Appendix D
Variable Infiltration Model

Connecticut River Basin Variable Infiltration Capacity Model

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9/11/2012

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Introduction

The need for a physically-based 'whole basin' hydrology model of the Connecticut River watershed is the primary driving force behind developing the Connecticut River Variable Infiltration Capacity (CRVIC) model. The CRVIC model has been developed in tandem with the Sustainable Yield Estimator (SYE) model (Archfield 2010). SYE is capable of producing spatially (sub basin) and temporally (daily) disaggregate estimates of streamflow using nearby index gages and local basin properties. A strongpoint and primary challenge of the SYE method is the reliance on historic, unregulated, streamflow time-series from nearby gages to create estimates of 'natural' streamflow at ungagged site. The process of generating streamflow estimates relies on heavily on basin properties and nearby basin response to storm events, but only cursorily includes basin climatology as a variable. Due to this, SYE is not directly suitable for climate change analyses. Instead, hybrid approaches have been proposed to couple SYE with physically based hydrology models to develop basin-wide streamflow estimates that reflect projections climate from the latest general circulation models.

This document provides an overview of the VIC model as applied to the Connecticut River Basin, a discussion of the forcing data for the model, highlights of calibration results, and lastly a basin-wide impact analysis of climate change.

VIC Model – Physics and Basin Setup

This section provides a brief overview of the VIC model as it pertains to the current setup of the Connecticut River model. A more in-depth explanation of the physics and algorithms that compose VIC can be found here: http://www.hydro.washington.edu/Lettenmaier/Models/VIC/

The Variable Infiltration Capacity model (VIC) is a macro-scale, semi-distributed, physically based, hydrology model. The land surface of a VIC model is composed of a series of grid cells. The spatial resolution of VIC is coarse relative to other hydrology models. VIC was originally developed to represent land surface processes within global circulation models, which tend to operate at scales larger than 2° latitude by 2° longitude, though VIC is now commonly applied to scales ranging from 1/16° to 2°. The spatial resolution of the CRVIC model is 1/8° latitude by 1/8° longitude, or approximately 54 square miles per grid cell (Gao et. al., 2010).

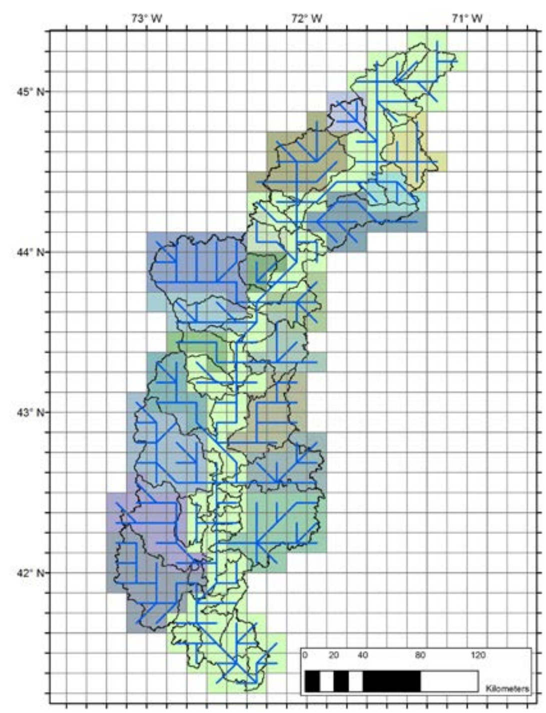


Figure 1: CRVIC Model Network

A distinguishing feature of VIC is its 'bathtub' nature, that is, each grid cell receives a separate meteorological forcing, and each grid cell has only one outlet from which water can leave and enter a stream network, thus the semi-distributed property. No water is passed between grid cells other than the water entering the stream channel network, and once water enters the channel network it is assumed to stay in the channel (it cannot flow back into the soil in another grid cell). A fundamental assumption of VIC is that the portions of surface and subsurface runoff that reach the local channel network within a grid cell are assumed to be significantly greater than the portions that cross grid cell boundaries. When VIC is executed, each grid cell is simulated independently of other grid cells, so in one simulation, each grid cell is simulated separately, one grid cell at a time. The output from each grid cell (water and energy balances, runoff, etc.) is written to a separate file. To generate streamflow within a basin, runoff from each grid cell is routed through the stream network of a basin using separate piece of software. The routing of water through a basin is described in more detail below.

Land surface characteristics are specified as distributions within VIC. Land cover within a grid cell can be subdivided into separate tiles, each representing a fraction of the total land cover within a cell (Liang et. al., 1994). Specific properties of vegetation and land cover classes are defined in the Vegetation Library File, which contains global parameterizations for each land cover type. Heterogeneity between grid cells, such as percent type of land cover present, leaf area indices, etc. are defined in the Vegetation Parameter file. Development of these files was part of a larger project (Land Data Assimilation System project), with values for grid cells in the Northeast provided by Dr. Andy Wood (Colorado Basin River Forecasting Center).

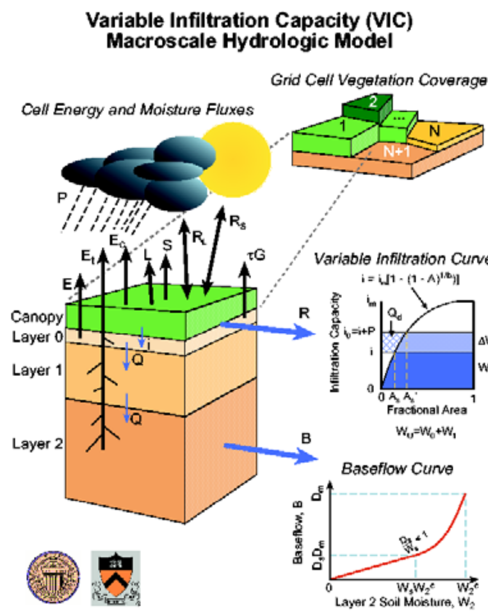


Figure 2: Schematic of VIC Model
(Taken from www.hydro.washington.edu)

Soil properties are specified by layer in the Soil Parameter file. The CRVIC model uses three soil layers in the current formulation of varying depth by grid cell (a calibration parameter). Soil properties, such as hydraulic conductivity, bulk density, porosity, etc. are also derived from the LDAS VIC parameterization. These data are primarily distilled from the CONUS-Soil database developed by Penn State University (http://www.soilinfo.psu.edu/). Elevation can be specified as a single value for a grid cell or as a distribution of elevations within a grid cell. For areas with little topography, one elevation value may be appropriate. In New England, snow melt processes are dependent enough on topographic variability that the distributional approach for representing elevation is chosen. In the CRVIC model, six 'snow bands' represent the elevation distributions within each grid cell, allowing snow to accumulate or melt based on the band in which it is located ().

VIC can simulate water and energy balances at time steps ranging from hourly to daily. It is common to allow the snow sub-model to be run at an hourly time step regardless of the time step chosen for the water and energy balance component to properly account for snow accumulation and melt physics. The current version of the CRVIC model is executed using energy, water, and snow budgets being calculated at a 3 hourly time step using daily meteorological data disaggregated to 3 hourly time steps as the forcing. More information about the forcing data is found in the Forcing Datasets section.

Within each grid cell, water and energy balances are calculated, translating the meteorological forcing into stored soil moisture, snow, evapotranspiration, and runoff. Detailed descriptions of the governing physics for each component are presented in Liang et al. (1994), with the most recent updates found in Gao et al. (In Review). Some of the algorithms used to compute water balances are: Penman-Monteith for evapotranspiration, Brooks-Corey relationship for determining soil water movement between layers, and baseflow from the bottom soil layer as described by the Arno model (Franchini and Pacciani 1991). Direct runoff is generated by those areas for which precipitation, when added to soil moisture storage at the end of the previous time step, exceeds the storage capacity of the soil, as formulated by the variable infiltration curve.

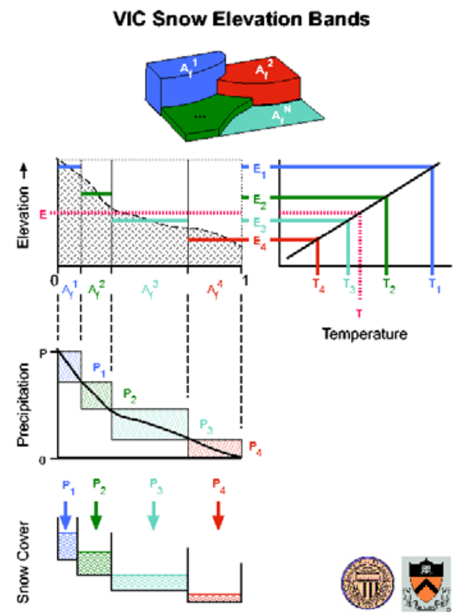


Figure 3: VIC Snow Elevation Model
(Taken from www.hydro.washington.edu)

Routing of runoff and baseflow within CRVIC uses a two-step methodology developed by Lohmann et al. (1996 and 1998). First, runoff and baseflow from a grid cell are routed to the 'outlet' of the grid cell using a unit hydrograph approach (Figure 4). Once total grid cell runoff reaches the outlet, it then enters the stream network for a given basin. Figure 1 contains the stream network for the CRVIC, with sub-basins of the Connecticut River watershed highlighted in different shades of blue, green, and purple. The unit hydrograph for the CRVIC was determined through trial and error matching of summer flows in small basins that are largely unregulated. Values for the diffusivity and convectivity parameters of the linearized Saint-Venant equation, which is used to route flood waves through the stream network, were selected as commonly suggested values.

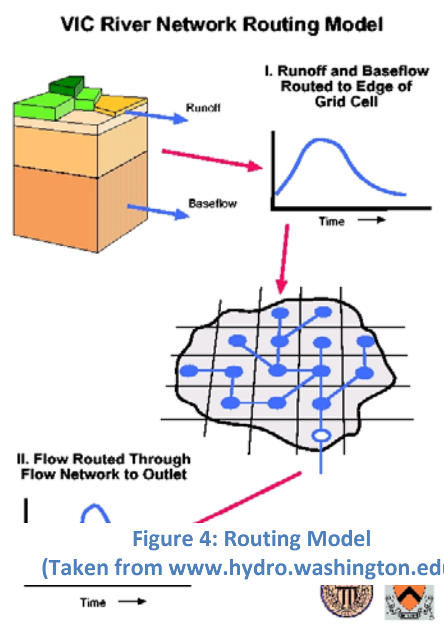


Figure 4: Routing Model
(Taken from www.hydro.washington.edu)

Forcing Datasets

The meteorological forcing data for the VIC model is a gridded 1/8th degree data product developed by Maurer et al (2005). This dataset contains daily estimates of maximum daily temperature (Tmax), minimum daily temperature (Tmin), and cumulative daily precipitation (Prcp), as well as wind speeds. Maximum daily temperature, minimum daily temperature, total daily precipitation, and wind speed are the meteorological forcings needed to drive the CRVIC model. The period of record for this dataset is 1949-2010. CRVIC is capable of disaggregating the meteorological forcings to sub-daily timesteps via various internal algorithms. As the model is currently setup, daily meteorological forcings are disaggregated to three-hourly timesteps.

