Forested Watersheds

Long-term Forest Management and Climate Effects on Streamflow

Shelby G. Laird, C.R. Ford, S.H. Laseter, J.M. Vose

Abstract

Long-term watershed studies are a powerful tool for examining interactions among management activities, streamflow, and climatic variability. Understanding these interactions is critical for exploring the potential of forest management to adapt to or mitigate against the effects of climate change. The Coweeta Hydrologic Laboratory, located in North Carolina, USA, is a 2,185-ha basin wherein forest climate monitoring and watershed experimentation began in the early 1930s. Extensive climate and hydrologic networks have facilitated research in the basin and region for over 75 years. Our purpose was (1) to examine long-term trends in climate and streamflow in reference watersheds, and (2) to synthesize recent work that shows that managed watersheds respond differently to variation in extreme precipitation years than reference watersheds. In the basin and in the region, air temperatures have been increasing since the late 1970s. Drought severity and frequency have also increased over time, and the precipitation distribution has become more variable. Reference watersheds indicate that streamflow is more variable, reflecting precipitation variability. Streamflow of extreme wet and dry years show that watershed responses to management differ significantly in all but a forest with coppice management. Converting deciduous hardwood stands to pine altered the streamflow response to extreme precipitation years the most. High evapotranspiration rate and increased soil water storage in the pine stands may be beneficial to reduce flood risk in wet years, but they create conditions that could exacerbate drought. Our results suggest that forest management can mitigate extreme precipitation years associated with climate change; however, offsetting effects suggest the need for spatially-explicit analyses of risk and vulnerability.

Keywords: climate, long-term monitoring, streamflow, forest management, watershed

Introduction

Climate change projections suggest significant changes in temperature (e.g., 2–9°F, or 1–5°C) and precipitation over the next several decades (U.S. Global Change Research Program 2009). Land managers and policy makers are challenged to develop adaptation and mitigation strategies to protect and ensure long-term forest health and sustained ecosystem services. Changing climate is one among many current and potential future threats to the sustainability of forest water resources. Population growth has increased demand for clean water, and pressures from sprawling metropolitan areas, interbasin transfers, and wastewater discharge are all threats to water quality and quantity (Sun et al. 2008). Other threats and stressors include changes in land use, invasive species, and fire. Often these stressors occur simultaneously, making it difficult to distinguish the effects of one single threat on streamflow (Vose et al. 2011). Long-term watershed research can offer valuable insights into the interactions among forest stressors and streamflow, as well as management options that might help forests adapt to or mitigate the effects climate change on water supplies.

Detecting climate change effects in streamflow data is complex, since simultaneous changes in land use (e.g., urbanization and development) can occur, and the signal of the latter can be greater in magnitude than the climate change signal. Long-term data from forested watershed that have undergone little to no change in land use can provide a robust way to detect the climate change signal in streamflow. Paired watershed studies that implement forest management regimes while accounting for climate variation are also a powerful means to investigate the effect of both management prescriptions and climate change on streamflow. Both

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approaches require sufficient length of records to allow effects to be detected. Without long-term data from research watersheds where land use is either constant or well documented, climate and management trends on streamflow may be difficult to identify (Burt 1994).

Streamflow responses to climate change are strongly related to changes in local precipitation, but are less so for temperature; however, the magnitude and timing of response may differ with different forest structure and species. Regions of the United States have shown both increasing precipitation and streamflow (Genta et al. 1998, Kiley 1999, Groisman et al. 2004) and decreasing water yield (National Research Council of the National Academies 2008), requiring further investigation and data collection to confirm regional effects. These large differences may be due to the highly variable precipitation changes. A larger portion of the available research attempts to predict streamflow, water supply, or water resources in relation to precipitation changes from various climate change scenarios (Milly et al. 2005, Moreau 2007, Seager et al. 2009). These models also show variable response to climate change due to a broad range of predictions for future precipitation.

Our objectives for this paper are (1) to present long-term climate trends from the Coweeta Hydrologic Laboratory, (2) to examine streamflow patterns in two control watersheds, varying in elevation, and (3) to discuss management strategies to adapt to or mitigate the effects of climate change on forested watersheds.

Methods

Site Description

Coweeta Hydrologic Laboratory is a U.S. Forest Service Southern Research Station Experimental Forest located in the Nantahala Mountain Range of western North Carolina, USA (Figure 1). Coweeta has been the focus of watershed experimentation since 1934. The Coweeta basin is 1,626 ha; elevations range from 675 to 1,592 m. Historic vegetation patterns have been influenced by human activity, primarily through both clearcut and selective logging, the introduction invasive species (Elliott and Hewitt 1997, Nuckolls et al. 2009), and fire (Hertzler 1936, Douglass and Hoover 1988). Forests on reference watersheds are relatively mature (approx. 85 years old) oak-hickory (at lower elevations) and northern hardwood species (at higher elevations) (Elliott and Swank 2008).

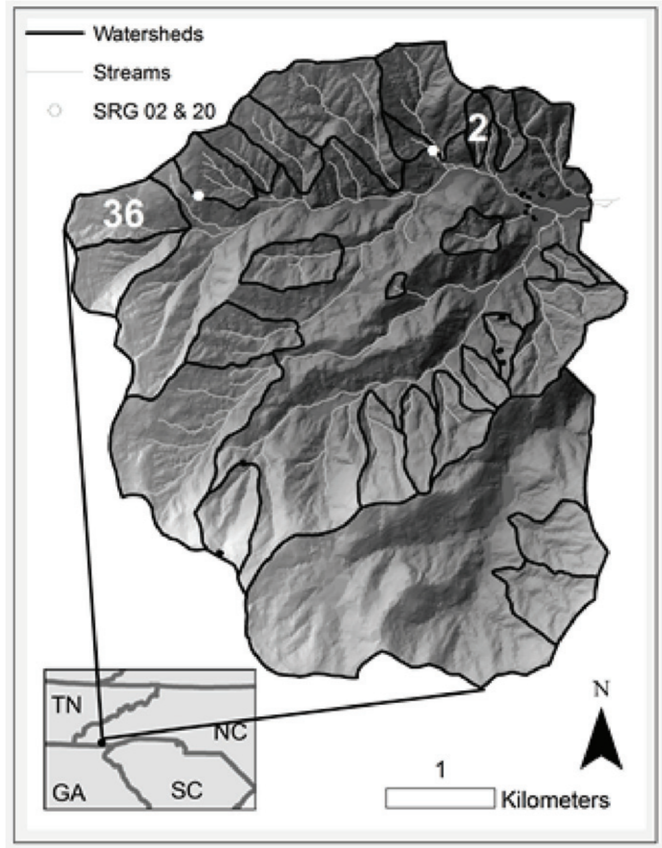

Figure 1. Map of the Coweeta Hydrologic Laboratory showing elevation gradients and main and subwatersheds. Watersheds 2 and 36 are labeled. (SRG, standard rain gauge)

Climate

Daily air temperature and precipitation have been recorded at the Coweeta main climate station (CS01) continuously since 1934. Temperature is recorded daily at 8 a.m. Eastern Standard Time using a National Weather Service (NWS) maximum, minimum, and standard thermometer. Total daily precipitation is collected in an 8-inch standard rain gauge (NWS). Recently, Laseter et al. (in review) presented the long-term trends in climate. We present a subset of those trends herein for comparison purposes.

Streamflow

To assess long-term trends in streamflow related to climate, we analyzed streamflow data from two control watersheds at Coweeta, watersheds 36 and 2 (Figure 1). The watersheds have similar aspects but varying elevations (Table 1). Streamflow data have been collected every 5 min since January 1936 for watershed 2 (WS2) and since May 1943 for watershed 36

(WS36). Both watersheds have remained undisturbed since the late 1920s, with the minor exception of a partial defoliation of WS36 by cankerworm from 1972 to 1979. Both watersheds consist of mixed hardwood forest, though the higher elevation watershed (WS36) contains northern hardwood community species.

Table 1. Characteristics of two control watersheds.

