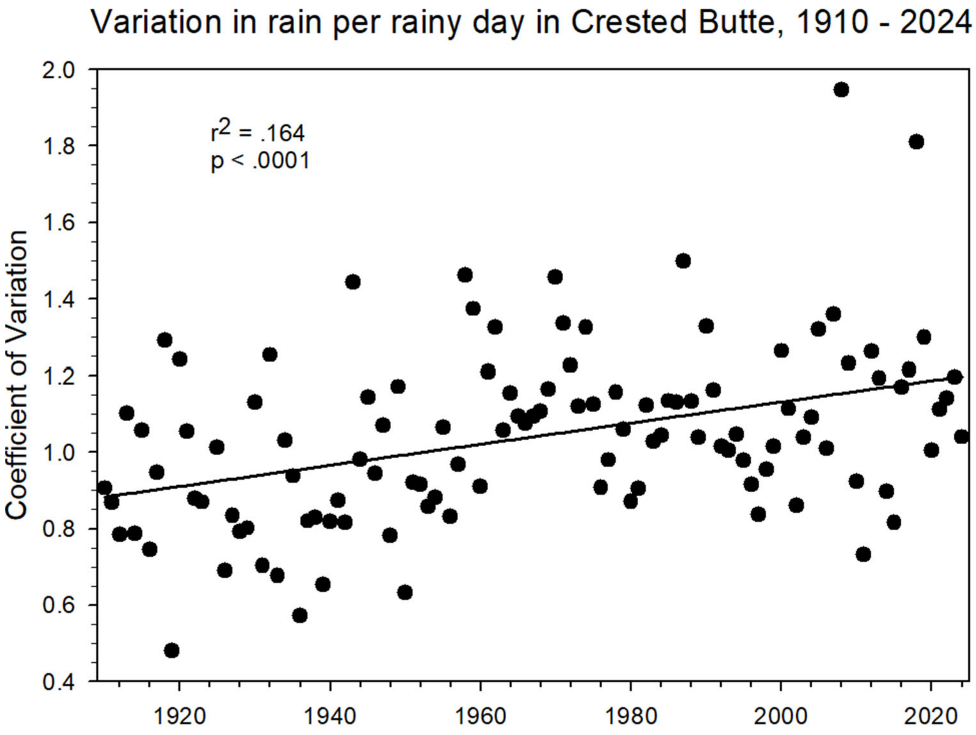


Figure SF15. Variation in amount of rain per rainy day in Crested Butte, as measured by the standard deviation, which is declining significantly (p = .057). Data from the Crested Butte weather station, 1910 – 1924.

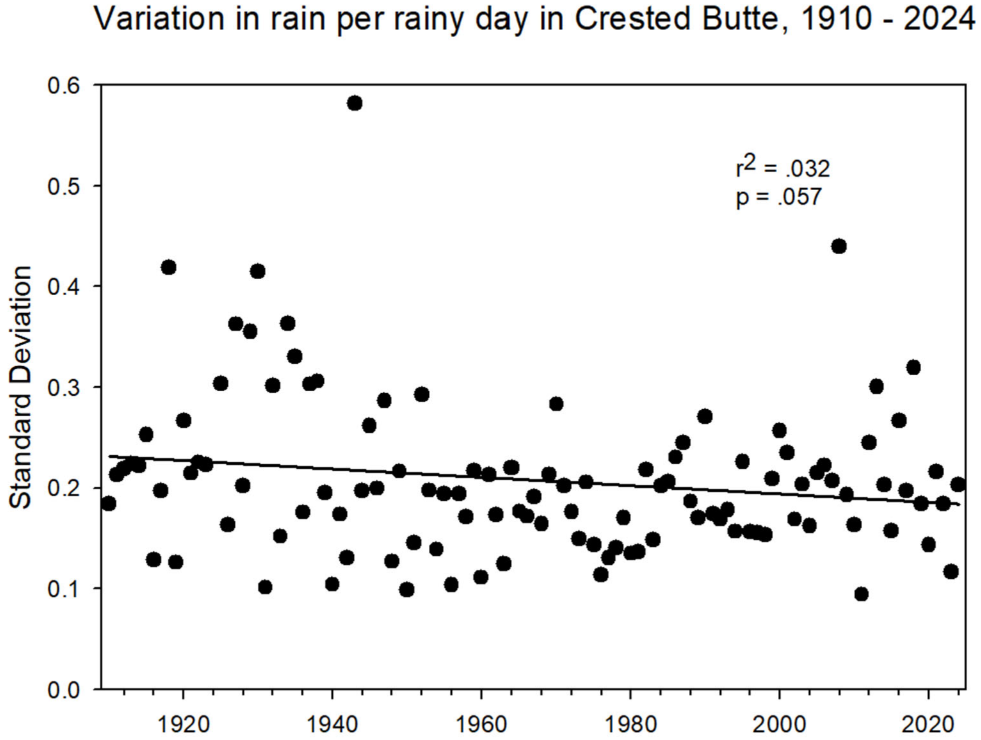


Figure SF16. Variation in the mean daily precipitation in summer (May – September) in Crested Butte, as measured by the standard deviation. There is a marginally significant trend. Data from the Crested Butte weather station, 1910 – 1924.

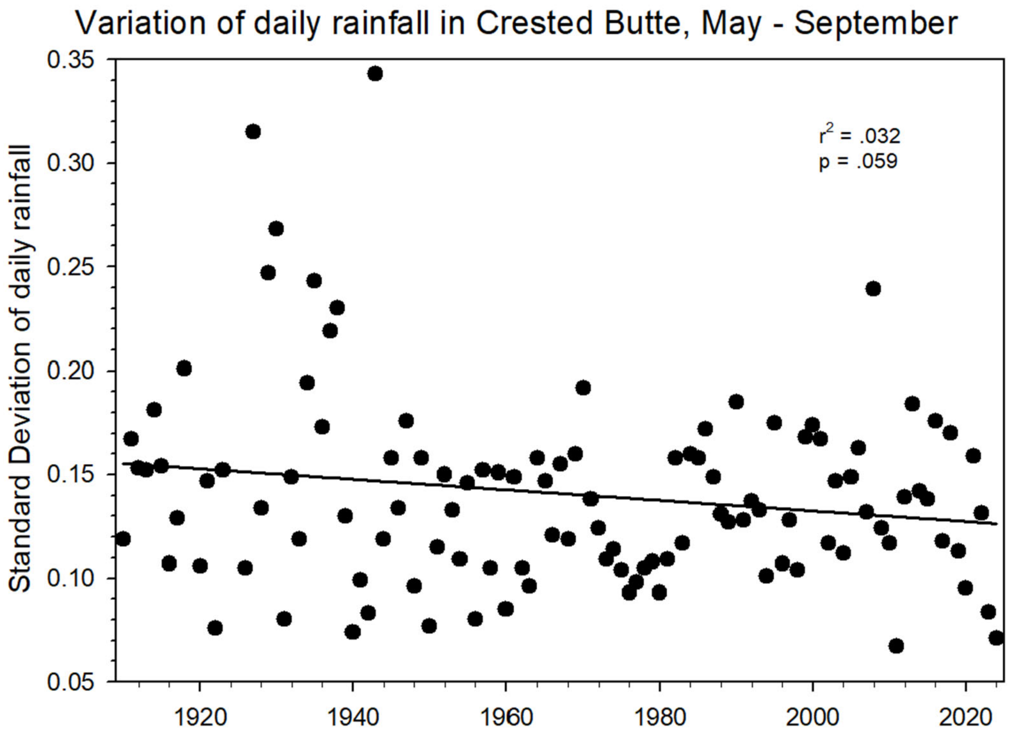


Figure SF17. Variation in the mean daily precipitation in June, as measured by the coefficient of variation, is trending lower (p = .06). Data from the Crested Butte weather station, 1910 – 1924.

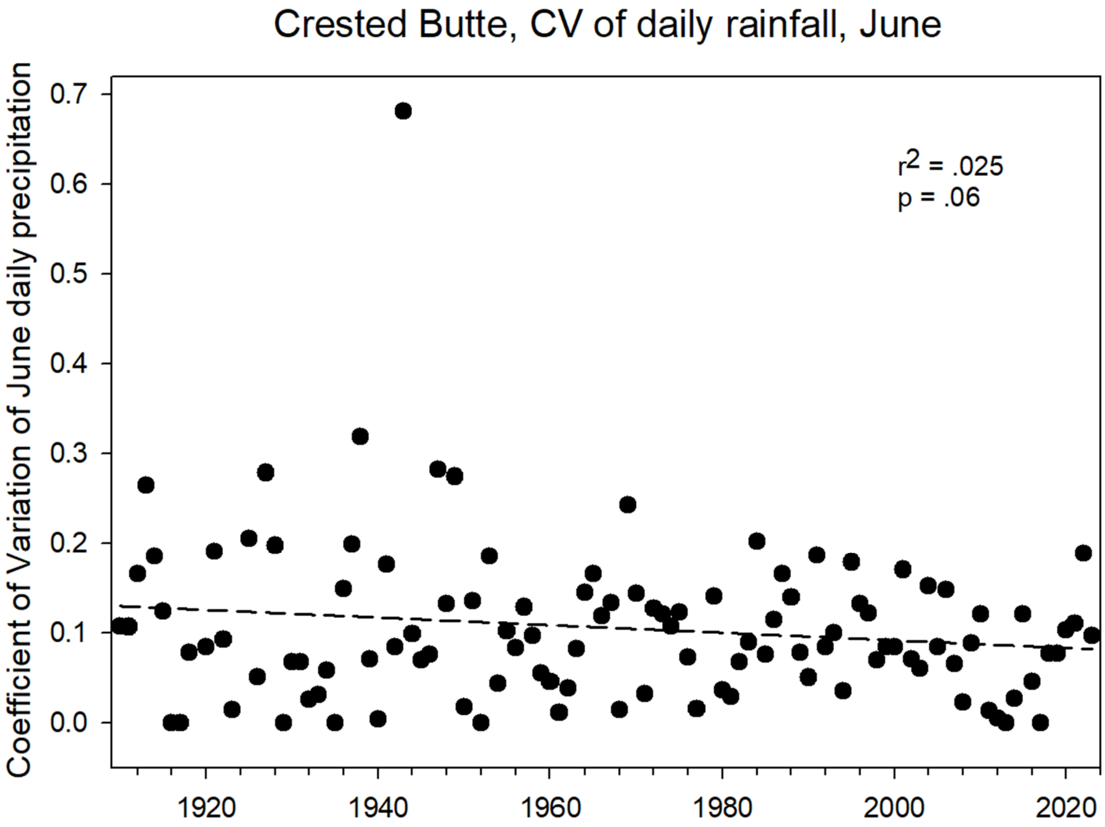


Figure SF18. Variation in the mean daily summer (May – September) precipitation in Crested Butte. There is no significant trend. Data from the Crested Butte weather station, 1910 – 1924.

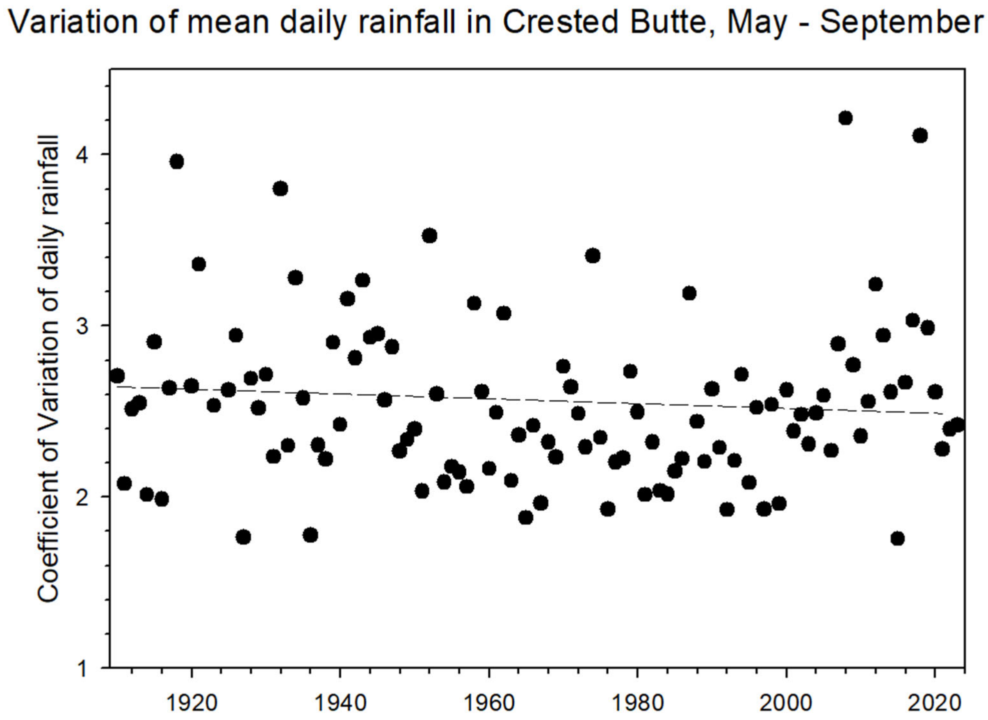


Figure SF19. Variation in the mean amount of precipitation per storm during summer (May – September) in Crested Butte. There is no significant trend. Data from the Crested Butte weather station, 1910 – 1924.