Parameter Estimation and Verification Results - Upper Third Basins

The Upper Third portion of the Connecticut River watershed consists of eleven major sub-basins, three of which model parameter estimation focused on: the Upper Ammonoosuc, the Ammonoosuc, and the Passumpsic (Figure 5).

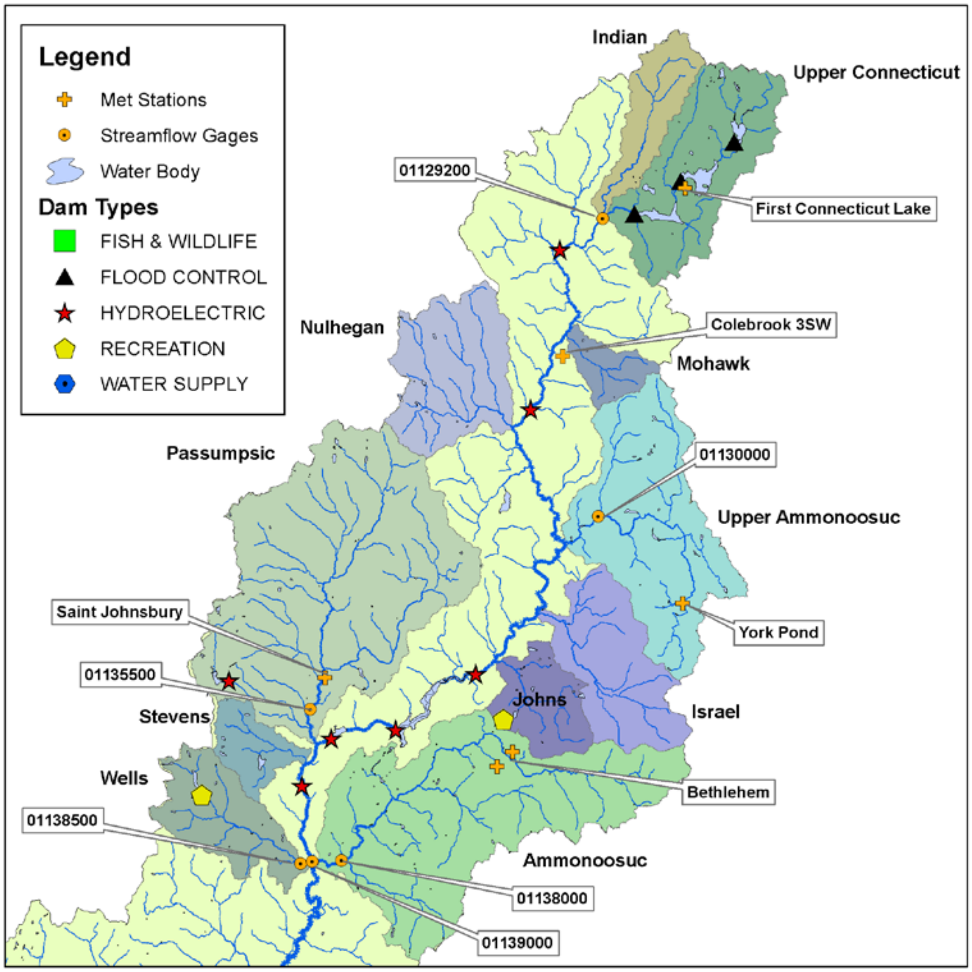


Figure 5: Upper Third Watershed Schematic

Ammonoosuc

The Ammonoosuc River is located in Northern New Hampshire, draining approximately 403 square miles and running nearly 55 miles long at the confluence with the mainstem of the Connecticut River. The river runs close to natural, unimpeded conditions, despite a number of small run of river dams within the system (~70). The east side of the basin is dominated by mountains, including the western half of the Mt Washington complex, whereas the far western portion of the basin is lower elevation and open floodplain as the river meets the Connecticut. The elevation range in the basin is the greatest in the Connecticut River watershed, with a high of 1912 meters above mean sea level and a low elevation of 121 meters. The mean elevation of the basin is ~500 meters above sea level. There are two USGS gages within the basin with long monitoring records. The first gage is in the upper watershed (Ammonoosuc River at Bethlehem Junction, NH - 01137500) monitoring runoff from approximately 88 square miles of basin, the second gage is located near the mouth of the river (Ammonoosuc River near Bath NH - 01138000) and monitors runoff from approximately 395 square miles.

Snow

When comparing a point measurement (observed) with an aerial estimate (model simulated) of snow depth, it is challenging to make direct comparisons between the values as there is significant topographic variation within a grid cell and the measuring station may be at an elevation that is much different than mean conditions (Figure 6). CRVIC can account for some of the sub-grid variability through use of a distributed elevation approach as described in the section above, so we expect the values to be relatively close, but not perfect as vegetation is not spatially variable, the model forcing dataset is less than perfect, and the final depth values that are used for comparison from the model are mean depth estimates for the entire grid cell (~54 square miles). Instead, we expect basic seasonal cycles and trends to match and the daily time series to reflect much of the weather variability that dominates local snow formation and accumulation.

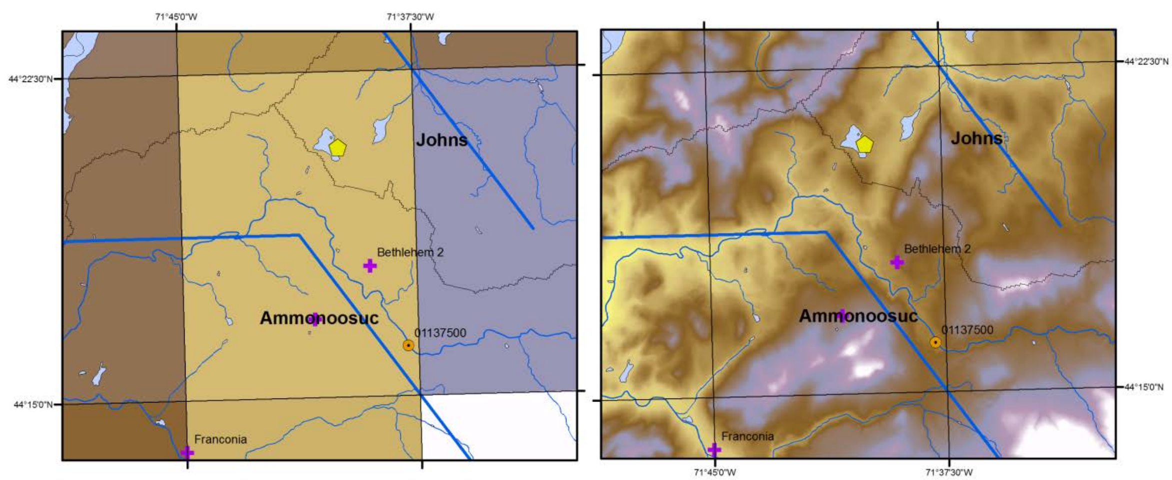


Figure 6: VIC Grid compared with Observed Elevation

To this extent, the CRVIC model replicates well the snow processes in the Ammonoosuc Basin. Given the northern latitude of the basin and the high elevations, it is no surprise that there is significant winter accumulation of snow relative to other sub-catchments within the Connecticut River watershed. At the meteorological gage in Bethlehem, NH, the mean and median daily snowpack occurs around March 1st, with approximately 60 cm of snow accumulation (Figure 7). Snow begin to accumulate in the end of November, increasing to a maximum around March 1st and then decreasing throughout the spring, with typical complete melt occurring around May 1st. The VIC model for the Ammonoosuc River basin tends to under-predict snow during the peak periods, resulting in earlier complete melt for the pixel centered over the Bethlehem station. The time series match well for most years, though tend to under predict two large snow years (1963 and 1969). In general, the effect of snow melt on runoff timing replicates observed conditions well as evidenced by the next section.

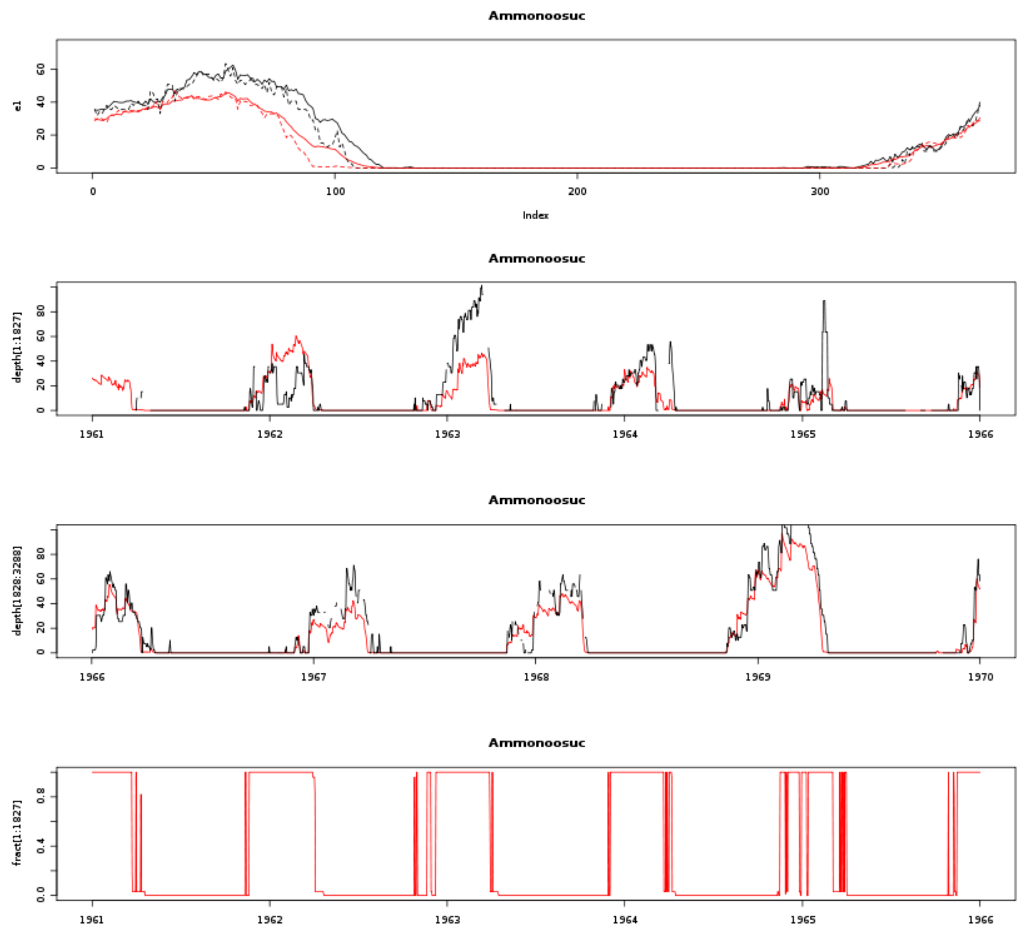


Figure 7: Snow Plots for the Ammonoosuc Basin at Bethlehem

Streamflow

This section contains three figures used to assess overall model skill and representation of physical water balances within the basin. The first figure (Figure 8) presents the mean and median Julian day flow for the Ammonoosuc River at gage 01180000. Both the median and mean streamflow match the observed hydrograph well, though flows are low in the early and mid-winter months. The seasonality of the snow signal is well represented as are summer low flows. However, winter flows are a chronic challenge with the CRVIC model across many of the basins. Whether this is due to meteorological input error or model structure is uncertain. In general, baseflow during these months tends to be lower than observed and many of the flashy one-three day events are completely missed by the model. The relevance to the overall mass balance is a negative bias in mean annual flow.