	WS 2	**WS 36**
Max elevation (m)	1,004	1,542
Elevation at weir (m)	709	1,021
Area (ha)	12	49
Aspect	SSE	SSE
Closest standard rain gauge (SRG)	20	02

Data from the closest standard rain gauge near each watershed were assumed to be representative of rainfall across the watershed. Standard rain gauge (SRG) 02 was used for watershed 36 and SRG 20 was used for watershed 02 (Figure 1).

We analyzed time trends in two ways. First, precipitation was regressed directly against streamflow using simple linear regression. The fit of the relationship was analyzed by looking at residuals and other possible variables of influence, including temperature (not discussed). Secondly, we calculated a runoff coefficient (RO/P) index for the two control watersheds by dividing annual streamflow (RO) by annual precipitation (P). From a simple mass hydrologic balance, streamflow output is the balance of precipitation inputs minus evapotranspiration (ET) losses: RO = P – ET. The ratio of RO to P is thus the fraction of rainfall that appears as streamflow. We used a penalized B-spline curve to analyze any possible trends in the runoff coefficient over time. Spline fit used default and custom settings for SAS 9.2, including 3 degrees, 10 control points (knots), and weight of 0 to characterize the spline curves.

Management

The interaction of management and climate was determined recently by Ford et al. (in press); they analyzed data from six paired treatment and reference watersheds throughout the Coweeta basin. They modeled the responses of streamflow to vegetation and climate. Management included species conversion, clearcuts on high and low elevation, coppice, and old field succession. We present some of the results of that study herein for comparison purposes.

Results and Discussion

Long-Term Climate

Climate data from Coweeta shows that average maximum, annual, and minimum air temperatures have increased significantly relative to the long-term mean, appearing to begin in the late 1970s (Figure 2). The rate of increase is about 0.5°C per decade beginning in the mid 1970s.

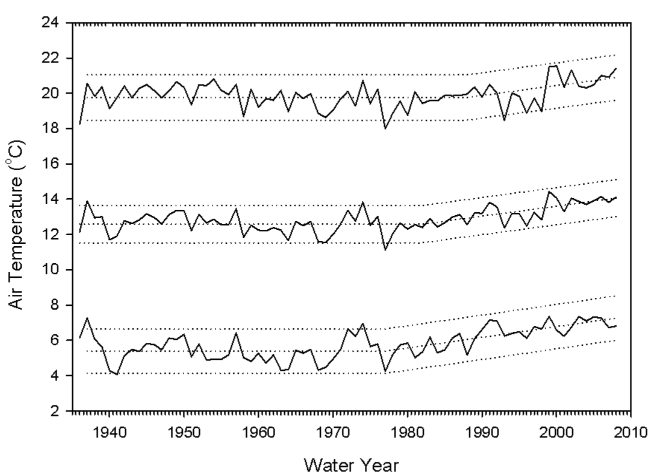

Figure 2. Long-term average maximum (top), annual (middle), and minimum (bottom) air temperatures at Coweeta Hydrologic Laboratory climate station CS01 in Laseter et al. (in review).

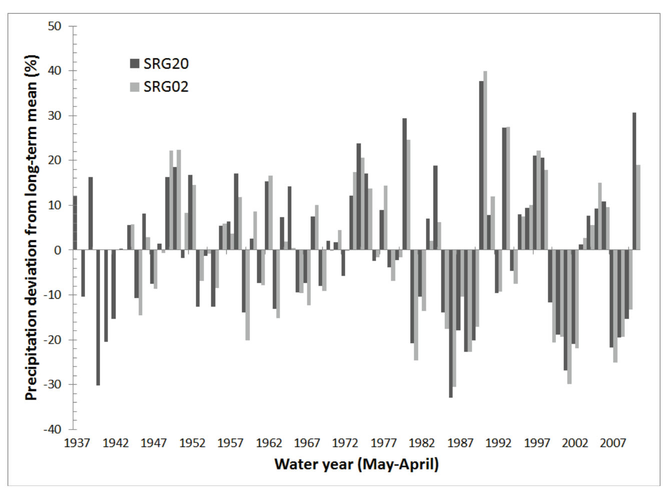

Figure 3. Deviation of annual precipitation totals from the long-term mean recorded at Coweeta Hydrologic Laboratory standard rain gauges 02 and 20.

Coweeta has some of the highest annual precipitation amounts in the eastern United States, averaging 1,794 mm/yr. Analyses of long-term precipitation suggest no significant change in mean precipitation at Coweeta (Ford et al., in press; Laseter et al., in review); however, the variability of precipitation is changing over time. For example, extreme annual precipitation (i.e., low and high rainfall) event years are occurring more frequently with time (Figure 3), which has resulted in increased recent drought severity and frequency (Laseter et al., in review).

Long-Term Streamflow

Streamflow data indicate similar trends to those found in precipitation data, including increased variability since the 1970s, largely due to the strong linear relationships between precipitation and streamflow for the two control watersheds (Figure 4). The higher elevation watershed 36 predicted streamflow or runoff is $RO'_{36} = 0.92P - 204.77$ (R^2=0.68, $p<0.01$), and the lower elevation watershed 02 predicted streamflow is $RO'_{02} = 0.71P - 626.87$ (R^2=0.78, $p<0.01$).

For any given amount of precipitation, annual streamflow on the higher elevation watershed (WS36) is at least 500 mm greater than that for the lower elevation watershed (WS2), and differences become even greater at higher amounts of precipitation. Greater streamflow per unit precipitation at the higher elevation WS36 is related to a combination of factors that reduce ET, including a shorter growing season, differences in species composition, and indirectly steep slopes and shallow soils.

Runoff coefficient analysis shows a clear upward then downward trend over time (Figure 5), which is most well defined in WS02. Simple spline curves were used for graphical display of the trends in data, which will be more completely analyzed in further study. Trend lines for both watersheds suggest decreases in the fraction of precipitation that ends up as streamflow, and hence increases in ET, over time. Our data show an increase up to the mid 1970s, followed by a leveling off or slight decrease thereafter. Decline in the runoff coefficient for WS02 is greater than that for WS36. The declining trend seems to begin in both watersheds around 1980 and may coincide with a drought that occurred at that time. More research is needed to determine the causes of this declining ratio.

Temporal variation in RO/P suggests that either biological or physical factors are changing the rainfall-runoff relationship in both WS2 and WS36. We suggest that most of this variation is due to changes in climatic driving variables and (or) structural and functional attributes that determine ET. For example, data from long-term vegetation plots indicate changes in species composition (Elliott and Vose 2011), with subsequent effects in transpiration (Ford et al., in press). Due to the nature of reference watersheds, the runoff coefficient

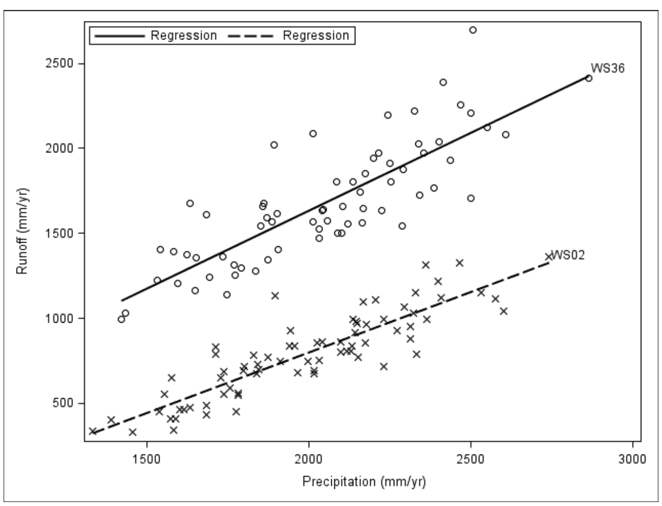

Figure 4. Precipitation versus streamflow for watersheds 36 and 2.

