Can forest management be used to sustain water-based ecosystem services in the face of climate change?

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Abstract. Forested watersheds, an important provider of ecosystems services related to water supply, can have their structure, function, and resulting streamflow substantially altered by land use and land cover. Using a retrospective analysis and synthesis of long-term climate and streamflow data (75 years) from six watersheds differing in management histories we explored whether streamflow responded differently to variation in annual temperature and extreme precipitation than unmanaged watersheds.

We show significant increases in temperature and the frequency of extreme wet and dry years since the 1980s. Response models explained almost all streamflow variability (adjusted $R^2 > 0.99$). In all cases, changing land use altered streamflow. Observed watershed responses differed significantly in wet and dry extreme years in all but a stand managed as a coppice forest. Converting deciduous stands to pine altered the streamflow response to extreme annual precipitation the most; the apparent frequency of observed extreme wet years decreased on average by sevenfold. This increased soil water storage may reduce flood risk in wet years, but create conditions that could exacerbate drought. Forest management can potentially mitigate extreme annual precipitation associated with climate change; however, offsetting effects suggest the need for spatially explicit analyses of risk and vulnerability.

Key words: climate; Coweeta basin, southern Appalachians, USA; drought risk; forest management; land use; paired watershed; precipitation; streamflow; warming; water supply.

INTRODUCTION

Forested watersheds are an especially important provider of ecosystem services related to domestic water supply (cf. Brown et al. 2008). Both managed and unmanaged forests provide the cleanest and most stable water supplies for drinking, aquatic habitat, and groundwater recharge compared to all other land uses (National Research Council of the National Academies 2008). Because of the combination of biological and physical controls on hydrologic processes, climate change has the potential to impact water resources directly and indirectly (Brian et al. 2004, Sun et al. 2008).

The direct impacts of climate change on water resources will depend on how climate change alters the amount, type (i.e., snow vs. rain), and timing of precipitation and the subsequent effects on baseflow, stormflow, groundwater recharge, and flooding (Karl et al. 2009). In the eastern continental United States, long-term USGS streamflow data suggest that over the past 100 years, the average annual streamflow has increased and is linked to greater precipitation (Karl and Knight 1998, Lins and Slack 1999, IPCC 2007); however, less than two-thirds of atmosphere–ocean general circulation

models (GCMs) can agree on the predicted change in direction of future precipitation for the eastern United States (IPCC 2007). A more certain aspect of future precipitation regimes is that the frequency of the extremes will increase (Fig. 1). Most GCMs predict that, as the climate warms, the frequency of extreme precipitation increases across the globe (O'Gorman and Schneider 2009). Indeed, many regions of the United States have experienced an increased frequency of precipitation extremes over the last 50 years (Easterling et al. 2000*a*, Huntington 2006, IPCC 2007). However, projections of the timing and spatial distribution of extreme precipitation are among the most uncertain aspects of future climate scenarios (Karl et al. 1995, Allen and Ingram 2002). Despite this uncertainty, recent experience with droughts and low flows in many areas of the United States indicate that even small changes in drought severity and frequency will have a major impact on ecosystem services, including drinking-water supplies (Easterling et al. 2000*b*, Luce and Holden 2009).

The indirect impacts of climate change on water resources are related to changes in temperature and atmospheric CO_2, and are likely to manifest as both short- and long-term responses. In the short term, increased temperature has the potential to increase plant water use via transpiration and evaporation, and hence decrease excess precipitation available for streamflow (i.e., change the rainfall runoff ratio) or groundwater recharge. The impacts of temperature may be offset (or
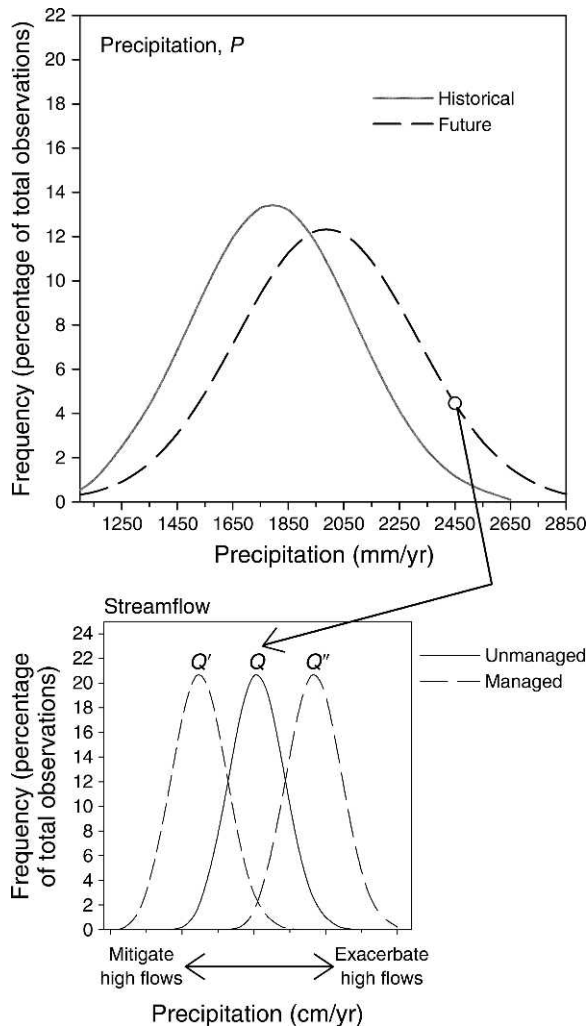
FIG. 1. Hypothetical frequency distribution of future and historical annual precipitation (P; upper panel) and resultant hypothetical mean catchment streamflow distribution (Q; lower panel) for the given precipitation amount. Future hypothetical P distribution indicates increasing frequency of extreme event years. Shifting the resulting mean Q to either Q' or Q'' (shifting solid line to broken lines) represents forest management (managed vs. unmanaged).

exacerbated) by changes in other factors that influence evapotranspiration (ET) such as vapor pressure, wind patterns, increases in CO_2, and changes in net radiation. In the longer term, climate warming will likely produce shifts in species distributions (Iverson et al. 2008). Many tree species differ considerably in the amount of annual and seasonal water use via transpiration and interception (Ford et al. 2010, Sun et al. 2011). For example, in some geographic regions with mild winters, a shift from deciduous to evergreen forests is likely to result in greater forest water use due to greater year-round transpiration and interception. Controlled laboratory and field studies have demonstrated that increased atmospheric CO_2 reduces transpiration in many tree

species that may translate into increased streamflow (Wullschleger and Norby 2001, Ainsworth and Rogers 2007); however, it isn't fully understood if these patterns will persist over the long term.

Forests are unique among land uses because they are long-lived, relatively stable, and respond to climate through ET; yet, their structure and function can be substantially altered by forest management (Vose et al. 2011). These structural and functional changes can be either transient or long term, depending on management intensity, and whether the biological or physical characteristics of the watershed are affected. Biologically, management activities that favor or replace one species (or several species) can alter ET through changes in albedo, canopy roughness, transpiration, and interception. Species vary considerably in the amount of leaf area, transpiration per unit leaf area, and overall whole-tree water use due to differences in rooting depth, tree age, tree height, leaf boundary layer resistance, leaf chemistry, leaf duration, and stomatal sensitivity to vapor pressure deficit (Stoy et al. 2006, Bond et al. 2007, Ford et al. 2010, Oishi et al. 2010). Stand density also influences the amount of water intercepted by and evaporation from surfaces through changes in live and dead leaf, branch, and stem area. Leaf area duration (i.e., evergreen vs. deciduous) influences ET through interception and transpiration, with annual ET being generally higher in evergreen compared to deciduous forests. The ET difference is greater in higher precipitation regimes compared with lower ones (Ford et al. 2010).

Physically, forest management can alter hydrology through activities that create soil disturbances or alter flow paths. While most physical soil disturbances are related to forest removal or cutting (e.g., skid trails, log sorting, and loading areas, etc.), these can be short-lived and have little impact on streamflow over the long term. In contrast, road construction and associated engineering related to road surfacing, drainage, culvert design, and their location are much longer lasting. Depending on design and the surface area impacted, these can permanently alter hydrologic flow paths of forested watersheds.

In this study, we aimed to quantify the interaction among forest management, climate, and streamflow. Because most GCMs disagree on whether the mean precipitation regime will increase or decrease for the southern Appalachians, but agree on the increase in frequency of precipitation extremes (Easterling et al. 2000b, Groisman et al. 2004, Huntington 2006), we focus on the latter instead of the former. Extreme events also present greater challenges to water resource managers than those presented by average conditions. We hypothesized that climate impacts may either be mitigated or exacerbated by forest management practices that alter land cover, depending upon how land cover changes impact hydrologic function. Our objectives were to identify if and by how much land use could affect the streamflow amount for any precipitation amount

(shifting Q to Q' or Q''; Fig. 1) and to identify forest management treatments that would mitigate against extremes in annual precipitation (shifting Q to Q'; Fig. 1) or exacerbate extremes in annual precipitation (shifting Q to Q''; Fig. 1). We used a retrospective analysis of long-term climate and streamflow data to explore whether streamflow from managed watersheds responds differently to variation in air temperature and extremes in annual precipitation than unmanaged reference watersheds. Land managers and policy makers are looking to forests and forest management as options to mitigate climate change (Pacala and Sokolow 2004) and create more resilient ecosystems (Baron et al. 2009). The results from this study provide insights for land use planners on potential responses of managed forests to climate change, and whether forest management can be a viable technique to sustain water-based ecosystem services under climate change.

