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Changes in climate and hydrochemical responses in a high-elevation catchment in the Rocky Mountains, USA

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Abstract

A continuous climate record since 1951 at Niwot Ridge in the Colorado Front Range shows a decline in mean annual temperature, an increase in annual precipitation amount, and a decrease in mean daily solar radiation for the summer months. The increase in precipitation amount explains about half of the 200% increase in annual wet deposition of NO_3^- to Niwot Ridge over the last decade. Differences in climate parameters between 1994 and 1995 (increased snow depth and decreased net energy flux to the snowpack) resulted in a 4–5-fold increase in the magnitude of solute release from the snowpack in the form of an ionic pulse. In turn, the high chemical loading of strong acid anions in the seasonal snowpack and release of these solutes from the seasonal snowpack in the form of an ionic pulse is causing episodic acidification (ANC < 0 μeq liter^{-1}) in headwater catchments at present deposition levels. Small changes in climate parameters may cause large changes in the hydrochemistry of alpine streams. The changes in climate at Niwot Ridge are not in synchrony with lowland warming in the Great Plains to the east and serve as a reminder that climate in alpine areas is driven by local conditions and may be asynchronous with regional and global climate trends.

Many alpine regions are susceptible to environmental damage that will affect both their ecological health and the regional economies. Small changes in the flux of energy, chemicals, and water to high-elevation ecosystems may result in large changes in the climate, ecosystem dynamics, and water quality of these catchments (e.g. Baron 1992). For example, field and laboratory experiments have demonstrated that initial stages of snowmelt often have ionic concentrations many times higher than averages for the whole snowpack—an ionic pulse (e.g. Johannessen and Henriksen 1978; Colbeck 1981). The magnitude of the ionic pulse may be increased or decreased by small changes in energy flux (Williams and Melack 1991*b*). In turn, the release of solutes from the seasonal snowpack may have a direct and large effect on the solute content of stream waters (Williams et al. 1993). However, to date in the central Rocky Mountains there has been little long-term climate data available to document how climate may be changing in high-elevation areas and to provide insight into how these changes in climate may affect the hydrochemistry of these catchments.

Our knowledge of the climatic characteristics of mountain regions is limited by both paucity of observations—short records that seldom span 100 yr and a sparse station network—and insufficient theoretical attention given to the complex interactions of spatial scales in weather and climate phenomena of mountains (Barry 1992). Climate-change scenarios under elevated greenhouse gases have been extrapolated from low elevations to the Rocky Mountains (e.g. von Katwijk et al. 1993), but the efficacy of such extrapolations is unknown. The coarse spatial resolution of most general circulation models (GCMs) ($4.5 \times 4.5°$ mesh of T-42 simulations have a grid size of $\sim 500 \times 500$ km) is far too low to provide any meaningful information for alpine areas. Local climate in mountain areas is often different from regional and global climate conditions because of the presence of precipitation as snow and orographic barriers to the movement of air masses, causing problems in modeling the climate of alpine areas. To illustrate, regional climate change scenarios over the U.S. created from a coarse grid GCM and a

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nested limited area model (LAM) produce precipitation change scenarios in mountainous areas that locally differ in magnitude, sign, and spatial and seasonal detail (Giorgi et al. 1994). The LAM (the NCAR/Penn State MM4) also overpredicts current precipitation in the western U.S. An analysis of long-term climate measurements in mountain areas is needed to better our understanding of current climate in these areas and to validate climate change scenarios forced by increases in greenhouse gases (e.g. Giorgi et al. 1994) or by regional changes in land use (e.g. Copeland et al. 1996).

Here we present a 40-yr climate record from alpine and subalpine climate stations located on Niwot Ridge in the Colorado Front Range. Our objective is to analyze trends in temperature, precipitation, and shortwave radiation. Furthermore, there has been a long-term warming trend in annual temperature in the central Great Plains east of the Rocky Mountains in Colorado (Kittel 1990; Lauenroth and Sala 1992), and we discuss the potential response to these climate variables at Niwot Ridge to lowland warming. We also present information on how changes in these climate variables may directly affect chemical loading from wet deposition, release of solutes from the seasonal snowpack in an ionic pulse, and changes in the chemical content of streamwaters.