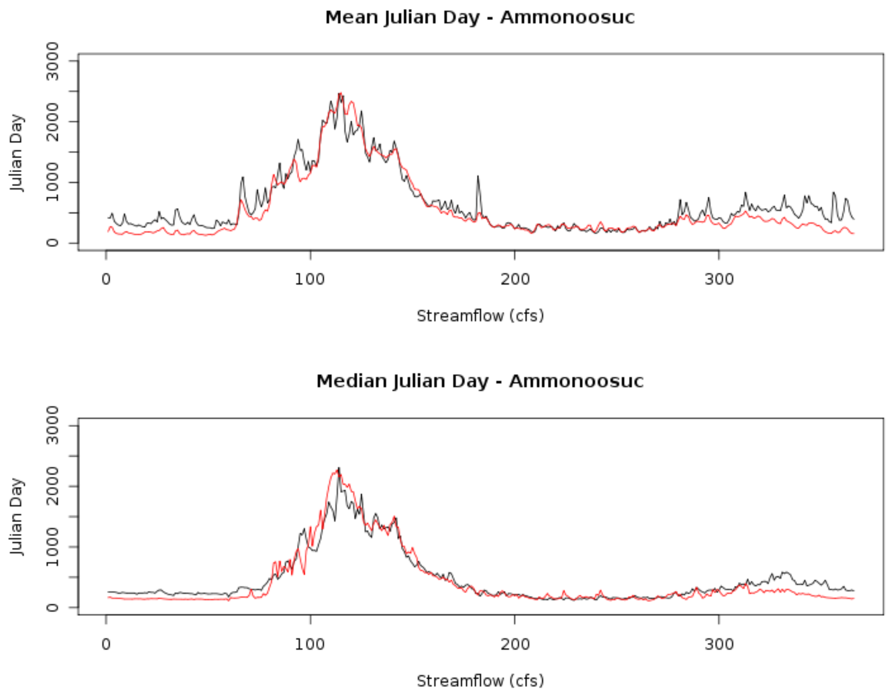


Figure 8: Mean and Median Julian Day Flow

The monthly and daily time series match the observed flow series well. The monthly Nash-Sutcliff (NS) value and daily NS value for the calibration period were 0.8 and 0.62 respectively. These numbers quantitatively reflect the good visual match apparent in Figure 9 and Figure 10. Some of the peak daily streamflow values are missed (Figure 10) though in general, streamflows are well represented between low and high flows.

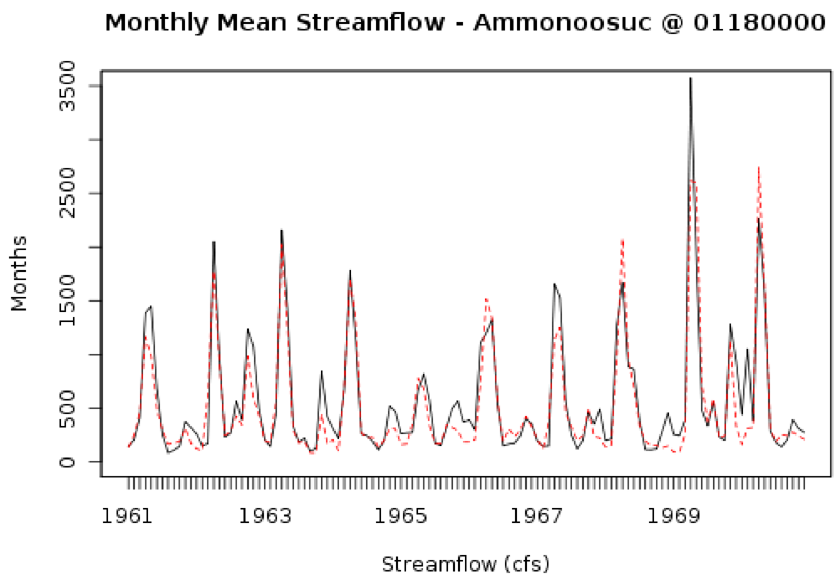


Figure 9: Timeseries of Mean Monthly Streamflow @ 01180000

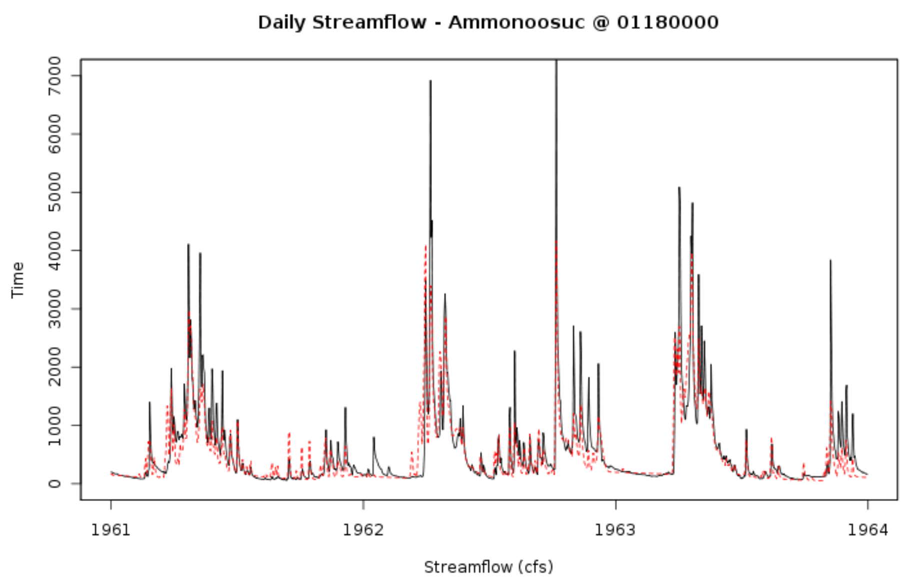


Figure 10: Timeseries of Daily Streamflow @ 01180000

Upper Ammonoosuc

The Upper Ammonoosuc River is located in the northern part of northeast New Hampshire. The total drainage area of the watershed is 252 square miles and the river is 38 miles in length. The river rises in the Pond of Safety (elevation of 1975 feet above the tide) in the town of Randolph, NH. It runs to the north before turning west as it heads towards the town of Northumberland before finally draining into the Connecticut River near the town of Groveton, NH (elevation of 805 feet from tide). There is an 1100 feet differential between the highest and lowest points of the river. The river flows through a couple of dams on its way to the Connecticut River. One major dam that the river flows through is the Godfrey Dam as it passes through Berlin, NH. The USGS maintains only one gage in the Upper Ammonoosuc basin. The location of this gage is on the river near Groveton, NH (Upper Ammonoosuc River near Groveton, NH – 01130000) where the river empties into the Connecticut River. This gage monitors a total drainage area of 232 square miles with discharge records from August 1940 to November 1980 and October 1982 to the current year.

Snow

Snow time series are compared from the meteorological gage at York Pond, NH and the grid cell that overlays the region in Figure 11. The mean and median daily snowpack occurs around April 1st, with approximately 60 cm of snow accumulation (Figure 11). Snow starts accumulating in the middle to end of November, increasing to a maximum around April 1st and then decreases throughout the spring, with typical complete melt occurring around May 1st. The VIC model for the Upper Ammonoosuc River basin predicts snow well across the seasons, though does over-predict snow during the January to April period. The time series match well for most years, though over predicts two snow years (1996 and 1998). In general, the effect of snow melt on runoff timing replicates observed conditions well as evidenced by the next section.

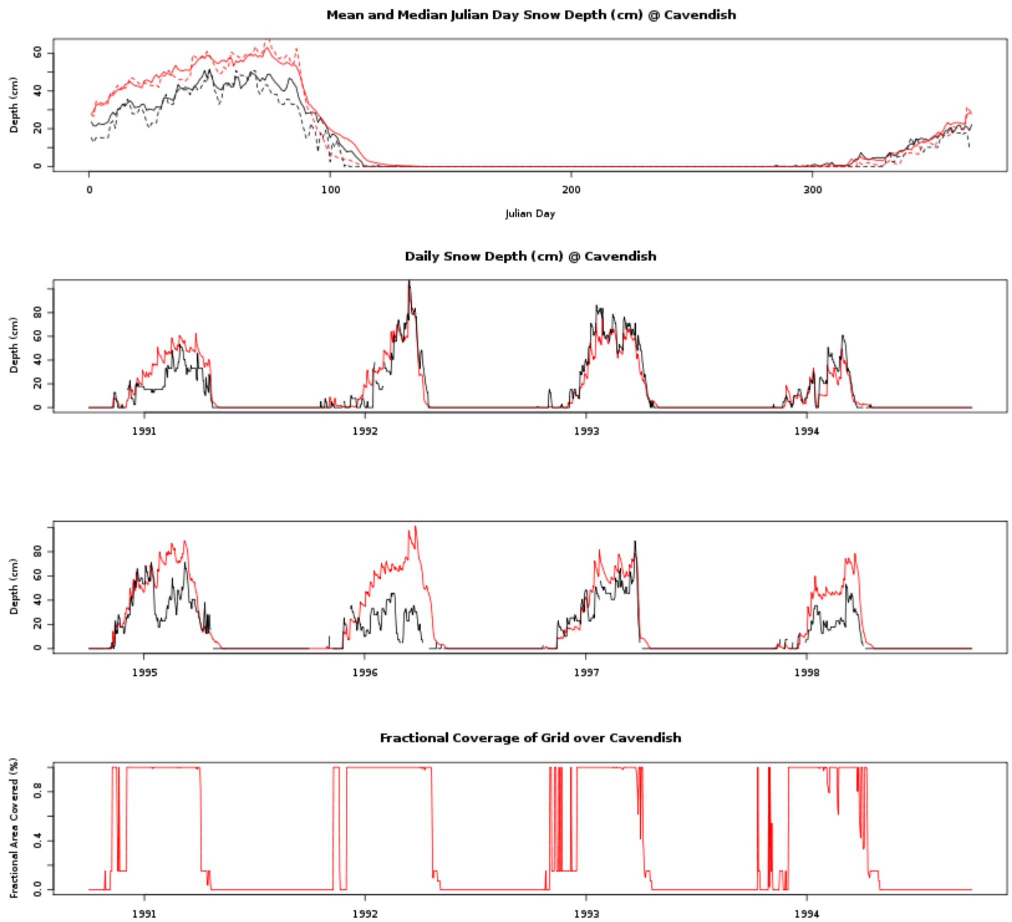


Figure 11: Snow Plots for the Upper Ammonoosuc Basin at York Pond

Streamflow

The CRVIC model replicates streamflow well for the Upper Ammonoosuc Basin. Figure 12 presents the mean and median Julian day flow for the Upper Ammonoosuc River at gage 01130000. Both the median and mean streamflow match the observed hydrograph well, with only slight under prediction occurring during mid-winter months. The seasonality of the snow signal is well represented as are summer low flows.