METHODS

Site description

The Coweeta basin is located in the southern Appalachian Mountains, USA (see Plate 1). Climate is classified as marine, humid temperate (Swift et al. 1988). Precipitation is characterized by frequent, small, low-intensity rainfall events and rare large storms. Average annual rainfall ranges 1800–2300 mm, depending on elevation (Swift et al. 1988). Historic vegetation patterns have been influenced by human activity, primarily through both clear-cut and selective logging, the introduction of chestnut blight (*Cryphonectria parasitica*; Elliott and Hewitt 1997), and fire management (Hertzler 1936, Douglass and Hoover 1988). The resulting unmanaged forests are relatively mature (~85 years old) oak–hickory (at lower elevations) and northern hardwood forests (at higher elevations) with an increasing component of fire-intolerant species (Elliott and Swank 2008).

Climate data and extreme events

Air temperature (T) and precipitation (P) have been recorded at the main climate station (CS01) continuously since 1934 (Swift and Cunningham 1986, Swift et al. 1988). For the long-term record, daily maximum and minimum T are recorded (National Weather Service [NWS] maximum, minimum, and standard thermometer) and then averaged to determine the mean daily T. Automated measurements of T began in 1983; however, they act as a supplement to the manual max min T measurements; for consistency we only display and analyze trends in the long-term record (the NWS maximum, minimum and standard thermometer). These automated measurements occur every five minutes (HMP45C; Vaisala, Helsinki, Finland) and hourly maximum, minimum, and average T are digitally recorded (CR10×; Campbell Scientific, Logan, Utah, USA). P measurement frequency is also every 5 min and includes amount (8-inch [20.32 cm] Standard Rain

Gage; National Weather Service, Indianapolis, Indiana, USA), and volume and intensity (Belfort Universal Recording Rain Gage, Belfort Instrument, Baltimore, Maryland, USA). A network of nine recording rain gages and 12 standard rain gages are also located throughout the basin; six of these have continuous measurements from 1936 to the present. Depending on the analysis, series from three (those co-located on the watershed of interest) to six P gages were used in this study. Although P increases with elevation at the site, P amounts among the rain gages are highly correlated ($0.96 < R^2 < 0.99$).

We tested the hypothesis that mean annual T has been increasing in the recent part of the record by fitting a time series intervention model to T data observed at the site and in the state's climate division that contains the site (NC Div 01; *available online*).[2] Candidate models were a simple level, or a mean level plus a linear increase starting at time t. Each potential starting time in the 1975–1988 range, which was the visual range of the temperature increase, was evaluated sequentially (PROC ARIMA, version 9.1; SAS Institute 2003a). We computed Akaike's information criterion (AIC) for each model, which is a statistic used to evaluate the goodness of fit and parsimony of a candidate model, with smaller AIC values indicating a better fitting and more parsimonious model than larger values (Johnson and Omland 2004). We used the differences in the AIC values among candidate models with all starting times ($\Delta_i = \mathrm{AIC}_i - \mathrm{AIC}_{\min}$) to compute a relative weight (w_i) for each model relative to all models fit:

$$w_i = \frac{e^{-0.5\Delta_i}}{\sum_{r=1}^{R} e^{-0.5\Delta_r}} \quad (1)$$

with the sum of all w_i equal to 1. The final model selected was the model with the highest w_i (Burnham and Anderson 2002, Johnson and Omland 2004).

We used the standard precipitation index (SPI) to identify extreme P years, both wet and dry extremes. The SPI value of a given year represents the probability of a 12-month precipitation amount occurring, based on the entire monthly precipitation record used. Conceptually, the SPI represents a z score, or the number of standard deviations above or below that an event is from the mean. To calculate the SPI, we fit a gamma probability density function to the frequency distribution of monthly precipitation totals from 1934 to 2009 for CS01 using maximum likelihood to estimate the shape (α) and scale (σ) parameters (Swift and Schreuder 1981). Goodness of fit of the Γ distribution was assessed with the Anderson-Darling test statistic (A^2). Fitted parameters were then used to find the cumulative probability of an observed precipitation event for a given year. The cumulative
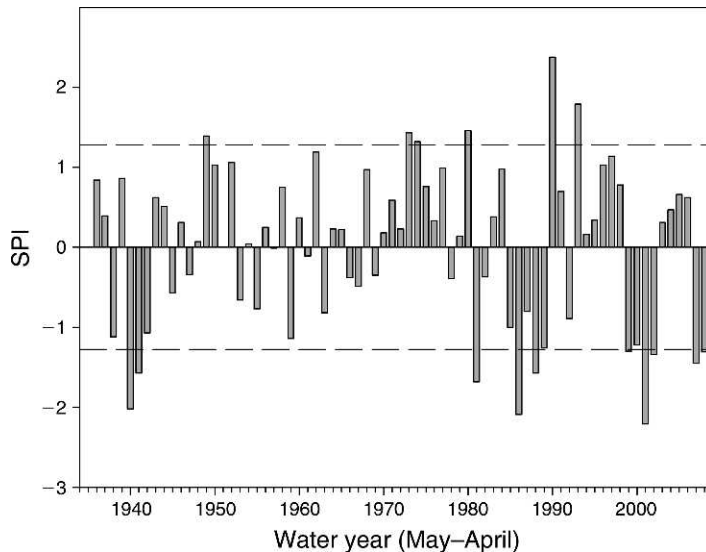
FIG. 2. Standardized precipitation index (SPI) for precipitation totals from a standard rain gauge (SRG19) at the climate station in the valley floor (CS01) in the Coweeta basin, southern Appalachian Mountains, USA. Bars represent the SPI for a 12-month water year. For example, the SPI for 2008 (−1.31) represents the 12-month SPI for the period starting May 2007 and ending April 2008; during this same period total precipitation was 1417 mm. Dashed lines represent the values of SPI that correspond to the 0.10 and 0.90 cumulative probability of the Γ distribution; values beyond dashed lines were considered "extreme" event years with respect to precipitation.

probability was then transformed to the standard normal random variable Z with mean of zero and variance of one, which is the value of the SPI. For precipitation, SPI values in theory are bounded by ±3, with positive values representing wet years and negative values representing dry years. SPI values outside the bounds of ±1.28, corresponding to probabilities of 0.10 and 0.90, were considered extreme (Fig. 2; Guttman 1999).

We tested the hypothesis that the distribution of precipitation is changing over time with a quantile regression approach (Cade and Noon 2003). We analyzed linear trends in all quantiles of P to quantify changes to the annual and monthly P distribution. To represent the basin spatially, we used the entire P series from six active standard gages initiated during 1934–1936. Our model predicted the precipitation amount in each quantile of the distribution from each gage as a function of water year, with elevation as a covariate to account for orographic effects. All models were fit using PROC QUANTREG in SAS (version 9.1; SAS Institute 2005). If the bootstrapped 95% confidence interval around the estimated coefficient for the quantile overlapped zero, we interpreted this as no significant time trend. We examined in detail the predictive models for only the 10th, 25th, 75th, and 90th quantiles.

Paired catchment–forest management studies

We used the long-term streamflow records from six watersheds (WS) that have different management and land use histories (Table 1). Streamflow measurements and rating equations have been described elsewhere (Swift et al. 1988). Management practices on all watersheds were for the primary purpose of elucidating their effects on streamflow as described and summarized by Swank et al. (1988). Our analysis updates the annual water yield responses with 25 years of additional data for each watershed. Two watersheds (WS1 and WS17) were species conversion experiments from southern

Appalachian deciduous forest to evergreen, eastern white pine plantations at 1.8×1.8 m spacing. Two watersheds at high and low elevations (WS37 and WS7, respectively) were clear-cut using different techniques (Table 1). One watershed was subjected to successive clear-cuts (WS13) separated by 23 years, resulting in a multispecies coppice stand in which the vegetation recovered via stump sprouting from existing, well-established rootstock (Leopold et al. 1985). The remaining treatment watershed is an old-field succession following conversion of a mixed-hardwood forest to a grass cover (WS6). All of these management activities are still widely practiced on both publicly and privately managed forests in the eastern United States, and thus represent viable land use strategies.

Modeling the effect of management on annual water yield

The effect of management treatments on annual water yield was determined by using the paired-watershed approach. This method uses the relationship between gauged streamflow from two closely located watersheds similar in size and pre-management cover conditions. In the subsequent treatment period, one watershed served as a reference and remained undisturbed, while a management treatment was applied to the other watershed. Successive observations in time are considered independent replicates. In all cases, streamflow from an unmanaged catchment (Q_C) served as a better model for the basin to be treated (Q_T) than precipitation (P), as inferred from pretreatment regressions of annual totals of Q_C vs. Q_T (mean $R^2 = 0.99$) and P vs. Q_T (mean $R^2 = 0.88$), although the relationship between precipitation and Q_C was highly significant ($P < 0.001$ in all cases) and consistent over time (no precipitation by time interactions, $0.43 < P < 0.87$). All annual periods in our analyses refer to May–April water years, defined as May of the previous calendar year through April of the

TABLE 1. Characteristics of paired watersheds (WS) in the study site in the Coweeta basin, southern Appalachian Mountains, USA.