Site description

The Colorado Front Range rises directly from the Denver–Boulder–Fort Collins metropolitan area. This geographical setting results in alpine basins of this portion of the Continental Divide being just west of large urban and agricultural areas. Green Lakes Valley is an east-facing headwater catchment, 700 ha in area, and ranging in elevation from 3,250 to ~4,000 m at the Continental Divide (Fig. 1). The catchment appears typical of the high-elevation environment of the Colorado Front Range and includes Niwot Ridge, where research has been conducted since the early 1950s (Caine and Thurman 1990). The Green Lakes Valley is a water source for the city of Boulder and is owned by the city. Public access is prohibited; hence the Green Lakes Valley does not have the recreational impact of other high-elevation sites in the Front Range. The catchment is a linear cascade of five lakes with only Green Lake 1, on the north wall below Niwot Ridge, tributary to this sequence. Seven sites, ranging in area from 9 to 700 ha, are routinely sampled for water quality along the axis of the basin. About 80% of the annual precipitation occurs as snow. Streamflows are markedly seasonal, varying from <0.1 $m^3 s^{-1}$ during winter to >1.5 $m^3 s^{-1}$ at maximum discharge during snowmelt just below Lake Albion at the lower end of the valley.

Several research sites are on Niwot Ridge, which is an alpine tundra ecosystem extending eastward from the Continental Divide and forming the northern boundary of the Green Lakes Valley watershed. Climate data have been collected since the early 1950s at the D1 and C1 stations on Niwot Ridge (Fig. 1). D1 is an alpine site at 3,750 m; C1 is in a heavily forested subalpine site at 3,048 m. The long-term ecological research (LTER) network operates a high-elevation tundra laboratory at the Niwot Ridge Saddle, located between D1 and C1 at an elevation of 3,500 m. Also located at the saddle is a subnivean laboratory where snowpack meltwater samples are automatically collected before contact with the ground.

Methods

Climate data have been collected continuously on Niwot Ridge since 1951 at D1 and since 1953 at C1. Precipitation amount was collected at D1 by an unshielded Belfort recording gauge from 1952 to 1964 and with an Alter-type shield from 1964 to the present; shielded and unshielded gauges were run concurrently for 2 yr and the pre-1964 data adjusted. Precipitation at C1 has been collected continuously with a Belfort recording gauge placed in a forested clearing with a diameter about half the height of the highest trees. Mean daily temperature has been measured continuously at both sites using a Belfort hygrothermograph with a chart recorder; modern electronic instruments and recorders have been in place since the mid-1980s, but to ensure comparability in the long-term temperature data set we used only the chart information. Incoming shortwave solar radiation (0.4–1.1 μm) as mean daily total during summer (June, July, and August) was measured from 1970–1987 with a Belfort actinometer, from 1988 to 1989 with a Kipp-Zonnen CM-11 pyranometer, and from 1990 to the present with a LiCor LI-200. Calibration of new instruments involved a minimum of 2 yr overlap between older and newer instruments. Accuracy of monthly precipitation totals is ±20 mm and accuracy of monthly mean temperature values is ±1°C. Missing data were treated with regression analyses between D1 and C1 and with other nearby climate stations as presented by Greenland (1989). Trends in climate data are considered significant at the $\alpha = 0.05$ level.

Wet deposition is sampled on the Niwot Ridge saddle (3,500 m) as part of the National Acid Deposition Program (NADP), which operates about 200 wet precipitation collectors throughout the continental U.S. Surface water samples were collected approximately weekly starting in 1985 along an elevation gradient from about 1 May through the end of September and about monthly during winter at Green Lake 4. Snowpack meltwater samples were collected in 1-m^2 snow lysimeters before contact with the ground in 1994 and 1995 following the protocol of Bales et al. (1993). Meltwater discharge was measured continuously in tipping buckets, conductance was measured continuously with an inline conductance meter, and daily grab samples were analyzed for major solute concentrations. Snowpack samples were collected on a weekly basis from a site located 20 m from the snow lysimeters, following the protocol of Williams and Melack (1991*a*). Since 1993, analysis for major solutes from stream, snow, and meltwaters followed the protocol of Williams and Melack (1991*a*); detection limits for most solutes was 1 μeq $liter^{-1}$ and precision was generally better than 2%. From 1985 to 1992, sample collection and analysis fol-