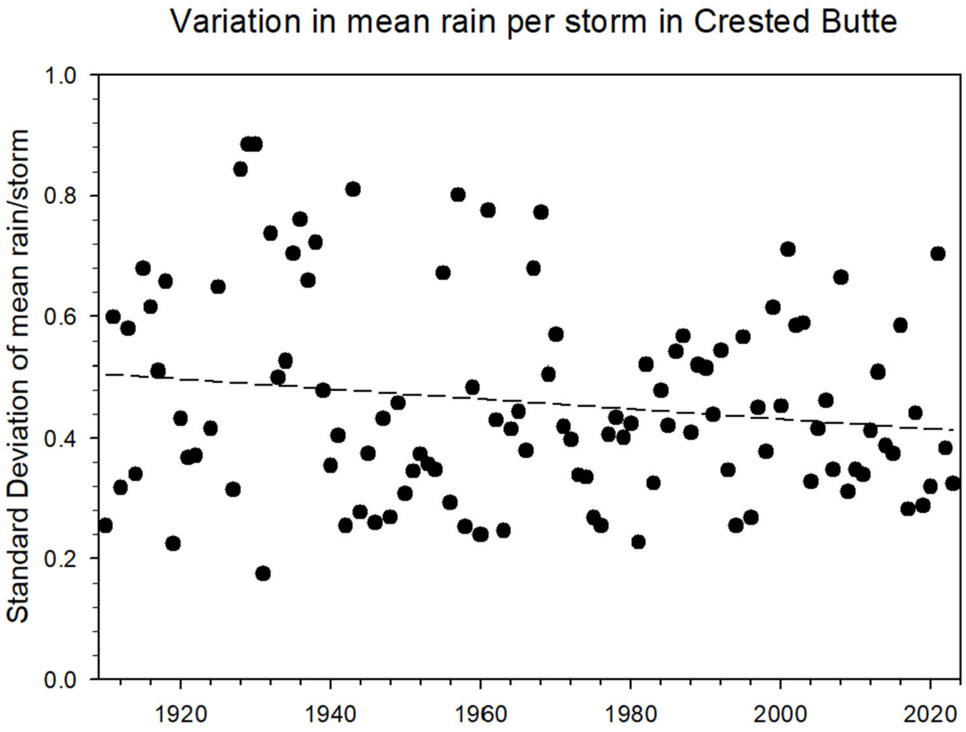


Figure SF20. Standard deviation of the number of consecutive days each May - September with (circles) or without (squares) rain. The decline for days without rain is significant, while the trend for an increase for days with rain is not.

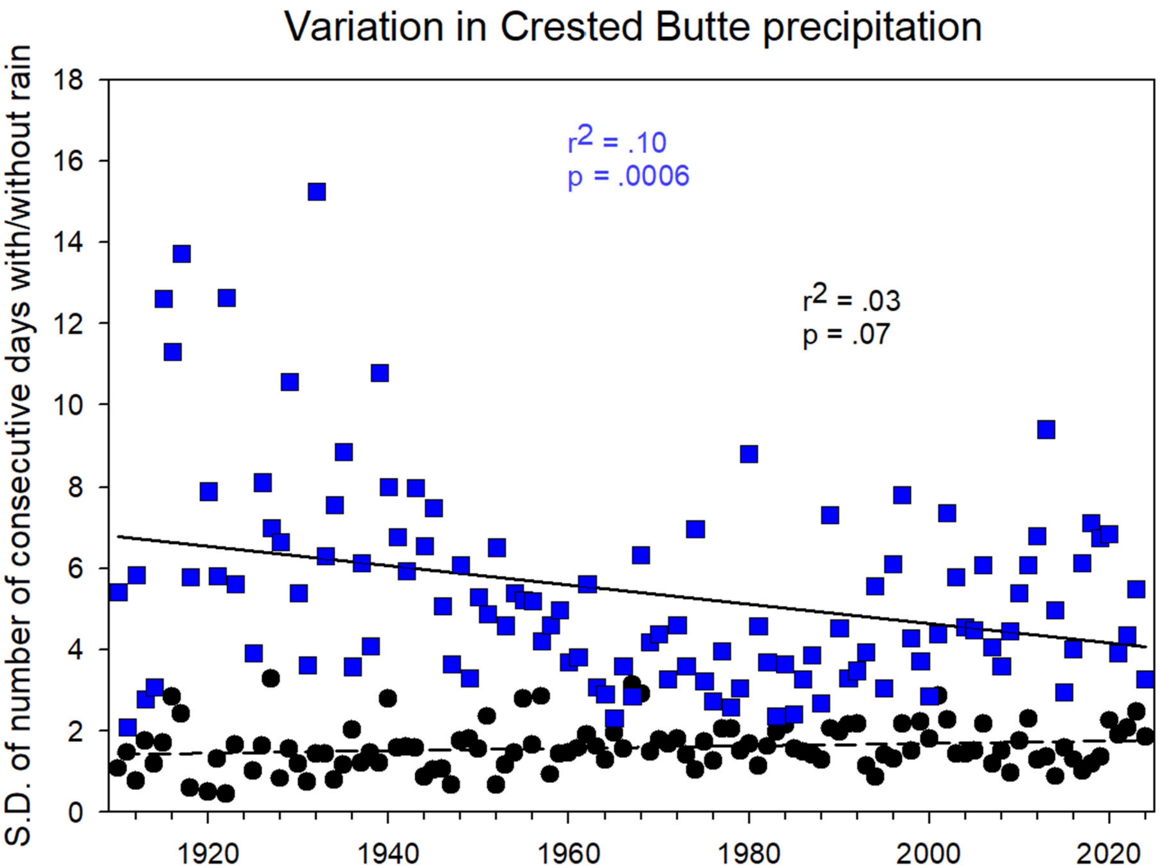


Figure SF21. The number of rainy days from May - September in Gothic has a declining trend but no significant change (range 39 - 89).

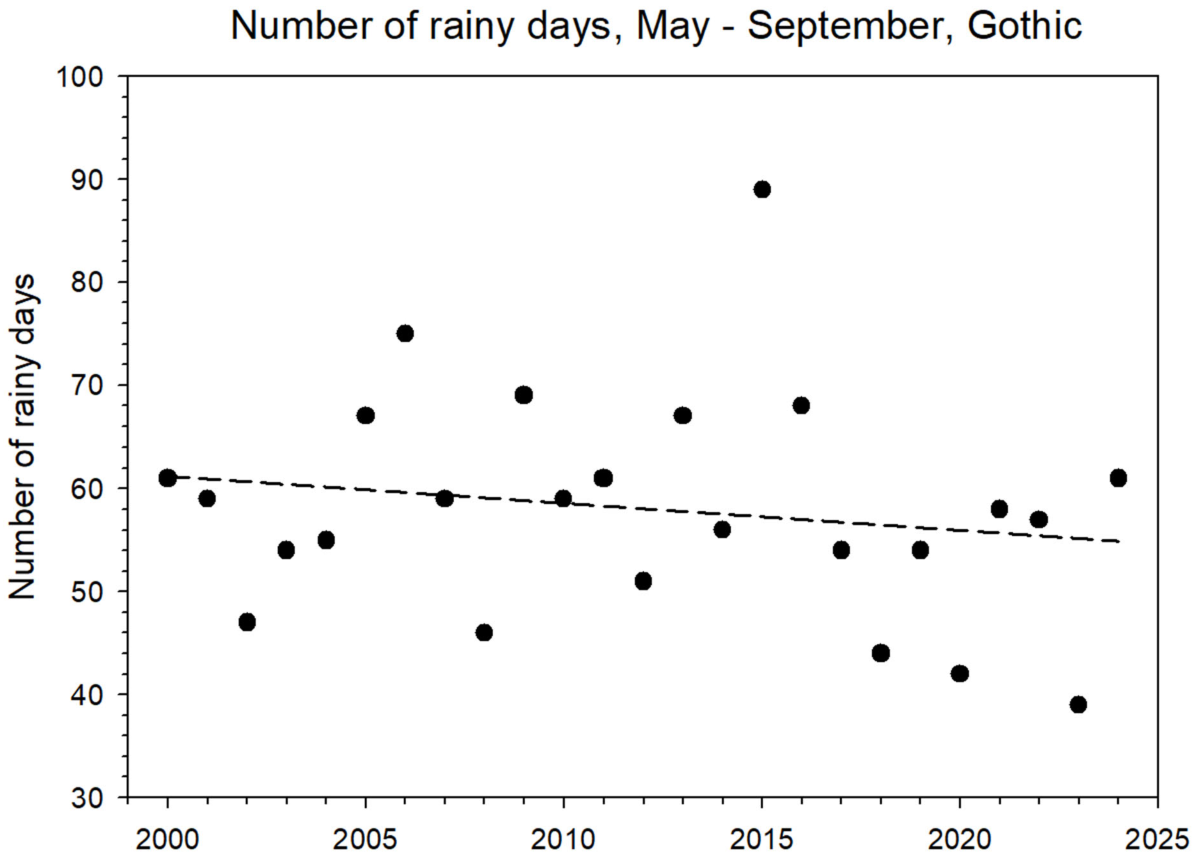


Figure SF22. The mean number of consecutive days with or without precipitation in Gothic, from May – September 2000 – 2024. The increasing (days with precipitation) and declining (days without precipitation) trends are not significant.

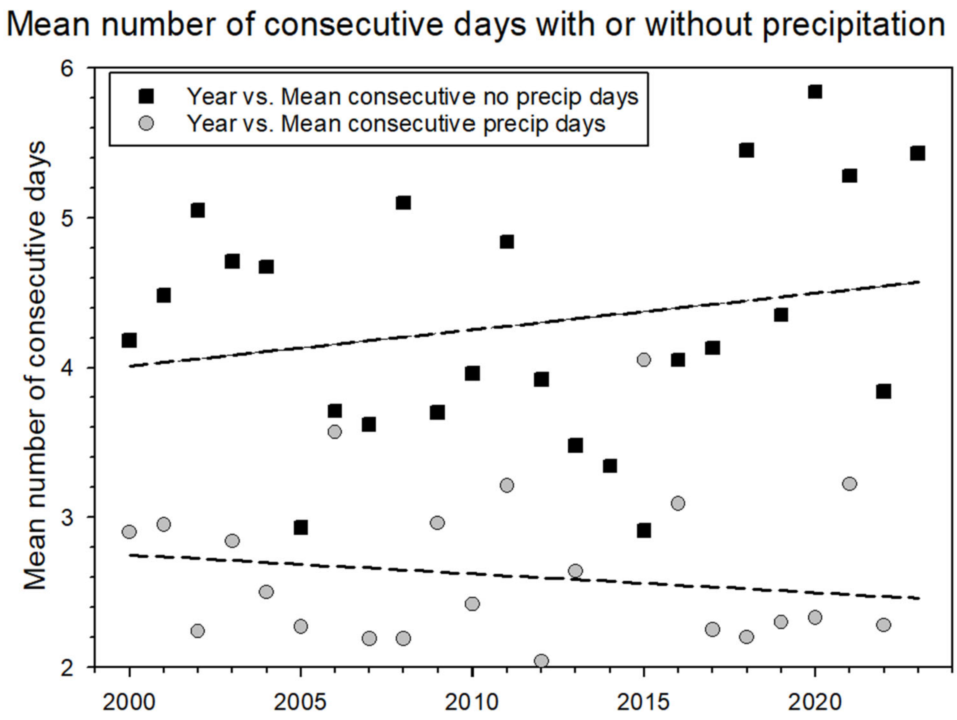


Figure SF23. Maximum consecutive days per summer (May – September) in Gothic with or without precipitation, 2000 – 2024. The declining trend in precipitation days and increasing trend in no-precipitation days are not significant.