The monthly and daily time series match the observed flow series well. The monthly Nash-Sutcliff (NS) value and daily NS value for the calibration period were 0.8 and 0.58 respectively. These numbers reflect the good visual match apparent in Figure 13 and Figure 14. Some of the peak daily streamflow values are missed (Figure 14) resulting in a smaller NS value, though in general streamflows are well represented between low and high flows.

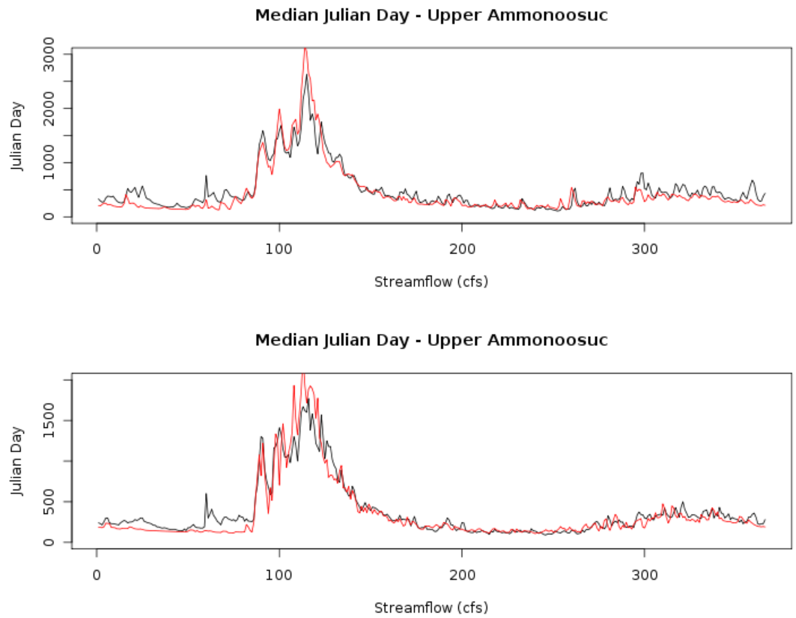


Figure 12: Mean and Median Julian Day Flow at Gage 01130000

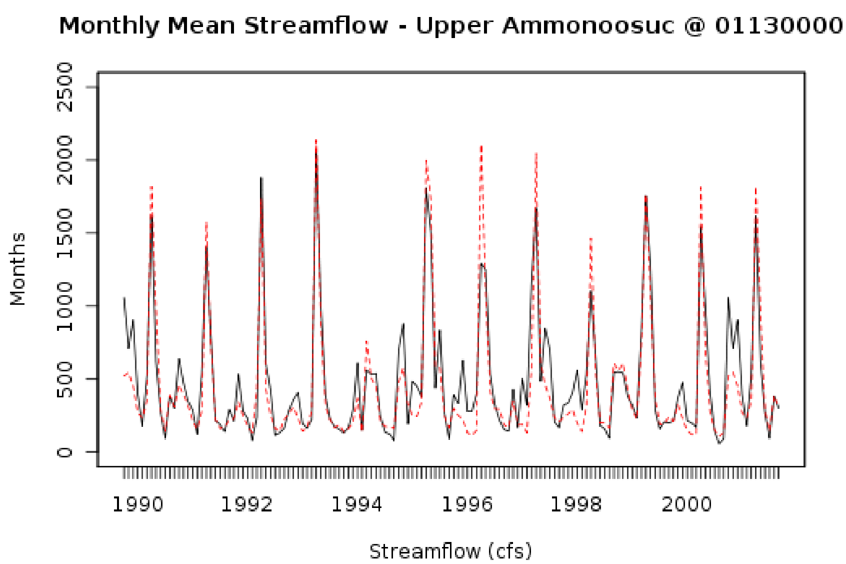


Figure 13: Timeseries of Mean Monthly Streamflow at 01130000

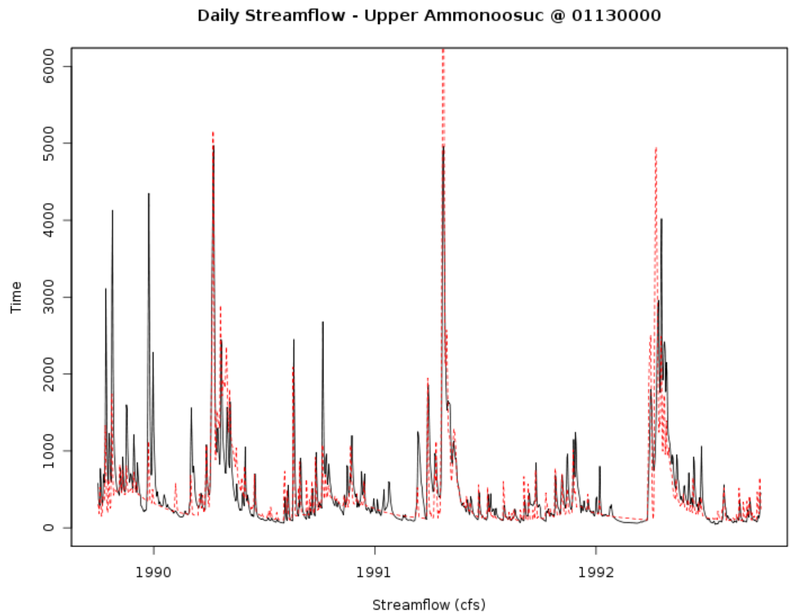


Figure 14: Timeseries of Daily Streamflow at 01130000

Passumpsic

The Passumpsic River basin is a 507 square mile watershed located in northeastern Vermont. The landscape of the basin is mostly forested with the occasional farm and a very small proportion of developed land. The mainstem of the river comes from its east and west branches. The east branch begins in the town of Brighton, VT and is 19.5 miles long while the west branch begins in the town of Westmore, VT and flows for 14 miles before combining with the east branch near East Burke road just northeast of where VT 114 intersects with US 5. The river supports several dams that produce hydroelectric power for the region. These dams include the Passumpsic, Pierce Mills, Arnold Falls, and Gage dams. The USGS maintains four gages along the Passumpsic. These four gages are located in East Haven, VT (VT – 01133000), at Pierce Mills near St Johnsbury, VT (VT – 01133500), at US 5 in St. Johnsbury, VT (VT – 01133550) and at Passumpsic, VT (VT – 01135500). The East Haven gage covers 53.8 square miles and is 943 feet above NGVD29 and has daily data from 7/9/1939 – 7/10/2012. The Pierce Mills gage covers a drainage area of 237 square miles and is 580 feet above NGVD29 and has daily data from 5/26/1909 – 7/24/1919 and is no longer being used. The St Johnsbury gage covers a drainage area of 244 square miles and is 575 feet above NGVD29. The final gage at Passumpsic, VT covers a drainage area of 436 square miles and is 500 feet above NGVD29 and has daily data from 11/15/1928 – current.

Snow

Snow time series are compared from the meteorological gage at Saint Johnsbury, VT and the grid cell that overlays the region. The peak mean and median daily snowpack occurs much earlier in this basin relative to others at approximately mid-February, with approximately 45 cm of snow accumulation (Figure 15). Snow starts accumulating in the middle to end of November, increasing to a maximum around February 15th and then sharply decreases by about 15 cm throughout the mid-winter period, staying at 30 cm until April 1st and then sharply decreasing until complete melt occurring around May

1st. The VIC model for the Passumpsic River basin predicts snow well across the seasons, though does over-predict median snow during the late February to April period. The time series match well for most years. As with other northern basins in the Connecticut River watershed, the effect of snow melt on runoff timing replicates observed conditions well.

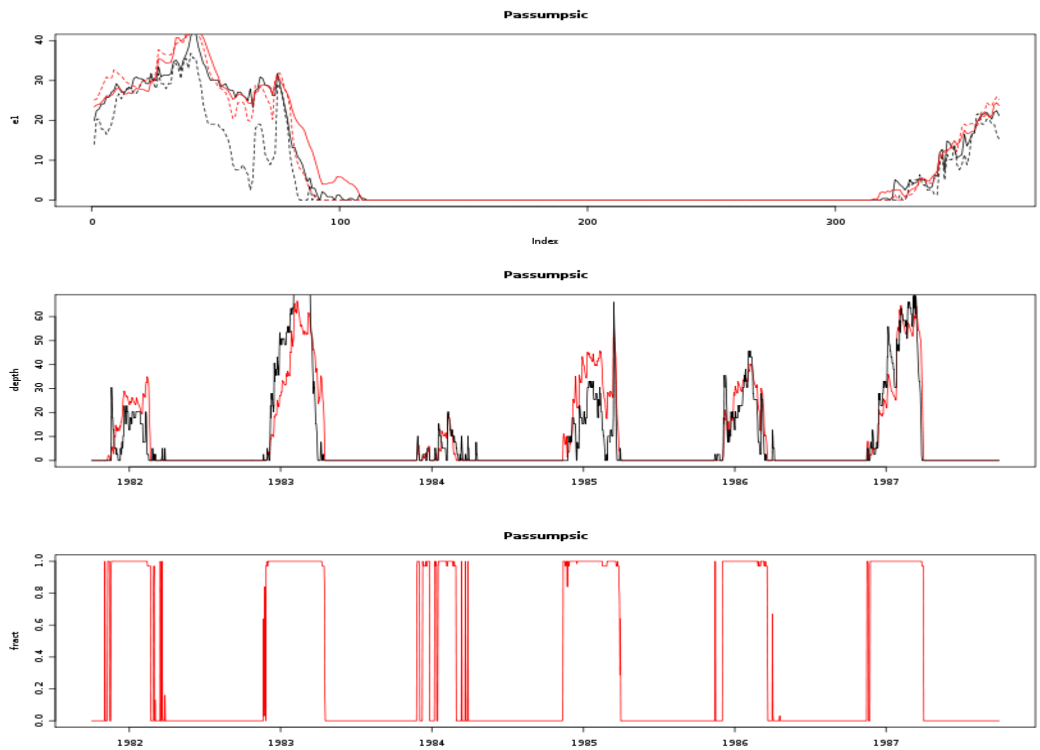


Figure 15: Snow Plots for the Upper Passumpsic Basin at Saint Johnsbury

Streamflow

The CRVIC model adequately simulates streamflow in the Passumpsic River. Seasonally, the model replicates streamflow well, though winter streamflow is under-predicted in January and February. Peak flows in late March are under simulated as depicted in Figure 16 though median flows are simulated well. Streamflow in late spring through early winter periods are simulated well in both the mean and median flow metrics. Monthly mean streamflow follows historic patterns, with over and under-prediction occurring about equally over the evaluation period (Figure 17). The daily time series (Figure 18) replicates the observed flows except at peak events where under prediction occurs. Low percentile flows are modeled well as is the seasonal cycle due to snow melt. The daily NS value for the Passumpsic River is 0.46 and the monthly NS value is 0.6, both lower than other northern basins but acceptable. Discussions with TransCanada have suggested that the gage used for calibration produces inaccurate streamflow data at certain times of the year, namely winter periods.