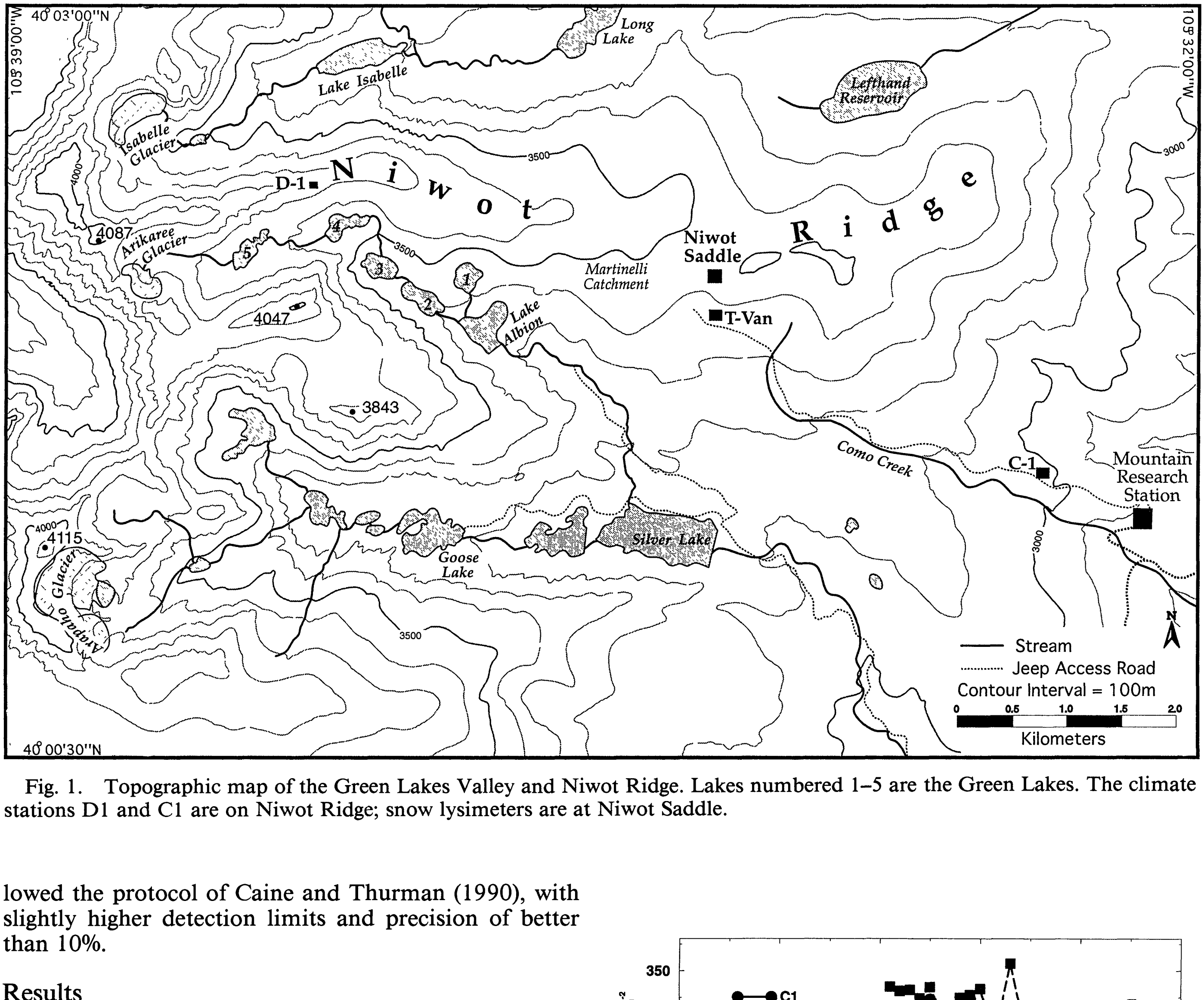

Fig. 1. Topographic map of the Green Lakes Valley and Niwot Ridge. Lakes numbered 1–5 are the Green Lakes. The climate stations D1 and C1 are on Niwot Ridge; snow lysimeters are at Niwot Saddle.

lowed the protocol of Caine and Thurman (1990), with slightly higher detection limits and precision of better than 10%.

Results

The long-term temperature record from Niwot Ridge shows an overall decrease in annual temperature with high interannual variability (Fig. 2). Mean annual temperature at the D1 climate station from 1951 to 1994 was −3.8°C (SD = 1.24°C). A simple linear regression analysis shows a slight decreasing trend in temperature (slope, −0.009°C yr^{-1}) that is not significant (r^2 = 0.01, P = 0.68). Mean annual temperature at the C1 forested site was 1.34°C (SD = 0.98°C). Linear regression analysis at C1 also shows a decreasing trend in annual temperature (slope, −0.025°C yr^{-1}), but one that is significant (r^2 = 0.10, $P < 0.05$); the decrease in mean annual temperature since 1953 is 1.0°C.