Characteristic	Treatment	Control	Treatment description
Species conversion (S)			
Watershed number	1	2	Entire watershed prescribed burned in Apr 1942. All riparian trees
Maximum elevation (m)	988	1004	and shrubs (25% of WS area) deadened with chemicals in 1954,
Elevation at weir (m)	705	709	retreated in 1955–1957 growing seasons. All trees cut and burned
Area (ha)	16	12	in 1956–1957; no products removed. Watershed converted to eastern
Aspect	S	SSE	white pine plantation (*Pinus strobus*) in 1957. In subsequent years, *P. strobus* released from hardwood competition by cutting and chemicals as necessary.
Species conversion (N)			
Watershed number	17	18	All woody vegetation cut in Jan–Mar 1941, regrowth cut annually
Maximum elevation (m)	1021	993	thereafter in most years until 1955; no products removed.
Elevation at weir (m)	760	726	Watershed converted to eastern white pine plantation (*Pinus strobus*)
Area (ha)	13	13	in 1956. In subsequent years, *P. strobus* released from hardwood
Aspect	NW	NW	competition by cutting and chemicals as necessary.
High-elevation clear-cut			
Watershed number	37	36	All woody vegetation cut in 1963; no products removed.
Maximum elevation (m)	1592	1542	
Elevation at weir (m)	1033	1021	
Area (ha)	44	49	
Aspect	ENE	ESE	
Low-elevation clear-cut			
Watershed number	7	2	Lower portion of WS grazed by ~6 cattle for 5 month/yr, 1941–1952.
Maximum elevation (m)	1077	1004	Commercially clear-cut and cable-logged in 1977: in Apr–Jun 1976,
Elevation at weir (m)	722	709	three roads occupying 2.5% of total WS area constructed for
Area (ha)	59	12	logging access. Timber cutting and yarding with mobile-cable
Aspect	S	SSE	system Jan 1977–Jun 1978.
Coppice			
Watershed number	13	18	All trees and shrubs on the entire drainage cut in 1939–1940 and
Maximum elevation (m)	912	993	again in late 1962. No trees or shrubs removed after either clear-cut;
Elevation at weir (m)	725	726	soil disturbance was minimal.
Area (ha)	16	13	
Aspect	ENE	NW	
Old-field succession			
Watershed number	6	18	All woody vegetation cut and scattered in 5-m zone vertically above
Maximum elevation (m)	905	993	stream in Jul 1941. Clear-cut in 1958, products removed, remaining
Elevation at weir (m)	696	726	residue piled and burned. Surface soil scarified, watershed planted
Area (ha)	9	13	with grass, limed, and fertilized in 1959, fertilized again in 1965.
Aspect	NW	NW	Grass treated with herbicide in 1966 and 1967; watershed subsequently reverted to successional vegetation.

current calendar year. Only years with complete annual data were used in the analysis.

Our goal was to predict the streamflow response to management given streamflow from an unmanaged catchment, a model of watershed response, and a model of the interaction of the watershed response and precipitation. Our approach was conceptually similar to the classic paired-watershed regression approach (Wilm 1944, 1949). The significant difference was that we only used one equation to model the following: (1) the pretreatment relationship between Q_T and Q_C, (2) the posttreatment relationship between Q_T and Q_C through time as influenced by management, and (3) the posttreatment relationship between Q_T and Q_C through time as influenced by the interaction between climate and management. The classic approach typically uses one equation each to represent 1–2 relationships, and the estimation errors from each equation are not propagated through to the others. The classic approach

can often yield watershed response results interpreted as significant when, in fact, the response is within the pretreatment error. Additionally, studies often suggest that the posttreatment relationship between Q_T and Q_C is indeed influenced by the interaction between climate and management, but have found it challenging to model and quantify this effect on streamflow (see Zhao et al. 2010 and references therein). For our one equation, we fit the following model to Q_T using all data, both prior to treatment or management ($M = 0$) and following treatment ($M = 1$):

$$\hat{Q}_T = bQ_C + \left[Mc \left(h - \frac{1}{1 + \exp^{k1-k2t}} \right) \right]$$

$$+ \left[MdP \left(h - \frac{1}{1 + \exp^{k1-k2t}} \right) \right] \quad (2)$$

which predicts annual streamflow from the managed catchment ($\hat{Q}_T$) from annual Q_C, annual P from the

nearest gage, and a logistic model of watershed vegetation recovery based on time since treatment (t in years, where $t = 0$ during the first year after the treatment is completed), and the interaction of the two latter terms. Coefficients for terms are b, c, and d. In the logistic function, the $k1$ and $k2$ parameters describe the slope and intercept of the decline in streamflow with vegetation recovery through time; and the h parameter sets the level of the function. The two bracketed terms were repeated in the case of multiple treatments during a management cycle, such as for the coppice stand with repeated cuts. If simplifying the logistic or linear parts of the model resulted in a better fit, terms were omitted; for example, omitting $k1$ simplified the logistic to an exponential. We tested for models that examined the importance of watershed vegetation recovery interactions with temperature, but they were never significant and not included in the final Q_T model.

We defined the observed management response, D (cm/yr), as the streamflow deviation from that predicted by the control without management,

$$D = Q_T - (\hat{Q}_T, M = 0) \qquad (3)$$

and the predicted management response was inferred from the partial model residuals,

$$\hat{D}_b = \hat{Q}_T - (\hat{Q}_T, M = 0). \qquad (4)$$

We explored the relationship between the observed vegetation response and precipitation by plotting partial model residuals (observed streamflow from the managed catchment [Q_T] and all the model terms except d) in the latter part of the series:

$$\hat{D}_c = Q_T - (\hat{Q}_T, M = 1, d = 0). \qquad (5)$$

Models were fit and solved numerically using an iterative procedure (PROC NLIN, version 9.1; SAS Institute 2003b). We estimated the percentage of variability explained by the model using the ratios of the error- to total-sum of squares, and total- to error-degrees of freedom:

$$R^2_{\text{adjusted}} = 1 - \frac{\text{SS}_E}{\text{SS}_T} \times \frac{\text{df}_T}{\text{df}_E}. \qquad (6)$$

We interpreted parameter estimates as statistically significant if

$$\Pr\left\{ t = \frac{\text{estimate}}{\text{SE}} > t_{dfE,2} \right\} \leq 0.10.$$

Autocorrelation among the residuals for each year with those from the preceding six years was also calculated and tested for significance as a test of independent observations in time.

Forecasting responses under future climate

We used statistically downscaled, forecasted P data for two scenarios of climate change (warm/wet and hot/dry) along with forecasted $\hat{Q}_C$ data (forecasts based on empirical models) to estimate watershed treatment responses from the end of the observed series (2009) out to the year 2050. Most atmosphere–ocean general circulation models (GCMs) agree that the southern Appalachians will be warmer. However, most GCMs fail to reproduce historical precipitation in our region; because of this, there are large uncertainties in forecasted precipitation from these models. As a result, most of the GCMs do not agree on the direction of change in precipitation, i.e., fewer than two-thirds consensus (IPCC 2007). For example, for the four GCM model runs that we had downscaled and processed for our region, the correlation with past precipitation ranged $-0.09 < R < 0.15$ and was never significant $0.37 < P < 0.85$. We therefore used two data sets (warm/wet and hot/dry) with the intent of bracketing the forecasted response. These two data sets also had the least disagreement with past precipitation.

Data for the warm/wet scenario were from the World Climate Research Program's Coupled Model Intercomparison Project phase 3 multi-model data set, which was referenced in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC 2007). We used a spatial subset of the Community Climate System Model version 3 (CCSM3) data (*available online*)[3] comprising nine grid points, centered on the latitude and longitude of Coweeta, each of which had a spatial resolution of approximately 12×12 km. We chose to use T and P data for the A1B greenhouse gas emission scenario, which is based on atmospheric CO_2 concentrations reaching 720 ppm in the year 2100.

Data for the hot/dry scenario were from the first version of the Canadian Centre for Climate Modeling and Analysis (CCCma) Coupled Global Climate Model (CGCM1) data set (*available online*; see footnote 2). The spatial extent of the forecasted data was for the Hydrologic Unit Code containing the Coweeta basin.

We empirically forecasted annual Q_C from 2009 to 2050 for each watershed pair as a function of modeled annual P from each scenario to use as an independent variable in our models. Models were simple linear regressions, and all models were significant at $P < 0.001$ with good fits ($0.88 < R^2 < 0.92$). We simulated management effects for both future climate scenarios for all catchments during 2009–2050. In these simulations, we assumed that forests on the treated catchments were comparable to the reference watershed in 2008, and then the original treatments were applied to the same catchments in 2009 (i.e., forest age was reset to 0 in 2009). Using this approach, we assumed that species composition and structure (e.g., stocking, leaf area index, and so on) recovered over the ~40 year posttreatment period comparable (both spatially and temporally) to what was observed in the actual treated watersheds.

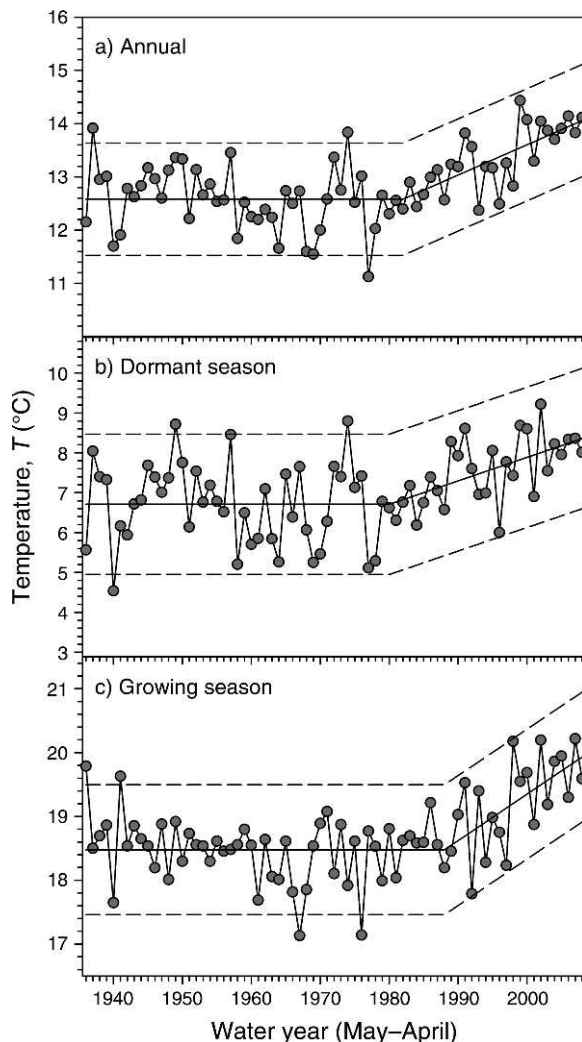[3] ⟨http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php⟩

FIG. 3. (a) Observed mean annual air temperature (T) at the climate station in the valley floor (CS01), calculated from the average of daily minimum and maximum air temperature, as well as (b) dormant and (c) growing-season average T. Solid lines correspond to a time-series intervention model with a significant ramp function at 1982, 1980, and 1988 for panels (a)–(c), respectively. Dashed lines are the upper and lower 95% confidence intervals about the modeled mean.