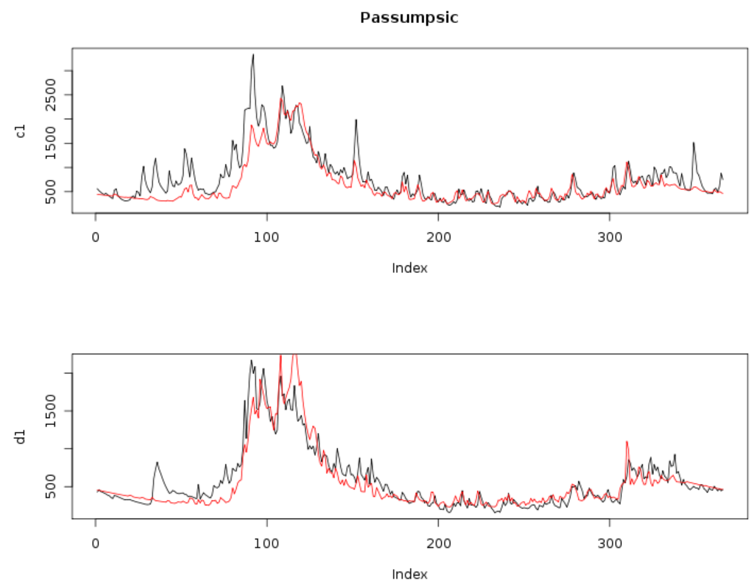


Figure 16: Mean and Median Julian Day Flow at Gage

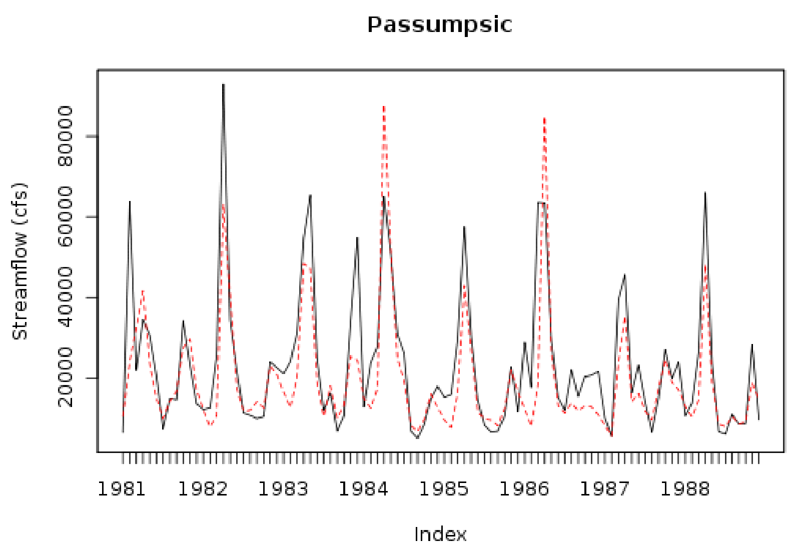


Figure 17: Timeseries of Mean Monthly Streamflow at

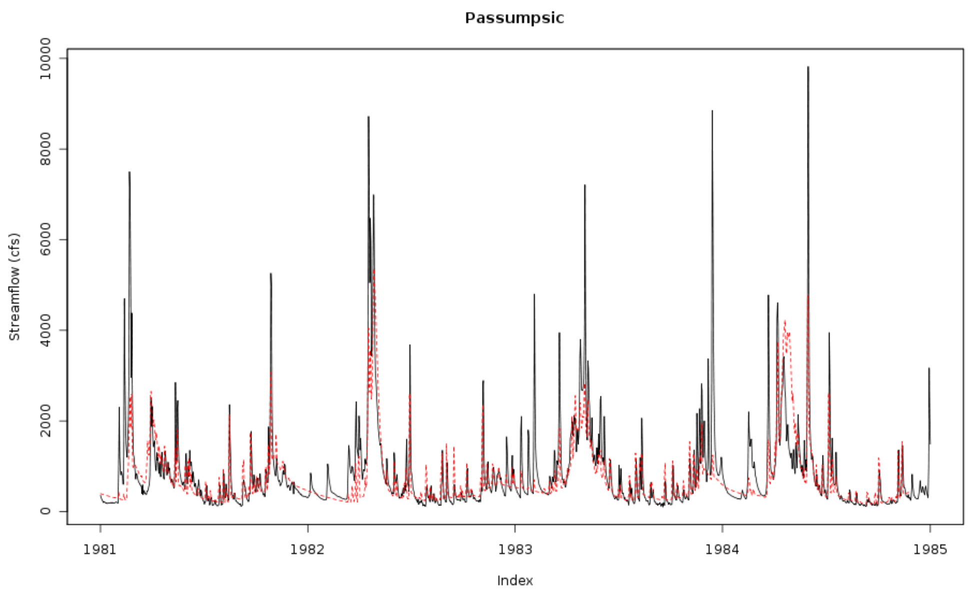
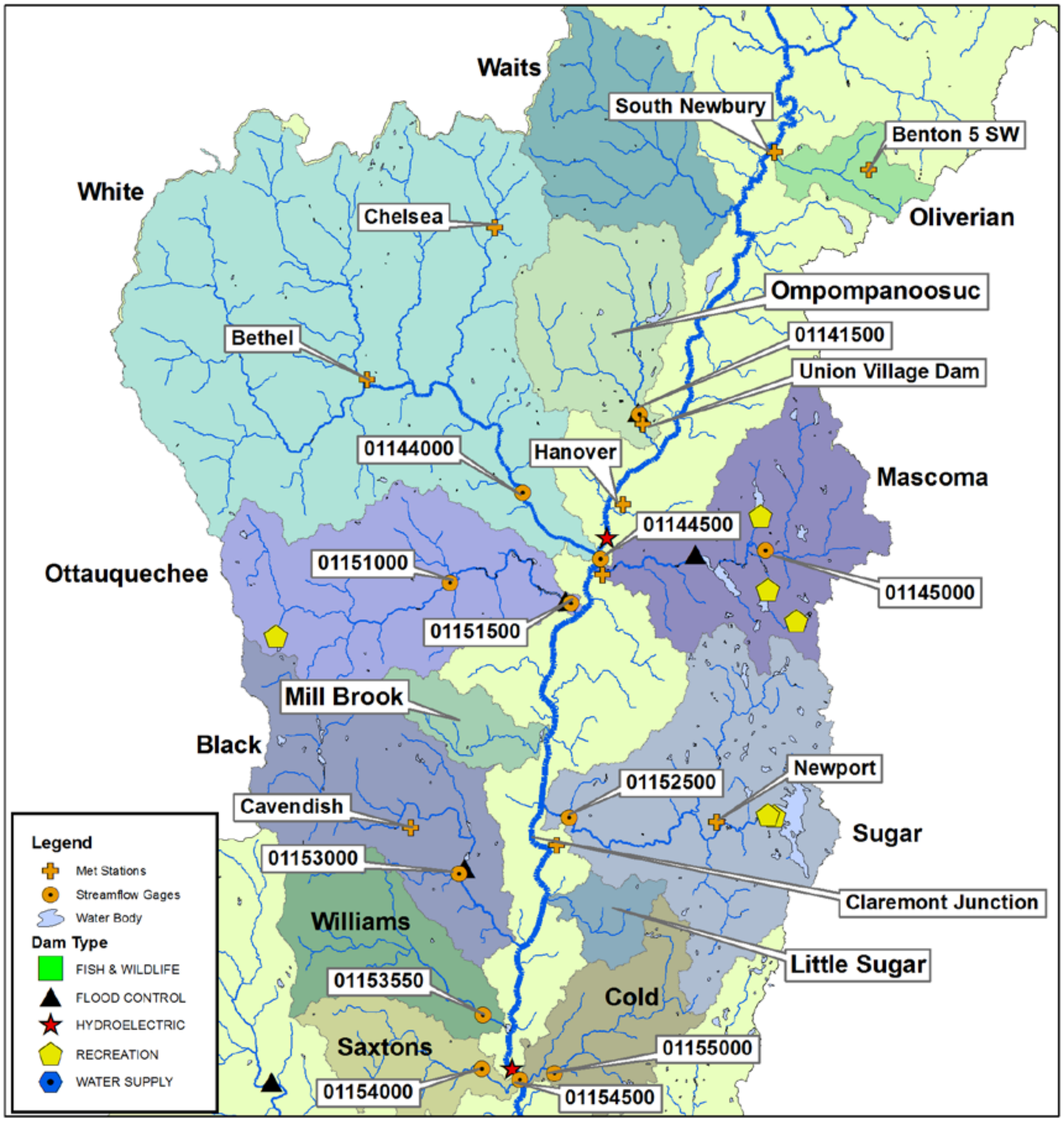


Figure 18: Timeseries of Daily Streamflow at

Parameter Estimation and Verification Results - Middle Third Basins



White

The White River Basin is located in central Vermont and covers a total drainage area of 710 square miles. The White River begins in Ripton, VT at the base of the Battell Mountain and flows to the south and east before merging into the Connecticut River near Hartford, VT. Unlike the five major tributaries of the White River, there are no dams that restrict flow along the mainstem of the river. Of the 17 major river basins in Vermont, the White River Basin has the fourth fewest lakes and ponds with only 39. Most of the surrounding landscape of the river is forest with some agricultural lands. USGS currently maintains one gage along the White River located in West Hartford, VT (VT – 01144000) immediately before the river merges with the Connecticut River. The gage covers a drainage area of 690 square miles and has an elevation of 374 feet above NGVD29. There is daily data available for this gage from 6/9/1915 to the current year.

Snow

Snow data for the Bethel gage in Vermont has variable consistency in measuring depth over the calibration time period. The model replicated historic data well. Both the mean and median snow fall match well (Figure 19), with both snow accumulation and melt replicated. The model accumulated slightly more snow during the mid-to-late winter months (February through early March). Snow pack begins accumulation in mid-to-late November and is typically completely melted by the beginning of April. The maximum snow depth (40 cm) for this part of part of the White River basin occurs around late January and holds steady until approximately the middle of March, where a rapid melt occurs until April.