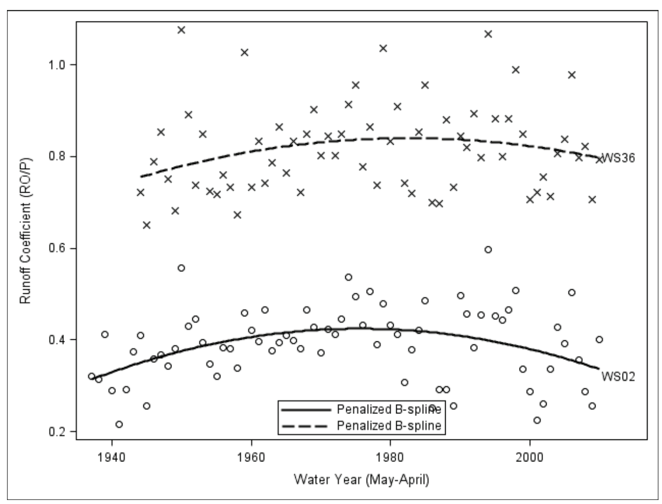

Figure 5. Runoff coefficient (RO/P) over time with penalized B-spline curves to show trends over time. RO/P increases through the first part of the period of record until about 1980, when a decrease is shown for both watersheds.

is a variation of ET adjusted for P. In an altered watershed where roads, compaction, altered flow paths and other interference factor in, the runoff coefficient then represents much more than ET.

Management Implications

In each of the management scenarios, management significantly altered the expected level of streamflow (Figure 6). All watersheds showed significant declines in streamflow excesses over time compared to what would have been expected, with most managed watersheds returning to near expected levels of streamflow within a decade.

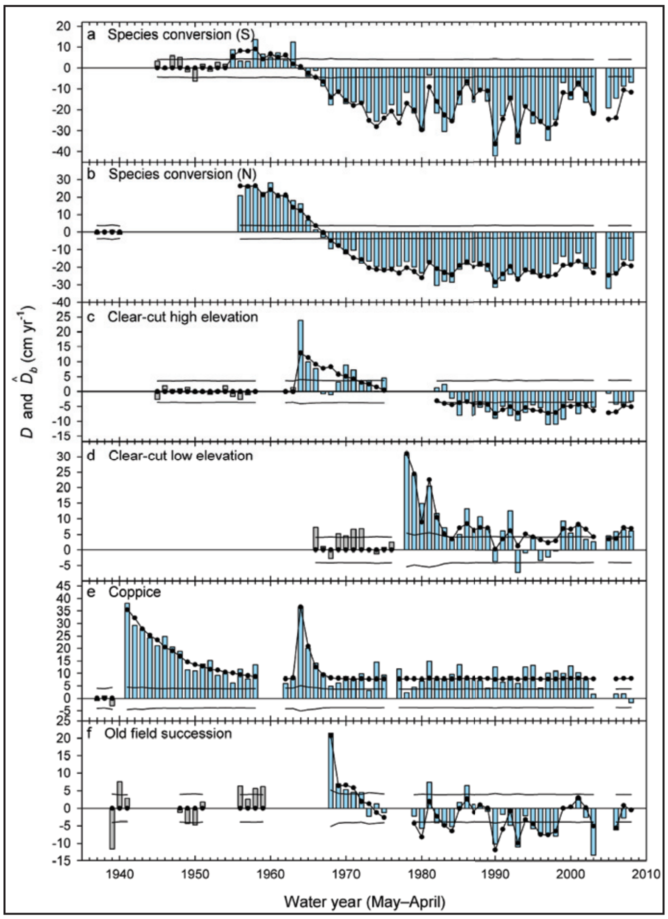

Figure 6. Streamflow response to six management treatments. Bars show observed streamflow excesses (*D*) during pretreatment (grey) and posttreatment (cyan) years. Solid lines show bounds of prediction interval. Filled symbols and lines show modeled streamflow responses ($\hat{D}_b$). From Ford et al. (in press).

Some management options, such as species conversion to pine, created persistently lower levels of streamflow than expected following canopy closure until the end of the record. A coppice forest management strategy allowed for a long-term higher streamflow than the expected levels. Other management strategies eventually returned streamflow to near those expected.

Conclusions

Long-term climatic records indicate warming and increased variability in annual precipitation over the past three decades. The combination of reference and managed watersheds provided a unique opportunity to examine streamflow responses to this variation and examine interactions between management activities and climate.

Precipitation explained a significant portion of the variation in streamflow response for the control watersheds. Runoff coefficients initially increased then declined over time, suggesting corresponding changes in ET over time. The change from an increasing trend to a decreasing trend with time coincided with drought increases in the 1980s and increasing temperature in the late 1970s.

Different forest management strategies could potentially mitigate or exacerbate effects associated with climate change. Forest management affects the vegetation structure and function of the watershed. Streamflow responses depended on the management treatment, and they could be used to mitigate climate change effects. Looking purely at water quantity shows forest management can mitigate for extreme precipitation events in a changing climate. However, these changes should be taken in context with other factors such as carbon sequestration, local climate, and water quality.

Long-term data such as those recorded at Coweeta Hydrologic Laboratory show the trends over time that can sometimes be difficult to resolve in shorter temporal datasets. When managing a forest over the long term, corresponding data collected over the time period of management is key to understanding the full scope of forest development on water resources.

Acknowledgments

This research was supported by the U.S. Department of Agriculture Forest Service, Southern Research Station and by National Science Foundation grant DEB0218001 to the Coweeta Long-Term Ecological Research program at the University of Georgia. We thank Ge Sun and Tony Crump for helpful review/comments.

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Effects of Forest Cover and Environmental Variables on Snow Accumulation and Melt

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Abstract

The goal of this study was to assess the effects of topography and forest cover resulting from different treatments on snow accumulation and melt in small watersheds in the western United States. A paired-watershed study was implemented at the Priest River Experimental Forest, ID, where ten small watersheds with an average area of 4.5 ha were treated by (1) thinning with mastication, (2) burning, (3) prescribed fire with salvage logging, (4) thinning with a prescribed fire, and (5) control (no treatment). At each watershed, a 30-m sampling grid was established for a total of 383 measurement locations. At each location, snow depth was measured between 2004 and 2010 from early February through April to characterize conditions near peak snow and during the midmelt phase. A total of 70 snow density measurements were made at several randomly selected points within each watershed and were used to determine the snow water equivalent. Forest canopy cover at each measurement location following the treatments was measured, and specific topographic variables (elevation, slope, aspect, and curvature) were derived. Correlations between forest cover, snow accumulation, snowmelt, topographic, and meteorological variables were obtained to determine the effects of these variables on snow accumulation and melt.

Keywords: snow accumulation, snowmelt, forest cover, topography, small watersheds

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Introduction

In arid and semiarid landscapes, such as the western United States, mountain snowpack represents a seasonal water storage reservoir that is the primary source of streamflow during the melt season. Both snow accumulation and melt processes are influenced by the interactions among topography, climate, and forest cover. A good understanding of these interactions is a challenge for forest hydrologists and water resource managers. Ideally, we need to manage watersheds to accumulate sufficient snowpack during the winter and slowly melt it when temperatures increase, yielding a long-duration runoff hydrograph with a low peak flow. Most managed forest watersheds receive different treatments. As a result, the snowpack dynamics within the watersheds change. As topography is constant, differences in spring runoff characteristics are mainly due to differences in the climate and forest cover. Climate is not constant and cannot be controlled; therefore, we need to understand the effects of forest management on forest cover and winter processes if we are to optimize streamflows in high-elevation forests in the western United States and elsewhere.

Currently, the processes that control snow accumulation and melt are well known, yet their interactions across varied terrain present modeling and prediction difficulties. Variables like elevation, aspect, slope, temperature, precipitation, solar radiation, relative humidity, wind speed, and canopy are known to influence snow accumulation and melt (Marks 1998, Anderton 2004, Watson 2006, Varhola 2010), but more studies are needed to determine interactions among these variables and how they change for different locations. Changes in elevation, for example, are often associated with changes in climate. Similarly, changes in canopy can change the wind patterns within the watershed and the snow cover energy balance under the canopy (Marks et al. 1998).

Varhola et al. (2010) reviewed 33 different studies from North America and Europe. Using forest cover alone as an independent variable, the authors were able to find significant relations between changes in forest cover and changes in snow accumulation and melt, yet no prediction models could be developed. The authors stated the need for studies that would measure multiple sources of variability as well as the meteorological conditions during the measurements.

