Reconstructing a Water Balance for North Crestone Creek: Streamflow Variability and Extremes in a Snowmelt Dominated Internal Drainage Basin

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Abstract:

The San Luis Valley in Colorado is a semi-arid region that relies on streamflow from the surrounding mountain ranges for agricultural productivity and to recharge the important aquifer systems of the basin. The (North) Crestone Creek watershed is characteristic of the many small watersheds that drain the Sangre de Cristo Mountains on the eastern side of the valley providing water for local water users, wildlife, and augmentation of Rio Grande flows through the Closed Basin Project reclamation efforts. This study investigates the range of hydrologic variability and extremes in this area over the last 426 years using readily available historic hydrologic and paleo-climatic data derived from tree-rings and other sources. Streamflow and precipitation reconstructions were generated and compared to the historic period of observation. Water balance modeling was performed using historic and paleo-derived model inputs. The results of this study show that the drought conditions experienced in the San Luis Valley over the last decade are not unusual in the context of streamflow and precipitation reconstructions spanning hundreds of years. Past droughts were at least as intense as those in 1950 and 2002 and several droughts in the paleo-record were of much longer duration than any recorded in the instrumented period. These results are similar to those demonstrated in other paleo-hydrologic research from the western part of the San Luis Valley in Colorado and throughout the western United States. The water balance modeling provided a means to examine monthly changes to runoff and other hydrologic and state variables output by the model under differing past climate conditions. Together, the climate reconstructions and water balance model provide insight into regional water use sustainability and future development issues for a highly variable natural system.

Keywords: hydrologic variability, drought, tree rings, San Luis Valley, Crestone Creek

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Introduction and Justification

Water resource availability in the western United States is affected by problems such as over-allocation, continued population growth and associated land use changes, and a warming climate (IPCC, 2007; McCabe and Wolock, 2007; Barnett *et al*., 2008). Drought conditions over the last decade have prompted much research related to streamflow variability over longer temporal scales than the instrumented record, and a desire to make that research available to water managers in the region (e.g. Jain and Eischeid, 2008; Barnett and Pierce, 2009; Rice *et al.*, 2009; Miller *et al.*, 2012). This is also true in Colorado, where two of the most important river systems in the western United States, the Colorado River and the Rio Grande have headwaters in snowmelt-dominated streams and have experienced some of the lowest flows in hundreds of years (Ray *et al.*, 2008; Gray *et al.*, 2011; Pederson *et al.*, 2011; Woodhouse *et al.*, 2012).

The Rio Grande originates in the San Juan Mountains of southern Colorado and flows east then south through the San Luis Valley, a geologically unique region of the state. The valley is surrounded by the San Juan Mountains to the north and west and the Sangre de Cristo Mountains to the east. The rainshadow effect of the mountains creates a high-desert landscape on the valley floor. The northern portion of the valley is a closed basin underlain by the Rio Grande rift, which is filled with several kilometers of alluvial sedimentary and volcanic deposits, forming an important aquifer system that is separated from the southern part of the valley topographically and hydrologically (Brister and Gries, 1994; Anderholm, 1996).

The waters of the Rio Grande play a significant role in the long history and agricultural industry of the San Luis Valley, with the Treaty of 1906 and the Rio Grande Compact of 1939 ensuring downstream water rights to the states of New Mexico and Texas and the country of Mexico (Athearn, 1975; USBR, 1987). Surface water rights in the valley were almost completely allocated by the turn of the century and groundwater usage and rights became increasingly important by the 1930's (Mix *et al.*, 2012). The valley population doubled in the 1940's and water availability for uses other than agriculture became a more common issue, both within the valley and downstream. In an effort to augment Rio Grande flows to meet compact obligations, the Bureau of Reclamation began a project in 1980 to "salvage" water that would otherwise be lost to evaporation using groundwater pumped from the closed basin portion of the valley into the Rio Grande (USBR, 1987). The Closed Basin project was mostly complete by the end of the 1980's with 170 salvage wells currently in operation supplying water to the river and to other recreational and wildlife needs in the area (CGWA, 1989; Mix *et al.*, 2012).

The source of water for the closed basin aquifer system is from snowmelt-fed streams that flow from the Sangre de Cristo Mountains into the valley. The flows replenish the aquifer for Rio Grande augmentation, and maintain water levels in local well fields for domestic and agricultural use, especially in and around the town of Crestone. Additionally, the flows support narrow riparian corridors and playa lakes forming valuable habitat for waterfowl, migrating neotropical songbirds, and other animal and plant species of special concern in a dry region (USFWS, 2005).

This study assesses the natural variability and extremes of streamflow in the (North) Crestone Creek watershed draining from the Sangre de Cristo Mountains into the closed basin of the San Luis Valley. Variability was examined over the instrumented flow period of 1948 through 2012, and the meteorological record period of 1896 through 2012. Runoff variability from 1900 to 2012 was examined using a monthly water balance modeling approach. Longer-term streamflow and climate variability was characterized through a tree-ring based streamflow reconstruction (1577-2002) and the creation of a monthly paleo-water balance using tree-ring based reconstructed precipitation and other paleo-proxy temperature data (1640-2002). The reconstructions only extend to 2002 because of the length of the available tree-ring chronologies for the area. The objectives of this study were met through comparing historical hydrologic variability to that seen in paleo-reconstructed records for the purposes of providing insight into regional water use sustainability and future development issues. Open-source R statistical software for Mac was used for all analyses with the exception of the use of the Thornthwaite Monthly Water Balance model on the Windows platform (McCabe and Markstrom, 2007; R Core Team, 2013).

Data and Methods

Stream Gage and Meteorological Data

The Crestone Creek watershed area used in this study has its outlet at the Colorado Division of Water Resources (DWR) NOCRESCO gage (original USGS gage 08227500-North Crestone Creek) in Saguache County (Figure 1). The watershed covers about 27.7 square kilometers and lies within the boundaries of the Rio Grande National Forest in the Sangre de Cristo Mountains. The primary vegetation type in the watershed is mixed conifer spruce-fir forest with some aspen and has elevations up to 4210 meters above sea level. The headwater stream has no known diversions upstream from the gage site and is in an area relatively free of human impacts. The gage is at an elevation of 2553 meters above sea level and has only been moved once (from a position 48 meters upstream in 2005) in the 65 continuous years of record analyzed here, WY 1948-2012. A water year (WY) spans October 1st of a preceding year through September 30th of the following year, and is a common measure of annual streamflow in locations such as the study area that receive much of their precipitation in the form of snow during the cool season (Oct.-Apr.). The movement of the gage resulted in a slightly larger contributing area in terms of runoff in the last 7 years of the record, but the change was considered negligible for the purposes of studying runoff from the entire watershed. The streamflow data was further analyzed for use as a calibration set for tree-ring based hydroclimatic reconstructions, and for several monthly water balance modeling scenarios for Crestone Creek.

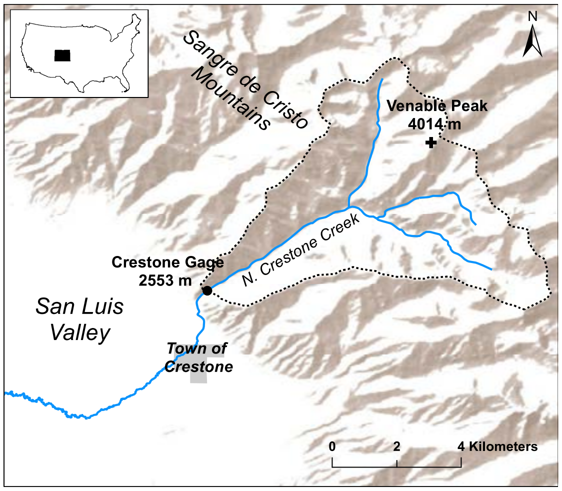

Figure 1. Location map of the Crestone Creek watershed in the San Luis Valley of Colorado. The gage used for watershed delineation is located north of the town of Crestone at an elevation of 2553 meters above sea level.