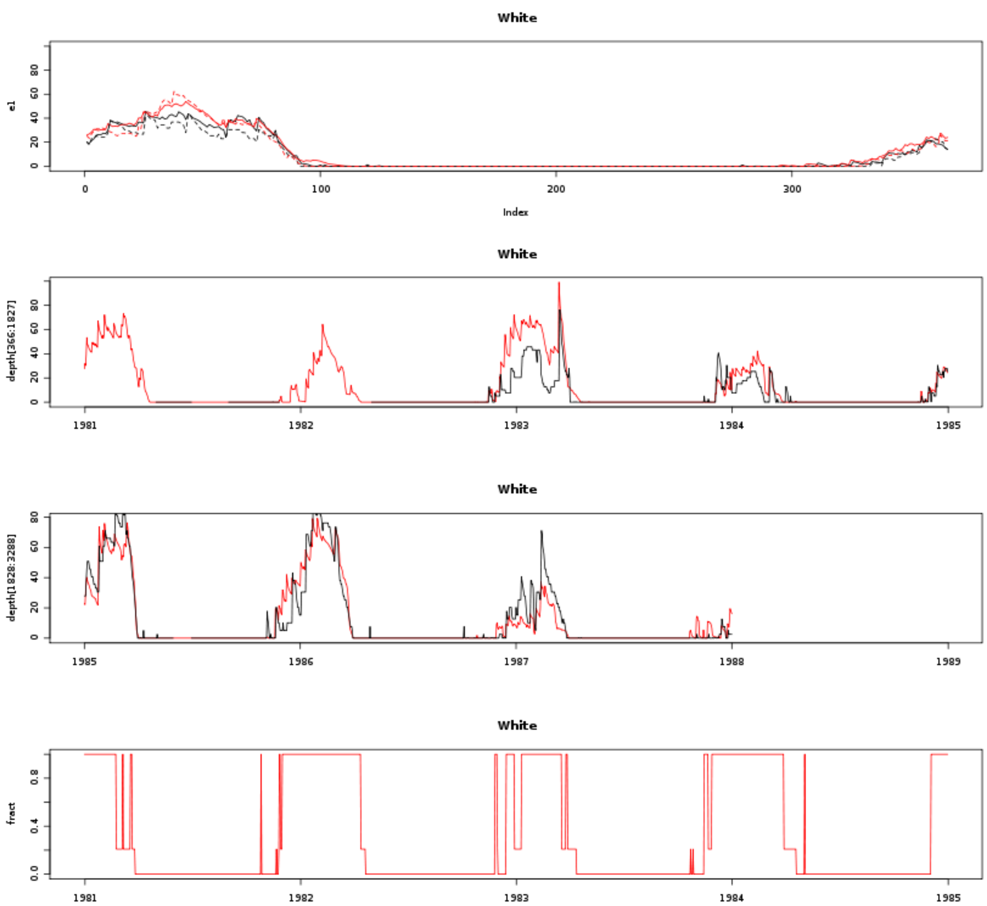


Figure 19: Snow Plots for the Upper White River Basin at Bethel

Streamflow

Streamflow in the White River basin is well replicated in the spring and summer months, with a negative model bias in the late fall, early and mid-winter months (Figure 20). Observed and modeled mean monthly flows are presented in Figure 21, and show good agreement. The monthly and daily NS-values for the White River basin are 0.77 and 0.62 respectively, suggesting a good model fit. The daily time series (Figure 22) is replicated well, though some high flows are over estimated in some years and under-estimated in others. The spring recession is captured well across all years and summer flows match well with the observed record, though heavy rainfall events in the summer tend to be slightly over-estimated.

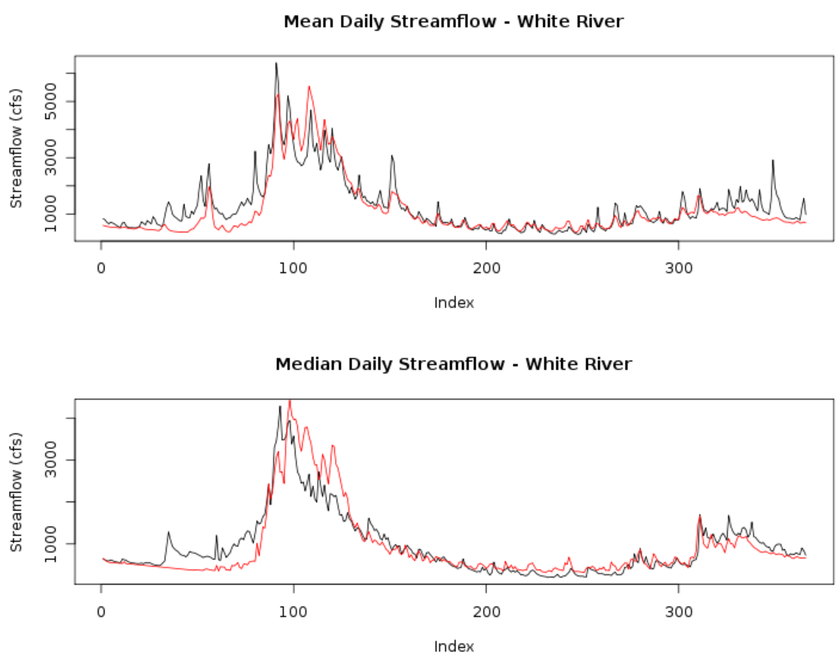


Figure 20: Mean and Median Julian Day Flow at Gage

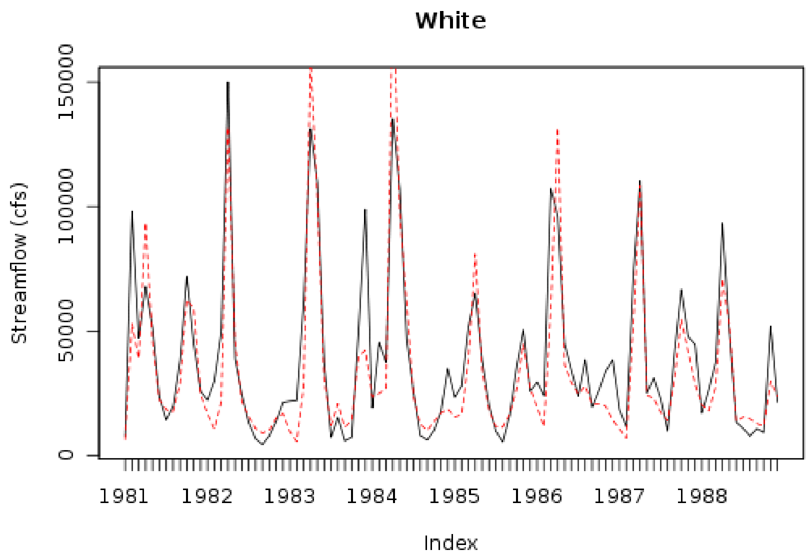


Figure 21: Timeseries of Monthly Streamflow at

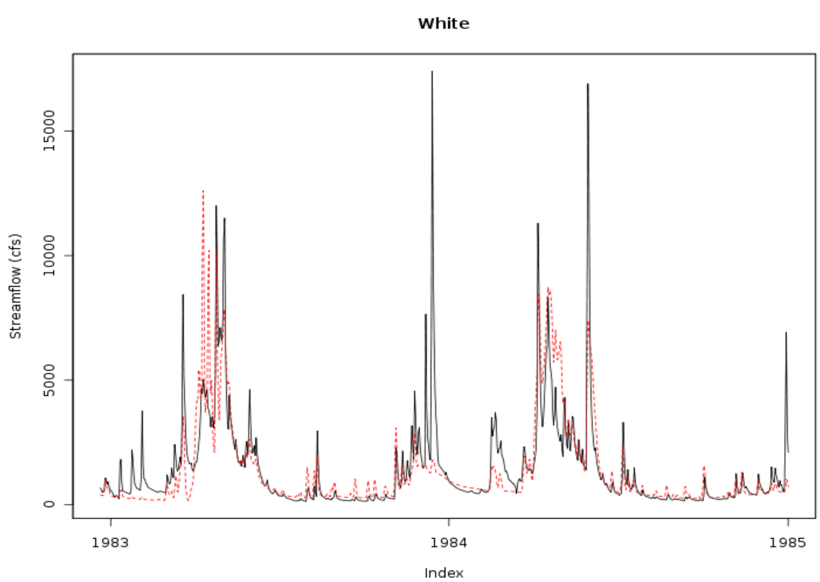


Figure 22: Timeseries of Daily Streamflow at

Black

The Black River Basin is a 40 mile river located in southern Vermont. The river begins in Black Pond in Plymouth, VT and combining with its tributaries, creates a total drainage area of 202 square miles. The surrounding landscape of the basin is mostly mountainous and hilly. The river finally empties into the Connecticut River a quarter mile downstream of the Chesire Toll Bridge in Springfield, VT. The Black River passes through many dams (~30), most of them small. However, the dam at North Springfield Lake is large and very important for flood control for towns in the neighboring area and for flooding into the Connecticut River. USGS currently maintains two gages along the Black River. One is located in North

Springfield, VT (VT – 01153000) and the second in Coventry, VT (VT – 04296000). The North Springfield gage covers a drainage area of 158 square miles and is 445 feet above NGVD29. This gage has collected historical data from 11/26/1929 – 9/30/1989 and it collecting data from 3/13/2012 – current. The Coventry, VT gage covers a drainage area of 122 square miles and is 710 feet above NGVD29. This gage has historic data from 10/1/1951 – 9/30/2011 and is collecting data from 10/1/2007 – current.

Snow

When comparing the observed snow data collected for the Black River Basin with the model's output it is found that overall the CRVIC does a decent job of predicting snowfall in the basin. Figure 23 (mean and median Julian day snow depth) shows the median and mean snow depth over the course of an average year. It is observed that both the mean and median CRVIC outputs are consistently roughly 10 cm over the historical data of the area. The VIC model does a great job of predicting the low to zero snow fall depths, which can be seen in the summer months and continue into the fall and very early winter. Figure 23 shows daily snow depth from the time period of 1981-1988. This figure reveals that again, the CRVIC model tends to over-predict snow fall totals compared to historical data. In 1983, the model over-predicts the historical record by almost double. For most of the other years, the model still over-predicted but was within a few centimeters of the historical data. In the early winter of 1987, the model under-predicted snow totals.