In contrast to temperature trends, annual precipitation amount at D1 is increasing. Mean annual precipitation was 1,006 mm (SD = 243 mm). A linear regression shows the increase in annual precipitation amount from 1951 to 1994 to be significant (r^2 = 0.15, $P < 0.01$), with a slope of 7.5 mm yr^{-1} and an increase of ~300 mm since

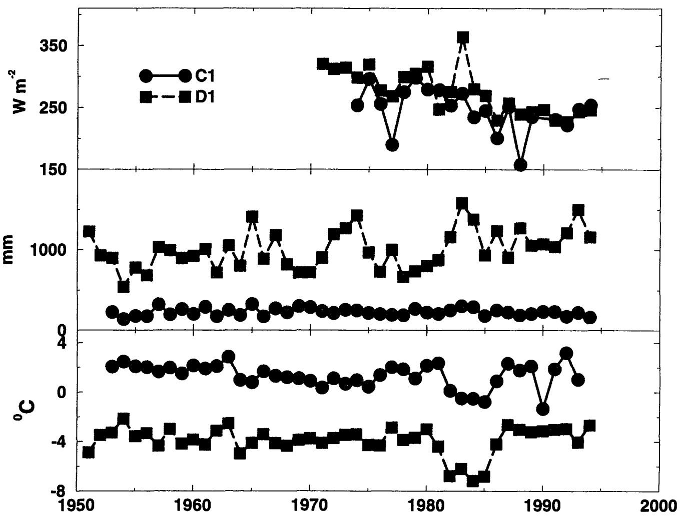

Fig. 2. Data from the D1 (3,750 m) and C1 (3,048 m) climate stations on Niwot Ridge. Annual temperature (°C) and precipitation (mm) amount from 1951 to 1994 at D1 and starting in 1953 at C1; mean daily solar radiation (W m^{-2}) during summer (June, July, and August) from 1971 to 1994.

the 1950s. Most of the precipitation increase has occurred since 1967 (a period during which the same gauge and screen were used), with a slope of 14 mm yr^{-1}. Mean annual precipitation at C1 over the 42 yr of record was 228 mm—about 25% of the annual precipitation at D1. There is no trend in precipitation at C1 ($r^2 = 0.001$). The reason for the difference in precipitation trends between D1 and C1 is unknown. The increase in annual precipitation amount at D1 of 300 mm is greater than the mean annual precipitation at C1 of 228 mm.

Daily mean solar radiation during summer at D1 has decreased significantly since continuous solar radiation measurements began in 1971 ($r^2 = 0.54$, $P < 0.001$), at a rate of -3.8 W m^{-2} yr^{-1} (Fig. 2). Mean daily shortwave radiation at D1 in summer since 1971 is 276 W m^{-2} (SD = 37 W m^{-2}). The decrease in incoming shortwave radiation over the last 25 yr has been almost 100 W m^{-2} or ~30% of the average daily incoming shortwave radiation in the early 1970s. Mean daily shortwave radiation at C1 also shows a decreasing trend (slope, -2.2 W m^{-2} yr^{-1}), but this decrease is not statistically significant ($r^2 = 0.15$, $P = 0.10$). The decrease in incoming shortwave radiation during summer means that atmospheric attenuation has increased during this timespan and suggests that there has been increased cloud cover during summer.

These changes in temperature, solar radiation, and precipitation may cause large changes in the solute concentrations of snowpack meltwater. In 1994, the maximum concentrations of NH_4^+, NO_3^-, and SO_4^{2-} in snowpack meltwater were ~4 times that of bulk concentrations in a colocated snowpit (Fig. 3); other solutes showed a similar pattern. Maximum concentrations of meltwater for these three solutes ranged from 25 to 33 μeq $liter^{-1}$. In 1995, the maximum concentrations of these solutes in meltwater were ~20 times that of bulk concentrations,

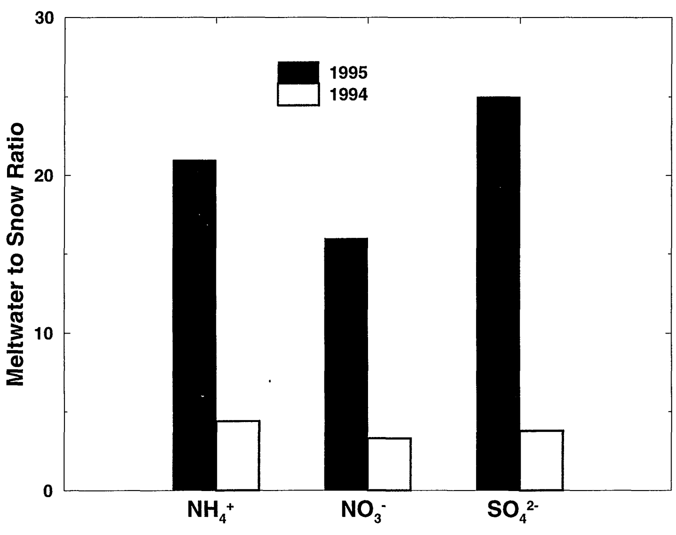

Fig. 3. Ratio of maximum concentrations of NH_4^+, NO_3^-, and SO_4^{2-} in snowpack meltwater before contact with the ground to bulk concentrations from a colocated snowpit sampled concurrently, 1994 and 1995.

with meltwater concentrations ranging from 130 to 200 μeq $liter^{-1}$ (Fig. 3).