The annual historical streamflow record for Crestone Creek captures a range of wet and dry conditions. The highest volumes of flow per year on record occurred in 1973 at 18.9 MCM (cubic meters x 10^6), while the lowest annual flow volumes occurred in 2002 at 2.0 MCM. Mean flows are 10.4 MCM, and the coefficient of variation (CV) was 0.37. Crestone Creek streamflow typically peaks in the month of June, with low monthly flows occurring in January and February. The Durbin-Watson statistic was used to test whether the autocorrelation present in the annual water year streamflow data (acf=0.02) was significantly different than zero (Kutner *et al.*, 2005). The test concluded that Crestone Creek flows are not significantly autocorrelated at the $p<0.01$ level. The distribution of the data was also tested for significant skew using the D'Agostino test (Kutner *et al.*, 2005). Skew was not significantly different from zero at the $p<0.10$ and $p<0.05$ levels (skew=0.22) and the data were used in regression analysis without transformation.

Temperature and precipitation data used to characterize the site and for model inputs were extracted from PRISM (Parameter-elevation Regressions on Independent Slopes Model) monthly data grids over the area of the watershed (Daly *et al.*, 1994). The grid cells are at about 4 km x 4 km resolution (~2.5 arc min), and two grid cells cover a majority of the study area. Due to the elevational gradient of the watershed, the values between the two grid cells are on average about 20-30% different from each other, with generally lower temperatures and higher precipitation in the grid cell covering the higher elevation portion of the watershed. Despite the differences in the values between the two grid cells, the basin was not subdivided for analysis since there was only one streamflow gage at the outlet near Crestone. The maximum and minimum temperatures from the gridded data were averaged to get an average air temperature, and then the temperature data from the two grid cells were averaged over the watershed to provide a single average temperature value for each month and year from 1895 to 2012. The precipitation data from the two cells were also averaged to provide a single, watershed-wide precipitation value for that time period. PRISM data were selected for use in these analyses since they have been shown to robustly estimate patterns of orographic precipitation and temperature in mountainous terrain better than averaging meteorological station values across an area of interest (Gray and McCabe, 2010; http://www.prism.oregonstate.edu).

Temperature during the PRISM period of record (averaged from the two grid cells) varied from a monthly high of 24.5 degrees C in August of 2002, to a low of -20.9

degrees C in January of 1919. Annual precipitation (averaged from the two grid cells) varied from a maximum of 1147 millimeters in 1938 to a minimum of 451 millimeters in 2002. Climate extremes are greater at Alamosa, CO, about 48 kilometers from the study area near the valley's semi-arid center, with an elevation of 2296 meters above sea level. The maximum monthly high temperature was 31.3 degrees C in July of 2003 and the minimum temperature was -27.5 degrees C in January of 1984. Annual precipitation amounts varied between a high of 293 millimeters in 1969 and 86 millimeters in 1956 (http://www.ncdc.noaa.gov/cdo-web/search).

The statistical properties of the precipitation data such as maximum, minimum, mean, median, standard deviation, and skew were explored to prepare the data for use in the tree-ring based precipitation reconstruction. The CV was 0.51 and the data were not significantly autocorrelated when tested with the Durbin-Watson test (acf=0.08, p<0.01 level). Skew was significantly different than zero at the p<0.10 level (skew=1.02) when checked with the D'Agostino test (Kutner *et al.*, 2005). The precipitation data were transformed using a square root function, which reduced the CV 0.26 and the skew to 0.23. This value is similar to the skew of the streamflow data (skew 0.22). The skewness was not significantly different than zero at the p<0.10 level with the transformed precipitation data. The square root transformed data were used in regression modeling and the resulting reconstructed values were back-transformed by squaring the model outputs for subsequent analyses and water balance modeling.

Tree-Ring Data

The tree-ring data used in the hydrologic reconstructions were obtained from the International Tree-Ring Databank (ITRDB). Nine tree-ring chronologies were selected as potential predictors to reconstruct streamflow and precipitation based on available chronology length and location (http://www.ncdc.noaa.gov/paleo/treering). The chronologies include a variety of moisture-sensitive species and are within about 100 kilometers of the study area with sites in the San Juan, Sangre de Cristo, and Wet Mountains (Figure 2). The selected chronologies have been used in previous streamflow reconstructions in the Upper Rio Grande and other basins (e.g. Woodhouse and Lukas, 2006; Woodhouse *et al.*, 2011; Woodhouse *et al.*, 2012), but have not been used to reconstruct flow on the eastern side of the San Luis Valley (Table 1).

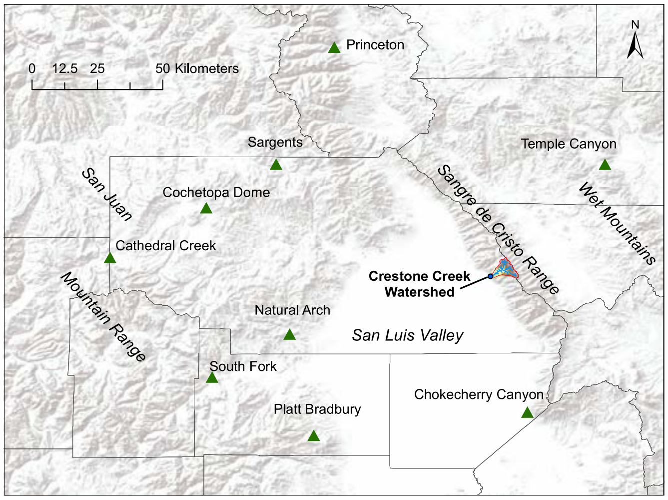

Figure 2. Map of the distribution of selected tree ring sites around the San Luis Valley and the Crestone Creek watershed.

Table 1. Description of the selected site chronologies.

Chronology	**Species**	**Elev. (m)**	**Time Span (yrs. A.D.)**	**Number of Dated Series**	**Interseries Correlation**
Cathedral Creek[a]	*Psuedotsuga menziesii*	2895	1366-2002	58	0.88
Chokecherry Canyon[b]	*Pinus edulis*	2575	1450-2003	41	0.85
Cochetopa Dome[a]	*Pinus ponderosa*	2835	1437-2002	62	0.85
Natural Arch[b]	*Pinus edulis*	2667	1508-2002	38	0.80
Platt Bradbury[c]	*Pinus ponderosa*	2835	1632-2004	20	0.79
Princeton PSME[a]	*Psuedotsuga menziesii*	2956	1169-2002	57	0.89
Sargents[a]	*Psuedotsuga menziesii*	2621	1275-2002	40	0.86
South Fork[a]	*Pinus ponderosa*	2591	1566-2002	39	0.79
Temple Canyon[b]	*Pinus edulis*	1900	1577-2003	28	0.86

[a] Woodhouse, *et al.*, 2006
[b] Woodhouse *et al.*, 2011
[c] Woodhouse *et al.*, 2004

Tree ring chronologies undergo a stringent series of collection, measurement, and standardization procedures to prepare the ring-width data for archiving in the ITRDB and for subsequent use (Stokes and Smiley, 1968; Cook and Kairiukstis, 1990; Fritts, 1976). The chronologies downloaded for this study were created from tree ring series that were standardized using negative exponential, linear (straight line), and smoothing splines (⅔ series length) to remove growth trends and were processed into standardized chronologies (Cook, 1985). All of the dated series had high interseries correlation values, which means that the rings within the individual trees (radii) have a high degree of similarity over the chronology site (Cook and Kairiukstis, 1990). Significant low order autocorrelation exists in the series, which can be attributed to biological factors such as food energy storage from year to year, fruiting, and climatic effects (Fritts, 1976). These series were then further processed, or pre-whitened, as a next step in the standardization procedure to remove the low-order autocorrelation using an auto-regressive (AR) modeling approach resulting in the creation of residual chronologies (Cook and Kairiukstis, 1990). The persistence in the residual chronologies better matched that of the streamflow and precipitation data and these were selected for use in the regression analyses (Kutner *et al.*, 2005).