In our study we evaluated a total of 33 topographic and climatic variables, including snow depth, snow water equivalent (SWE), snowmelt, and forest cover. Our main objectives were to observe (1) differences in treatment effects on SWE and snowmelt, and (2) correlations of SWE and snowmelt with 31 variables. Our study was conducted in watersheds with an elevation range of 800–1200 m using a multitude of variables in contrast to most of the current studies conducted at higher elevations or alpine regions. We anticipate our findings will contribute much needed information to the knowledge base regarding snow accumulation and melt processes.

Methods

Study Site

This research was located in ten small watersheds at the Priest River Experimental Forest (PREF), ID (48°21′24 N., 116°48′26 W; Figure 1). The watersheds have an average area of 4.5 ha and range in elevation from 843 to 1,236 m. The average slope of the watersheds is 29 percent and increases in elevation from watershed 1 to 10. Within this elevation range, up to around 1,000 m, the soils are shallow Saltese with deeper Jughandle soils only on drainage concave areas. Above 1,000 m the Jughandle soils are predominant and are characterized by a zone of maximum soil water and temperature effectiveness (McConnell 1965). The watersheds were intentionally chosen for their predominant southern aspects, and gneiss is the predominant bedrock type.

Climate

The climate is transitional between northern Pacific maritime and continental types. Average annual maximum and minimum temperatures are 14°C and 0°C, respectively, and precipitation is 805 mm based on the 2004–2010 average values measured at the PREF weather station. Snowfall depths at the same weather station were approximated into snow water equivalent values using a snow density of 100 kg/m^3. Twenty-two percent of the total average precipitation for 2004–2010 was in form of snow. The years with the largest amount of snow were 2008 (47 percent of annual precipitation) and 2007 (33 percent), while the driest years were 2005 (13 percent) and 2010 (13 percent). Year 2005 was discarded from our analysis as the snowpack did not accumulate enough to conduct the field measurements. At the study site snowfall accumulates from December through February and may persist on the ground until late April or even May, depending on the amount of precipitation received during the snow season.

Field Methods

A paired-watershed sampling design was used in this study where each pair received one or two of the following treatments in the years noted: (1) thinned (2007)/mastication (2008), (2) burned (2006), (3) burned (2006)/salvage (2007), (4) thinned (2007)/burned (2008), and (5) controlled. For simplicity, we will refer to the treatments as 1, 2, 3, 4, and 5 hereafter. The treatments were not random; they were prescriptive in that those plots that were unlikely to experience wildfire and did not require thinning were controls, whereas those plots with the greatest fire risk were selected for the burning treatments. The remaining plots were assigned the thinning treatment. The ten watersheds were instrumented in summer of 2003 and monitored for seven years starting in 2004. The simulated wildfire treatments were applied in October 2006, the salvage and thinning in October 2007, and the post-thinning prescribed fire in October 2008.

For each watershed, a sampling grid with a 30-m spacing was applied (Figure 1). The number of points within each watershed ranged between 19 and 62 depending on the size of the watershed. In total, there were 383 sampling locations that were marked with a stake. Between 2004 and 2010, from early February through April or May, snow depth was measured at each sampling location using a 1-m marked iron rod to characterize conditions near peak snow and during the midmelt phase. A total of 70 snow density measurements were made using a 1-m marked tube at randomly selected points within each watershed.

An average snow density was obtained for each watershed on each sampling day and multiplied by the snow depth at each location to estimate SWE.

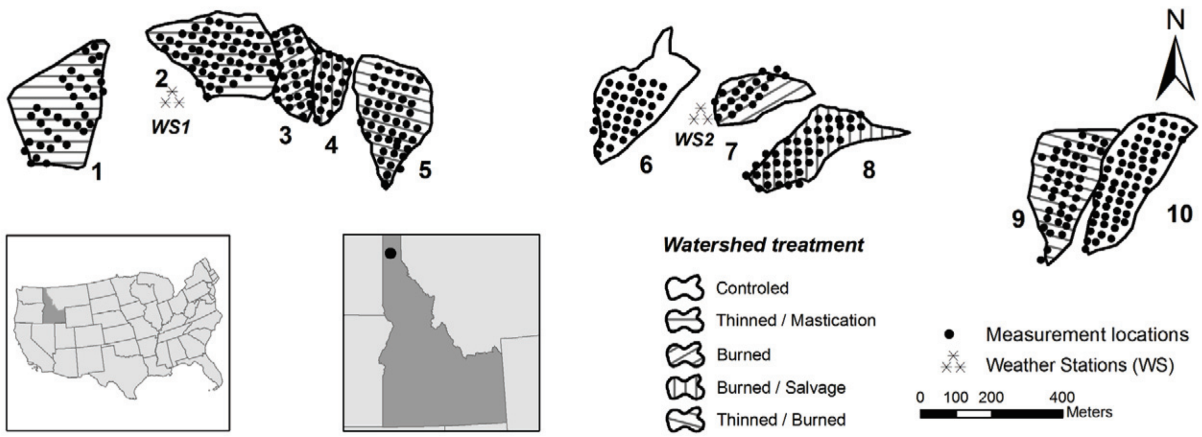

Figure 1. Location of the study site and sampling scheme.

The sampling was initially twice per month, distributed throughout the winter season until the snowpack was completely melted. It took 2–3 days to complete a sampling event. On three occasions half of the watersheds were measured at the end of the month while the other half were measured at the beginning of the next month. In such situations, for analysis, we considered all watersheds as being surveyed at the end of the month.

Snowmelt rates were calculated as the difference in SWE divided by the number of days between two consecutive measurements.

The snowpack melted by the beginning of March in all years except 2008, when it lasted until May. To evaluate the means of SWE and snowmelt for the whole snow season (February–May) in all years, we averaged the SWE and snowmelt values across the four months, considering a value of 0 for the months without snow.
Forest canopy cover at each measurement location following the treatments was measured using airborne laser mapping technology, also referred to as Light Detection And Ranging (LiDAR). Specific topographic variables (elevation, slope, aspect, and curvature) were derived from a 10-m digital elevation model (DEM) for the same locations. Two hydrometeorological stations were installed in the vicinity of watersheds 2 and 7 (Figure 1), and continuous climatic variables were measured for the entire period of the study. Weather station 1 was situated at a lower elevation and the climatic variables at this location were assigned to the first five watersheds, while weather station 2 was at a higher elevation and the climatic variables here were assigned to the last five watersheds.

In total, we used 33 variables: SWE; snowmelt; site; year; month; Julian date; treatment; elevation; aspect; slope; curvature; plan curvature; profile curvature; solar radiation derived from DEM; percent canopy cover; snow depth; average, maximum, and minimum temperatures; average, maximum, and minimum relative humidity; solar radiation measured at the two nearby weather stations; average, maximum, and minimum wind speed; wind direction; average soil water content; precipitation; degree day (number of days with temperature above 0°C between two consecutive measurements); average temperature between two consecutive measurements; actual vapor pressure; and dew point temperature.

Data Analysis

Descriptive Statistics

We were interested in observing the effects of the treatments on SWE both by year and by month. The effects, however, were not obvious as SWE and snowmelt are influenced not only by treatments but also by topographic and climatic variables. Another reason is that vegetation tends to recover in the years following the treatments, and therefore observing the effects of the treatments is difficult in a single study as the weather effects in the first year after treatment will be greater than in the subsequent recovery years.

Proc Mixed Procedure (1999, SAS) was used to assess the least squares means (LS means) and the differences between LS means between treatments across all years and months at a significance level of $\alpha = 0.05$.

Correlation Analyses

The statistical analyses were conducted with SAS 9.1 at $\alpha = 0.05$ (1999, SAS). The Spearman correlation coefficient was used to assess the correlations between SWE and snowmelt with 33 variables. SWE and snowmelt were the response variables while the topographic, climatic, temporal, and canopy cover variables were considered independent variables.

Results and Discussion

Descriptive Statistics

Before any treatments, we observed more snow in the control plots, and this trend continued throughout the study. We observed a greater decrease in the amount of SWE for treatments 1 and 2 compared to the other watersheds for year 2006 (Figure 2). There was no statistically significant difference between the burned watersheds in 2006 and the watersheds in 2007, following the harvesting and thinning treatments.