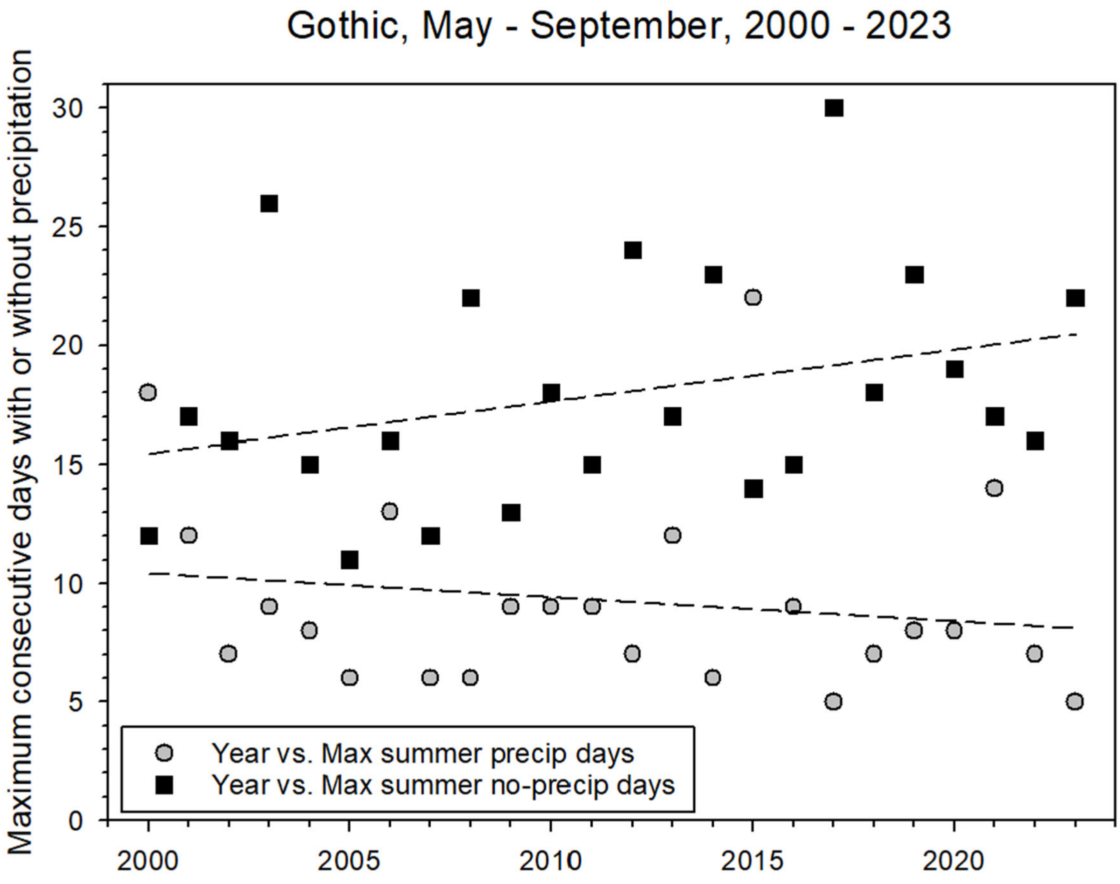


Figure SF24. Mean precipitation (rain + SWE) per summer (May – September) day in Gothic, 2000 – 2024. The declining trend is not significant.

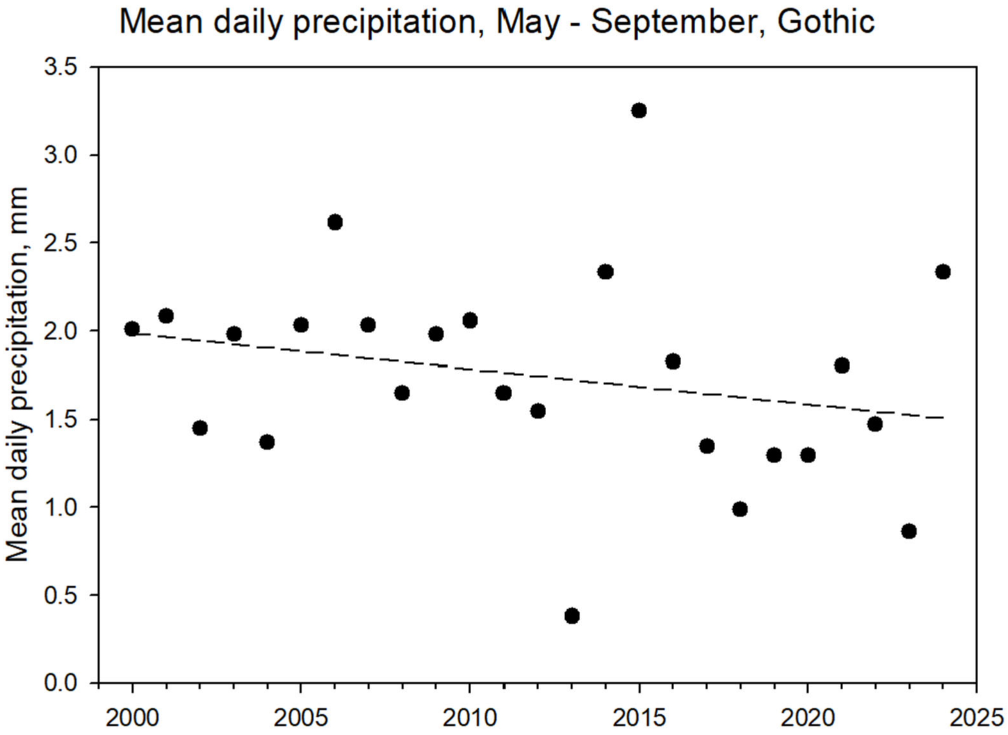


Figure SF25. Mean rain per rainy day (including SWE) from May - September in Gothic. The trend toward lower means is not significant (p = .18).

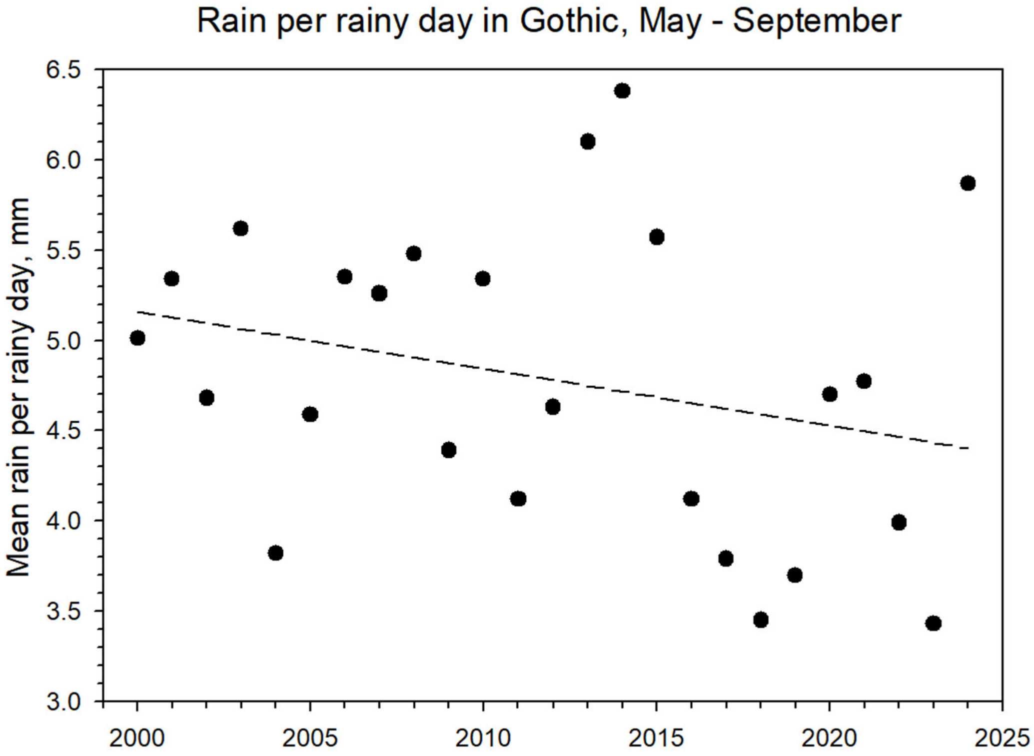


Figure SF26. Maximum cm of precipitation (including SWE) per day during the summer in Gothic. There is no significant trend.

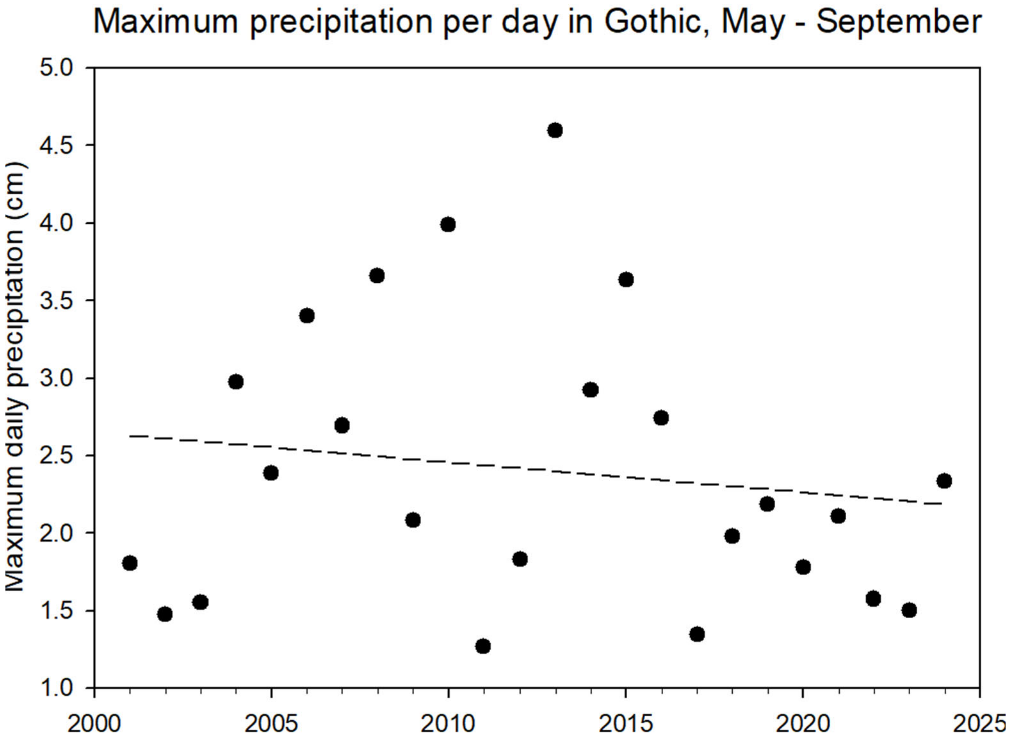


Figure SF27. Maximum amount of precipitation per storm from May - September in Gothic. The trend is not significant.

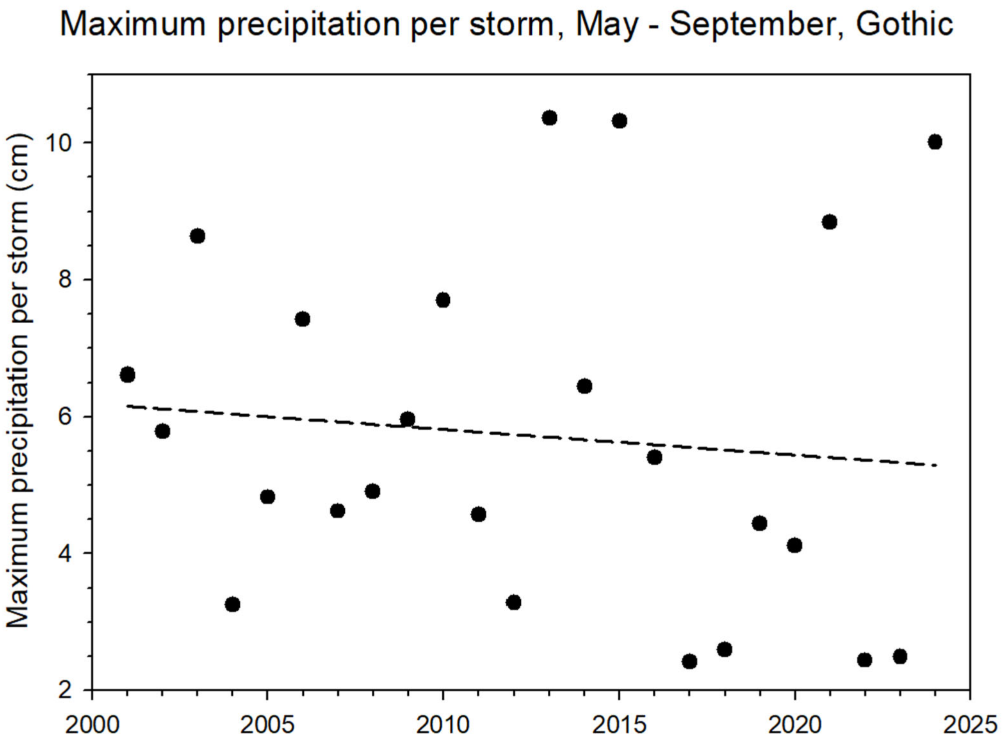


Figure SF28. Mean monthly rain and snow (measured as Snow Water Equivalent) in Gothic, 1999 - 2024. Summer rain data only started in 2000.