Each chronology was correlated with the water year streamflow (common period WY 1948-2002) and precipitation data (common period WY 1896 to 2002). All chronologies were significantly correlated with streamflow and precipitation at the $p<0.05$ level, but correlation values varied widely depending upon the selected chronology. Streamflow correlations ranged from a high of $r=0.78$ for the Chokecherry Canyon site to a low of $r=0.43$ for the Sargents site. For precipitation, the best correlation was $r=0.59$ for the Chokecherry Canyon site, and the lowest was $r=0.26$ for the Sargents site.

Additionally, correlations were examined from positively and negatively lagged tree ring records to assess growth effects, and correlations were also performed with seasonal precipitation defined as winter (December-February), spring (March-May), summer (June-August), and fall (September-November) (Fritts, 1976). Cool and warm season cumulative precipitation values were also considered (Dec-May) and (June-Nov), respectively. These time periods were generally not as well correlated as water year cumulative values. This result implies a direct relation between current year's ring growth and soil moisture conditions preceding tree growth from cool season precipitation (snowpack) in this area (Fritts, 1976, Woodhouse *et al.*, 2004).

The number of tree ring series used in a chronology typically reduces as the record extends back in time. Therefore, fewer trees contribute to the overall site climate signal, which increases the uncertainty in the signal as a result (Wigley *et al.*, 1984). Expressed population signal (EPS) and sub-sample signal strength (SSS) were considered to determine appropriate lengths of the tree-ring record for reconstruction, with SSS>0.85 as a threshold value (Wigley *et al.*, 1984). The length of the shortest predictor in the final pool also limited the potential length of the reconstructions.

Reconstructing Streamflow and Precipitation

Stepwise regression is a common method used with tree-ring data and was used for creating the reconstruction models in this study (Woodhouse and Lukas, 2006). The chronologies were entered as predictors into a forward stepwise linear multiple regression in order of their explained residual variance to develop the final models. Akaike's Information Criterion (AIC) was employed to initially search for parsimonious models that maximized the fit between the predictors and the predictand (streamflow or precipitation) (Kutner *et al.*, 2005). The results of this analysis were compared to the use of a LEAPS regression model, which is an exhaustive search best subsets technique using a branch-and-bound algorithim, and were also compared to a model using the best-correlated chronologies (Kutner *et al.*, 2005; R Core Team, 2013). Diagnostically, there was very little difference between the three models, but there were differences between the predictors selected to reconstruct streamflow and those selected for reconstructing precipitation.

The resulting models were examined regarding multiple and adjusted R^2 values (coefficients of determination describing the proportion of the variance of the measured data explained by the model), the F-level, and the Variance Inflation Factor (VIC- to assess possible multicollinearity) and Mallow's Cp (a measure of multicollinearity and bias) (Kutner *et al.*, 2005). Residuals were checked graphically for normality and constancy of variance and also with the Kolmogorov-Smirnov test (Kutner *et al.*, 2005).

The best-fitting models for the common period were selected for validation using two different methods. First, basic cross-validation was used by dividing the modeled time periods into halves and fitting the model on each half then assessing the model diagnostics. A second method selected was *k*-fold cross-validation. This method involves randomly splitting the data into *k* equal sized subsets and retaining one subset for the validation set and using the remaining data in the entire set for calibration. The process is repeated *k*-number of times (folds) using all the subsets once for validation sets and then statistically summarizing the results of the repetitions (Kutner *et al.*, 2005). Ten-fold cross-validation, as well as 6 and 3-fold validations were performed and compared. Reconstructions were completed for a WY 1577-2002 period of streamflow for Crestone

Creek and a WY 1636-2002 period of precipitation for the Crestone Creek watershed. Both reconstructions were constrained by the length of the shortest predictor chronology used in the final models, and the tree-ring data collection dates (through 2002).

Runoff Estimates Using Water Balance Modeling

The WY 1948-2012 period of observed streamflow was used with the PRISM derived temperature and precipitation values to calibrate a water balance for Crestone Creek using the Thornthwaite Monthly Water Balance Model (TMWB) (McCabe and Markstrom, 2007; http://wwwbrr.cr.usgs.gov/projects/SW_MoWS/Thornthwaite.html). Water balance models are widely used and are ideal for understanding basic hydrologic principles and process interactions in a particular area (Thornthwaite, 1948; Hay and McCabe, 2002). This model is fairly simple to implement, and has been found suitable for use in mountainous areas in the western U.S. that receive a significant amount of their annual precipitation in the form of snow (McCabe and Wolock, 1999; Hay and McCabe, 2002).

The TMWB model employs a monthly accounting method to allocate water among the various components of the model (Figure 3). Inputs are monthly precipitation and temperature. Parameters include runoff factor, direct runoff, soil-moisture storage capacity, latitude of the location, rain temperature threshold, snow temperature threshold, and maximum snow melt rate of stored snow. Outputs include runoff, which is streamflow discharge over the area of the watershed, potential evapotranspiration (PET), precipitation, precipitation minus PET, soil moisture, actual ET, PET-AET, snow storage, and moisture surplus.

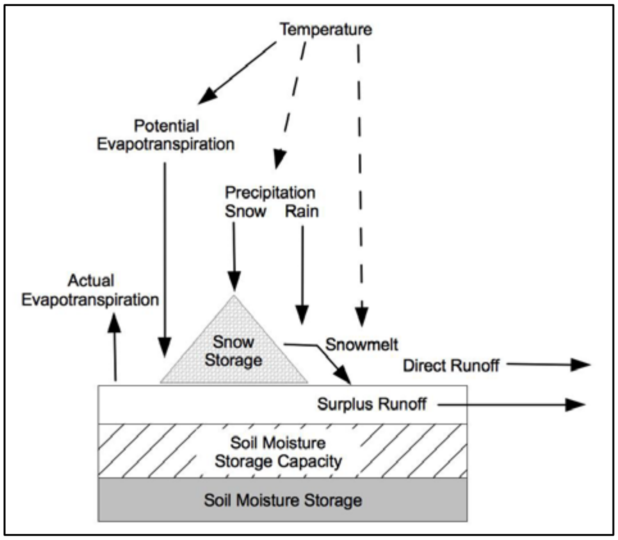

Figure 3. Schematic representation of the elements of the Thornthwaite Monthly Water Balance model. Adapted from McCabe and Markstrom, 2007.

The model has components to calculate actual and potential evapotranspiration, soil moisture storage, and runoff available from any surplus from total monthly precipitation after evapotranspiration effects and soil moisture storage conditions have been calculated. A key element of this model is a conceptual snow accumulation and melt model that can compute the occurrence and storage of snow based on specified temperature thresholds provided in the parameters. More details of the TMWB model are provided in the McCabe and Markstrom (2007) model user's guide. The resulting calibrated water balance parameters were a runoff factor of 40%, a direct runoff factor of 2%, soil moisture storage of 900 mm, latitude of 38 degrees, a rain temperature threshold of 5 degrees, a snow temperature threshold of 0 degrees, and a maximum snowmelt rate of 97%. The calibrated model parameters were then applied to the PRISM historical record (1895-2012) to examine a scenario of variation in flow extending from the previous half century before the instrumented streamflow record through 2012.