The difference in the magnitude of the ionic pulse in meltwater between 1994 and 1995 was caused by differences in climate parameters and not by differences in bulk snowpack concentrations. In both years, volume-weighted mean concentrations of these three solutes in the snowpack at maximum accumulation were about the same, ranging from 6 to 10 μeq $liter^{-1}$. Snowpack depth in 1995 was about twice that of 1994, with a snow depth at the initiation of melt in 1994 of 1.50 m and of 3.05 m in 1995. Spring weather in 1995 was much cooler and cloudier than in 1994, delaying the initiation of snowmelt by 20 d compared to 1994 (Fig. 4). Melt rates are shown as an index of climate conditions during snowmelt in 1994 and 1995 because snowmelt integrates all energy fluxes. Once snowmelt began in 1995, melt rates compared to 1994 were much lower and more discontinuous for the first 20 d of melt. As a consequence of the slower melt rate and deeper snowpack at the initiation of melt in 1995, solute concentrations in meltwater were much greater in 1995 than in 1994 (Fig. 3). To further illustrate the difference in meltwater chemistry between years, maximum measured conductance in 1995 was 58 μS cm^{-1}, slightly more than 4 times the maximum conductance of 13 μS cm^{-1} measured in 1994 (Fig. 4). Because daily conductance values are highly variable, maximum daily conductance values are shown for ease of presentation. Conductance remained elevated in 1995 through the first 20 d of melt compared to 1994. Differences in climate parameters between 1994 and 1995—increased snow depth and decreased net energy flux to the snowpack—resulted in a 4–5-fold increase in the magnitude of the ionic pulse.

Changes in climate may also influence the chemical content of stream water in high-elevation catchments by changing chemical loading from atmospheric deposition. NADP results show that there has been an increase of ~200% in NO_3^- loading from wet deposition at Niwot Ridge over the last decade, from ~8 kg ha^{-1} yr^{-1} on average for 1985–1987 to 16.5 kg ha^{-1} yr^{-1} for 1990–1992 (Fig. 5). Earlier and comparable measurements of annual NO_3^- deposition extend the record back to 1982 and indicate an even larger increase in NO_3^- loading, with a mean of 5.7 kg ha^{-1} yr^{-1} for 1982–1986 (Reddy and Caine 1988). A simple linear regression analysis shows that increases in precipitation amount account for about half the increase in annual NO_3^- loading ($r^2 = 0.56$, $P = 0.01$) and about half comes from increases in the annual volume-weighted mean concentration of NO^{-3} in precipitation ($r^2 = 0.59$, $P = 0.01$); a multiple linear regression analysis shows that the combination of precipitation amount and concentration explains annual wet deposition of NO_3^- ($r^2 = 0.99$, $P \ll 0.0001$). Increased precipitation to mountain areas will directly result in an increase in wet deposition of atmospheric pollutants, even with no increase in ambient concentrations of those pollutants.

The increase in NO_3^- loading from wet deposition has caused a fundamental change in the NO_3^- concentrations of stream water in the Green Lakes Valley. Paralleling