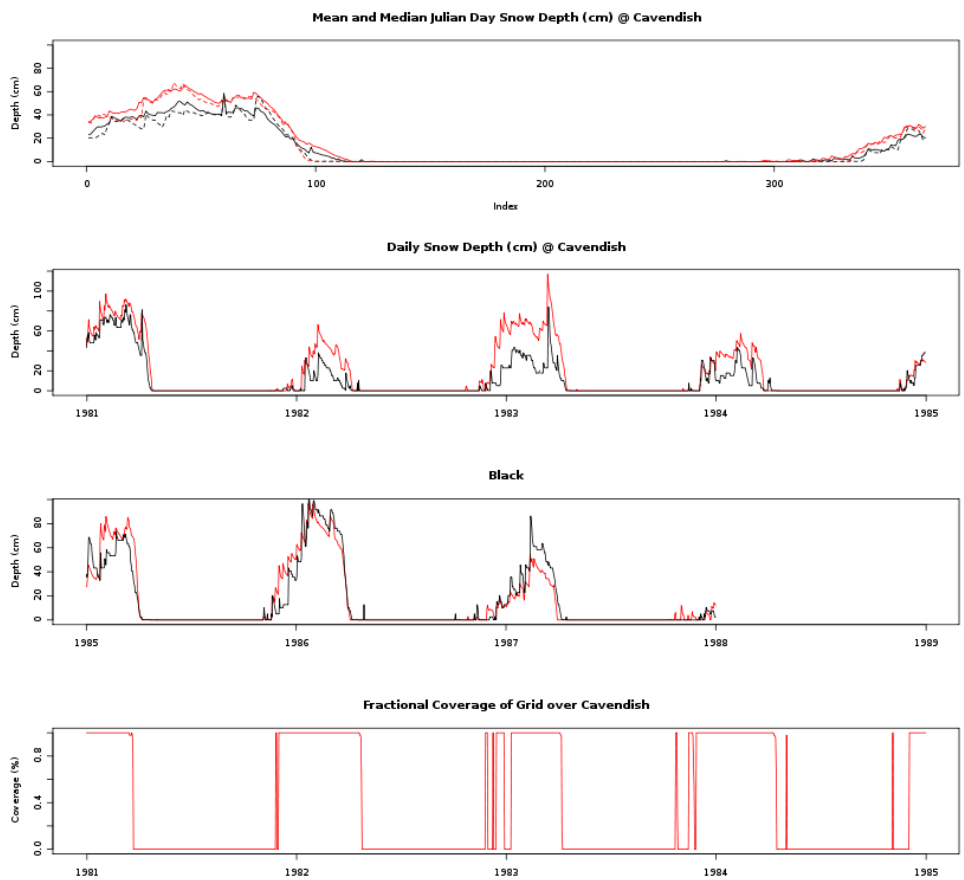


Figure 23: Snow Plots for the Upper Black River Basin at Cavendish

Streamflow

The CRVIC model does a good job of predicting streamflow data for the Black River Basin. When looking at the Nash-Sutcliffe numbers for the model, the daily value is 0.44 but the monthly value is 0.81. The high monthly Nash-Sutcliffe value attests to the accuracy of the model in monthly response and quantity. The lower daily NS-value may arise from the regulation of dams upstream of the gage.

Figure 27 shows the mean Julian daily streamflows at gage 01153000. The model over-predicts higher mean and median flows during the spring months compared with the observed record. This may be due to the operation of USACE facilities within the basin regulating floods. The observed snow signal in the winter months and the summer low flows both accurately represent the seasonality of the streamflow. Figure 25 demonstrates the monthly mean streamflow of the Black River. The model fit represents the range of mean monthly flows well over a long time period, with minimal bias. Daily streamflow is also well represented despite the lower NS-value. This may be due to the fact that the regulation impacts high and low flow significantly and can shift daily flow timing. When the timing of a storm event is shifted or peak flows are altered relative to a 'natural' flow regime, we can expect large decreases in the NS-value derived from comparing a flow model generating 'natural' flow and a human-altered observed record. Overall, the modeled streamflow for the Black River replicate the observed data well.

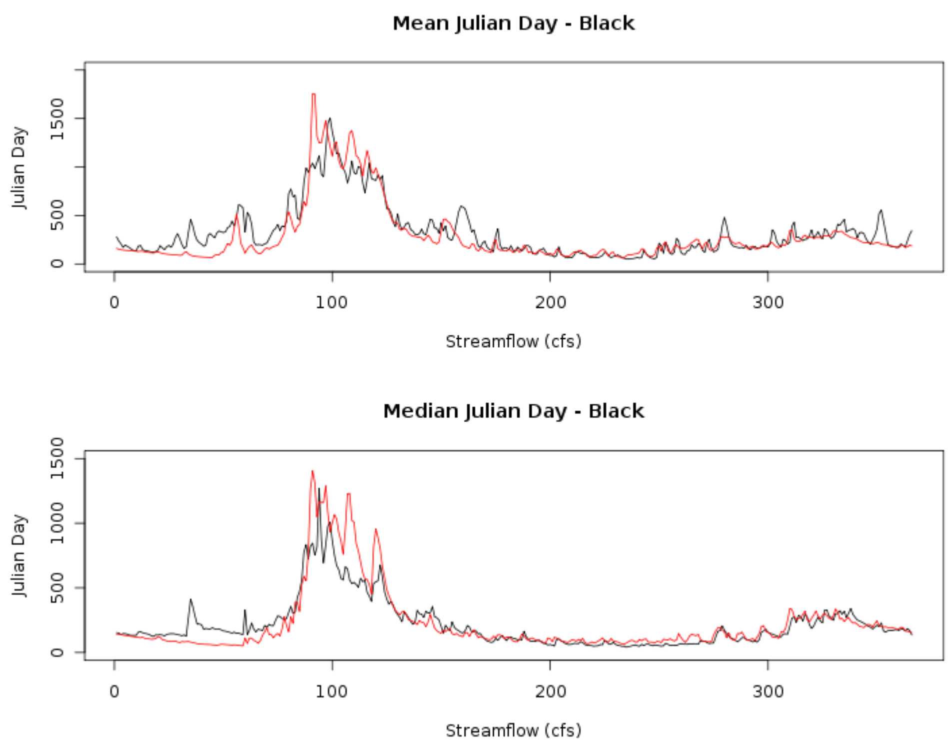


Figure 24: Mean and Median Julian Day Streamflow at

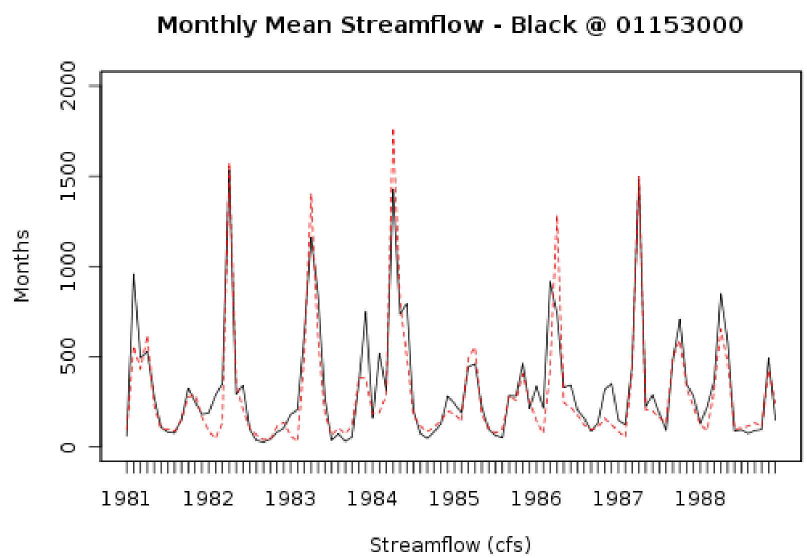


Figure 25: Timeseries of Monthly Streamflow at

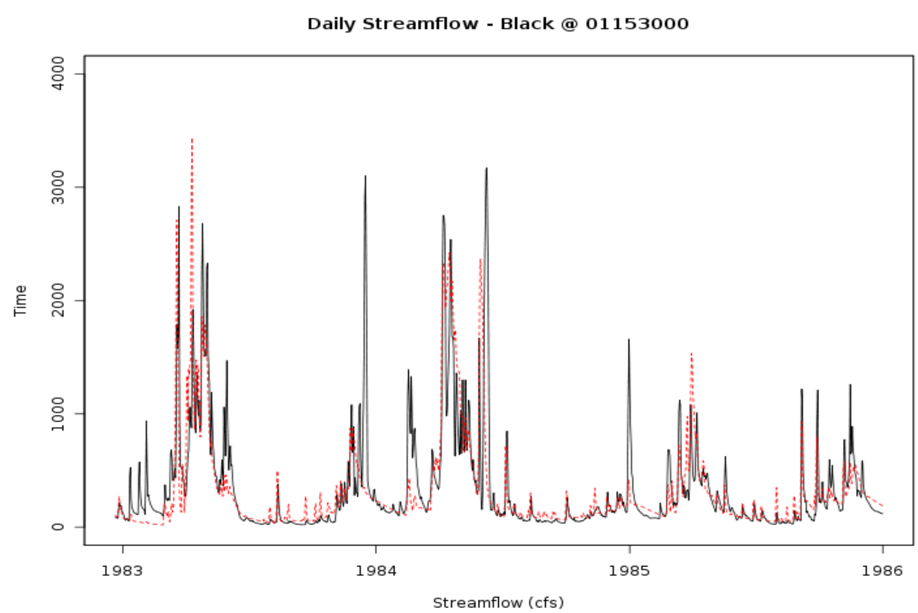


Figure 26: Timeseries of Daily Streamflow at

Mascoma

The Mascoma River Basin is a 153 square mile river basin that is located entirely within New Hampshire. The headwaters of the Mascoma originate from Cummins Pond and flows west towards the Connecticut River. The river passes through the Mascoma Lake and then numerous small hydroelectric dams before entering into the Connecticut River in West Lebanon, NH. USGS currently maintains only one gage in the Mascoma River Basin where the Mascoma River meets the Connecticut River in West

Lebanon, NH (NH - 01144500). This gage covers a drainage area of 4092 square miles at an elevation of 321 feet above NGVD29 and has collected daily data from 11/1/1911 – current. USGS also had previously operated two gages along the Mascoma River to monitor flow in this basin. The first gage was located on the Mascoma River at West Canaan, NH (NH – 01145000). The gage covered a drainage area of 80.5 square miles and had an elevation of 815 feet above NGVD29. The gage had collected daily discharge data from 7/26/1939 – 9/30/1978. The second gage was located on the Mascoma River at Mascoma, NH (NH – 01150500). The gage covered an area of 153 square miles at an elevation of 740 feet above NGVD29 and collected daily data from 8/16/1923 – 9/30/2004. Due to no immediate meteorological station within the basin, modeled snow outputs could not be compared against observed values.

Streamflow

The CRVIC model adequately represents streamflow within the Mascoma River Basin. The daily NS efficiency coefficient is 0.50 and the monthly NS efficiency coefficient is 0.842, both values deemed acceptable. As with many of the basins in the Middle Connecticut, the daily values tend toward lower values as upstream impoundments tend to either mute flood peaks, increase/decrease summer flow, or shift the timing of runoff. The high monthly NS-values indicate a good mean response from the model over long time periods, suggesting the overall physical mechanics of runoff processes are being captured correctly.

Figure 27 presents the mean daily streamflow data from the CRVIC model and compares it to the historical mean daily streamflow data. Overall, the model does a good job of replicating the seasonality of streamflow within the basin (Figure 28). The model does, however, under-predict the daily streamflows during the mid-winter months. Figure 29 presents the daily modeled and observed streamflow. Both the high and low flows are well replicated over the calibration period, as is the spring recession. There is no consistency in the model of either over-predicting or under-predicting the high peaks of flow. Overall, the CRVIC model matches observed streamflow in the Mascoma River Basin well.