The calibrated parameters are a compromise of attempting to properly represent physical watershed properties and obtain model results that best represent the data. For example, a soil moisture storage of 900 mm is unrealistic for thinner mountain soils, but other processes such as the effects of vegetative cover, varying terrain, and the fact that there are only 7 parameters to adjust for this model necessitated setting this value higher than the average level of 150 mm recommended for most locations (McCabe and Markstrom, 2007). Additionally, the rain temperature and snow temperature thresholds were set partly based on model results, but also based on recent work suggesting that the probability of snow occurring at temperatures warmer than 0 degrees C is much higher in the semi-arid and high-elevation areas of Colorado than in other parts of the country (Fassnacht *et al.*, 2013). This finding also supports a higher rain temperature threshold. The upper snow threshold limit of the TMWB model is 0 degrees C and the rain threshold is 5 degrees C (McCabe and Markstrom, 2007). A maximum snowmelt rate of 97% was selected to meet the highly peaked nature of the streamflow hydrograph for Crestone Creek.

To assess longer scenarios of flow variability using a water balance approach, precipitation and temperature inputs were derived from tree-ring records. The annually derived reconstructed precipitation values were decomposed into monthly inputs using a percentage approach. The distribution of precipitation by month varies depending upon wet or dry year conditions, and was defined by those years where total annual precipitation was above or below the period of record mean.

Of the 118 years of data available from the PRISM dataset, 52 years were considered wetter than average and 65 were drier. The monthly precipitation values for each group (wet and dry) were averaged through time from 1895-2012 to derive monthly percentages of total annual precipitation. These monthly percentages (for wet and dry years, respectively) were then multiplied by the reconstructed annual precipitation values to give approximated monthly precipitation conditions for above and below average precipitation years through the reconstructed period of 1636-2002 for input into the water balance model.

Reconstructed northern hemisphere temperature anomaly values for the period 1500-1973 were downloaded from the NOAA Paleoclimatology site to provide temperature inputs into the paleo-water balance model (http://www.ncdc.noaa.gov/paleo/recons.html). The temperature anomaly reconstruction was derived from 91 sources of multi-proxy data from around the globe, many of which were tree-ring sources (Christiansen and

Ljungqvist, 2012). Other sources were ice cores, lake sediments, speleothems, and pollen records.

The reconstruction was created using northern hemisphere calibration data from HadCRUTv3, a gridded global temperature dataset created from land-surface and sea-surface air temperature datasets (http://www.metoffice.gov.uk/hadobs/hadcrut3). The calibration data were from the period 1880-1960. To create a paleo-temperature dataset scenario for input into the TMWB, the reconstructed temperature anomaly values were converted to a mean annual temperature. This was accomplished by combining the reconstructed anomaly values with the annual mean temperature for the period of 1895-1960 of the PRISM data for the Crestone watershed to yield a time series of temperature values from 1500 through 1973 that followed similar trends to the global temperature variations, but reflected the local temperature conditions of this region. Comparison of the overlapping time period data values between the temperature anomaly reconstruction and the PRISM temperature set (1895-1973) justified filling the last years (months) of the time-series from 1974-2002 with the original PRISM temperature values to match the end date of the precipitation reconstruction.

The annual mean temperatures provided in the reconstructed series were decomposed to monthly values by finding the difference between each of the PRISM monthly temperature values and the annual mean temperature for each year in the PRISM dataset (1895-2012). This gave an “anomaly” value for each month in the time series from the mean annual value. A mean monthly anomaly value was calculated through time from the individual monthly anomaly values through all the years in the PRISM dataset and the difference of this value was taken with the reconstructed annual temperature values to yield reconstructed monthly temperature values that tracked local average monthly changes in temperature and retained some of the variability of the reconstructed northern hemisphere anomaly values for the period of 1500-2002. These were then truncated to 1636-2002 to equal the length of the reconstructed precipitation data. The decomposed reconstructed temperature and precipitation values were used to run two paleo-water balance scenarios from 1640-2002, one for the precipitation derived from the average wetter years and one derived from the drier years.

Results

Variability and Extremes in the Historical Hydrologic Record

The historical streamflow record for Crestone Creek and the PRISM precipitation values for the watershed show high amounts of interannual variability. Averaging differences in year-to-year streamflow over the instrumented record resulted in a length of record average of 20%. Differences between individual years were as great as 70%. For precipitation, year to year variability is less than for streamflow. Maximum interannual variability is nearly 30%, with a mean variability of approximately 8%.

Duration of drought is characterized by the number of years of streamflow (or precipitation) falling below the long-term mean or median value, while drought intensity is a measure of the how much the year's streamflow falls below the long-term average flow. Median and mean streamflow values for the historical Crestone Creek record are similar, at 10.2 and 10.4 MCM respectively. Mean values are used here as an estimator of wet and dry years, defining wet years as those with annual water year flows above the mean and dry years as those with annual water year streamflow below the mean.

Historically, Crestone Creek has experienced dry conditions 34 of the 65 years of record. Lower than mean flows occurred in every decade, but the most pronounced low flows were in 1950, 1963, 2002, and 2011. The most intense droughts (lowest flows) were in 2002 and in 1950. Low flow duration is generally measured as a sequence of several years, such as from 1950-1956, and 1970-1978, and 2000-2004. However, in many cases, more sustained periods of low flows were interrupted by a higher flow year, as in 1971-1974 where flows were about 6 MCM for all years except 1973, which was the highest flow on record (18.9 MCM). Alternatively, single drought year events can be significant, with the low flows of 2002 (2.0 MCM) bracketed by closer to average flows in 2001 (11.6 MCM) and 2003 (9.6 MCM).

In the PRISM precipitation record, low amounts of rainfall are recorded in years prior to the streamflow record, such as in the late 1890's and mid 1920's. Other periods of lower than average precipitation match well with times of lower than average streamflow, as in 1950, 2002, and 2011. Higher than average precipitation also correlated well with periods of higher than average streamflow, as in 1965 and 1992 (Figure 4).

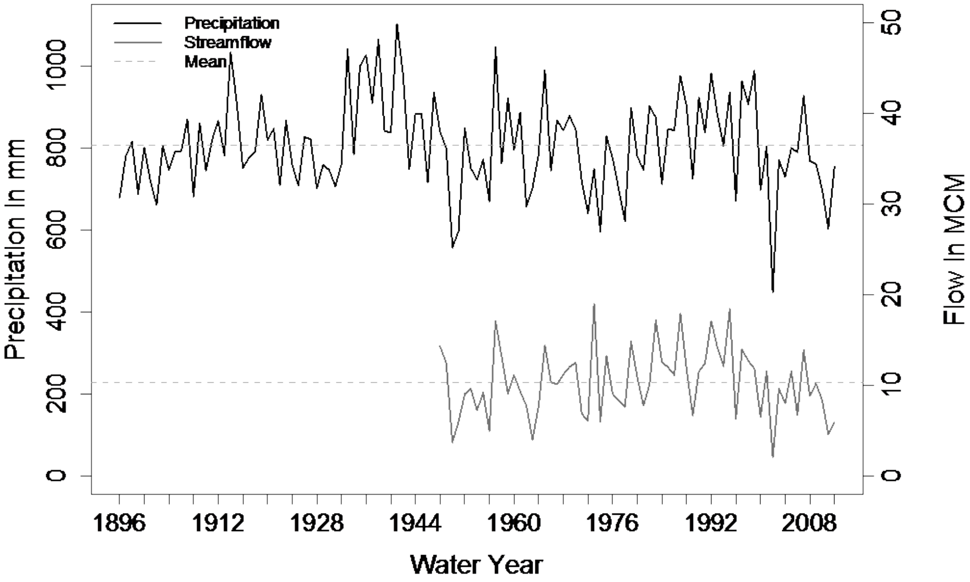

Figure 4. Comparison of precipitation and streamflow for the WY 1896-2012 period.