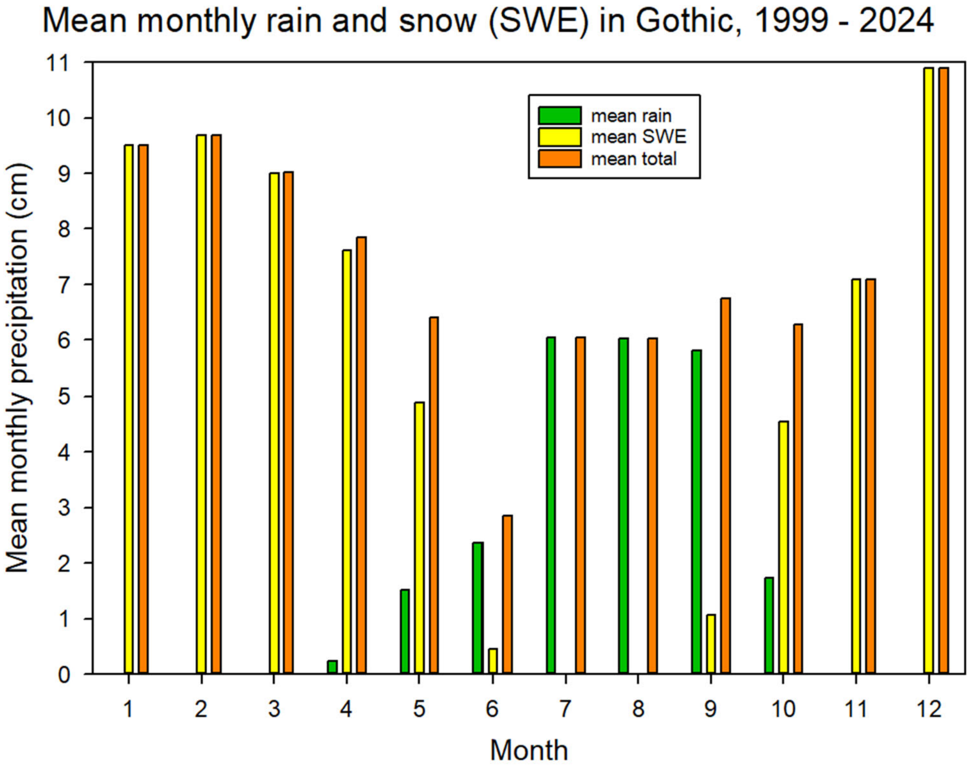


Figure SF29. July and August rainfall in Gothic, 2000 - 2024.

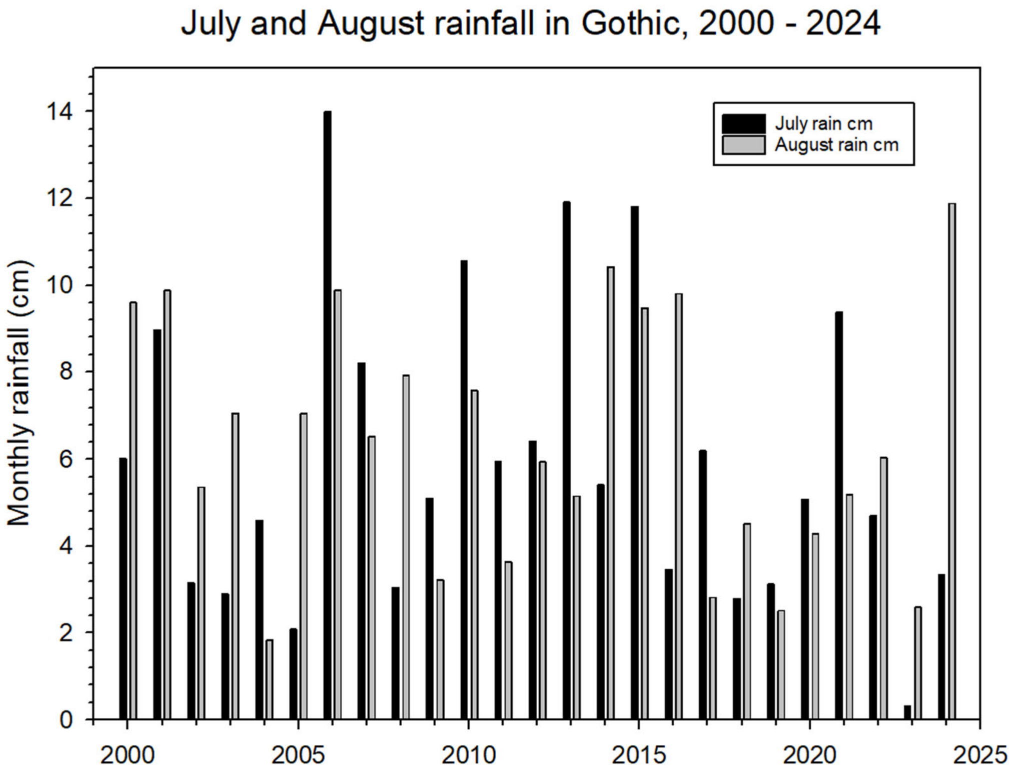


Figure SF30. Precipitation (rain and SWE) in April in Gothic, 2000 – 2024. The declining trend in SWE is not significant.

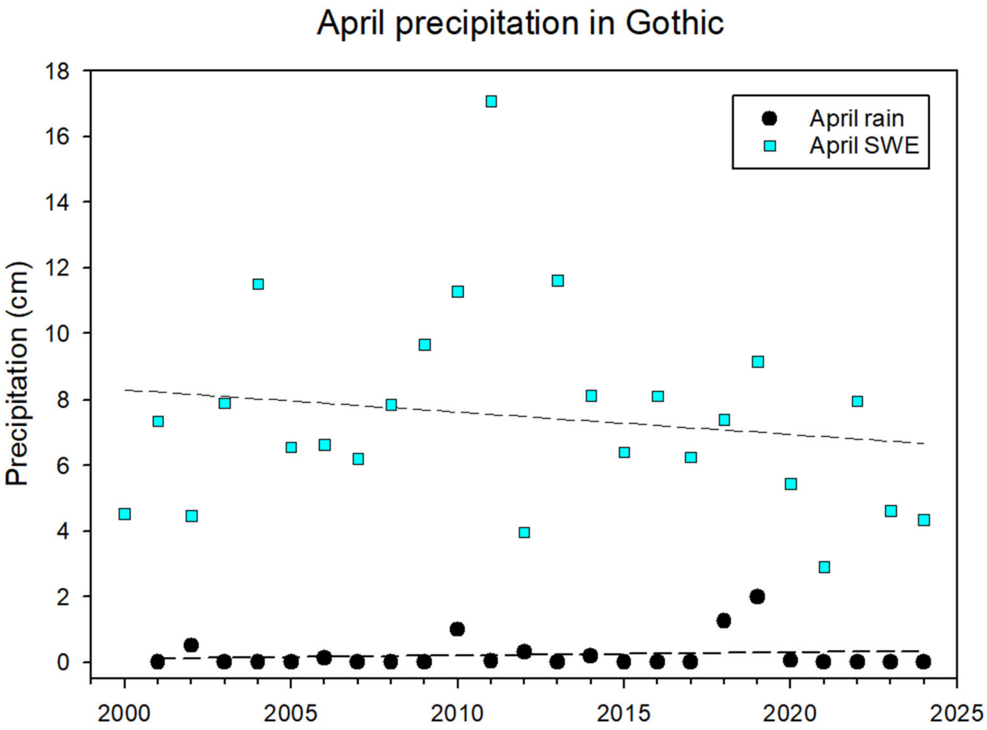


Figure SF31. Precipitation (rain) in September in Gothic, 1999 – 2024. The declining trend is not significant (p = .15).

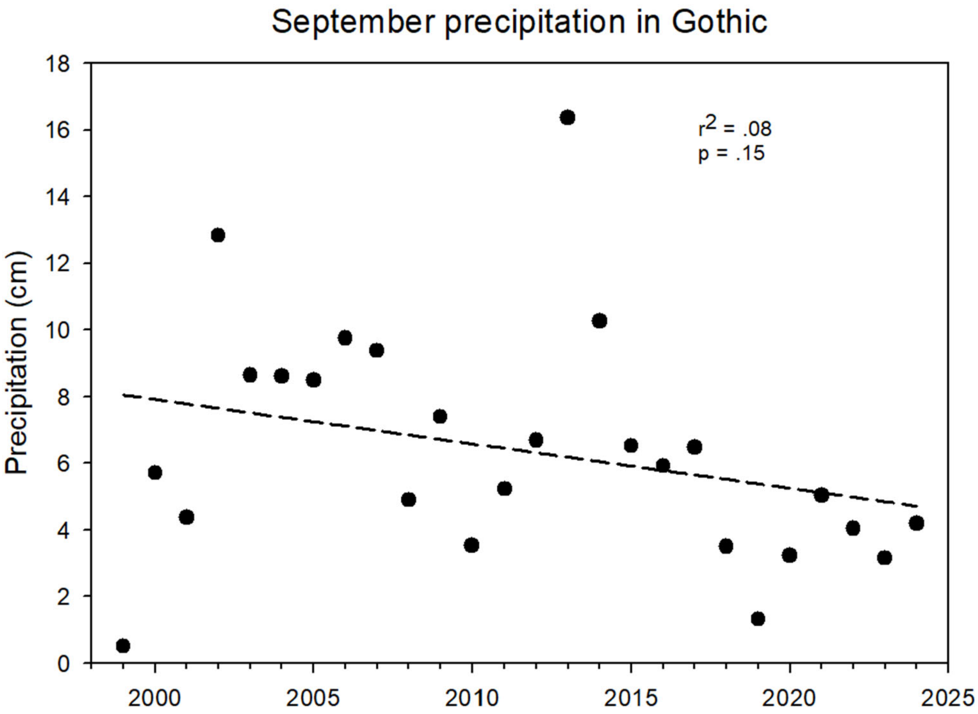


Figure SF32. Standard Deviation of the number of consecutive days with (black circles) or without (pink squares) precipitation in Gothic, by water year (September - August). Neither trend is significant.

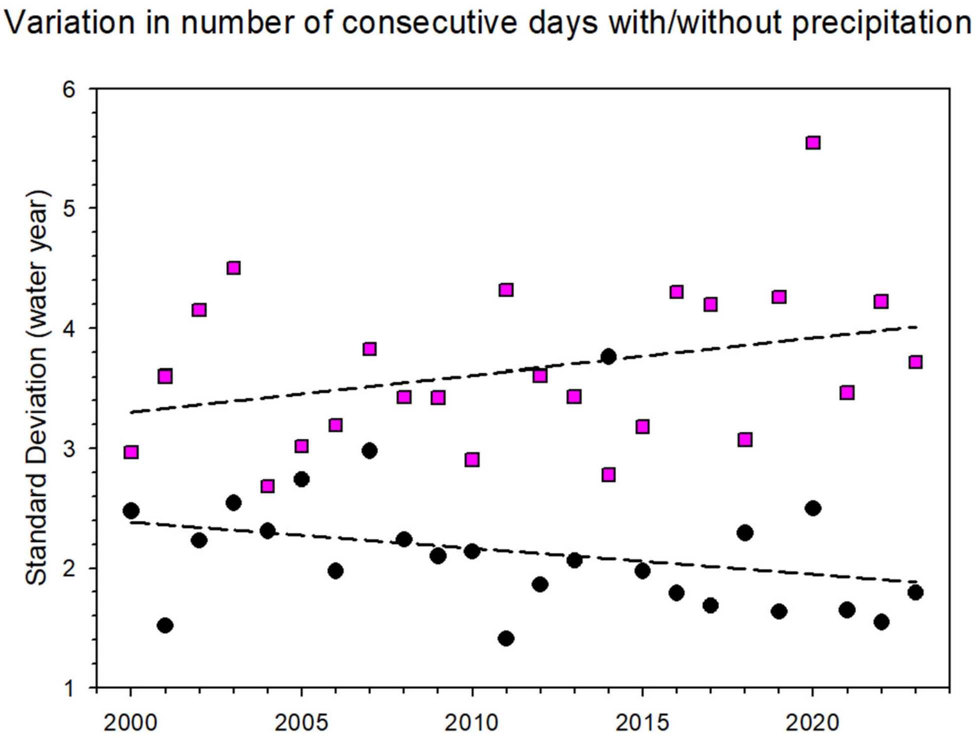


Figure SF33. Variation in the number of consecutive days with (black circles) or without (pink squares) precipitation in Gothic, May – September 2000 – 2024.