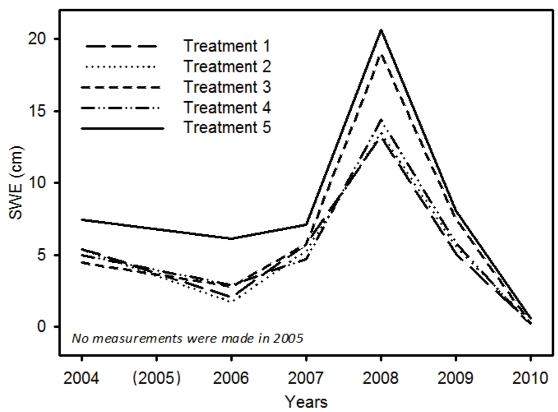

Figure 2. Differences in yearly SWE across all treatments.

Similarly, all the treatments had less SWE compared to the control in year 2008, except treatment 3 which, following the burn in 2006 and the salvage in 2007, had a similar SWE to the control treatment, although they differed statistically.

The difference is smaller in the other years, with generally less amounts of SWE on all treatments. Treatment 3, burn followed by salvage, resulted in a vegetation condition similar to the control watersheds, which, in fact, was the silvicultural goal for the prescription. The other treatments appeared to have little effect on SWE. The difference in SWE is greater between years than among treatments within any given year, likely as a result of variability in weather.

The slope of the curve SWE versus month in Figure 3 reveals the rate of snowmelt. The snowmelt rates changed with time similarly as SWE. Treatments 1, 2, and 4 experienced a similar trend in the snowmelt rate from February until May. For treatments 3 and 5, melt was slower from February to March but faster from March to April. Both treatments 3 and 5 statistically differed from the other treatments in terms of snowmelt rate and from each other.

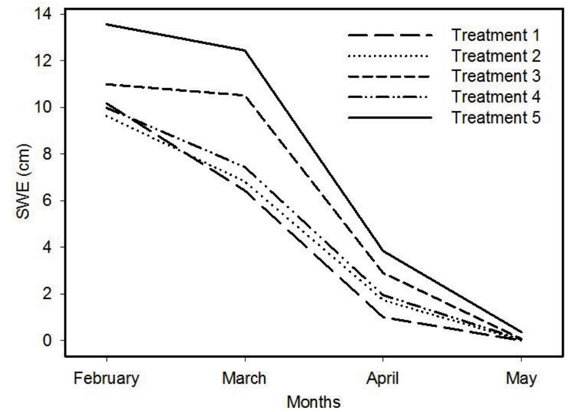

Figure 3. Differences in monthly SWE across all treatments.

Correlation between SWE (snowmelt) and Other Variables

Temporal, weather, and vegetation variables used in this study were correlated with both SWE and snowmelt, but topographic variables were not. We expected topographic variables to have an influence on snow processes. Varhola et al. (2010) submit that slope, together with aspect, should have an influence on snowmelt rates, although Anderton et al. (2004) found no significant terrain control, including slope, in an alpine study. The reason is that the majority of the variables were climatic, representing the main variables influencing both SWE and snowmelt. As shown in other studies (Watson et al. 2006, Jost et al. 2007), elevation and canopy should be strongly related to snow accumulation and melt as we originally expected. Yet we found both variables correlated with SWE for only the years after the treatments were applied, and among the two, only elevation influenced the snowmelt. Our study was focused on ten small watersheds with approximately 400-m difference in elevation, and substantial differences in snow accumulation and melt did not occur during our study. In regard to the canopy cover, most studies assessed effect of this variable under two conditions: open versus forested areas (Jost et al. 2007). In our study, we used a range of canopy amounts. Since the variation in SWE was high, similar to findings in previous studies (Jost et al. 2007), the correlation between SWE and canopy cover was also low.

We further evaluated the correlations separated for each year and month. Among the selected correlations, we present only those that were significantly different from the correlations obtained using the entire dataset. Unless specified otherwise, all the correlations presented were for $p < 0.0001$.

SWE

The strongest correlations were between SWE and month before ($r_s = -0.82$) and after the treatments ($r_s = -0.67$), suggesting that as month increases, SWE decreases. No correlations were observed for SWE versus elevation before treatments and for the year 2010. All the other years had correlation coefficients between 0.05 and 0.12 for 2007 and 2009, and 0.30 and 0.36 for years 2006 and 2008, respectively. The overall correlation after the treatments was 0.14. The higher correlation with elevation in 2008 suggests that elevation was more important for years with larger amount of snowfall.

Canopy cover had a low correlation with SWE for years 2006 ($r_s = 0.06$, $p = 0.0009$) and 2008 ($r_s = 0.08$, $p = 0.0001$). No correlation was found for all the other years, including the year before the treatments.

The correlations between SWE and wind speed or wind direction varied for the different years. All the correlations with maximum wind speed are negative and range from −0.26 for 2006 to −0.40 for 2008; there are no correlations when considering the dataset with all the years after the treatments. There is a positive correlation before the treatments ($r_s = 0.10$). The correlations were negative for all years except for year 2008 when the correlations with average wind speed ($r_s = 0.23$) and wind direction ($r_s = 0.34$) were positive. A negative correlation means that SWE decreases with increasing wind speed, while a positive correlation means that SWE increases with increasing wind speed. As wind speed was shown to have a great influence on the redistribution of SWE (Luce et al. 1999, Anderton et al. 2004), our results are agreeable with these findings.

There was no correlation between SWE and soil water content for the years before treatments, but there was a positive correlation for the years after the treatments ($r_s = 0.41$). The positive correlation is observed for the individual years 2006 ($r_s = 0.42$), 2007 ($r_s = 0.28$), 2009 ($r_s = 0.55$), and 2010 ($r_s = 0.40$); the correlation was negative for year 2008 ($r_s = -0.29$). The positive correlation suggests an increase in soil water content with increasing snow accumulation. This result may be interpreted as a higher SWE and could imply a warmer soil temperature at the beginning of the snow season, causing fresh snow in contact with soil to melt and increasing the soil water content. The negative correlation for 2008 might be due to the late melting of the snow in late April and early May for this year.

Actual vapor pressure was the only variable correlated with SWE after the treatments ($r_s = -0.27$, p = 0.05), and dew-point temperature was negatively correlated with SWE for the years both before ($r_s = -0.08$, $p = 0.03$) and after ($r_s = -0.25$) the treatments.

Snowmelt

The correlation between snowmelt and solar radiation increased from 0.26 before the treatments to 0.42 after the treatments, with a maximum for the year 2008 ($r_s = 0.45$). The stronger correlation after the treatments is possibly due to an increase in the open areas within the forests.

Average, maximum, and minimum temperatures were in general positively correlated with snowmelt after the treatments and for each year. Among the various forms of relative humidity, average relative humidity had the strongest correlation after the treatments ($r_s = -0.58$) compared to before the treatments ($r_s = -0.34$).

There was no correlation between soil water content and snowmelt rates before the treatments, but there was a negative correlation after the treatments ($r_s = -0.16$). The highest correlation between snowmelt and soil water content was for the year 2009 ($r_s = -0.68$), and there is no correlation for 2006. The highest monthly correlations were for February ($r_s = -0.22$) and March ($r_s = -0.42$) and the lowest for April ($r_s = -0.07$).

Conclusions

We were able to observe the effects of the treatments on SWE and snowmelt, although in the third year following the treatments, the effects were less pronounced for normal years, suggesting a high variability of both SWE and snowmelt with years but less with treatments. Greater differences in the rates of snow accumulation and melt among the five treatments were noticed for the more extreme years, such as 2008. With the exception of treatments 3 and 5, there was no clear difference among the means of SWE and snowmelt for the other three treatments. Although the treatments can increase the SWE by creating more open areas within the canopy, this increase can be offset by the increased wind speed and solar radiation, which in turn can result in higher sublimation rates

(Woods et al. 2006, Varhola et al. 2010). No difference was observed in the snowmelt rates in any of the treatments in the two years after the disturbance. As in the case of snow accumulation, the differences in the rates of snowmelt were substantial only in 2008, again underlying the importance of climate in these watersheds.