Reconstruction Models

The streamflow and precipitation reconstructions captured the general state of the hydrologic systems (i.e. dry or wet) even though the models tend to over and underestimate the extremes of the observational record (Figures 5 and 6). Overestimation was most noticeable during periods of low flow and underestimation generally occurred during periods of higher observed flows. This effect is partly due to the compression of variance inherent in the models from the regression process and how tree growth responds to soil moisture conditions (Fritts, 1976; Meko and Woodhouse, 2011).

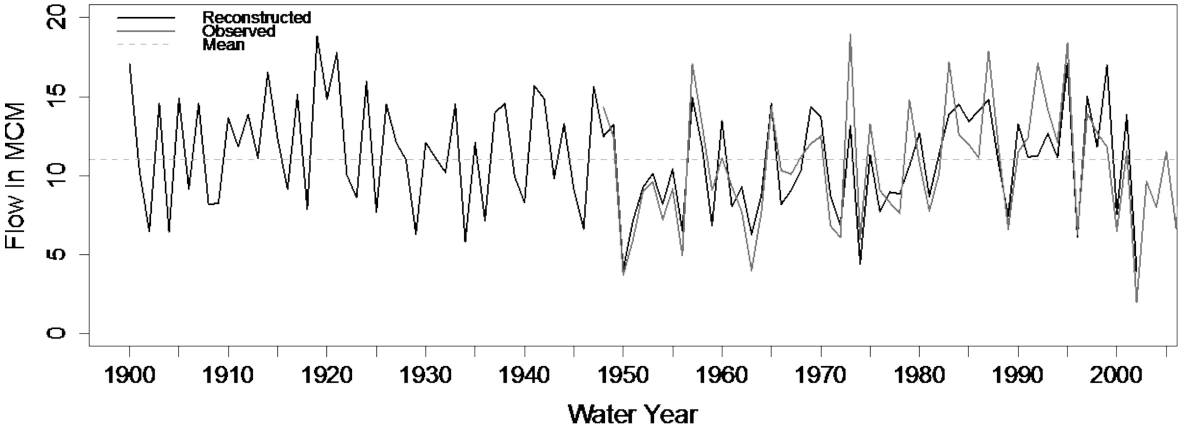

Figure 5. Subset of reconstructed (black line) and observed (gray line) annual water year streamflow for Crestone Creek. Mean flow for the entire reconstructed period (WY1577-2002) is shown as a dashed line.

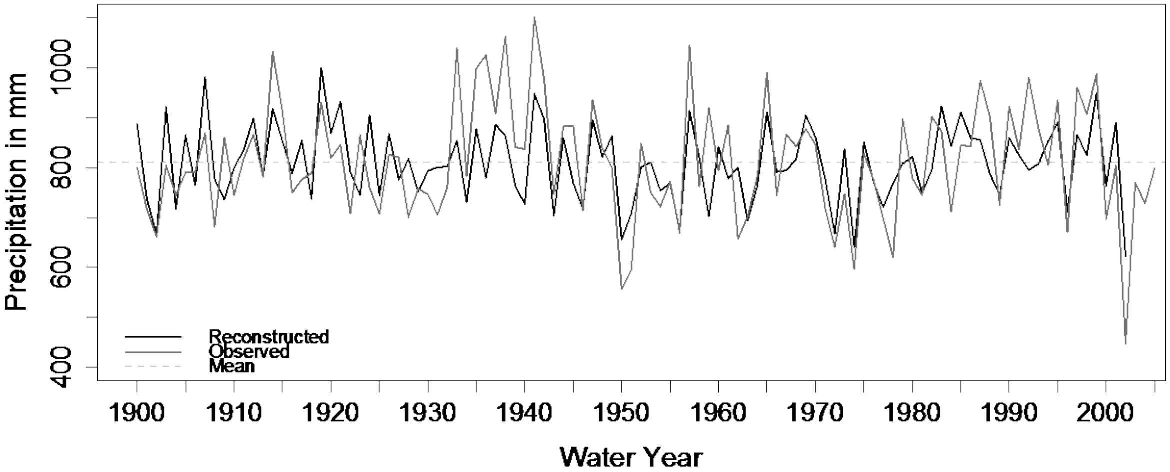

Figure 6. Subset of reconstructed (black line) and observed (gray line) annual water year precipitation for Crestone Creek. Mean precipitation for the entire reconstructed period (WY1636-2002) is shown as a dashed line.

The 426-year streamflow reconstruction from WY 1577 -2002 explains 70% of the variance in the gage record and the 367-year precipitation reconstruction from WY 1636-2002 explains 46% of the variance in the PRISM record (Table 2). Other evaluations of model robustness and fit are the Reduction in Error (RE), which can be thought of as a validation equivalent of the calibration R^2, and the Root Mean Squared Error (RMSE), a measure of the association between modeled and observed values (Fritts, 1976; Kutner *et al.*, 2005; Woodhouse and Lukas, 2006).

Table 2. Model calibration and verification statistics for the reconstructions.

Reconstruction	**Period**	**R^2**	**Adj. R^2**	**RE**	**RMSE[a]**	**RMSE[b]**
Streamflow	WY 1577-2002	0.70	0.68	0.30	2.17	2.33
Precipitation	WY 1636-2002	0.46	0.43	0.56	86.1	85.4

[a] Results of calibration in MCM and mm, respectively.
[b] Results of 3-fold cross validation, units as above.

The results of testing indicate that the residuals are not distributed significantly different than the normal distribution (at the $p<0.05$ level). The variables selected do not exhibit high levels of multicollinearity. The variance of the residuals does not deviate systematically and is relatively constant suggesting a linear relationship between the predictors is appropriate (Kutner *et al.*, 2005). The RE statistic provides a test of association between the modeled and observed data and positive values are desirable (Fritts, 1976). The RMSE values suggest that the errors calculated between the calibration modeled and observed datsets and the sets that were created in the cross-validation process are similar, and the RMSE values are within one standard deviation of the observed data, indicating relatively low model error (Moriasi *et al.*, 2007).

The final model selected for the streamflow reconstruction includes residual chronologies from Chokecherry Canyon, Cathedral Creek, and Temple Canyon. These three sites are to the west, south and east of Crestone Creek and are taken from *Pinus edulis* (Pinyon pine) and *Pseudotsuga menziesii* (Douglas-fir) trees. The final precipitation reconstruction predictors selected were Chokecherry Canyon, Princeton PSME, Natural Arch, and Platt Bradbury. These sites are located closest to the Crestone Creek watershed to the south, north, west, and east respectively. The tree species represented in these chronologies are *Pinus ponderosa* (Ponderosa pine), *Pinus edulis* (Pinyon pine), and *Pseudotsuga menzesii* (Douglas-fir).

Long-Term Streamflow Variability and Extremes

A streamflow reconstruction is essentially a sample record that is drawn from a population of possible hydrologic records for a site (Stockton, 1971). As such, it is constrained by the biases and uncertainties that are inherent in the statistical method used for the analysis, but the results provide a “window” into a past that would otherwise be difficult to study at an annual resolution. The main goal of reconstructing climate variables such as streamflow and precipitation (and temperature) with proxy data is to assess variability and extremes on a scale longer than the historical, instrumented record.