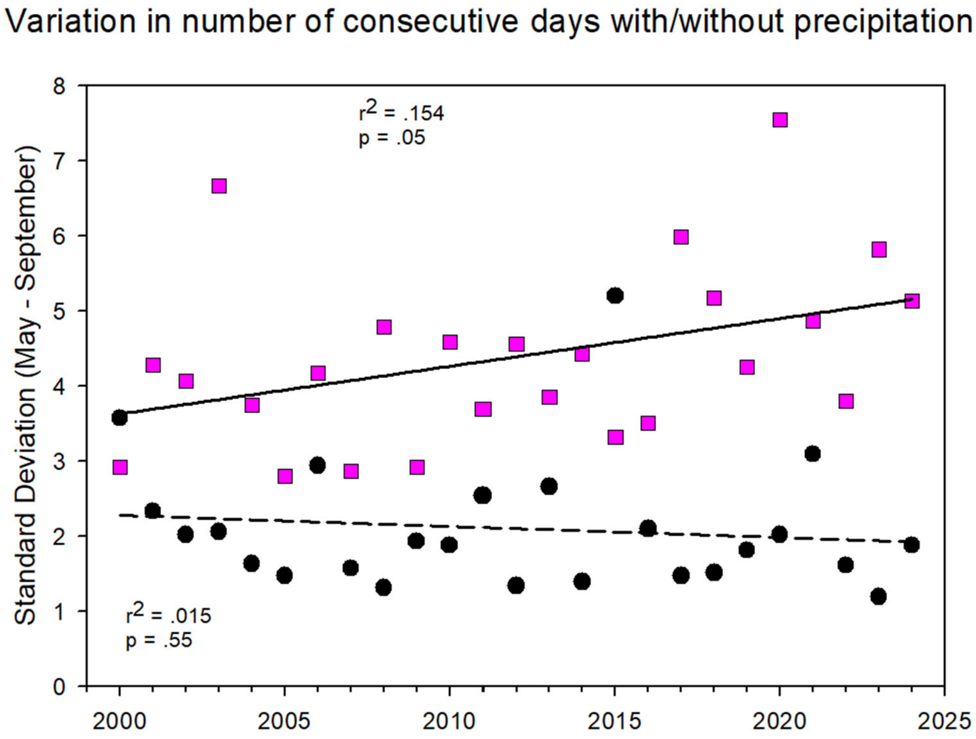


Figure SF34. The ratio of monthly precipitation between Crested Butte and Gothic, for May and September, 2000 - 2024. Neither trend is significant; typically there is more precipitation in Gothic.

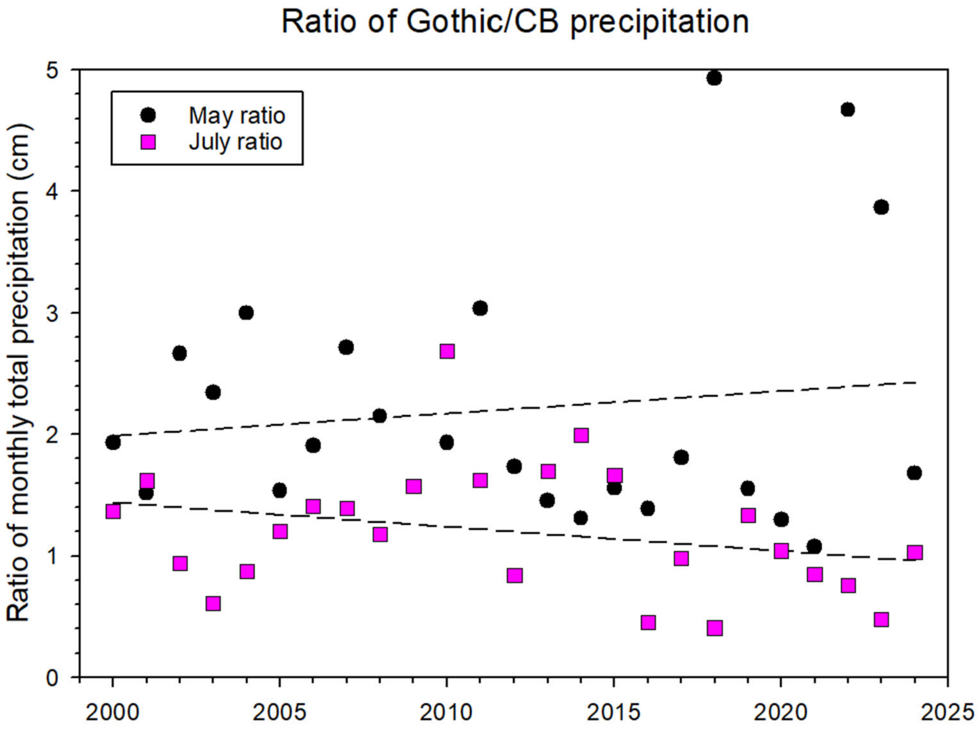


Figure SF35. The ratio of annual precipitation per water-year (September - August) between Gothic and Crested Butte, 2000 - 2023. In all years but 2017 there has been more precipitation at the higher elevation in Gothic.

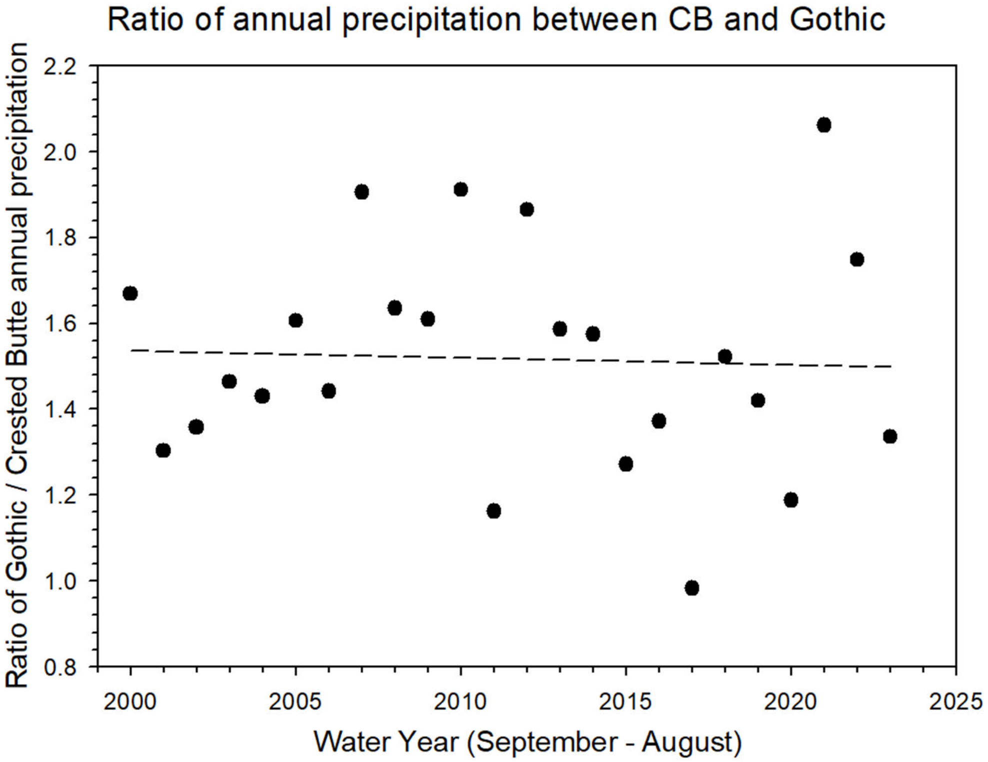


Figure SF36. The mean amount of snow per snowstorm (consecutive days with snow) in Gothic, 1974 - 2024.

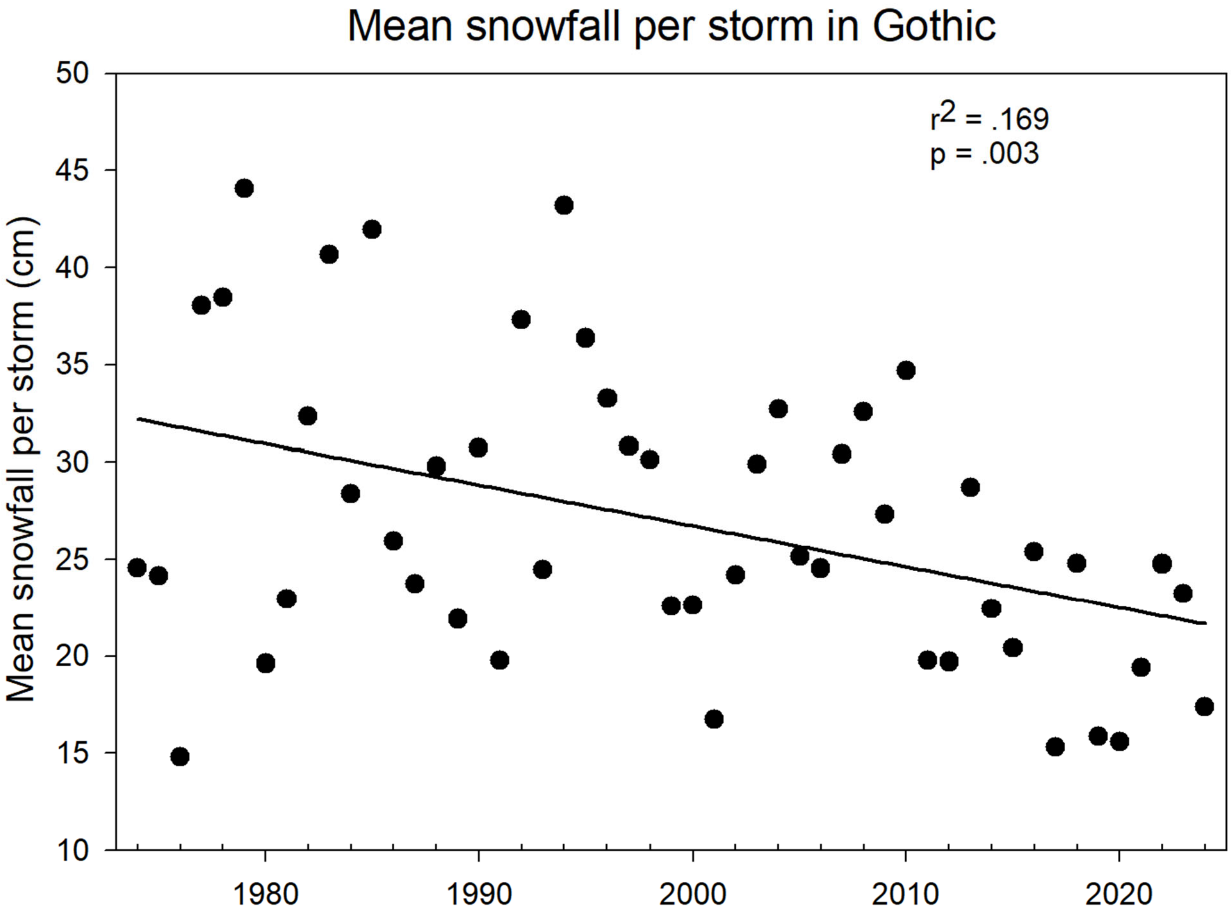


Figure SF37. The temporal pattern of annual snowfall in Gothic, calculated as 3-day totals, and shown for two time periods, 1974 - 1999 and 2000 - 2024.

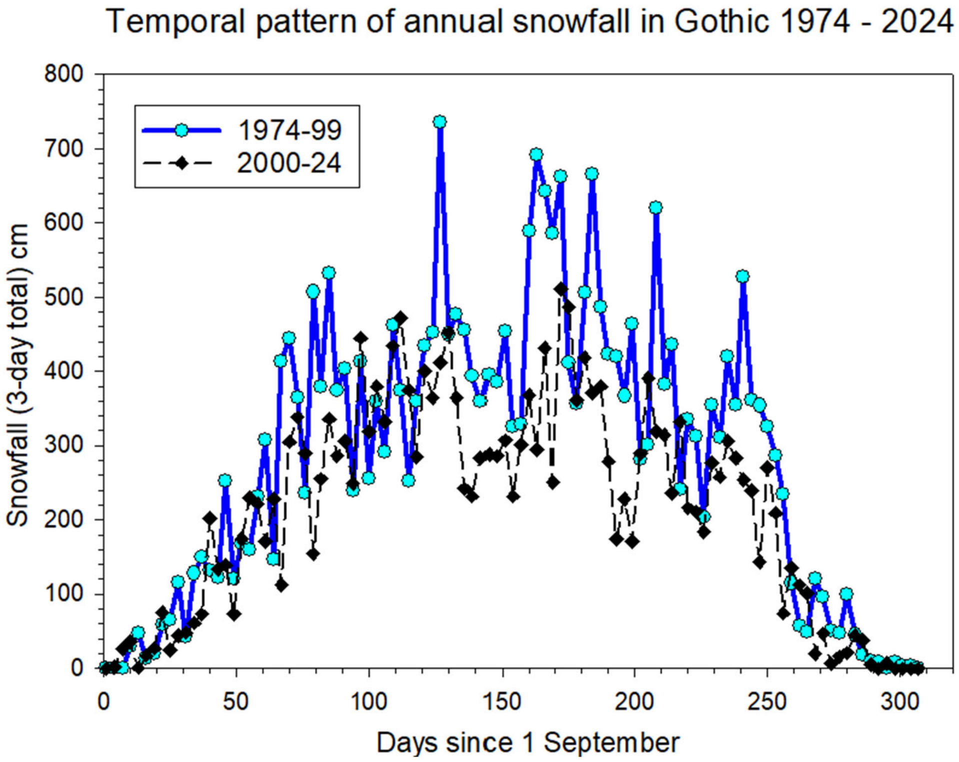


Figure SF38. Monthly mean snow density in Gothic from November - May, 1975-76 to 2024-25. Lines calculated by a quadratic fit to data.