The large number of correlations in our study is another indicator of the complex interactions among the climatic and topographic variables in forests. At high elevations, as in alpine regions, and in areas with large variations in elevation, the variables responsible for snow accumulation and ablation might be easier to detect. In contrast, these interactions are more pronounced in lower elevation areas or in areas with insignificant variations in elevation.

The strongest correlations for SWE were between month, soil water content, temperature, relative humidity, and actual vapor pressure, and for snowmelt they were between month, relative humidity, temperature, solar radiation, average temperature between two consecutive measurements, and wind speed.

Although our study shows differences in SWE and snowmelt due to treatments, these differences are minor as the changes in canopy structure were not dramatic. This can be seen especially for treatments 1, 2, and 4.

The results presented in this paper are preliminary and exploratory. A more in-depth statistical analysis is needed and will be conducted to better understand the relations of SWE and snowmelt with the other environmental and physical variables as well as years used in this study.

Acknowledgments

We thank the two reviewers Dr. Pete Robichaud and Brandon Glaza for their valuable comments and suggestions. We also thank Brandon Glaza for the watersheds instrumentation and data collection during the first five years of project.

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Headwater Variability across the Rain-Snow Transition in California's Sierra Nevada: Stream Discharge, Runoff Timing, and Sediment Yield

C.T. Hunsaker

Abstract

The hydrologic response of eight headwater catchments located at and above the rain-snow transition, 1,500–2,500 m elevation, was investigated over five years (2004–2008) using hourly streamflow, precipitation, snowpack, and weather-station data. The Kings River Experimental Watersheds (KREW) is a watershed-level integrated ecosystem project for long-term research on headwater streams in the southern Sierra Nevada. It is designed to evaluate forest restoration treatments—mechanical thinning and understory burning—and addresses knowledge gaps identified in the Forest Service's adaptive management strategy for the Pacific Southwest Region. KREW also is ideally suited to address climate change because of its location; rain-dominated lower elevation watersheds, that also receive snow, provide a surrogate for how snow-dominated, higher elevation watersheds, could function with climate change.

The annual runoff ratio (discharge divided by precipitation) in eight headwater catchments located across the rain-snow transition increased about 0.1 per 300 m of elevation. Higher elevations have lower vegetation density, coarser soils, and a shorter growing season when compared with lower elevation lands. Average temperature across the 600-m average elevation range was only 1 to 2^{o}C warmer in the lower versus upper elevation catchments, with annual precipitation being 20–50 percent snow at the lowest elevations versus 75–95 percent at the highest. Peak discharge lagged peak snow accumulation on the order of 60 days at the higher elevations and 20 to 30 days at the lower elevations. Snowmelt dominated the diel streamflow cycle over a period of about 30 days in higher elevation catchments, followed by a 15-day transition to evapotranspiration dominating the diel streamflow cycle. Discharge from lower elevation catchments was rainfall dominated in spring, with the transition to evapotranspiration dominance being less distinct. Base flows ranged from <1 to 10 liters per second (L/s), but during spring snowmelt flows of 400 L/s occur for a month or more. Maximum peak flows of more than 1,000 L/s were measured for a single rain event. Annual sediment yield varied from year to year. One of the managed watersheds in the rain-snow transition zone produced 1.8, 15.2, and 18.7 kg/ha/yr for water years 2004, 2005, and 2006, respectively. The increase in sediment accumulation correlated with an increase in annual precipitation. The snow-dominated and managed watersheds and the one undisturbed watershed produced similar, and sometimes higher, sediment loads for these same years. This is an interesting finding since snow-dominated areas are expected to produce less sediment without rain-drop and rapid surface runoff erosion.

Climate warming that results in a longer growing season and a shift from snow to rain would result in earlier runoff and lower water yield. KREW measurements indicate that about one-third less water could flow in the streams from high elevation headwaters in the southern Sierra as temperatures increase 1–2°C. Sustaining or enhancing water yields in these mixed conifer forests will require management of evapotranspiration by reducing vegetation density and using thinning patterns that attenuate snowmelt. KREW hosts the National Science Foundation's Southern Sierra Critical Zone Observatory and is designated for inclusion in their National Ecological Observatory Network.

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Contrasts in Carbon and Nitrogen Ecosystem Budgets in Adjacent Norway Spruce and Appalachian Hardwood Watersheds in the Fernow Experimental Forest, West Virginia

Charlene Kelly, Stephen Schoenholtz, Mary Beth Adams

Abstract

We constructed watershed mass-balance budgets of carbon (C) and nitrogen (N) and measured seasonal net N mineralization in an attempt to account for nearly 40 years of large discrepancies in stream NO_3-N export in two adjacent, gauged watersheds at the U.S. Department of Agriculture Forest Service's Fernow Experimental Forest, WV. These watersheds have similar management histories, varying primarily by vegetation cover, where one watershed is a monoculture of Norway spruce (*Picea abies*) and the other has regenerated to native Appalachian hardwoods. Long-term stream chemistry indicates that the hardwood watershed has approached N-saturation, with relatively high stream export of nitrate-N (15 kg NO_3^--N/ha/yr), whereas the spruce watershed exports virtually no nitrate-N. We estimated the pool size of C and N within the mineral soil, forest floor, litter, above- and below-ground tree biomass, and stream dissolved organic N. We were unable to account for long-term differences in NO_3-N export via streamflow by estimating these pools. Total C and N pools were 28 percent and 35 percent lower in the spruce watershed, respectively. Though historic organic C and N were never measured in the long-term stream chemistry, the discrepancy in C and N budgets between the two watersheds suggests that the spruce watershed may have been subjected to a period of large losses of C and organic N from deeper subsurface soils. Such large losses suggest that species conversion has the potential to significantly alter ecosystem C and N budgets, with implications for long-term productivity, C sequestration, and water quality.

Keywords: Fernow Experimental Forest, Norway spruce, carbon, nitrogen

Introduction

Stream chemistry at watershed outlet weirs integrates ecosystem functions (chemical, biological, and physical) and displays responses of the total watershed to alteration. Studies of nutrient mass-balance budgets have been utilized to account for differences in stream chemistry (e.g., lower stream nitrate export) and to identify effects of management regimes on ecosystem processes influencing such differences (e.g., Triska et al. 1984). Conversion of native vegetation to monocultures of conifer may disrupt biogeochemical cycling of carbon (C) and nitrogen (N) (Guo and Gifford 2002).

Two adjacent watersheds within the U.S. Department of Agriculture (USDA) Forest Service's Fernow Experimental Forest (FEF) in West Virginia provide an excellent opportunity to investigate the specific role of tree species in ecosystem N cycling and retention. Long-term stream chemistry of these watersheds indicates divergent export of stream NO_3-N at the outlets. Mean annual stream NO_3-N exported from an experimental 39-yr-old hardwood stand (watershed 7) is nearly 15 kg/ha, whereas stream NO_3-N exported from a nearby experimental 37-yr-old Norway spruce stand (watershed 6) has been nearly zero for 20 years (mean = 0.18 kg/ha/yr).

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The present work is an attempt to account for nearly 40 years of large discrepancies in stream NO_3-N export in two nearly adjacent, gauged watersheds at the FEF using estimates and comparisons of key components of ecosystem C and N budgets. It was hypothesized that because NO_3-N export has been negligible from the spruce watershed, and because inputs to the two watersheds from atmospheric deposition are equal, then C and N pools will have accumulated to a greater extent in vegetation, forest floor, and soil horizons in the spruce watershed because of slower decomposition of organic material and slower nutrient cycling, resulting in low NO_3-N export to the stream.

Specific objectives were to (1) measure selected pool sizes of C and N within each watershed to ascertain if significant differences in these pools occur after nearly 40 years of influences from contrasting forest vegetation and (2) measure rates of net N mineralization to determine if this measure of current N flux was associated with differences in size of selected C and N pools in the two watersheds.