Of the 426 years of paleo-based reconstructed annual water year flow 208 years had flow conditions below the mean value of 11.0 MCM. The long-term reconstructed mean was only slightly higher than the observed mean flow of 10.4 MCM. The lowest streamflow year was in 1645 (3.7 MCM) with a flow that was less than that reconstructed for 2002

(3.9 MCM). The observed streamflow in 2002 was 2.0 MCM, illustrating the tendency for the reconstruction to underestimate low flows. The year of 1950 was the third lowest, with 1748 (4.3 MCM) and 1880 (4.3 MCM) being the next driest years in the reconstructed record, these were followed by the low flows of 1974 and 2011.

From a duration of drought perspective, runs of six to seven years of consecutive drought are common in the reconstructed record, but are exceptional in the historical record. One long dry period recorded in the reconstruction lasted from 1727 to 1745. A duration of nearly 20 years with only five breaks in the sequence (1732, 1734, 1736, 1739, and 1743). Other reconstructions of streamflow in Colorado also show this period as one of extended drought, though it is not a period of intense low flows as seen in 1950 and 2002 (Woodhouse and Lukas, 2006; Gray *et al.,* 2011). However, one of the lowest flows in the reconstruction occurred only five years later in 1748.

Another long drought sequence occurred in 1578 through 1592, a fifteen-year period interrupted by only marginally above mean flows in 1582 and 1588-89. Reconstructions of tributary flow in the Upper Colorado basin suggest that this period was dry in that region as well (Gray *et al.*, 2011). A third period of longer drought occurred around 1873-1883. This decade of drought had a few higher years of flow in 1876, 1878, and 1882, but these flow amounts were only slightly greater than the mean flow for the reconstruction. The low flows recorded in the tree-ring widths in 1880 and those recorded nearly 20 years later in 1899, were some of the most intense (flows well below the mean) in the reconstructions. Droughts in the historical observation periods of the 1950's, 1970's, and 2000's were also significant, however, there were generally more above mean flow breaks within these periods than those in the reconstructed record (Figure 7).

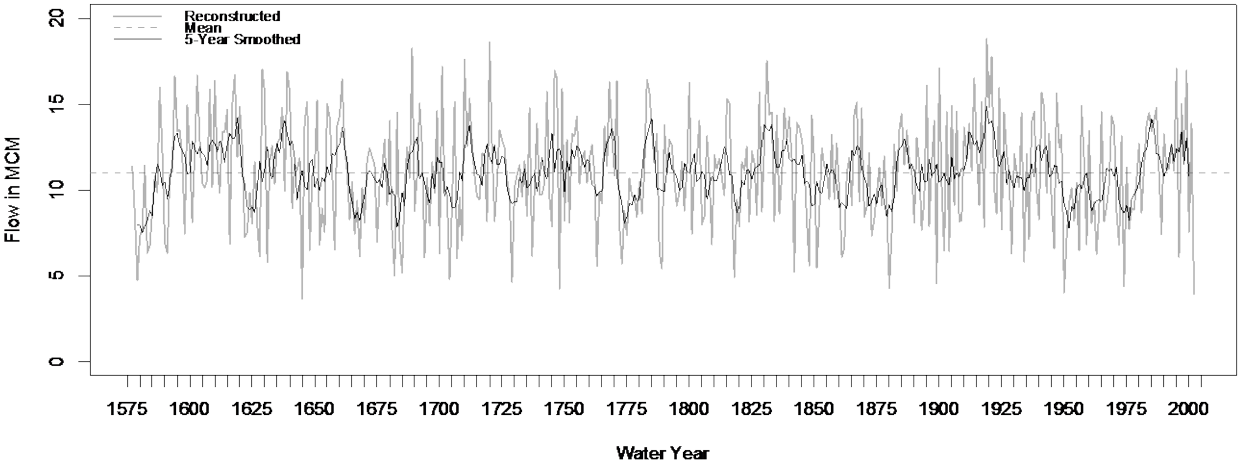

Figure 7. Full reconstruction of water year flows from 1577-2002 for Crestone Creek. Annual flows (grey line) were smoothed using a five-year moving average for display purposes (black line).

Pluvials, or periods of higher than average precipitation resulting in higher than average streamflow, were also represented in the reconstruction. Some of the highest reconstructed flows were captured during the wet sequence of years from roughly the turn of the century through the early 1920's. While higher flows are not well reconstructed due to the limitations of tree growth and soil moisture relations, the early 20th century pluvials are well documented in other Colorado reconstructions (Fritts, 1976; Woodhouse and Lukas, 2006). The extremely unusual wet climate conditions of that period played a pivotal role in water allocation for the West, particularly for the Colorado River (Stockton and Jacoby, 1976).

The variability and extremes captured in the Crestone reconstruction compare favorably with other streamflow reconstructions in the San Luis Valley. Two other reconstructions, the Rio Grande gage at Del Norte and the Saguache Creek gage at Saguache show similar patterns of wet and dry years. While the patterns are not exact between the three reconstructions, there is general hydrologic state (eg. wet or dry) agreement between them (Woodhouse *et al.*, 2004; Woodhouse *et al.*, 2012). Extreme dry years correlate well between the three sites with similarly ranked intense low flows in the selected years of 1590, 1685, 1729, 1748, 1773, 1805, 1881, 1934, 1950 and 2002. Pluvials match up well with similar rankings in the selected years of 1608, 1655, 1720, 1784, 1835, 1907, 1914, 1921, 1965, 1983, 1995 and 1999 (Figures 8a and 8b).

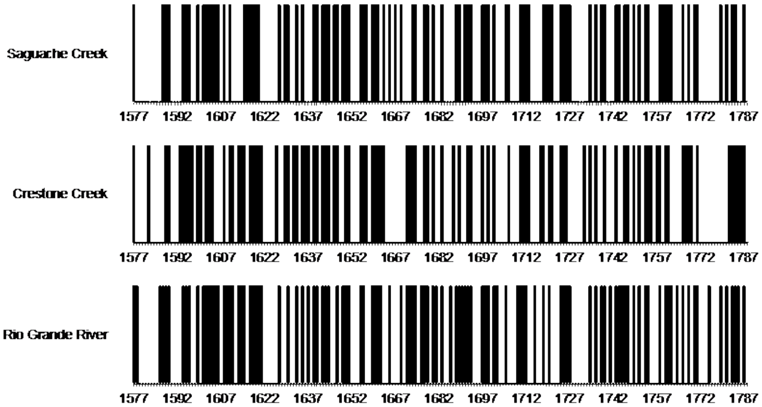

Figure 8a. Comparison of three streamflow reconstructions from the San Luis Valley. Wet (black lines) and dry years (white lines) are departures above and below the long-term mean flow.