RESULTS

Observed climate

Mean annual air temperature has been significantly increasing at Coweeta since 1983 (Fig. 3a). Long-term mean annual air temperature was 12.6°C (before 1982) and since 1982, it has been increasing at 0.5°C per decade (AIC = 118.5, $w_i = 0.12$). This warming trend in the local record was found in the dormant and growing season temperature series (Fig. 3b) and with maximum and minimum annual temperature series (not shown). The local temperature increases also agree with the more regional trends which represent not only a longer term record, but also a spatially averaged one (Fig. 4); a

recent increase since 1982 from the 12.8°C mean was detected, with an increase of 0.2°C per decade (AIC = 182.1, $w_i = 0.12$).

Using SPI, the frequency of monthly precipitation data fit a gamma distribution ($A^2 = 2.52$, $P < 0.001$), and was described by the shape and scale parameters $\alpha = 2.96$ and $\sigma = 50.29$, respectively. Monthly precipitation during the 73-year record was 149 ± 87 mm (mean ± SD). Mean annual precipitation during the 73-year record was 1794 ± 295 mm. Years in which precipitation was <1419 mm were defined as extreme drought years ($P < 1419$ mm, SPI ≤ −1.28, $P ≤ 0.10$). Years in which precipitation was >2200 mm were defined as extreme wet years.

The frequency of extreme dry years was greater in the latter part of the precipitation record compared to the early part of the record (Fig. 2). Since 1936, 10 extreme drought years were identified; eight occurring since 1980. The most extreme dry year was 2000, in which below-average precipitation occurred in 11 of the 12 months. The frequency of extreme wet years did not increase with time, and an extreme wet year was more defined by numerous rainfall events (frequent events) rather than short-duration, intense meteorological events (i.e., hurricanes). Of the six extreme wet years, three occurred during the 1970s. The most extreme wet year was 1989, in which 10 of the 12 months were above average.

Annual precipitation totals are becoming more variable over time, with wetter wet years and drier dry years (Fig. 5a, b, Table 2). Low quantiles (50th and lower) had a significant negative slope over time, i.e., parameter estimates for quantiles < 0.5 in Fig. 5b have a negative slope (fall below 0 line) with time. In contrast, the highest quantiles (90th and higher) had a significant
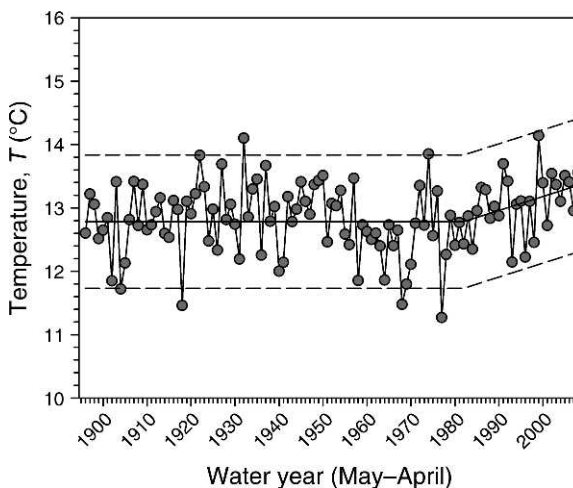

FIG. 4. Mean annual air temperature (T) data spatially averaged for North Carolina Climate Division 01 (division containing the study site). Data are available online: ⟨http://www7.ncdc.noaa.gov/CDO/CDODivisionalSelect.jsp#⟩. Solid lines correspond to a time series intervention model with a significant ramp function at 1982. Dashed lines are the upper and lower 95% confidence intervals about the modeled mean.
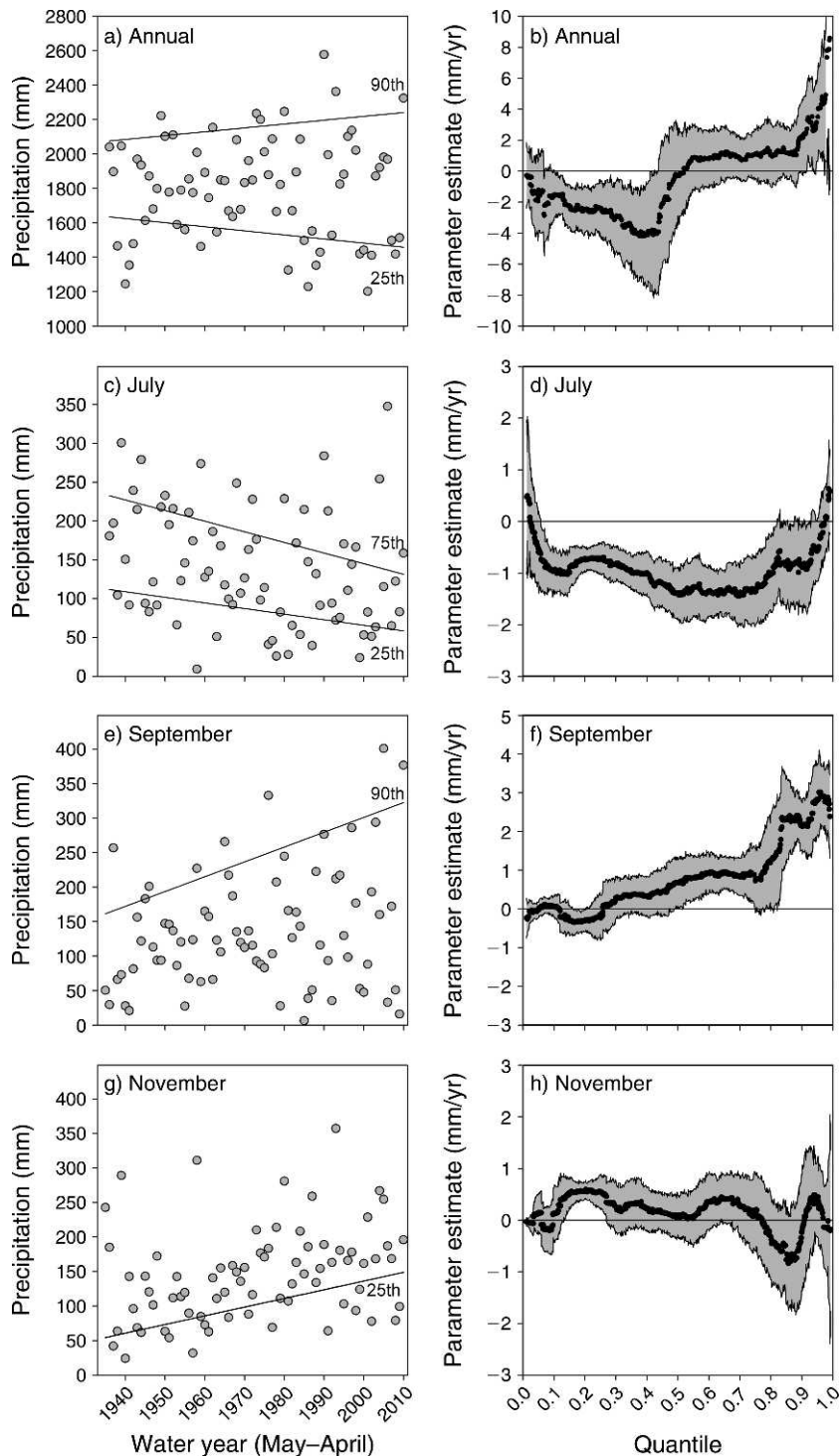
FIG. 5. (a) Annual and (c, e, g) monthly precipitation totals from gage at CS 01. Lines in each panel show the modeled qth quantile, conditional on water year. Full models for lines shown are in Table 2. All parameter estimates for each quantile for (b) annual and (d, f, h) monthly precipitation totals (solid symbols) are shown with bootstrapped 95% confidence intervals (gray area).

positive slope over time, indicating that the low and high ends of the annual precipitation distribution in the basin are changing during the period of record. This was consistent across all gages with differing aspects and elevations. The summer months, particularly July, are becoming drier over time (Fig. 5c, d), while the fall months are becoming wetter (Fig. 5e–h). In September, only the most extreme part of the distribution is

TABLE 2. Summary statistics and parameter estimates for quantile regression models on annual and monthly precipitation (P) totals.

Variable period	Quantile	Intercept (mm)	Water year (mm/yr)	Elevation (mm/m)	Predicted P in qth quantile in 1936 (mm)	Predicted P in qth quantile in 2010 (mm)	Predicted change in quantile, 1936–2010 (%)
Annual 90th quantile	0.90	−2852.85	2.24	0.87**	2075	2241	+8.0
Annual 25th quantile	0.25	5793.06**	−2.38**	0.64**	1634	1458	−10.8
July 90th quantile	0.75	2809.70**	−1.37**	0.10**	233	131	−43.5
July 25th quantile	0.25	1486.48**	−0.72**	0.03	112	58	−47.7
September 90th quantile	0.90	−4099.65**	2.15**	0.16**	164	322	+97.1
November 25th quantile	0.25	−2411.82**	1.26**	0.04**	56	149	+167.5

Note: Asterisks designate parameter estimates that are significantly different from zero at $\alpha = 0.05$ (*) and $\alpha = 0.01$(**).

increasing over time, i.e., parameter estimates for quantiles > 0.85 in Fig. 5f have a positive slope (fall above 0 line) with time, while low quantiles describing November precipitation increase over time. Extreme precipitation in fall increased 97–167% over the period of record (Table 2). In general, temporal changes in the P distributions for winter months were negligible and spring months were subtle (data not shown).