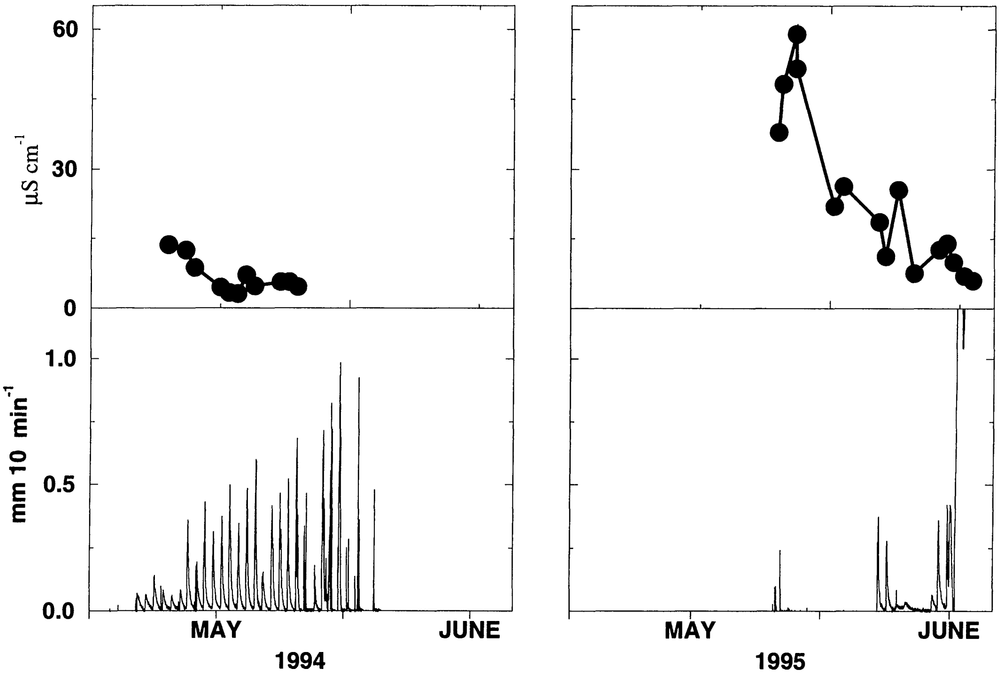

Fig. 4. Time series of snowmelt discharge [mm (10 min)$^{-1}$] and maximum daily conductance (μS cm^{-1}) for 1994 and 1995. The entire 1994 time series is shown and only the first 21 d are shown for 1995; after the first 21 d of melt in 1995, discharge volumes were off-scale and conductance values were similar to 1994.

the large increase in NO_3^- loading from wet deposition in the late 1980s, there has been an increase in the annual minimum concentrations of NO_3^- during summer in Green Lake 4, from below detection limits in 1985 to ~10 μeq liter^{-1} in 1990 (Fig. 5). Annual NO^{-3} export from Green Lake 4 in stream waters increased by 50% from 1985–1988 (5.0 kg ha^{-1} yr^{-1}) to 1989–1992 (7.5 kg ha^{-1} yr^{-1}). Mass balance analysis shows that currently ~50% of the NO_3^- loading from annual wet deposition is exported in stream waters (Williams et al. 1996). This increased export of NO_3^- in stream waters during the growing season may be the result of interactions between increased atmospheric deposition of inorganic nitrogen and decreases in incoming shortwave radiation that reduce the ability of vegetation to assimilate inorganic nitrogen during the growing season.

These environmental changes that have resulted in an increase in chemical loading and an increase in the magnitude of the ionic pulse are now causing episodic acidification in the headwaters of the Green Lakes Valley (Fig. 6). We define acidification as occurring when acid-neutralizing capacity (ANC) values decrease below 0 μeq liter^{-1}. Caine (1995) has shown that there has been a decrease in ANC of these headwater streams over the last decade, coincident with the increased loading of strong acid anions. On 9 June 1994, ANC values were −6.1 μeq liter^{-1} at the 9-ha Arikaree site and −1.6 μeq liter^{-1} at the 42-ha Navajo site (Fig. 6). ANC values increased with basin area, but were still <50 μeq liter^{-1} at the inlet to Lake Albion on this date—an area of 380 ha. ANC values were below 0 μeq liter^{-1} at the Arikaree site for much of June–September in 1994, and at the Navajo site there were five sampling dates with negative ANC values. The pH measurement on this date at the Arikaree site was 4.86 and pH at the Navajo site was 5.37. The pH values

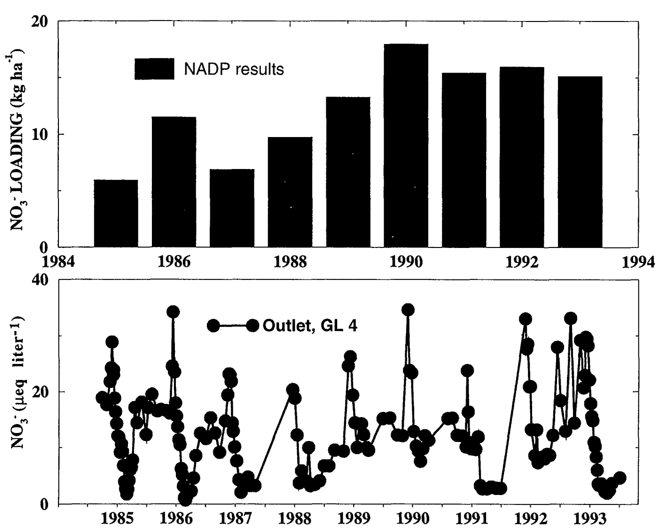

Fig. 5. Time series of NO_3^- concentrations from the outlet stream of Green Lake 4 and annual NO_3^- loading measured at the Niwot Ridge NADP site.