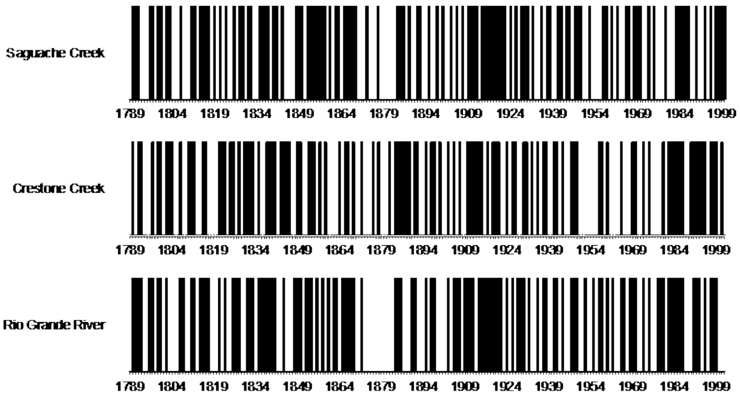

Figure 8b. Comparison of three streamflow reconstructions from the San Luis Valley. Wet (black lines) and dry years (white lines) are departures above and below the long-term mean flow. The Saguache reconstruction extends to 2000 and the Crestone Creek and Rio Grande extend to 2002 (Woodhouse *et al.*, 2004; Woodhouse *et al.*, 2012).

Historical and Paleo-Water Balance Results

The water balance model was calibrated to the WY1948-2012 observed streamflow record using the historical PRISM precipitation and temperature data as inputs. Results of the historical modeling show that the model consistently underestimated peak flows and often slightly over-estimated lower flows (Figure 9). These effects are noticeable during the pluvials of the 1980's and 1990's and the low flows of 1950 and 2002.

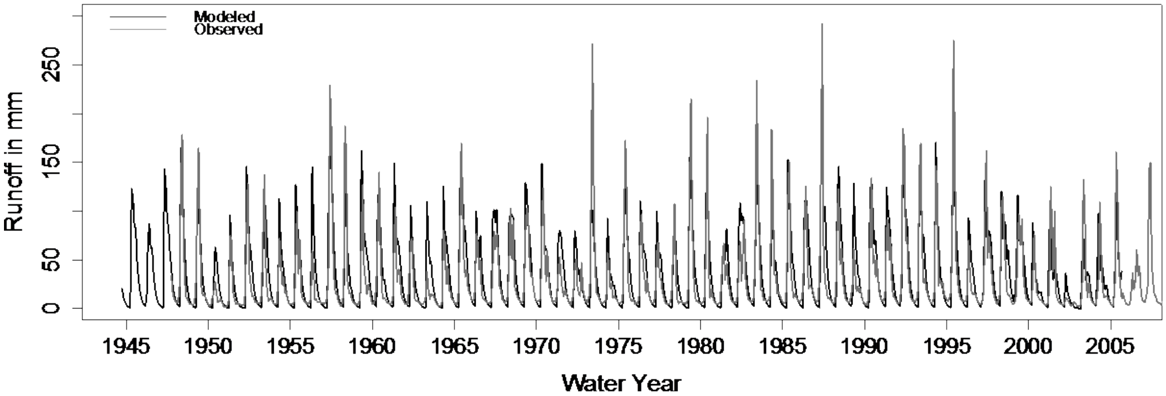

Figure 9. Subset of modeled (black line) and observed (gray line) annual water year runoff for Crestone Creek taken from the historical period model run of WY1895-2012.

The model run using the paleo-data (using wet and dry year inputs) with the same parameters as the original calibration resulted in similar output, but tended to overestimate monthly runoff in drier periods more than the water balance run using the historical data only. This is particularly noticeable during the early 1920's to mid 1930's, and in the late 1940's to early 1950's. The paleo-model also underestimated periods of higher flow compared to the historical water balance model run. This is seen in the mid 1900's and again in the late 1930's- mid 1940's. Similar to the historical water balance model, the paleo-data run model overestimated low flows in the 1950's but underestimated high flows in the late 1940's (Figure 10).

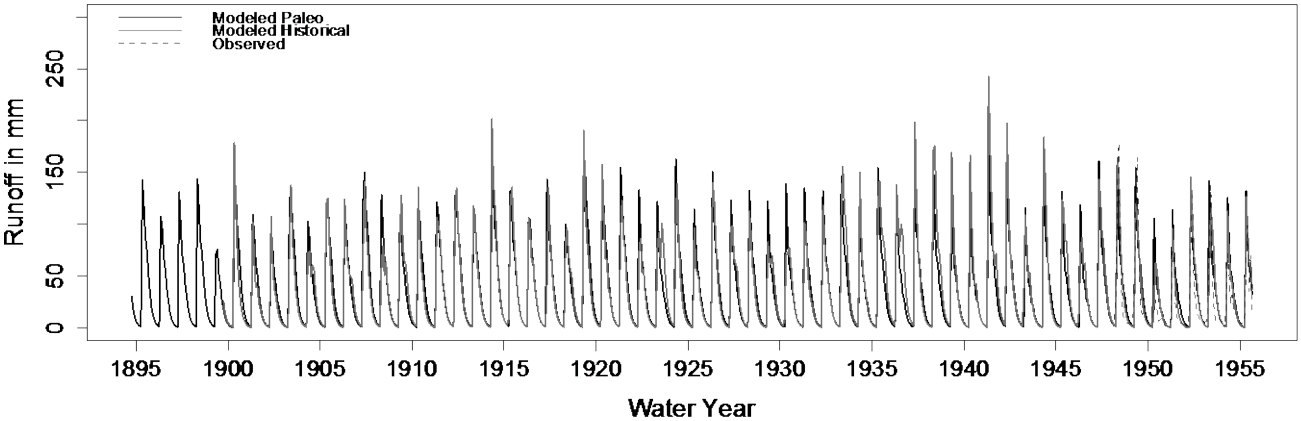

Figure 10. Subset of paleo-modeled (using drier than average precipitation years- black line) and historical modeled (gray line) annual water year runoff for Crestone Creek with a portion of the observed record (dashed line) included for reference.

Goodness-of-fit statistics for the model runs were calculated to assess model fits to the observed data during the 65 year calibration period of WY 1948-2012 (Table 3). These include the Nash-Sutcliff Efficiency (NSE), which is a measure of the relative magnitude of the residual variance (modeled values) compared to the measured data variance (observed values), the R^2 of the model, the RMSE, and the percent bias (PBIAS) which measures the tendency of the simulated data to be on average larger or smaller than the observed data (Legates and McCabe, 1999; Moriasi *et al.*, 2007).

All of the statistics are acceptable and represent the general state of the system at least as well as the reconstructed models (Moriasi *et al.*, 2007). The NSE values are much greater than zero, which means that the simulated runoff values are better than just using the mean of the data (Legates and McCabe, 1999). The models explained between 61 and 72% of the variance in the streamflow gage data (converted to runoff) and the RMSE values are within less than one standard deviation of the observed data (Moriasi *et al.*, 2007). The calculated runoff was on average greater than that of the observed data. This effect can be seen in the inability of the model to match the attenuation of the falling limb of the observed runoff for each water year and the tendency for the models to over-predict flow during drier years.

Table 3. Calibration statistics for the historical and paleo-water balance model runs.

Water Balance Run	Years	NSE	R^2	RMSE (mm)	PBIAS
Historical	65	0.68	0.72	23.67	0.21
Paleo-Dry[a]	65	0.56	0.61	28.20	0.18
Paleo-Wet[b]	65	0.58	0.62	27.82	0.18

[a] Uses an average of the years with precipitation below the mean for decomposition of annual reconstructed precipitation for a model input (see Figure 9).

[b] Uses an average of the years with precipitation above the mean for decomposition of annual reconstructed precipitation for a model input (not shown).

A simplified validation scheme of using alternate years of data (even and odd) to run the water balance model was used to assess performance. With the exception of the PBIAS measure for the odd year paleo-models, using even and odd years did not result in large differences between the calibration and verification statistics. The smaller value of the PBIAS in odd years shows that the average difference between the observed runoff data and the paleo-modeled runoff is less than when using the full dataset. The RMSE values however, are larger for the odd years, but are still within one standard deviation of the observed runoff values (Table 4).