Modeled watershed recovery and climate effects on streamflow

The general model (Eq. 2) with streamflow from the control catchment and precipitation explained almost all of the variability in streamflow from the treated catchment; in all cases the adjusted $R^2 > 0.99$ (Table 3, Fig. 6). In five of the six treatments, no serial autocorrelation in the residuals remained. This indicated that successive years of streamflow could be used as independent observations. Relationships between streamflow from the catchment pairs were reasonably similar, with a range of ±15% from unity (parameter b).

In all cases, management significantly altered the expected level of streamflow (Fig. 7a–f), and this response was nonlinear with time such that streamflow excesses declined significantly over time. The initial increases in streamflow compared to that expected from the catchment had it remained unmanaged ranged from 21.4 to 38.1 cm/yr, or 19% and 70% greater than expected. Increases in streamflow after treatment persisted generally only in the species conversions stands (3.5–5 years), and were associated with controlling competition so the new forest could establish. In general, streamflow returned to pretreatment levels in 6–7 years.

The rate of watershed recovery after initial cutting towards that expected from the catchment had it remained untreated also differed significantly depending on management (Fig. 7, Table 3). The smaller the $k2$ parameter, the slower the recovery rate of streamflow toward that expected (discussed by Swank and Helvey 1970). Watershed recovery was faster (larger $k2$ parameter) for several low-elevation treatments compared to the high-elevation treatment, and also faster for south-facing stands compared to north-facing stands. The second cutting in the coppice stand experienced the fastest watershed recovery rate among all management treatments, while the first cutting was among the slowest to recover.

Physical and biological factors that determine streamflow from managed catchments were affected differently by climate than from unmanaged catchments. The

TABLE 3. Summary statistics and parameter estimates for streamflow models.

Model parameter	Species conversion (S)	Species conversion (N)	High-elevation clear-cut	Low-elevation clear-cut	Coppice (first cut, second cut)†	Old-field succession
b	0.94[b]**	0.84[c]**	0.94[b]**	1.15[a]**	0.92[b]**	0.85[c]**
c	−29.43[c]**	13.13[b]	1.59[abc]	76.08[a]**	62.35[ab]**, 62.88[ab]*	−141.10[d]**
d	0.31[b]**	0.18[bc]**	0.18[bc]**	−0.29[d]**	−0.05[cd], −0.03[bcd]	0.86[a]**
$k1$	5.16[a]**	3.47[a]**	⋯	3.23[a]	⋯	⋯
$k2$	0.46[ab]**	0.34[b]**	0.14[c]**	1.21[abc]	0.22[bc]**, 1.23[a]**	0.37[b]**
h	0.29[d]**	0.56[c]**	0.83[b]**	1.21[a]**	1.14[a]**, 1.14[a]**	0.88[b]**
F	2991.60	5034.88	22 691.4	4552.15	6149.53	5185.51
P	<0.001	<0.001	<0.001	<0.001	<0.001	<0.001
Adjusted R^2	>0.99	>0.99	>0.99	>0.99	>0.99	>0.99
P (to lag 6)	0.92	0.09	0.02	0.12	0.59	0.99

Notes: Asterisks designate parameter estimates that are significantly different from zero at $\alpha = 0.05$ (*) and $\alpha = 0.01$(**); ellipses (⋯) denote where model terms were omitted for simplification. Within rows, parameters not sharing the same lowercase superscript letters differ significantly. See Table 1 for a description of species conversion (S) and species conversion (N). See *Methods: Modeling the effect of management on annual water yield* for a description of the model parameters.

† Some model terms occurred once for each cut (for b). Values for the two (first and second) cuts are separated by a comma. The terms F, P, R^2_{adj}, and P (to lag 6) are overall model parameters (and apply for the overall model for the coppice stand).
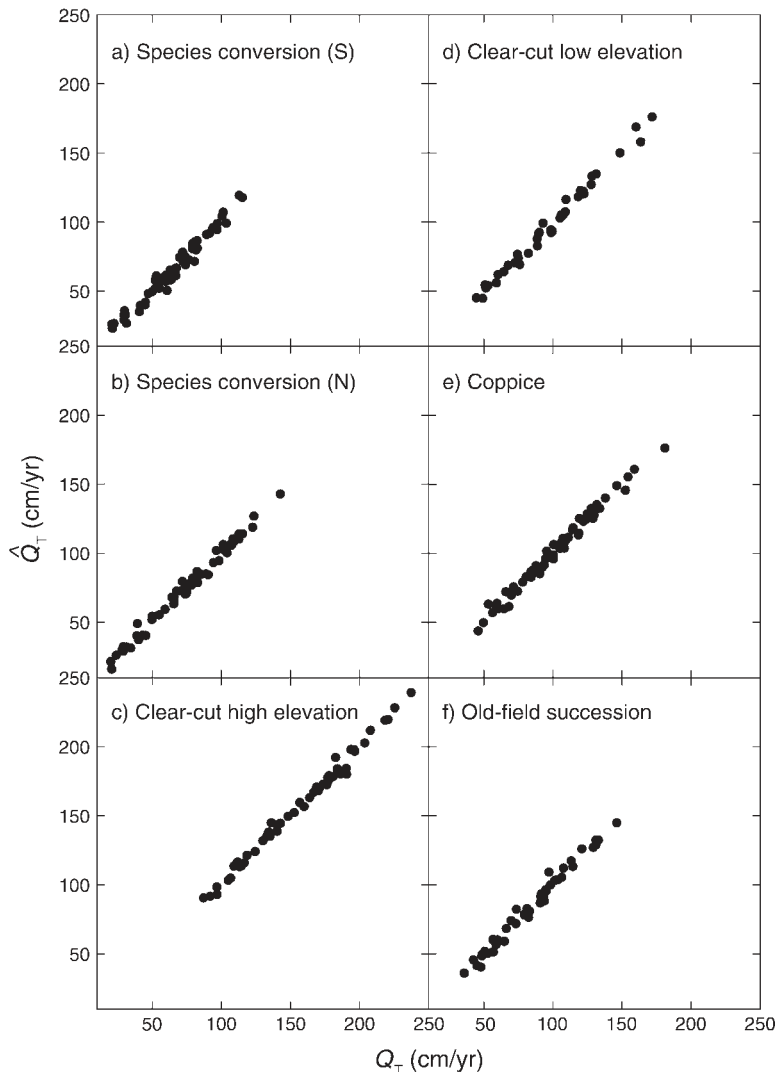
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FIG. 6. Predicted streamflow in treatment catchment ($\hat{Q}_T$) compared to observed data Q_T. Management treatments are described in Table 1, and model parameters are given in Table 3.

significant climate variable was precipitation; in no cases did temperature explain a significant amount of variability in the streamflow response. The vegetation by precipitation interaction was described by parameter d. This parameter describes the shift up or down from the logistic curve describing the watershed response with time due to precipitation (see Fig. 7a), and a significant d parameter implies that physical and biological components controlling hydrologic processes in the treatment watersheds were responding significantly different to variation in precipitation than the reference watershed.

The observed watershed response differed significantly in wet and dry years in all but the coppice managed stand (parameter d; Fig. 8, Table 3). In all other watersheds, as precipitation increased, the streamflow difference from that predicted using the reference

watershed increased. This effect was most pronounced in the old-field succession, south-facing species conversion, and the low-elevation clear-cut. Not only did these watersheds respond significantly different to climate compared to unmanaged stands, these three land uses differed significantly from one another in the magnitude of the effect on streamflow deficit. The greatest vegetation by climate interaction was seen in the old-field succession, followed by the two species conversions and the high-elevation clear-cut, then by the low-elevation clear-cut.