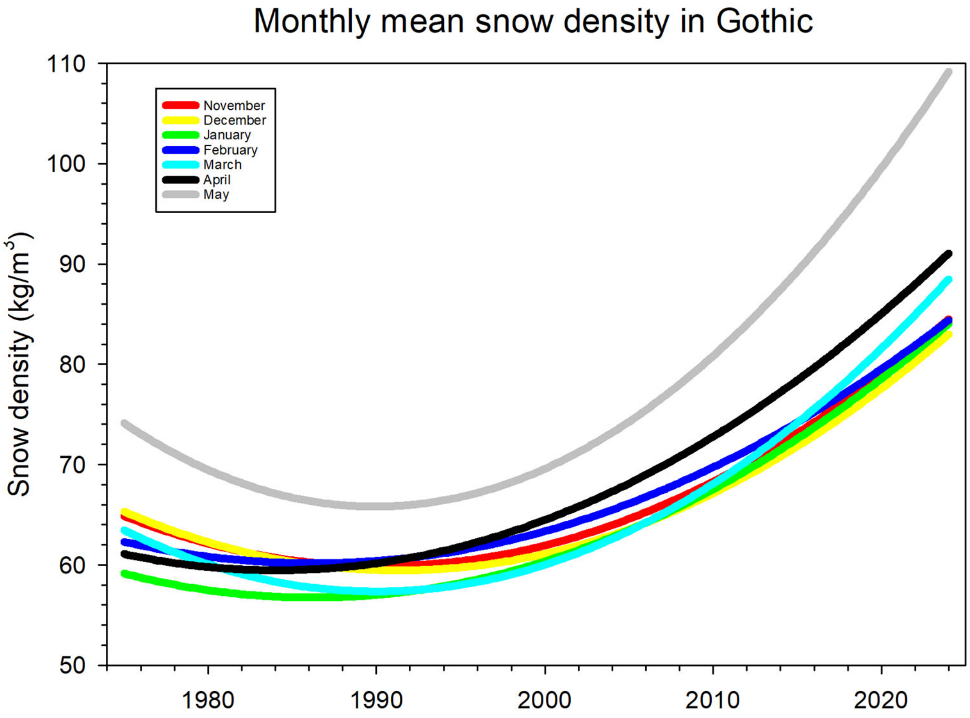


Figure SF39. Monthly mean, maximum, and minimum snowfall density in Gothic in the winter of 1992 - 1993. The fit of a quadratic equation to the mean is significant, but the other lines are not.

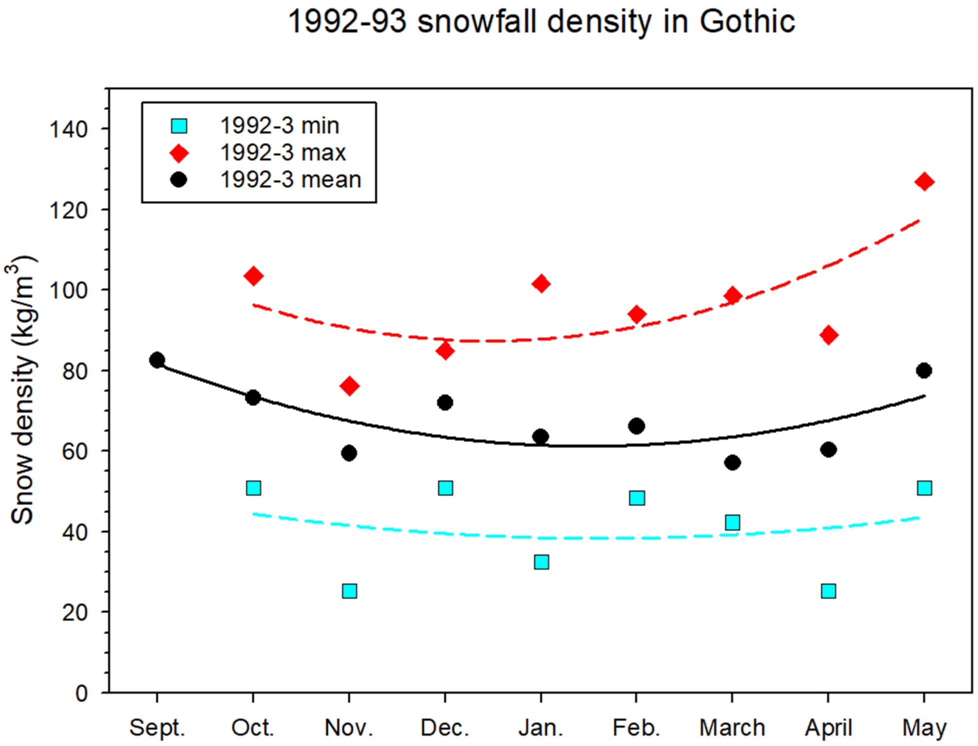


Figure SF40. The relationship between heating degree days (base 18° C) and snow density in Gothic, 1994-95 water year.

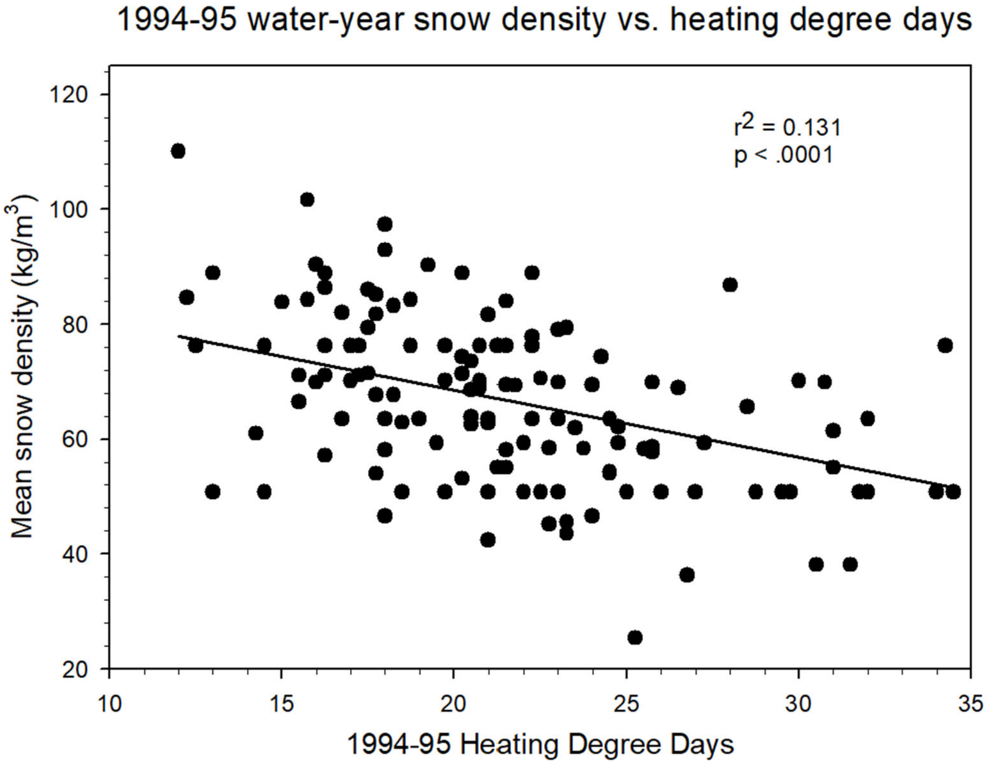


Figure SF41. The relationship between maximum daily temperature and snow density in Gothic.

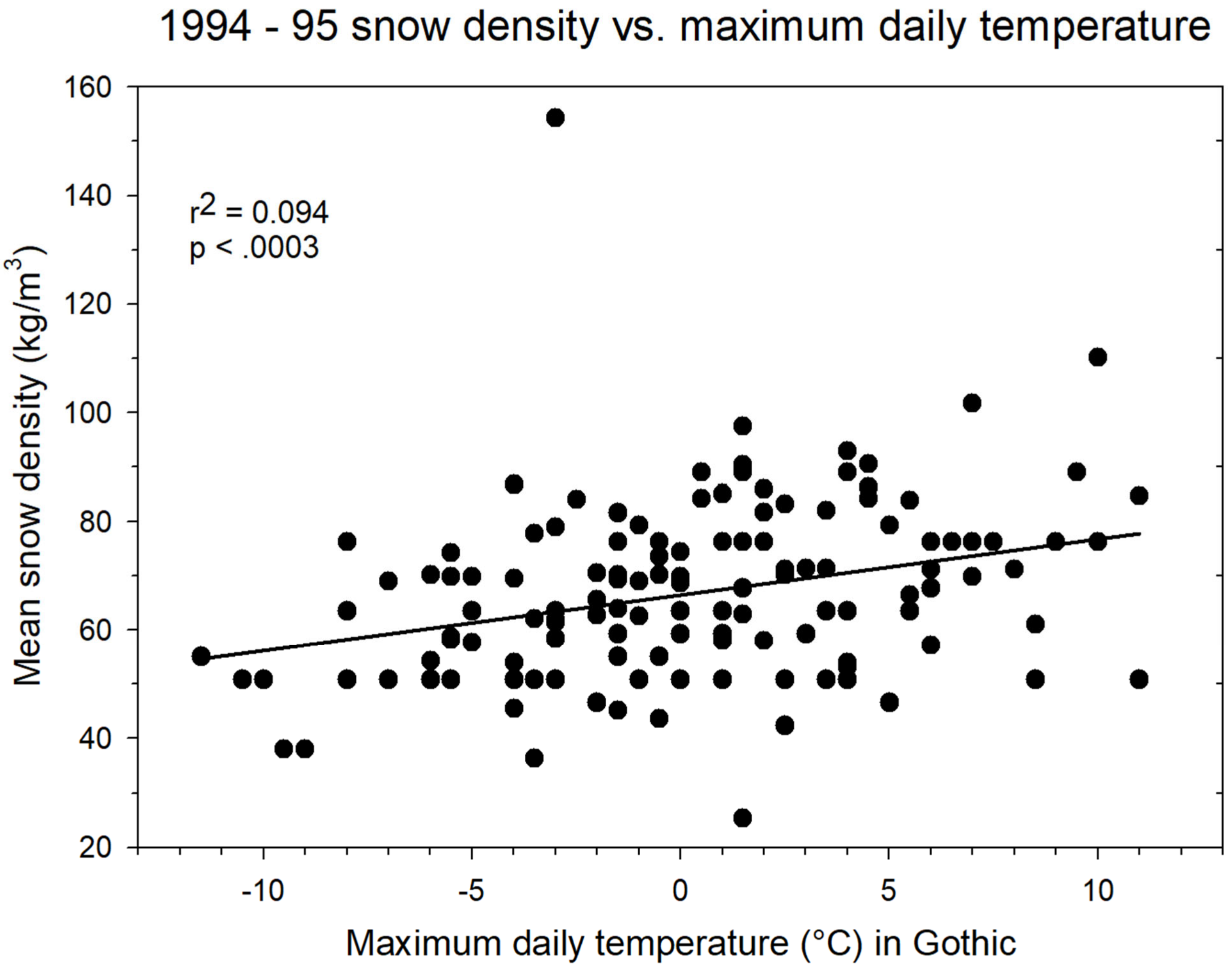
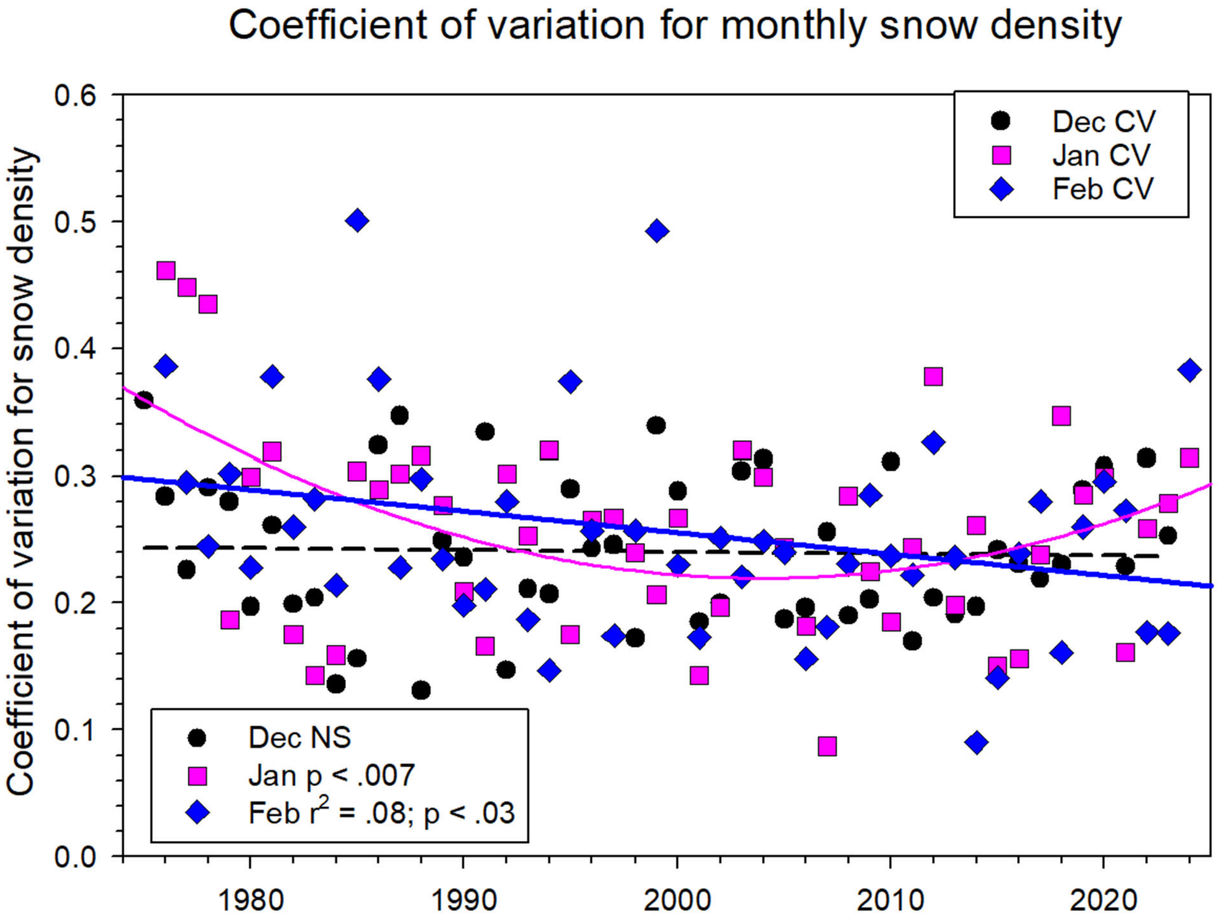