Table 4. Verification statistics for the historical and paleo-water balance model runs.

Water Balance Run	Years	NSE	R^2	RMSE (mm)	PBIAS
Historical Odd	32	0.63	0.65	28.90	0.12
Historical Even	33	0.56	0.65	23.48	0.18
Paleo-Dry Odd	27	0.51	0.53	33.65	0.06
Paleo-Dry Even	28	0.49	0.59	25.96	0.18
Paleo-Wet Odd	27	0.52	0.54	33.28	0.06
Paleo-Wet Even	28	0.51	0.60	25.67	0.18

Discussion

Though Crestone Creek is a relatively small watershed, it is representative of many such creeks that flow into the closed basin from the Sangre de Cristo Mountains supplying much-needed surface flows and water for aquifer recharge. The streamflow (and precipitation) reconstructions provide a localized view of the long-term hydrologic conditions at Crestone Creek within the context of other robust reconstructions created for rivers such as the Saguache and Rio Grande (Woodhouse *et al.*, 2004; Woodhouse *et al.*, 2012). As a result of these comparisons and analyses, it is clear that Crestone Creek streamflow is highly variable on annual, decadal, and century time scales.

Extreme variability in year-to-year streamflow is managed on larger rivers with in-stream or diversion storage of several years' worth of water, but in a headwater system, even single year drought events can be devastating to local water supplies (Woodhouse and Lucas, 2006). This effect was chronicled during the 2002 severe drought that affected streamflow and wells in the Crestone area (Nicholas, 2002). Locals stated that it was the worst drought in generations and many ranchers and farmers did not receive their annual surface water allocations as no ditch flows were expected through the town that year. About 2467 MCM of mountain snowmelt is expected to recharge the local streams and aquifer in normal years, but during 2002, water levels in the unconfined aquifer were down nearly 4.5 meters by May, leading local water engineers and managers to apply for emergency well-drilling funds (Nicholas, 2002).

While the historical flow record captures some of the longer-term hydrologic variability in terms of intensity of low flows (i.e. the well-below average streamflows of 1950 and 2002), only one other comparable extremely low flow event of similar magnitude (1645) was captured in the reconstruction. It is possible that even more intense droughts are not well represented because the reconstructed record is based on a calibration with historical flows and uses analysis methods known to under-represent extremes (Meko and Woodhouse, 2011). The historical record also does not adequately represent the possibility of extended duration drought events. Longer duration events than any of those seen in the historic record exist in the reconstructed paleo-flow record. This condition is found in a majority of the reconstructions for other river systems in Colorado and is particularly seen in longer timescale reconstructions than those performed here (e.g. Gray *et al.*, 2011; Meko, *et al.*, 2007; Woodhouse and Lukas, 2006).

Changing climate conditions that may cause or exacerbate drought conditions such as seasonal warming and drying, and/or changing cool season precipitation patterns and snowpack accumulation and melt conditions will affect this system and the many other streams like Crestone Creek that feed into the closed basin (e.g. Jain and Eischeid, 2008; Pederson *et al.*, 2011). Hydrologic models such as the TMWB have been used to examine flow variability in the pre-observational period and in future climate scenarios. These studies used specially developed or existing hydrologic models at annual timescales and/or decomposition of the paleo-data to other time steps needed for modeling (Saito *et al.*, 2008; Gray and McCabe, 2010; Solander *et al.*, 2010). Other research has incorporated tree-ring and other proxy data into hydrologic models with stochastic approaches, such as using the tree-ring record to inform the magnitude of synthetic flows based on hydrologic state in pre-observational periods or for examining drought severity (Burn *et al.*, 2004; Prairie *et al.*, 2008).

In this study, the monthly water balance model outputs resulted in a higher-resolution examination of possible streamflow (runoff) conditions in the recent and distant past. In addition, the model outputs allowed comparison of hydrologic variables and states other than runoff. As an example, the tree-ring reconstructed annual streamflow values for the drought years of 2002 and 1645 were very similar, with a slightly lower amount of flow for 1645 than 2002. The paleo-water balance results (dry years) suggested that reconstructed runoff in 2002 was lower than that of 1645. Other paleo-modeled output variables such as snow storage amounts, and soil moisture were lower for 2002 than 1645, and actual evapotranspiration and potential evapotranspiration amounts were higher. If only the historic record (PRISM data-based) is modeled for 2002 using the TMWB, runoff is significantly lower than what is output by the paleo-based water balance model. The different analysis methods (reconstruction and/or hydrologic modeling) result in slightly different scenarios of low flow conditions, but agree on hydrologic state and the general severity of drought events (Prairie *et al.*, 2008). Also, the ability to compare additional modeled output variables could prove useful in studies directed at understanding watershed processes and landscape change besides streamflow or runoff amounts (e.g. Saito *et al.*, 2008; Vittori, 2011).

Regardless of the modeling or reconstruction method chosen, it is necessary to be aware of the uncertainty inherent in the modeling process (Solander *et al.*, 2010; Meko *et al.*, 2012). Quantifying model uncertainty remains a challenge in reconstructions and hydrologic models (Woodhouse and Lukas, 2006; Moriasi *et al.*, 2007). The intent of using paleo-data for modeling in this study was to evaluate longer-term flow variability and extremes. An outstanding goal of this project and others like it is to find simple and robust ways of providing these combined historical and proxy data-based scenarios of flow for water managers to use in making resource allocation and use decisions (Burn *et al.*, 2004; Rice *et al.*, 2009; Gray and McCabe, 2010).

Of future interest is to better and more realistically calibrate several of the water balance parameters, possibly altering the model to reflect snow accumulation and melt conditions in this high elevation, semi-arid region. It may be advantageous to further divide the basin by elevational gradient or other condition to better simulate runoff processes and soil storage parameters. For the tree-ring reconstructions, other analysis methods such as principle components analysis (PCA) should be compared to the results of the stepwise regression used here. It is possible that better precipitation and/or streamflow model fits could be gained with other methods. Finally, the scope of this project should be extended to include not only an examination of streamflow variability and extremes for Crestone Creek but also to link these changes through time to atmospheric processes and influences such as the westerly systems that deliver winter precipitation via the jet stream and the effects of the El Niño/Southern Oscillation (ENSO) on precipitation in this area while considering possible impacts from climate change (Woodhouse *et al.*, 2012).

Conclusion and Summary

The drought conditions experienced in the San Luis Valley over the last decade are not unusual in the context of the hydrologic reconstructions spanning hundreds of years and are similar to climate conditions experienced throughout Colorado and the western United States as seen in other paleo-hydrologic research. Past droughts were at least as intense as those in 1950 and 2002 and several droughts in the paleo-record are of much longer duration than any recorded in the historic period. Water balance modeling provided a means to examine changes to runoff and other hydrologic and state variables output by the model under these differing past climate conditions.

While the population of the Crestone area is small, the agricultural heritage of the area and future population expansion opportunities rely on the availability of water from snowmelt runoff and recharge to the closed basin aquifer system. Water managers and users in the region need to understand the variability inherent in the hydrologic system to make sound management decisions and lifestyle choices.

This study used readily available historic and paleo-data to examine hydrologic variability and extremes in streamflow and precipitation for Crestone Creek and the surrounding watershed over time periods ranging from 65 years up to 426 years. Streamflow and precipitation reconstructions were generated using stepwise regression methods and results were compared to hydrologic variability and extremes in the historic period of observation. Water balance modeling at a monthly timestep was performed using historic and paleo-derived model inputs. The study also contributed to the development of hydrologic and dendroclimatologic data analysis proficiency for the lead author and an increased understanding of water management issues in Colorado.

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