Categorizing all responses in the extreme wet and dry years and averaging them also showed this pattern (Fig. 9a, c). In the old-field succession, responses ranged from slight streamflow excess during extreme dry years to deficits in extreme wet years. In the two species conversions, most observed extreme precipitation years
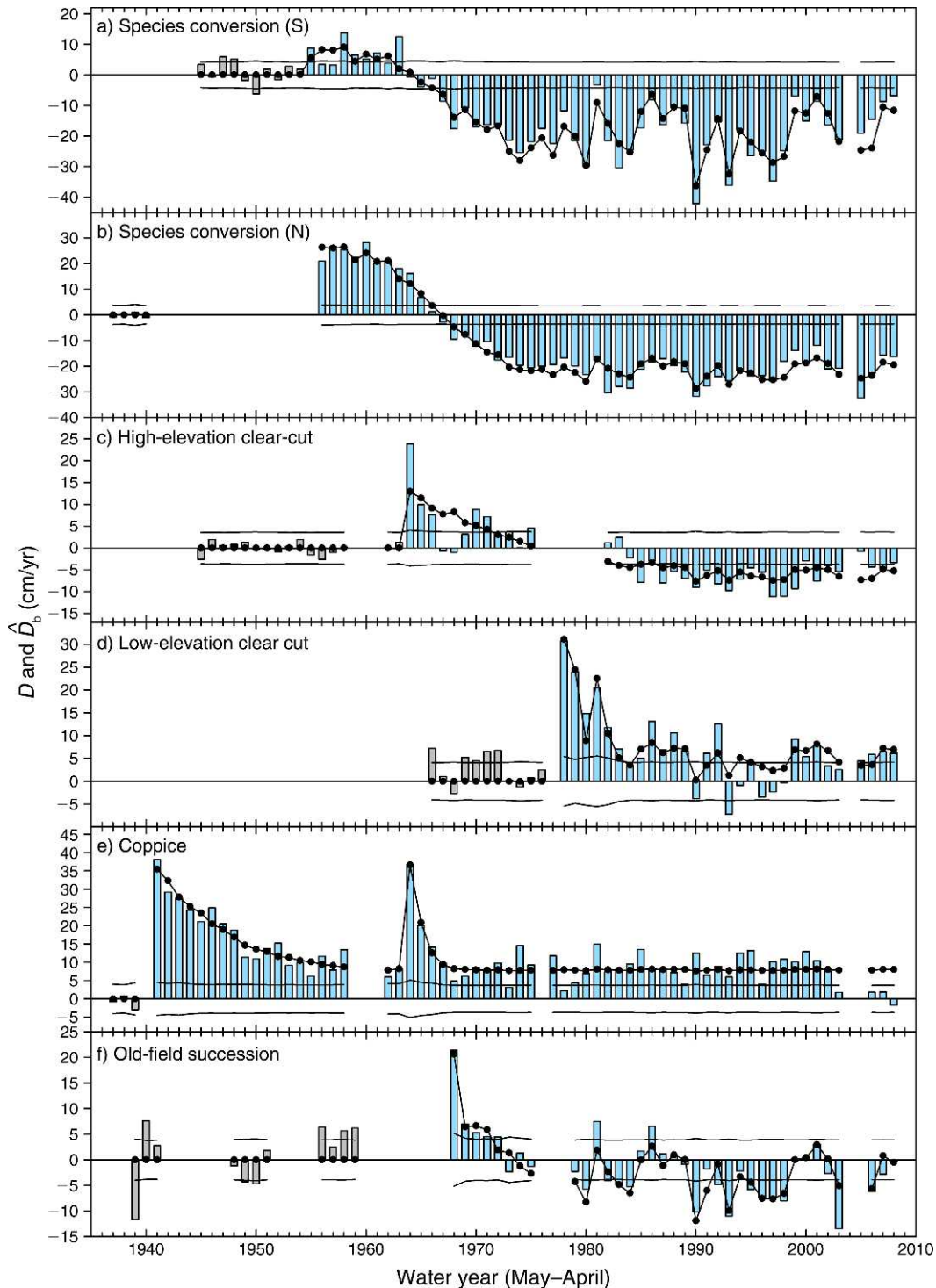
FIG. 7. Streamflow response to six management treatments. Solid symbols with lines are $\hat{D}_b = \hat{Q}_T - (\hat{Q}_T, M = 0)$ (see Eq. 4); lines alone are the standard error of the prediction. Vertical histogram bars are $D = Q_T - (\hat{Q}_T, M = 0)$ (see Eq. 3); gray fill indicates years before treatment ($M = 0$) and cyan fill is after treatment ($M = 1$). Management treatments are described in Table 1. D (cm/yr) is the observed management response and is calculated as the streamflow deviation from that predicted by the control without management. $\hat{D}_b$ is the predicted management response. M is a dummy variable with a value of 0 prior to treatment or management and 1 following treatment.
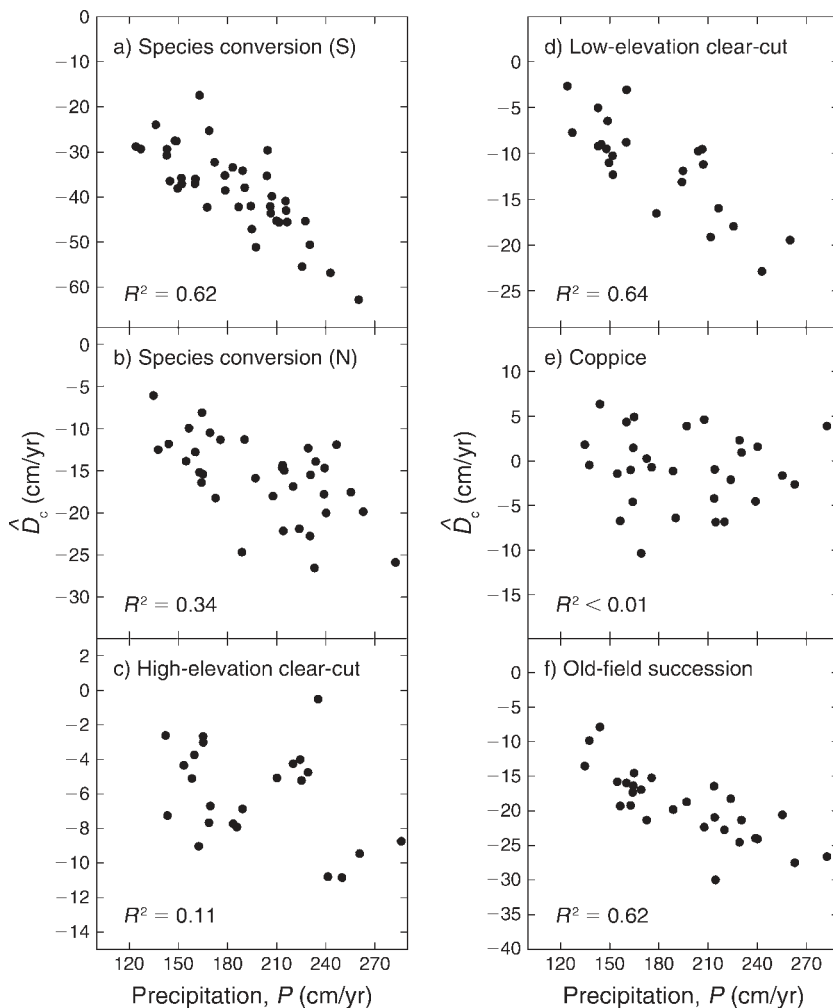
FIG. 8. (a–f) Relationship between precipitation and the vegetation response to different management ($\hat{D}_c$) from all treatment catchments during the latter part of the series. Partial model residuals ($\hat{D}_c$) are the difference in observed streamflow from the treatment catchment (Q_T) and the all model terms except d (i.e., $\hat{D}_c = Q_T - (\hat{Q}_T, M = 1, d = 0)$; see Eq. 5). Coefficient of determination (R^2) is also shown. Management treatments are described in Table 1, and model parameters given in Table 3.

coincided with fully developed stands; thus, streamflow responses ranged from moderate deficits (8–17 cm/yr) during extreme dry years to even greater deficits (24–30 cm/yr) in extreme wet years. While the high-elevation clear-cut also had similar interactions with climate as the species conversions (moderate deficits in droughts to more pronounced deficits in wet years), the average responses during extreme wet years appeared similar to those during extreme dry years. This was because half of the wet years coincided with the years soon after the watershed was cut (cf. Figs. 2 and 7c). In the low-elevation clear-cut, responses ranged from moderate streamflow excesses during extreme dry years (10 cm/yr) to slight excesses in extreme wet years (<4 cm/yr).

Forecasted climate

Forecasted temperature and precipitation out to year 2050 were markedly different from the mean of past