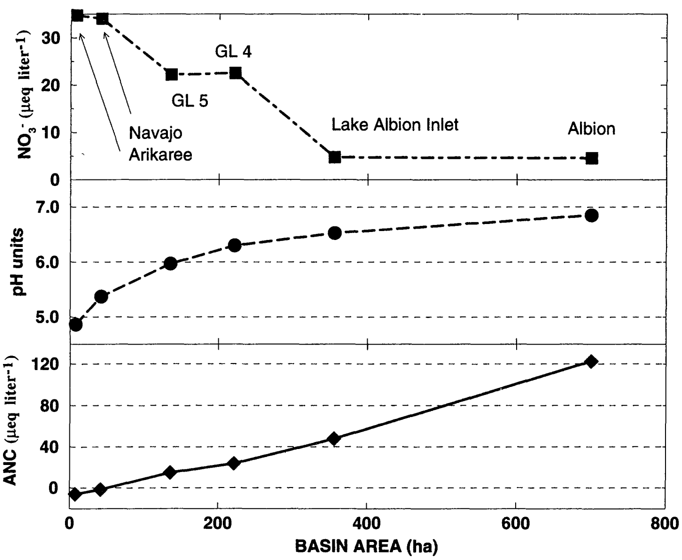

Fig. 6. ANC, pH, and NO_3^- values as a function of basin area, 9 June 1994, in Green Lakes Valley. Episodic acidification is occurring at present deposition levels in the headwater catchment.

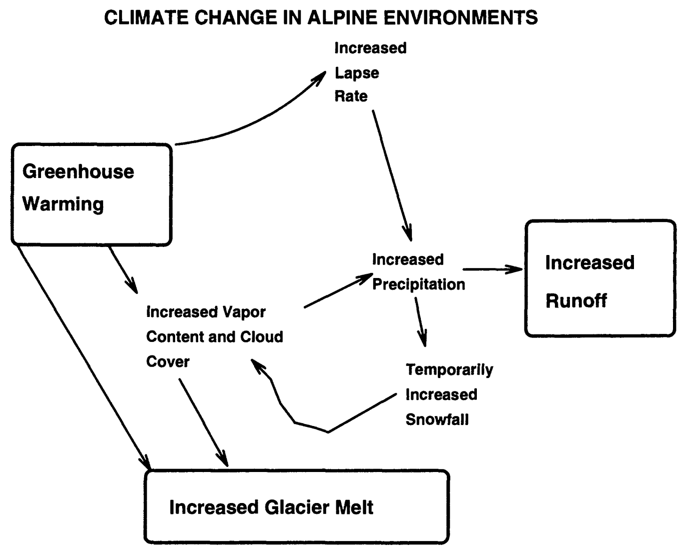

Fig. 7. Conceptual model of the effects of climate warming in mountain areas, adapted from Barry (1990).

then increased rapidly to ~6.0 at the 135-ha Green Lake 5. The decrease in pH and the low values for ANC were associated with concentrations of NO_3^- as great as 35 μeq liter^{-1} (Fig. 6). Changes in climate that increase chemical loading of acidic species and the magnitude of the ionic pulse will have a direct and negative impact on the chemical content and acid-base status of high-elevation streams.

Discussion

Modeling scenarios based on $2 \times CO_2$ GCM results have been applied to the Rocky Mountains using the nested LAM MM4 with a 60-km grid spacing. Under this scenario, climate responses for the Rocky Mountains include increased cool season precipitation and increased annual air temperature (Giorgi et al. 1994). Increasing annual temperatures in the Great Plains east of Niwot Ridge provide a natural experiment with which to test the applicability of these modeling scenarios to Niwot Ridge. Temperature and precipitation have been measured since 1939 at the Central Plains Experimental Range (CPER), a companion LTER site located ~100 km northeast of Niwot Ridge at an elevation of 1,650 m. Annual temperature at CPER has been increasing since 1967 and each year from 1974 to 1990 had temperatures above the 52-yr mean (Lauenroth and Sala 1992). Annual temperature in 1990 was ~2.5°C greater than annual temperatures in the early 1960s, consistent with the increase in annual temperature simulated for the continental interior of the U.S. by the doubled CO_2 MM4 modeling scenario. There is no trend in annual precipitation at CPER.

Additionally, long-term measurements of atmospheric water-vapor profiles for altitudes ranging from 9 to 27 km at Boulder (40 km east of Niwot Ridge) from 1981 to 1994 show a significant increase in water vapor of about 0.5% yr^{-1} (Oltmans and Hofmann 1995), consistent with the prediction of increased advection of water vapor to higher altitudes under both doubled CO_2 scenarios and land use changes in the Great Plains to the east (Copeland et al. 1996). At the D1 climate station on Niwot Ridge, the increase in annual precipitation and decrease in summer shortwave radiation (Fig. 2) is consistent with the $2 \times CO_2$ modeling scenarios of increased annual precipitation and increased water vapor simulated by the nested regional model MM4 (Giorgi et al. 1994). However, the lack of an annual temperature increase on Niwot Ridge is inconsistent with MM4 predictions of increasing annual temperatures.