Methods

Description of the Watersheds

The watersheds used in this study are located within the USDA Forest Service FEF near Parsons, WV (USA). See Kochenderfer (2006) and Kelly (2010) for complete site descriptions. Both watersheds 6 and 7 (WS6 and WS7) were clearcut-logged in sections (1964–1967) and maintained barren with herbicides until 1969. Watershed 6 (22 ha) was planted with Norway spruce in 1973, but WS7 (24 ha) was allowed to regenerate naturally beginning in 1970. After nearly 40 years of growth, the spruce has a closed canopy and dense stand structure with mean basal area of 23 m^2/ha.

Soils in WS6 are mapped as Calvin series (Soil Survey staff, USDA Natural Resources Conservation Service) derived from shale, siltstone, and sandstone parent material. Soils in WS7 are mapped as both Calvin and Dekalb series derived from acidic sandstone parent material. This watershed is dominated by yellow-poplar, red oak, and sugar maple, with mean basal of 17 m^2/ha.

Historic NO_3-N export and specific conductivity data from the spruce and hardwood streams indicate close similarity in ecosystem biogeochemical activity at the time of conversion to a Norway spruce stand (Kelly 2010).

Atmospheric Deposition and Stream Export

Data for wet and dry deposition of NO_3-N were attained from annual records from the National Atmospheric Deposition Program monitoring site WV18 and the CASTnet monitoring site PAR107. Streamflow and weekly NO_3-N concentration data were attained from the USDA Forest Service Timber and Watershed Lab, Parsons, WV. These data were used to calculate total inputs and exports of NO_3-N from the two watersheds for 1973–2009. Total dissolved N was analyzed from 2007–2009 in monthly stream samples from both watersheds in order to determine export values of dissolved organic N (DON) that had not previously been measured in these watersheds.

To determine the potential flux of inorganic N resulting from N mineralization, intact soil cores (0–10 cm) were collected at 12 sampling sites in each watershed. Incubations were performed seasonally for two years to determine temporal differences in net ammonification, nitrification, or immobilization.

Soil C and N Pools

To estimate C and N pool sizes within surface soil horizons and the forest floor, soil samples were collected in July 2007 from 30 sampling sites within each watershed at each horizon. Soil horizons are defined in these watersheds as A (0–10 cm) and B (10–46 cm).

Forest floor samples were collected in October 2007 and 2009 to characterize depth, oven-dry weight, and total C and N at each sampling location. Freshly deposited litter materials were collected monthly from 2008 to 2009.

Biomass C and N Pools

Diameters of trees were measured by the USDA Forest Service in 2003. Each individual tree was converted to biomass (kg) using allometric equations for both above- and below-ground (see Kelly 2010 for description of equations). From total above- and below-ground biomass estimates, tree compartment mass and percent C and N were estimated using values published by Whittaker et al. (1974) for the hardwood watershed and by Feng et al. (2008) for the spruce watershed.

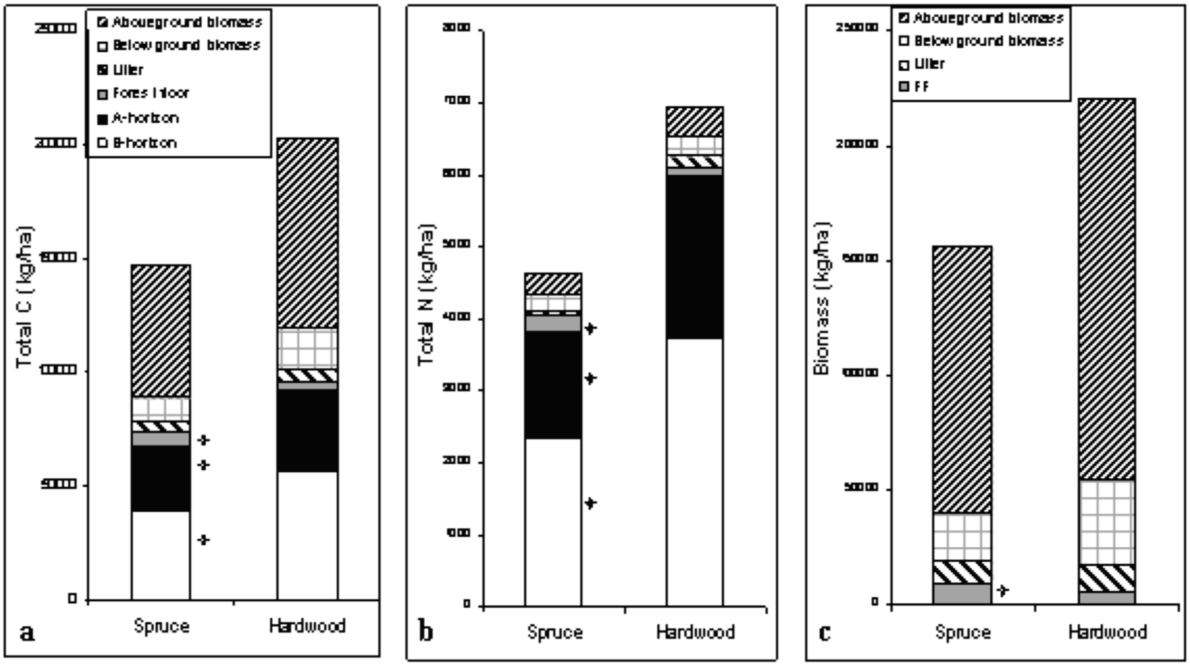

Figure 1. Watershed budgets depicting the mass of (A) carbon, C; (B) nitrogen, N; and (C) biomass contained within each soil and biomass compartment in the spruce (WS6) and hardwood (WS7) watersheds in the FEF. Asteriks denotes significant differences between spruce and hardwood pools at the α=0.05 level.

Data Analysis

One-way analysis of variance (ANOVA) was performed to determine differences in C and N pool size within each compartment by watershed and in in situ net nitrification and net ammonification annually and within each season by watershed.

Results

Inputs and Exports

Since 1973, combined wet and dry atmospheric deposition of NO_3-N is estimated to be 320 kg N/ha for each of the two watersheds. Total stream export of NO_3-N from the spruce watershed since 1973 is 45.93 kg N/ha, which is only 13 percent of the stream export of NO_3-N that occurred during the same time period in the hardwood watershed, which exported 341.07 kg NO_3-N/ha, exceeding the value of atmospheric deposition by approximately 21 kg N/ha. Annual DON export from the spruce watershed was also very low (mean = 0.477 kg DON/ha/yr) during the three years of measurement. In contrast, annual DON export (mean = 10.03 kg DON/ha/yr) from the hardwood watershed was nearly equal to export of NO_3-N (mean = 10.70 kg NO_3-N/ha/yr).

Potential Flux from N Mineralization

Total net N mineralization annual flux was approximately three times greater in the hardwood watershed than in the spruce watershed. Mean annual net N mineralization was approximately 182.50 and 64.0 kg N/ha/yr from the hardwood and spruce soils, respectively.

Soil and Forest Floor C and N Pools

Carbon and N pools within the mineral soil horizons were significantly lower in the spruce watershed relative to the hardwood watershed (Figure 1A). Within the A-horizon, spruce soil contained about 20 percent less C (kg/ha) than the hardwood soil (Figure 1A). Within the B-horizon, spruce soil contained approximately 30 percent less C (kg/ha) than the hardwood soil.

A-horizon soil in the spruce watershed contained nearly 35 percent less N than the hardwood soil (2.03 g N/kg in spruce versus 2.93 g N/kg in hardwood) (Figure 1B). Within the B-horizon pool, spruce soil contained nearly 38 percent less N content (kg/ha) than the hardwood soil (0.73 g N/kg in spruce soil versus 1.17 g N/kg in hardwood soil).

Forest floor C content (kg/ha) was significantly greater in the spruce watershed than the hardwood watershed (3,813 and 2,343 kg C/ha in the spruce and hardwood watersheds, respectively) (Figure 1A). Forest floor N (kg/ha) was significantly greater in the spruce watershed than in the hardwood watershed (132.25 kg N/ha versus 74.68 kg N/ha, respectively) (Figure 1B).