observed conditions, depending on the GCM data set used (data not presented). The CGCM1 forecasted warmer and drier conditions for the southern Appalachians. Average annual air temperature and precipitation during the 42 forecasted years by CGCM1 was 14.3°C and 1370 mm, or 11.5% greater temperature and 23.6% lower precipitation than the current long-term averages. Drought frequency increased to 6.2 years per decade, compared to the 1.4 drought years per decade in the observed record, which is more than a fourfold increase, and extreme wet years were completely eliminated. The CCSM3 forecasted warmer and wetter conditions for the southern Appalachians. Average annual air temperature and precipitation during the 42 forecasted years by CCSM3 was 12.9°C and 2092 mm, or 1.1% greater temperature and 16.6% greater precipitation than the current long-term average. Frequency of extreme wet
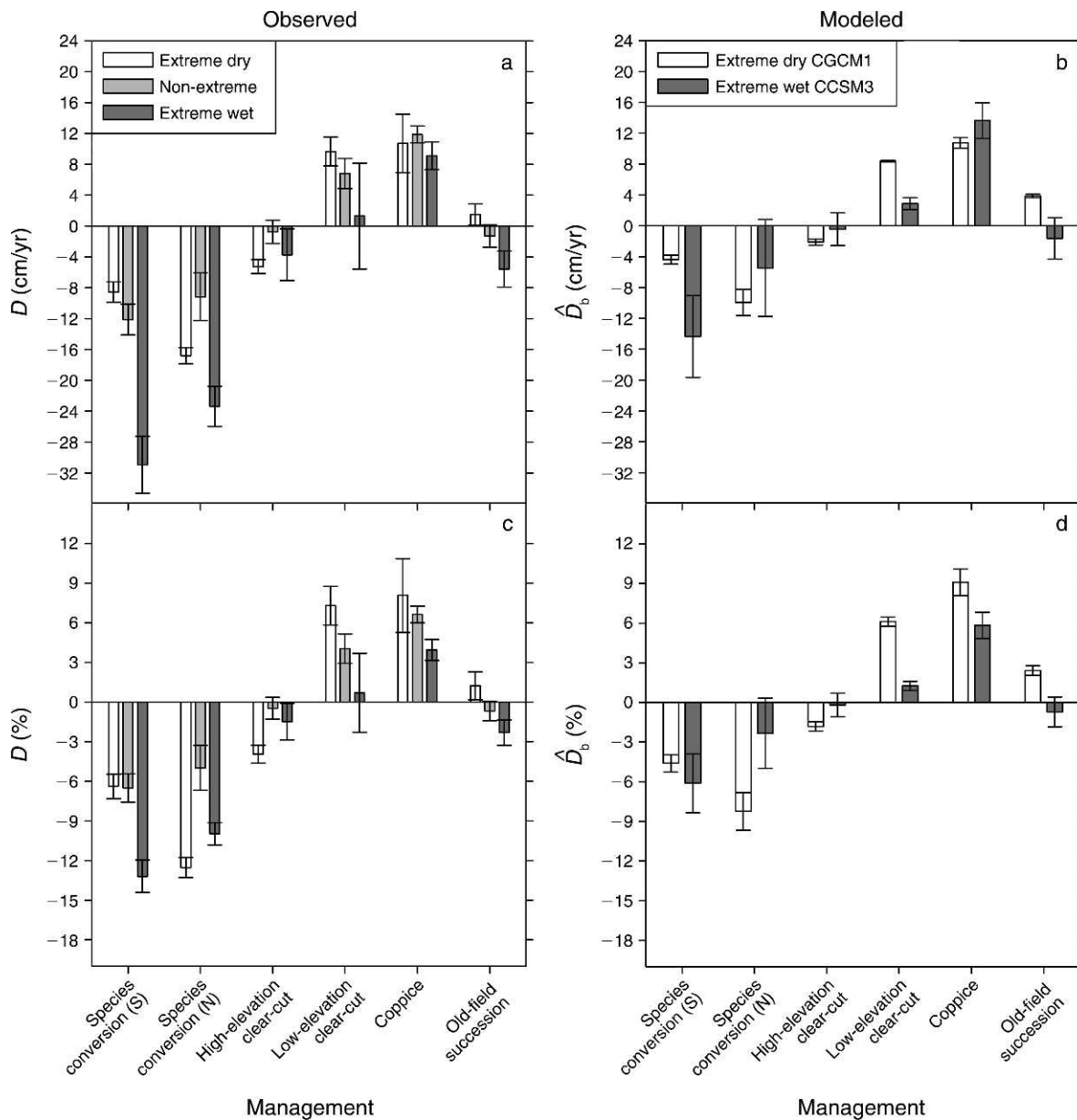
FIG. 9. (a) Observed vegetation response $D = Q_T - (\hat{Q}_T, M = 0)$ in six management treatments during extreme wet and dry years and non-extreme years (beginning of record to 2008); and (b) mean of extreme wet and dry years modeled repeating the entire management cycle in all watersheds (reset $t = 0$ in 2009, forecasting out to 2050) for both climate change scenarios (CGCM1 and CCSM3). (c) Observed and (d) modeled vegetation responses expressed as a percentage of observed or forecasted precipitation. Error bars are ±SE. Management treatments are described in Table 1.

years increased to 2.8 years per decade, compared to the 0.8 years per decade in the observed record, which is more than a threefold increase; and droughts were completely eliminated.

Management simulated under climate change scenarios

We simulated watershed responses to future climate scenarios by assuming that the same set of management treatments were repeated in 2009 and extrapolated

responses to 2050 using forecasted climate. The most notable differences again were observed for the species conversions (Fig. 9b, d). Although streamflow remained lower than the reference watershed (i.e., $D_b < 0$), the magnitude of the deficit was considerably lower than the steady-state simulations under both dry and wet extremes. In all other managed catchments, repeating the entire cycle under future climate was similar to that observed in the long-term record.

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PLATE 1. Study site in the Coweeta Basin, in the southern Appalachian Mountains, USA. From left to right, the outlined watersheds (WS) are: WS7, the low elevation clear-cut; WS2, the low elevation, south-facing reference watershed; and WS1, the south-facing, species conversion. Photo credit: W. T. Swank.

DISCUSSION

Interaction with climate

We found that the streamflow response to climate was affected by nearly all land uses examined, supporting our hypothesis that climate impacts may either be mitigated or exacerbated by forest management practices. The only exception was the repeatedly cut coppice stand. Converting to this land use resulted in a forest with lower ET, and thus higher streamflow than what would have been expected otherwise; however, despite the shift in the level of streamflow, the vegetation responded as expected (similar to reference vegetation) to climate. In all other treatments, the significant d parameter in the regression models indicated that using only streamflow from the reference watershed to adjust for the effects of climate coupled with the modeled vegetation response over time did not explain all of the variation. In other words, the treated watershed is responding differently to climate (in this case, variation in annual precipitation) than what would have been expected if it had been left untreated. Causal factors could include changes in physical conditions (roads, skid trails, and so on), changes in species composition and resulting canopy physiology (transpiration, stomatal sensitivity to vapor pressure deficit), and changes in vegetation structure (e.g., interception by stem and leaf area, tree height).

Physical changes are an unlikely causal factor, because the study watersheds were logged without a substantial road network, and in some cases, no roads were constructed and the vegetation was cut and left in place. For example, the low-elevation clear-cut managed watershed has 2.5% of the watershed area in roads, and in the first several years following harvest, small but measurable increases in peak flow rates and stormflow volumes were measured but were short lived (Swank et al. 2001). Road networks comprising 12% (Harr et al. 1975) and 6% (Alila et al. 2009) have been shown to affect storm hydrographs (i.e., peak flow frequencies); however, road networks comprising <6% of the watershed area appear not to change storm hydrographs significantly (Harr et al. 1975). We interpret the longer term responses to management in the low-elevation clear-cut watershed, as well as the other clear-cut and old-field succession watersheds to reflect a combination of species composition, physiological, and structural changes of vegetation rather than the effects of the road network or other physical factors.

Forest canopy species composition in the southern Appalachians is a function of topographic, edaphic, and climatic factors (Elliott et al. 1999), and both disturbance and time since disturbance (Elliott and Swank 2008). Permanent plots inventoried on reference watersheds from 1934 to 1993 showed a basal area

FIG. 10. Mean observed daily water use (DWU) estimated from sap flux density in trees of varying species (dbh, diameter at breast height) in reference watersheds at Coweeta. The key denotes the first two letters of scientific binomial: BELE, *Betula lenta*; NYSY, *Nyssa sylvatica*; COFL, *Cornus florida*; LITU, *Liriodendron tulipifera*; ACRU, *Acer rubrum*; PLOC, *Platnus occidentalis*; CASP, *Carya* spp.; QUPR, *Quercus prinus*; and QURU, *Q. rubra*. Symbols represent the mean DWU of replicate trees in each species during the growing season, days of year 128–280. LITU, QURU, QUPR, and CASP data are from Ford et al. (2010). BELE, NYSY, COFL, ACRU, and PLOC data are from C. Ford and J. Vose (*unpublished data*) but follow the methods in Ford et al. (2010). Symbols are as follows: circles are species with diffuse porous xylem anatomy, diamonds are species with semi-ring-porous xylem anatomy, and triangles are species with ring-porous xylem anatomy.
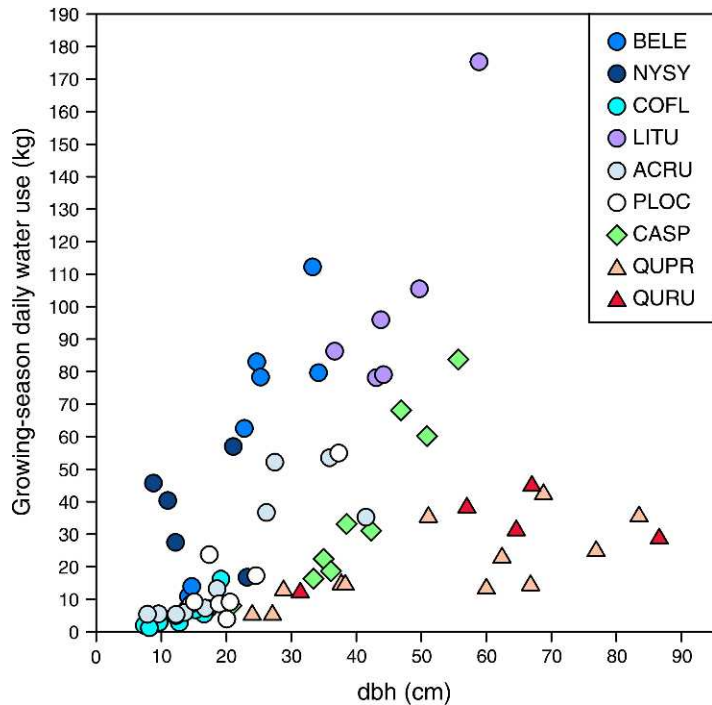
increase of only 27.9 to 30.7 m^2/ha. Over the same period, tree diversity (Shannon's H') also increased slightly from 1.76 to 1.84 (Elliott and Swank 2008). In contrast, management activities in the Coweeta basin have produced a shift from hardwood forests dominated by species with ring-porous xylem anatomy, to those with diffuse porous xylem anatomy. For example, the dominant species of the old-field succession watershed shifted from scarlet and chestnut oak to yellow poplar (Elliott et al. 1998); the high-elevation clear-cut dominant species shifted from northern red and chestnut oaks to sugar maple and black cherry, a decrease of 30% of the former two species and an increase of 18% of the latter two (W. T. Swank, *unpublished data*); the low-elevation clear-cut dominant species shifted from scarlet and chestnut oaks to dogwood, red maple, and yellow poplar (Boring et al. 1981); and the coppice stand-dominant species shifted from scarlet and chestnut oaks and pitch pine to yellow poplar, chestnut oak, and red maple (Elliott and Swank 1994). Large differences among these species in their stomatal conductance, transpiration per unit leaf area, and whole tree water use for any given diameter exist in the southern Appalachians (Wallace 1988, Wullschleger et al. 2001, Ford and Vose 2007, Ford et al. 2010); and this is especially the case in hardwoods at this study site between these two xylem functional groups (Fig. 10). Yellow poplar is shade intolerant and has among the highest transpiration rates of forest trees in the southern Appalachians (Fig. 10; Ford et al. 2010). While red maple and flowering dogwood have relatively high transpiration rates compared to the oak species,

their response to drought is much more plastic (Wullschleger et al. 1998, Wullschleger and Hanson 2003), i.e., anisohydric, than that observed by oak species (Bush et al. 2008), i.e., isohydric. These two functional groups also vary in their response to climate variability (Ford et al. 2010). The species comprising the managed catchments are more responsive to climate variability than the dominant species in the control catchments, as seen by the interannual variability in transpiration rates (Ford et al. 2010). Our results suggest that there will be measurable differences in forest ET, and an interaction between vegetation and climate in both conifer and hardwood aggrading stands in the southern Appalachian region (cf. Ford et al. 2010, Oishi et al. 2010). This finding is in contrast to studies in the Piedmont region of the southeast that have hypothesized and demonstrated that forest ET can be largely invariant among years due to the dynamics of precipitation, transpiration, interception, and the role that understory and overstory strata play (Roberts 1983, Oishi et al. 2010).