Precipitation as snow may confound the temperature prediction of MM4 for high-elevation areas. The climate patterns at Niwot Ridge are consistent with a conceptual model of climate change in mountain areas proposed by Barry (1990) (Fig. 7). Warming at lower elevations may result in the advection of increased water vapor to higher elevations and increased orographic precipitation as snow. Precipitation as snow increases the albedo, lowers net radiation, and is compatible with lower air temperature and an increase in lapse rate between low- and high-elevation areas. Late-lying snow on the ground may provide the moisture source for increased cloud cover during summer months, resulting in a decrease in direct shortwave radiation. In the short term (decadal time scale), there may be a positive feedback loop in high-elevation areas, with atmospheric cooling in turn lowering saturation vapor pressure, increasing precipitation, and continuing the cooling trend. The key factor here is precipitation as snow in spring and early summer and the resulting feedback with local climate. We recognize that there may be other processes that explain these observations and suggest that a search for such processes represents fertile ground for further research. Giorgi et al.

(1994) suggested that a negative temperature signal, caused by relatively high evaporative cooling and cloud shading, may occur over the mountainous west in locations of maximum precipitation.

An open question is whether the climate trends at Niwot Ridge are site-specific or are representative of high-elevation areas in general. There are few climate records at high-elevations sites with which to compare the Niwot climate record. A comparison of temperature records from 1953 to 1991 for three Colorado Rocky Mountain climate stations (elev., 3,000–3,500 m) paired with three stations in the Colorado plains (elev., 1,250–1,300 m) show a cooling trend for the past four decades, with an increase in the lapse rate and an increase in diurnal temperature ranges, in contrast with a decrease in the diurnal temperature range across much of the U.S. and Canada during 1941–1990 (Brown et al. 1992). These results are consistent with the NWT-CPER pattern and with the conceptual model of Barry (1990). Climate records from the Canadian Rocky Mountains show a decreasing trend in annual temperature since the 1940s at the highest elevation station at Lake Louise (1,534 m) and an increase in annual temperatures at Banff (1,397 m) over the same time period (Luckman 1990). Long-term climate records at Davos, Switzerland (1,540 m), show no consistent pattern in temperature change, but records of snow depth show that on 1 January snow depth has increased from ~300 mm in the 1930s to ~500 mm up to 1990 (Föhn 1990). Furthermore, records of snow water equivalence (SWE) on 15 April show that SWE has substantially exceeded that on 1 January since about 1975, whereas the two values were similar from 1950 to 1965 (Föhn 1990), indicating more snowfall in late winter.

The hydrochemical responses to changes in precipitation amount and energy flux at Niwot Ridge appear typical of other high-elevation catchments in the Rocky Mountains and other ranges such as the Sierra Nevada. Bales et al. (1990) have shown that initial solute concentrations of snowpack meltwater were 5–10 times bulk concentrations from a 2-m snowpack at Glacier Lakes in the Snowy Range, southern Wyoming. Williams et al. (1996) reported that increases in the annual wet deposition of nitrogen have caused a change in the nitrate concentration-discharge relationship throughout high-elevation catchments of the Colorado Front Range. Hydrochemical modeling scenarios conducted for the high-elevation Emerald Lake catchment in the Sierra Nevadas by Wolford and Bales (1996) show that increases in the chemical loading of snow result in a more-pronounced ionic pulse. The same modeling scenarios by Wolford and Bales also show that increases in the magnitude of the ionic pulse caused corresponding changes in the chemical content of stream waters during snowpack runoff. Clearly, changes in the climate of alpine basins that cause an increase in the amount of precipitation as snow or decreases in air temperature or net radiation can cause large changes in the chemical content of stream waters draining these basins.

Examination of the glacial record places the climate record of Niwot Ridge into a temporal perspective. Alpine glaciers advance and retreat in response to local, not global, climate (Gillespie and Molnar 1995). An exhaustive review of mountain glaciers shows that the maximum advances of mountain and continental glaciers were often asynchronous (Gillespie and Molnar 1995). The asynchronism of mountain and continental glaciations serves as a reminder that climate in mountain terrains need not be synchronized with global and continental climate. What is important about the climate trends at Niwot Ridge is that they do not conform to the predictions of some regional and global GCM scenarios of a warmer and drier climate in the near future. The climate of high-elevation catchments is influenced by regional synoptic air patterns, but in ways that are not always similar or intuitive. Extrapolation of GCM results to mountain areas must be done with care. More empirical data on climate and ecosystem processes of high-elevation areas are needed before meaningful predictions can be made about responses of high-elevation ecosystems to altered regional and global forcings caused by increases in greenhouse gases and land use changes.

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