Figure SF42. Coefficient of variation for monthly snow density in Gothic. (a) December, January, and February. (b) March, April, and May.



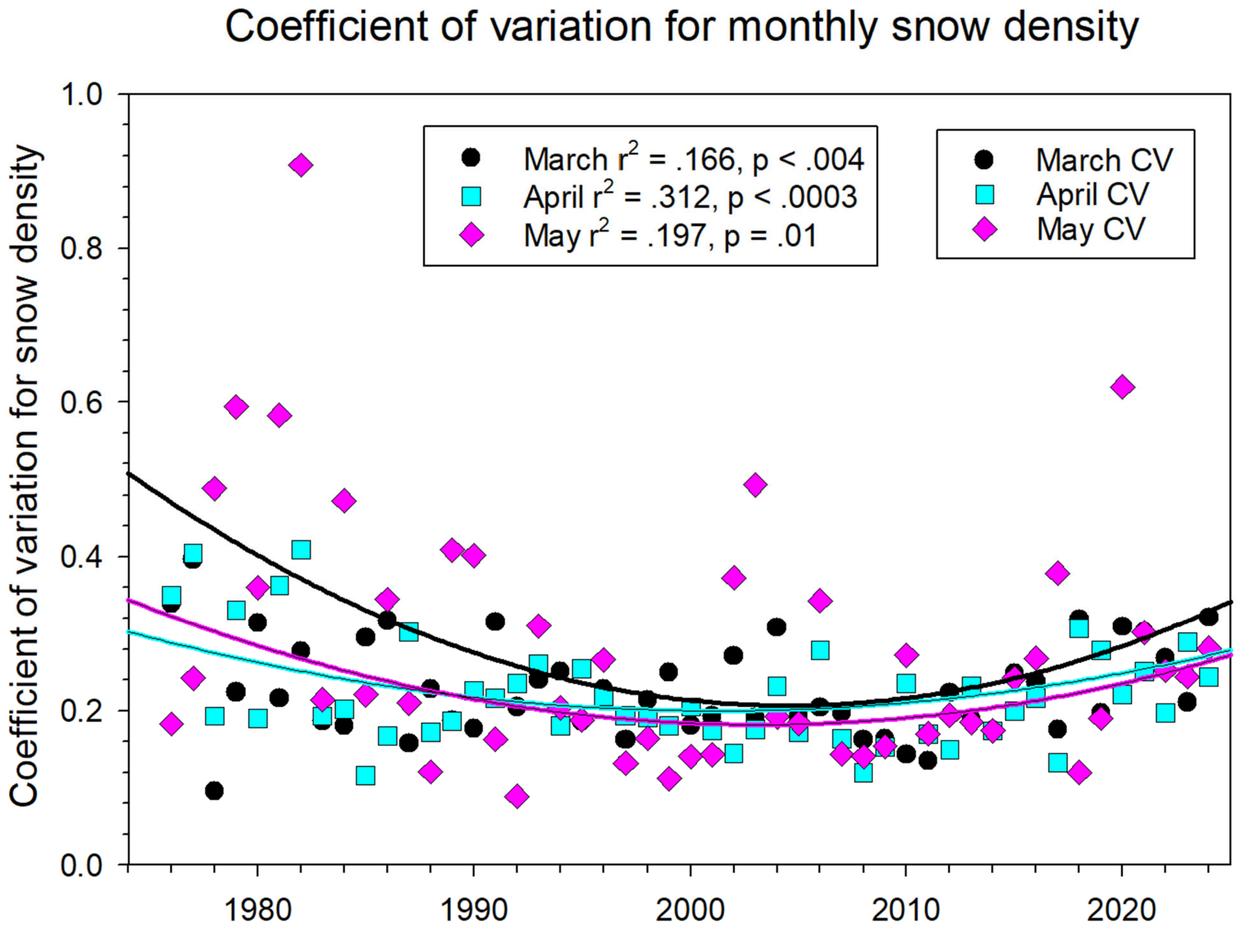

Coefficient of variation for monthly snow density
Coefficient of variation for snow density
1.0
0.8
0.6
0.4
0.2
0.0
1980
1990
2000
2010
2020
March r² = .166, p < .004
April r² = .312, p < .0003
May r² = .197, p = .01
March CV
April CV
May CV

Figure SF43, The day of year of the beginning of the winter's permanent snowpack in Gothic.

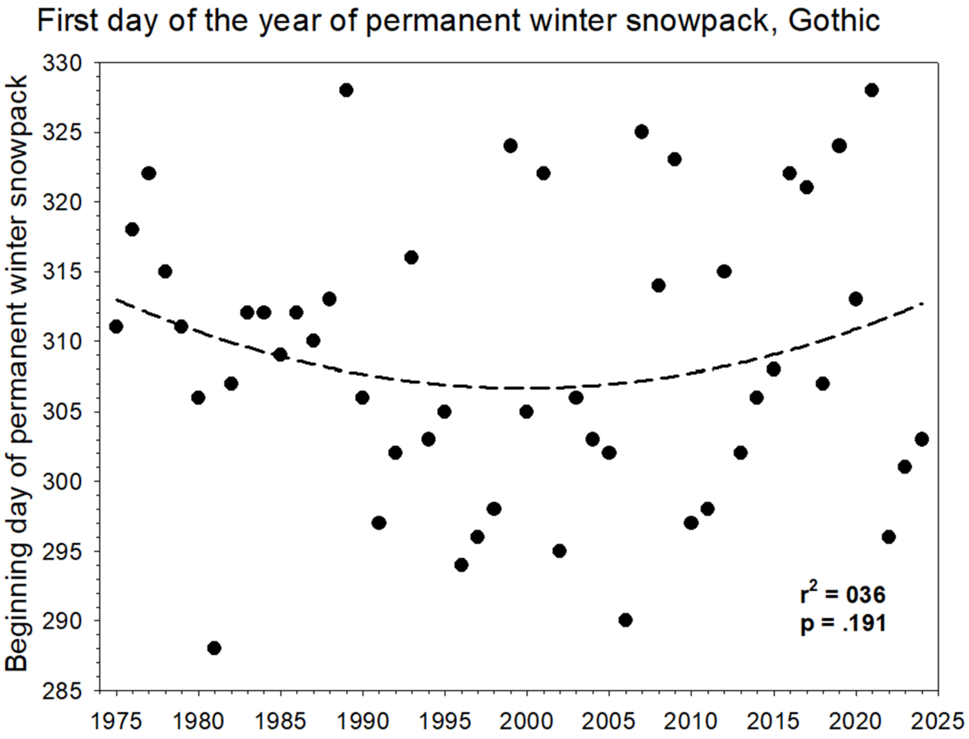


Figure SF44. The first and last dates in Gothic with ≥ 10cm of snow on the ground.

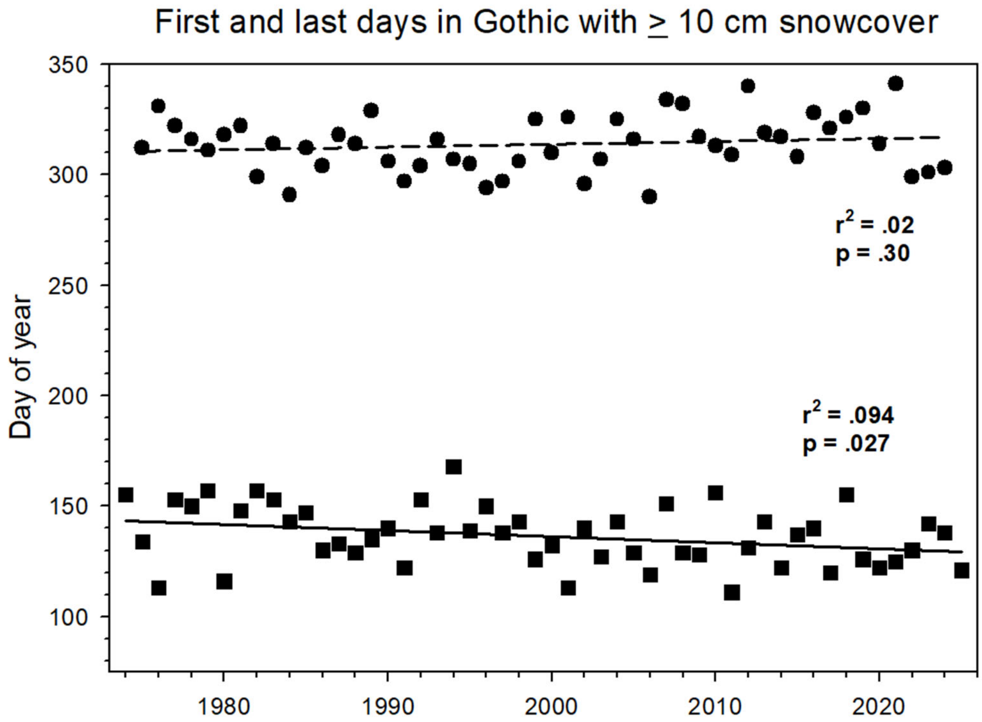


Figure SF45. The number of days of permanent snow cover in Gothic. The trend is not significant.