Tree Biomass Pools

Both above- and below-ground biomass in trees, as estimated by allometric equations, were higher in the hardwood watershed (Figure 1C). Above-ground biomass estimates were approximately 30 percent less in the spruce watershed (116,800 kg/ha) relative to the hardwood watershed (166,000 kg/ha). Below-ground biomass estimates were approximately 45 percent less in the spruce watershed (21,000 kg/ha) relative to the hardwood watershed (38,000 kg/ha). This greater biomass in the hardwood watershed equated with larger pool sizes of C and N in the hardwood trees than the spruce (Figure 1A, B).

Discussion

Contrasting Vegetation and C and N Pools

The goal of this study was to quantify selected ecosystem C and N pools in two watersheds that exhibit large differences in long-term stream export of NO_3-N measured since establishment of contrasting forest types in 1973. Results of this study indicate that we were unable to account for these differences in NO_3-N export via streamflow through estimation of the size of C and N pools within the forest floor, mineral soils, above-ground tree biomass, and below-ground tree root biomass in the two watersheds.

Total C and N pools were lower in the spruce watershed in nearly every compartment measured (Figure 1), as was total N mineralization. Total C pools were 28 percent less in the spruce and total N pools were 35 percent less in the spruce relative to the hardwood watershed. The B-horizon soil compartment exhibited the largest difference in both C and N stores (32 percent less C and 38 percent less N in the spruce watershed). These results were contrary to the hypothesis that soil and forest floor C and N stores would be higher in the spruce watershed, thereby accounting for 40 years of relatively high atmospheric N input and very low stream export of NO_3-N from the spruce watershed.

Total N pools in the mineral soil and in tree biomass of two additional watersheds (watersheds 4 and 10) within the FEF are shown in Figure 2. Watersheds 4 and 10 are often used as references because they have been left to natural recovery since being logged in 1905. Comparing the spruce watershed (WS6) to watersheds 4, 7, and 10, which are characterized by native hardwood forests, illustrates that the spruce watershed has considerably less N in the measured pools (Figure 2). It is also noteworthy that the native hardwood watershed of this study (WS7) has similar estimated N pool sizes in soil and biomass as reference watershed 10 (Figure 2).

What Accounts for Differences in C and N Pools?

Three possible mechanisms have been identified that could explain why, after 40 years of much lower NO_3-N export and high atmospheric deposition, the spruce watershed does not exhibit patterns of C and N accumulation comparable to the hardwood watershed.

1. The watersheds are intrinsically different, and the spruce watershed has always had much smaller storage of C and N than the surrounding watersheds;
2. The spruce watershed has been losing N via denitrification at a much greater rate than the hardwood watershed for the past 40 years; and (or)
3. The spruce watershed underwent a phase of organic matter degradation when the hardwood stand was replaced by the Norway spruce stand, causing a large amount of organic forms of N to be leached from the system (e.g., Guggenberger et al. 1994).

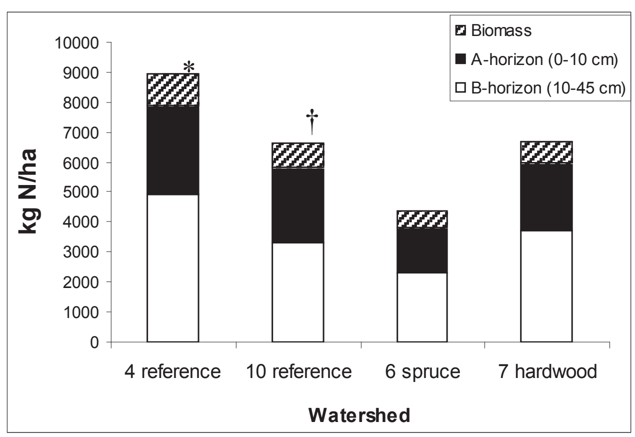

Figure 2. Nitrogen contained within A and B soil horizons and within total biomass (above- and below-ground) for four watersheds within the FEF.
[*] from Adams et al. (2006); sampled in 2002
[†] from Christ et al. (2002); sampled in 1997

It is unlikely that the spruce watershed is intrinsically different from the surrounding hardwood watersheds within the FEF to the degree observed in the present study (Figure 2). The soils within all of these watersheds are of the same soil series and have similar historic land use and atmospheric inputs. Historic NO_3-N export and specific conductivity of streamwater indicate close similarity in ecosystem biogeochemical activity at the time of conversion to a Norway spruce stand. When analyzed by decade after conversion, it can be seen that in the first decade following treatment (1971–1980), patterns of stream NO_3-N values were very similar between the watersheds (R^2=0.96). Furthermore, the divergence in specific conductivity did not occur until after the Norway spruce canopy closure occurred (R^2=0.20 after 1981 and R^2=0.0005 after 1991) (Kelly 2010).

Large losses of NO_3-N were not detected in the long-term stream chemistry data for the spruce watershed, suggesting that N might have been lost through fluxes in gaseous phase of N_2O or NO (Reddy and Patrick 1975) and (or) through stream export of DON (Campbell et al. 2000) (also not measured). It is unlikely that denitrification processes can explain the relatively small N pools in the spruce watershed because (1) large fluxes of denitrification usually result in accumulation of C in the organic horizons of soils, which was not observed in the current study, and (2) N losses via denitrification usually account for a small portion of total ecosystem N (Mohn et al. 2000). Additionally, soils in these watersheds are relatively well-drained upland soils with relatively low potential for significant denitrification.

It is more likely that soils within the spruce watershed underwent a phase of organic matter degradation or mass transport of sediment prior to spruce stand establishment, inducing a loss of C and N that was not detected in the long-term stream data that only measured losses of NO_3-N and that did not measure DON or particulate N. This concept of organic matter degradation or mass transport is strongly supported by a chronosequence study of soil C stocks beneath red spruce (*P. rubens*) forests in northeastern North America (Diochon et al. 2009). Soils within these forests exhibited increasingly smaller stocks of C from 1-, 15-, and 45-yr-old stands, reaching a minimum of approximately 76 Mg C/ha in the soil profile at 45 years. Soil C values in the 45-yr-old red spruce stand were very similar to the Norway spruce soils in the present study, which contain 74 Mg C/ha. Carbon loss (decrease in C concentration and content) from young stands in the Diochon et al. study (2009) was reported to occur through enhanced mineralization of organic compounds (verified with stable C isotopic analysis), especially in the deeper soil horizons. Similar declines in soil C were also observed in spruce forests of similar age by other authors (e.g., Parker et al. 2001, Tremblay et al. 2002), indicating that this rate of C loss from spruce soils is a common phenomenon.

Forest clearing may result in decreases in soil C, but C stores generally recover to original levels after several decades, especially if the stand is allowed to regenerate (Harrison et al. 1995). Loss of soil C upon hardwood conversion to conifer can be attributed to both disturbance and changes in amount and composition of plant material returned to the soil via litter and root turnover (Lugo and Brown 1993). Additionally, the presence of ectomycorrhizal fungi introduced upon vegetation conversion have also been documented to induce a 30 percent soil C depletion within 20 years of establishment of an exotic Radiata pine (*Pinus radiata*) plantation (Chapela et al. 2001).

Spruce vegetation has high lignin content in litter materials and shallow rooting architecture. High lignin content results in larger proportions of soil organic N relative to inorganic forms because of slower decomposition and mineralization (Berg and Theander 1984). Shallow rooting architecture may result in leaching of soil C following decomposition of deep roots of the native hardwood that were present prior to conversion, with little subsequent vegetative uptake or stabilization deep in the spruce soil profile. Thus, the spruce features of slowly decomposable organic matter and shallow rooting may help explain the apparent large mass losses of soil C and N relative to the native hardwood in this watershed study.

Conclusions

Results of this study suggest that a significant loss of C and N from ecosystem pools likely occurred following conversion from native hardwoods to a monoculture of Norway spruce in the FEF. Consequently, species selection should be taken into account when managing forests for future C sequestration, for provision of high-quality water, and for effects of high atmospheric inputs of N, especially when relatively short rotation times are implemented between harvests.

Acknowledgments

The authors appreciate the reviews of Jack Webster and James Burger.

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