In addition to physiological changes associated with management, changes in vegetation structure were also important determinants of the interaction between management and climate on the streamflow response. The two management actions that resulted in the longest term changes in vegetation structure were the coppice and species conversion stands. Although coppicing facilitated a more diffuse porous species composition than the reference stand, this management treatment resulted in significantly higher streamflow under all climate regimes. We hypothesize that
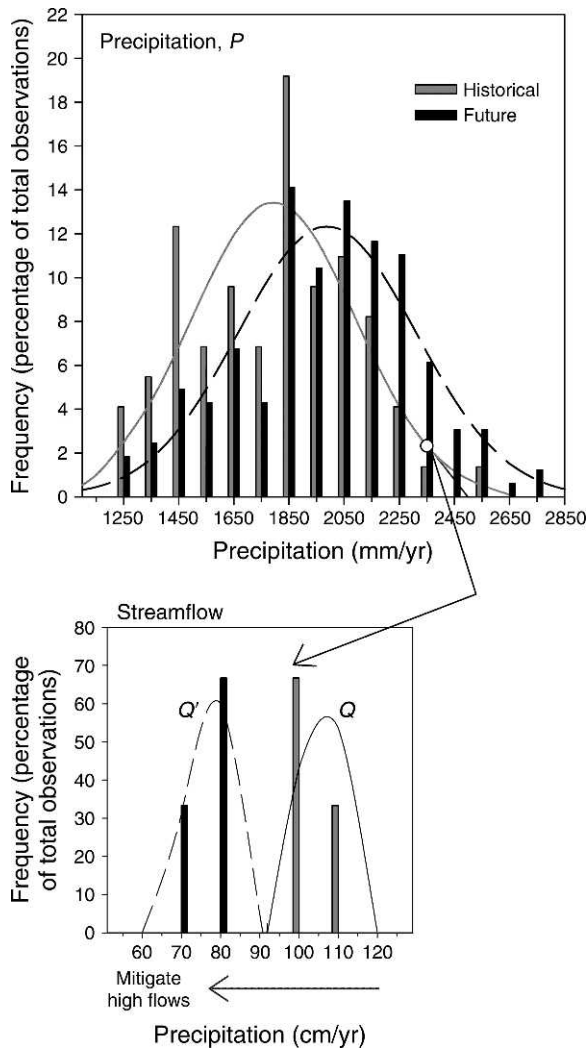
FIG. 11. Observed frequency distribution of future (CCSM3) and historical annual precipitation (bars, upper panel), and observed streamflow frequency distribution resulting from management with a species conversion (WS17, Q') or no management (Q) (bars, lower panel) for the given precipitation amount. Solid and dashed lines in upper and lower panels represent the expected frequency distribution fit to the historical (gray bars) and future (black bars) data, respectively.

repeated cutting has altered stand structure (seed vs. sprouting), and later in forest succession, reduced leaf area due to the high stem competition within sprouts. The vegetation response to climate in this stand was similar to the reference watershed; thus, the only apparent effect of management was to increase the magnitude of streamflow above that expected if the forest had been left as a native deciduous mixed-hardwood forest. In the watersheds managed for species conversions, interception is greater than in the control stand because of greater leaf area index (LAI), and branch and stem area (Swank and Schreuder

1973). Across three growing seasons spanning both relatively wet and dry years, interception by the north-facing pine stand was estimated to be 92% higher than that of the reference stand, i.e., 27% and 14% of precipitation was intercepted on the pine and reference watersheds, respectively (Ford et al. 2010). In the growing season, canopy transpiration can also be 131% greater in plots on the north-facing pine stand compared to those in the reference hardwood stand (Ford et al. 2010). Winter transpiration can also be significant in southern Appalachian forests that have evergreen leaf habit (Ford et al. 2007). Hence, the combination of greater interception and higher transpiration in the pine stands has a significant impact on soil moisture dynamics and resulting streamflow.

Can climate impacts on streamflow be mitigated or exacerbated by forest management practices that alter land cover?

Management affected the resulting vegetation structure and function, and the vegetation responded differently to climate than the reference watersheds. Whether the effects of extreme wet- or dry-precipitation years exacerbated or mitigated the streamflow response depended on the management treatment. For example, converting native deciduous catchments to dense pine monocultures reduced annual streamflow during both extreme wet- and dry-event years, which may exacerbate low flows and drought, but it also may potentially mitigate high flows and flood risk (Fig. 1). For example, compared to an unmanaged catchment, managing a catchment with a species conversion treatment decreased the apparent frequency of observed extreme wet-event years on average by a factor of seven (Fig. 11). Higher ET from these treated catchments means that soils have a greater capacity to store excess soil water (and soil water is the principle source of streamflow in most southern Appalachian headwater catchments) during wet years. This may be a good option under future climate scenarios of increased precipitation, but a poor choice under drier scenarios, because the higher ET also means that less soil water is available during drought conditions. The choice of species in this type of management treatment is a critical determinant of the magnitude of response. For example, tropical and subtropical species conversions to Eucalyptus hybrid plantations might exacerbate streamflow responses to extreme dry years even more than a pine plantation. In a review of more than 20 catchment conversion studies, Farley et al. (2005) showed that converting existing vegetation to Eucalyptus plantations reduced streamflow by 20% more than converting it to a pine plantation. This review also showed that the loss of low flows were more complete for Eucalyptus plantations compared to pine plantations (100% vs. ~80% reduction of low flows). In contrast, converting land use to either coniferous forests or bamboo forests from native forests would maintain

similar streamflow, as ET from coniferous forests and bamboo forests are similar (Komatsu et al. 2010).

Precipitation explained significant variation in streamflow response in all but one of the land uses examined. However, our results suggest that forest management activities that do not result in a forest type conversion will not differentially affect the streamflow responses to extreme precipitation events. This suggests that catchments with similar forest management activities will not substantially exacerbate or mitigate the effects of extreme annual precipitation. This is an important finding because it implies that many conventional management practices will not create watersheds that are more vulnerable to future climate conditions. However, we caution that other management actions associated with overall forest operations (especially design and density of access roads) could have important implications for hydrologic response.

Applications

Decades of coupled streamflow and climate data from managed and unmanaged forested watersheds provided a unique opportunity to examine the wide range of potential streamflow responses to extreme climatic conditions. Although temperature at our study location and within our region has been increasing over the past three decades, temperature was never a significant variable explaining streamflow response. This does not mean that temperature did not affect streamflow or ET, it only means that within the range of temperatures observed in the period of record of our study, temperature impacted biological and physical processes in both managed and reference watersheds in a similar way. Future climate scenarios indicate that temperature may increase by as much as 0.75°C above the current mean by 2050, and this may have both direct (i.e., interact with managed watershed response) and indirect (i.e., influence posttreatment species composition) impacts on streamflow response. For example, globally, local climates are moving at an average rate of 0.42 km/yr; for 29% of the globe, this rate of climate migration is faster than the rates of tree migration estimated during the late Holocene (Loarie et al. 2009). The suggestion is that for large areas, species will have to migrate (or acclimate) at just as rapid a rate to survive.

Precipitation explained significant variation in streamflow response in all but one of the managed watersheds. Although often measurable and statistically significant, differences among streamflow responses during average climate conditions vs. either observed or projected extreme annual drought or high annual precipitation were small, and would result in changes in annual streamflow ranging from +8% to −3% relative to average precipitation. These results suggest a limited capacity to create watershed conditions more (or less) resilient to extreme annual precipitation than native hardwood forests using traditional forest management practices that rely on natural regeneration. In contrast, manage-

ment activities that converted deciduous hardwood species to pine monocultures substantially altered the streamflow response to extreme annual precipitation, which may reduce flood risk but also exacerbate drought. This suggests that the trade-off between managing forests for opposite extremes should be carefully considered by water resource managers for contingency land use planning.

Currently, land managers and policy makers are both looking to forests as an option to offset the effects of climate change (Pacala and Sokolow 2004), and to forest management to create ecosystems that are more resilient to extremes and changing climate (Baron et al. 2009). We have shown here that, among the management activities we studied, that changing forest cover (e.g., species conversions) affects streamflow, and thus, downstream water supply in ways other than that expected from unmanaged forests; however, forest cover change will also affect many other ecosystem services, including carbon sequestration (Liao et al. 2010). Forests cannot be managed solely for water resources without affecting carbon sequestration, and vice versa (Jackson et al. 2005). While it is still uncertain whether increasing forest cover, or converting deciduous to evergreen forest cover, will mitigate against global climate change in the southeast United States (Jackson et al. 2008), our study shows the potential of forest management to mitigate against extreme annual precipitation associated with climate change.

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