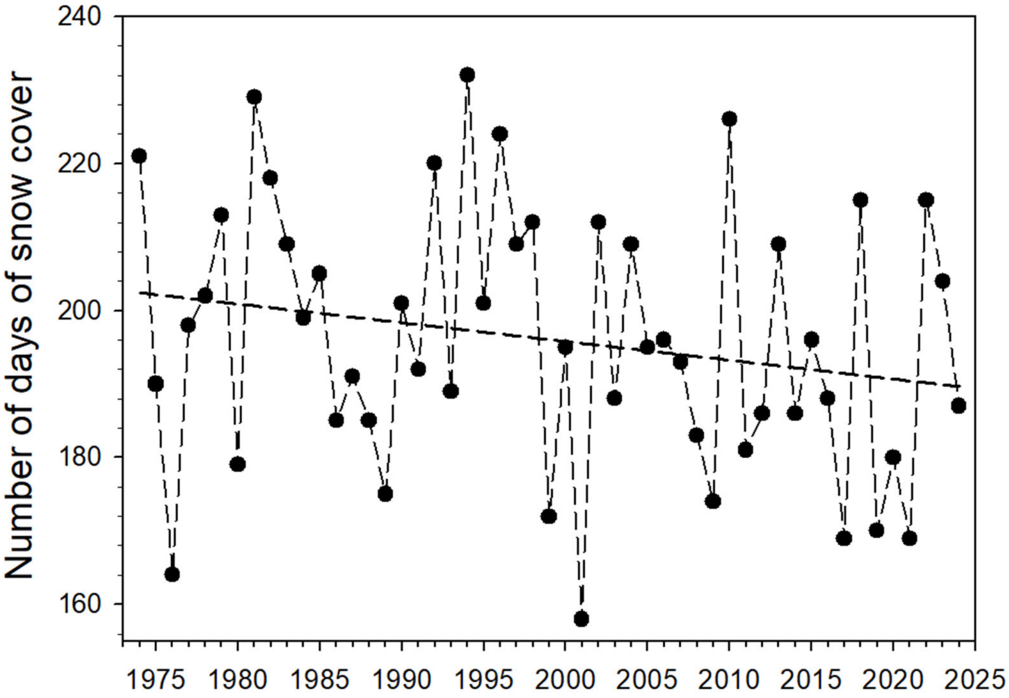


Figure SF46. The number of days in each year with ≥ 10cm of snow in Gothic.

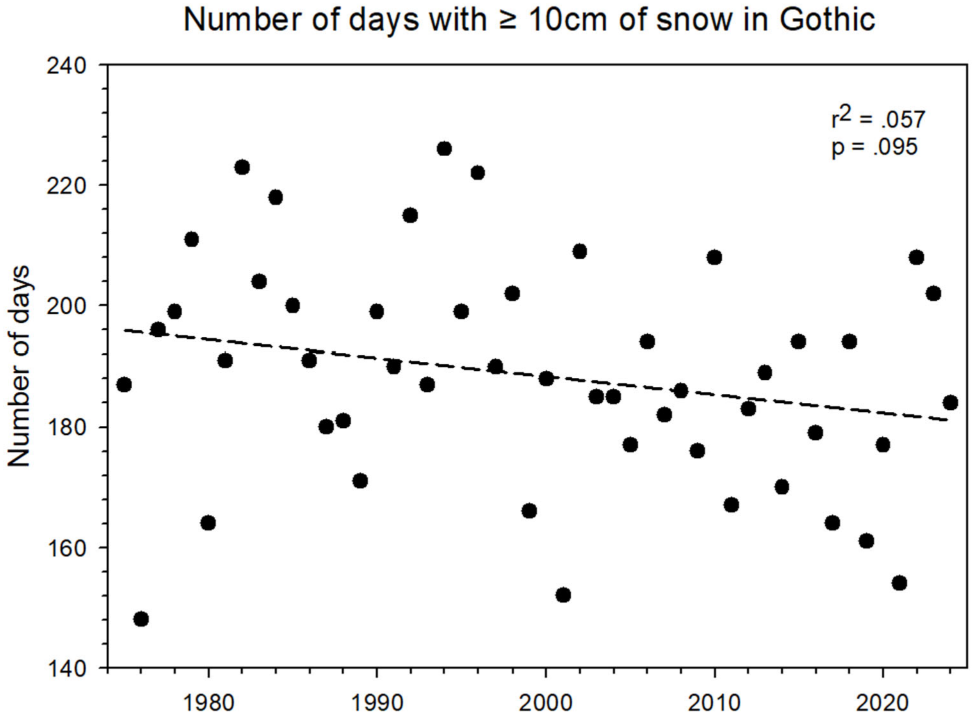


Figure SF47. The day of the year when the last snow melted from barr's snow observation site in Gothic each spring.

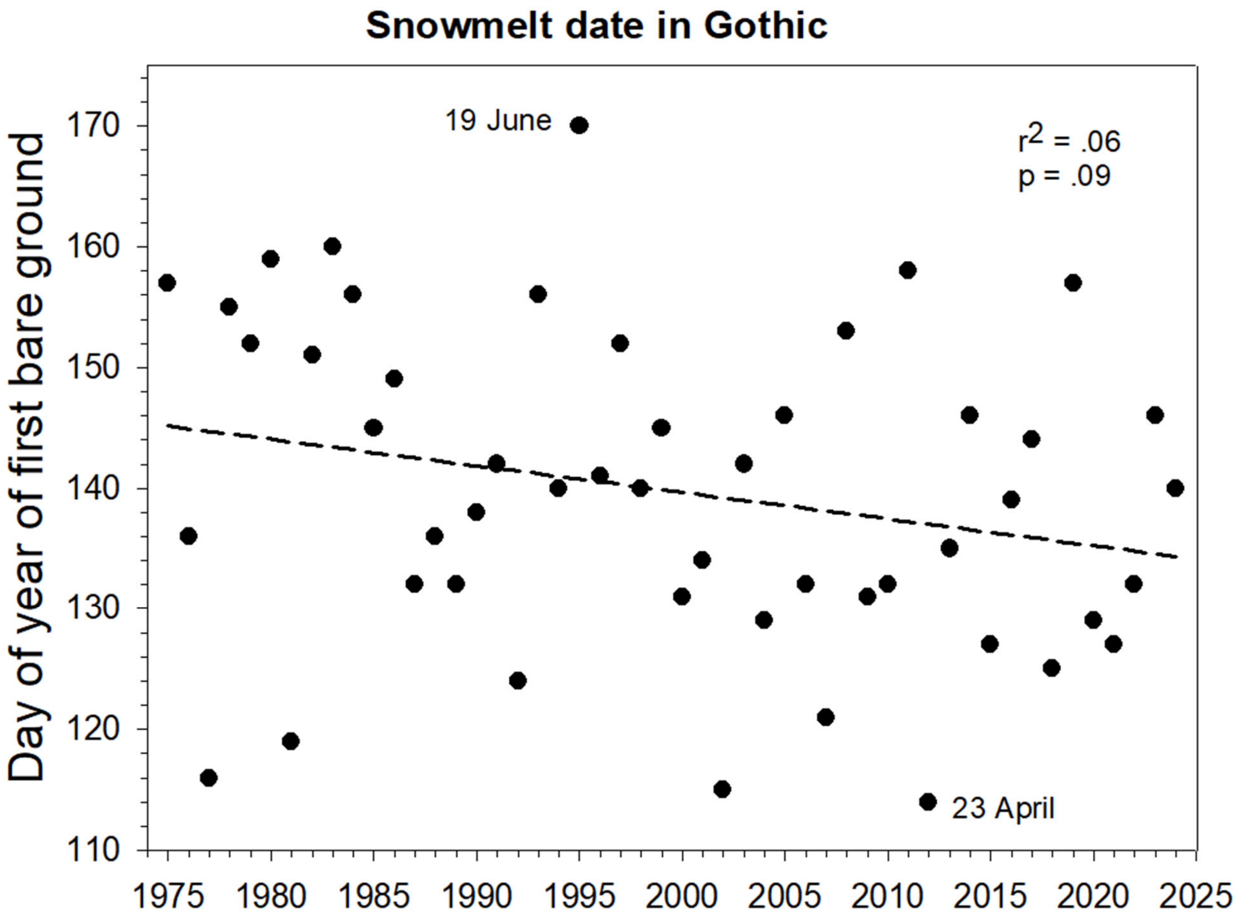


Figure SF48. The number of days between snowmelt in the spring and the last frost in Gothic.

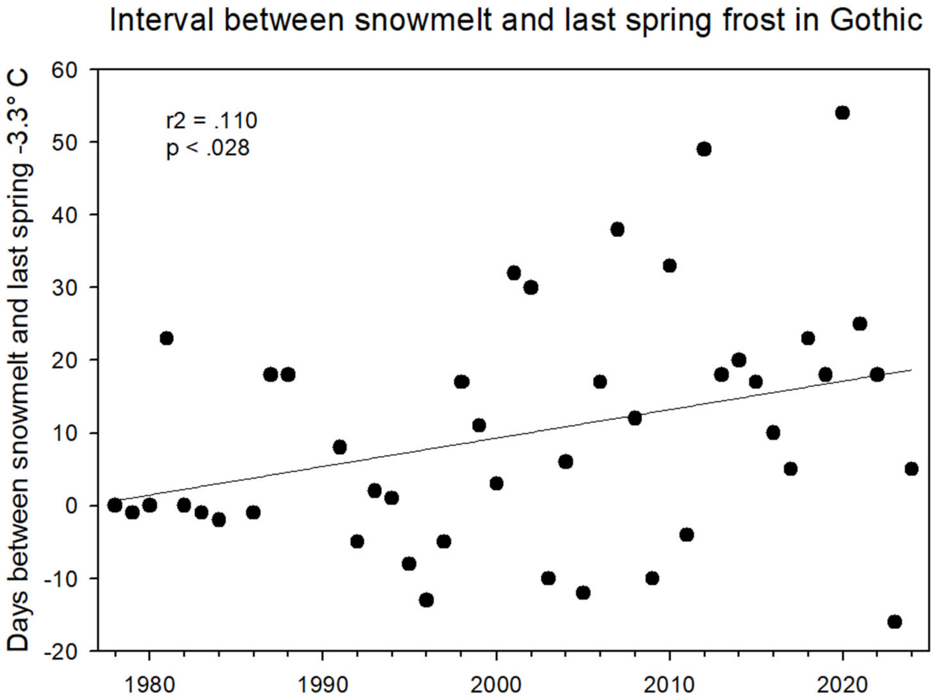


Figure SF49. The annual pattern of soil moisture in Gothic, showing the peak at snowmelt. Sensors at 10 and 15 cm.

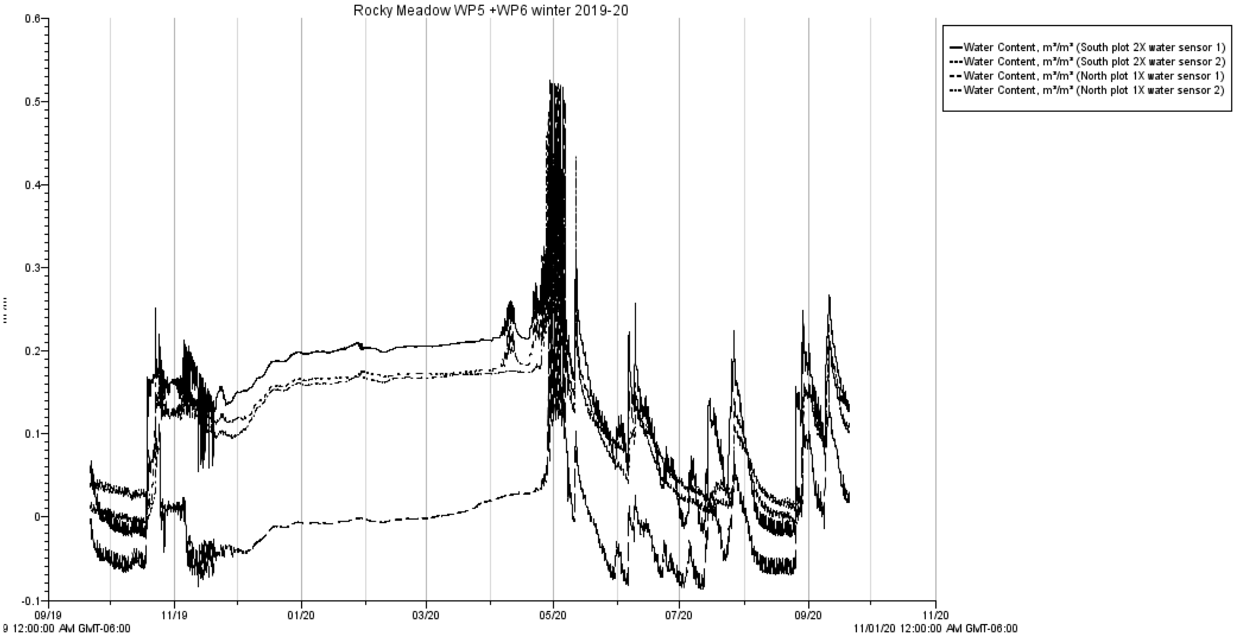


Figure SF50. The correlation between precipitaton in July and August in Crested Butte from 1910 - 2008 and the width of latewood from tree cores from Rio Pueblo (data from Connie Woodhouse).