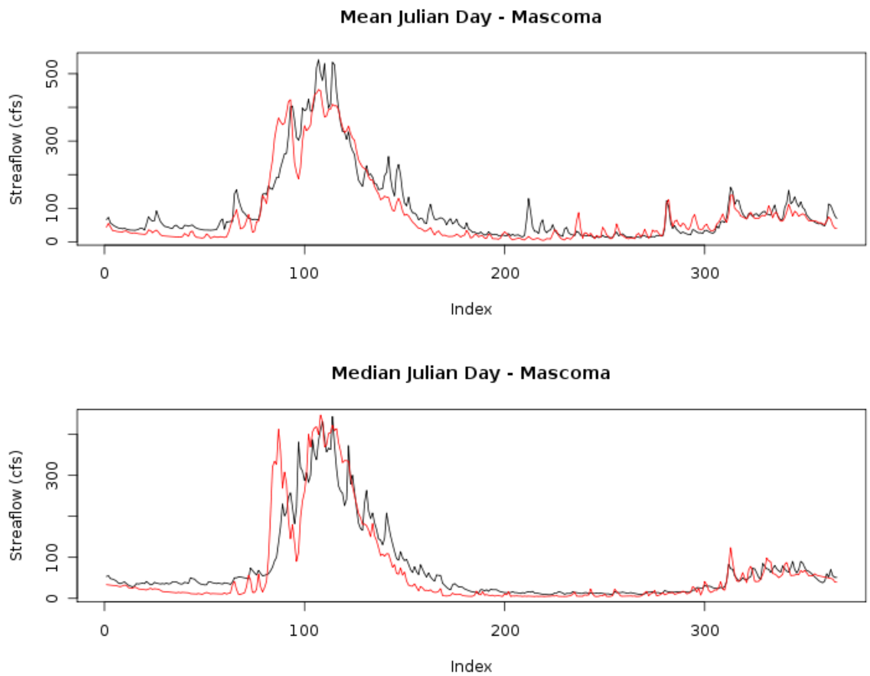


Figure 27: Mean and Median Julian Day Streamflow at Gage 01145000

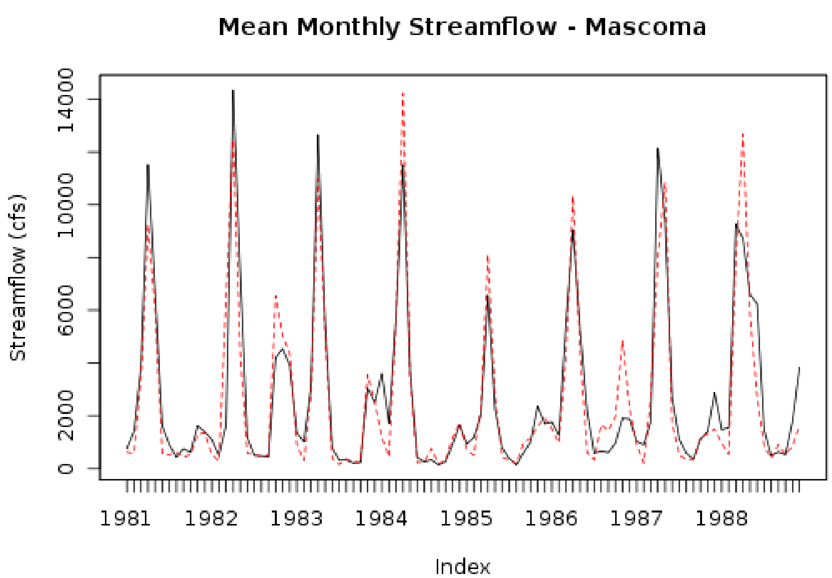


Figure 28: Monthly Time Series for Mascoma River at Gage 01145000

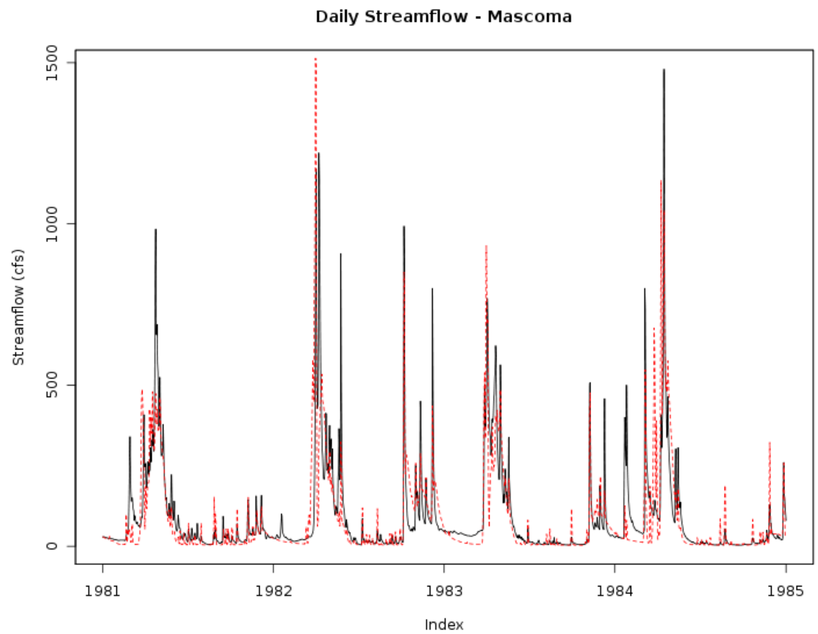


Figure 29: Daily Time Series of Mascoma River at Gage 01145000

Ottaquachee

The Ottauquechee River Basin is located in east central Vermont with a mainstem length of 38 miles and drains an area of 223 square miles. The landscape of the watershed is rugged and hilly with a few large meadows. Most of the land is forested with equally small percentages of agricultural land and developed land (roughly 7,000 - 8,000 acres each). The river originates in the Green Mountain Range in the town of Killington, VT. From there it flows northeasterly before flowing to the south and reaching the North Hartland Dam before finally emptying into the Connecticut River. USGS currently maintains two gages along the Ottauquechee River. In North Hartland, VT roughly 0.3 miles downstream of the North Hartland Dam is gage VT – 01151500. This gage covers a drainage area of 221 square miles at 336 feet above sea level. The period of record for this gage is from October 1930 to the current year. The other current gage is located in West Bridgewater, VT (VT – 01150900). This gage covers a drainage area of 23.4 square miles at an elevation of 1148 feet above sea level. The gage has collected data from February 1985 to the current year.

Streamflow

The CRVIC model does a decent job of modeling the streamflow of the Ottaquachee River Basin. When judging the model based on Nash-Sutcliffe numbers, it is found that the daily and monthly NS numbers are 0.23 and 0.70 respectively. From a day-to-day perspective, the NS number is low and demonstrates that the model is not incredibly accurate; but it is worth noting that there are many more factors that go into day-to-day streamflow than the ones that are used in the model. When looking on a broader perspective, like on a monthly basis, the NS number is in a very acceptable range. This higher NS number demonstrates that on a broader scale the model does a great job of modeling the streamflow patterns of the Ottaquachee River.

Figure 30 portrays the average historical daily streamflow data of the Ottaquachee and compares it to the models average output. Overall, the model does a great job of following the seasonality of the river basin. It follows the seasonal pattern of lower flows in the summer and winter with higher flows in the spring because of the wet spring climate and the melting of the snowpack. When inspected more closely, it is noted that the model tends to under-predict the winter stream flows in January and February while over-predicting in the early stages of the spring before matching up with the historical data in the later spring and early summer months. Figure 31 and Figure 32 show the monthly and daily modeled and observed data over the calibration period. From these figures it is evident that the model over-predicts flood peaks in most years relative to the observed record (early 1982, mid 1983, mid 1984, mid 1986), especially when the peaks are most extreme. The model replicates less extreme peaks and the overall seasonality of the system well. Overall, the model does a decent job of predicting streamflows for this river basin over a long period of time despite the low daily NS-value.

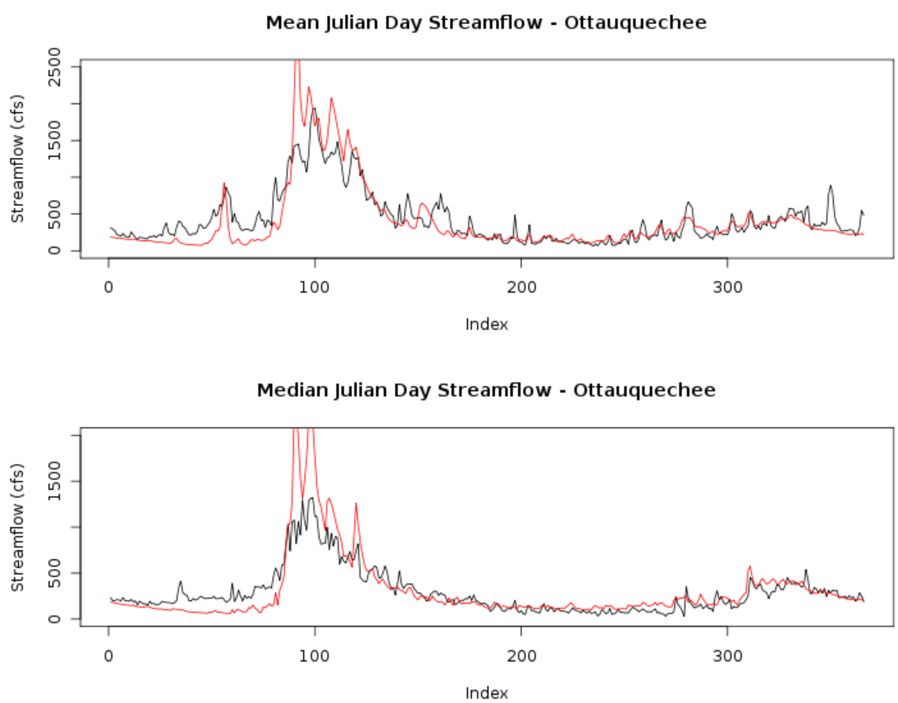


Figure 30: Mean and Median Julian Day Streamflow at Gage 01151500

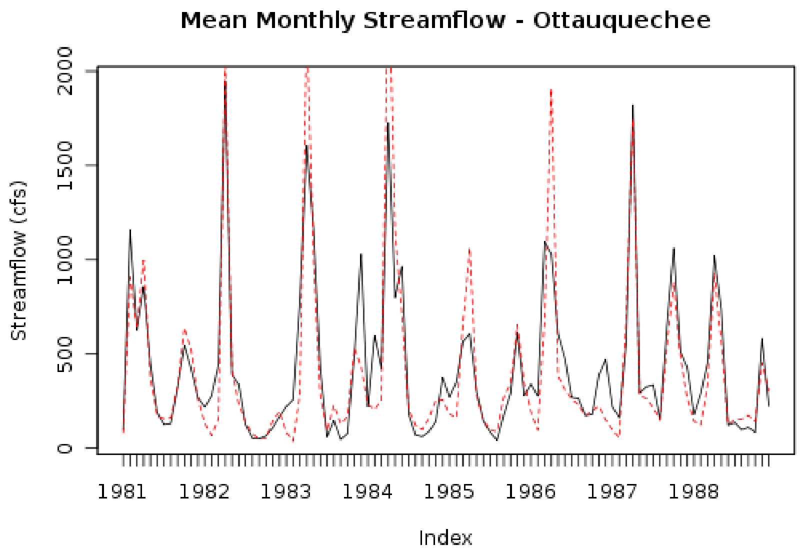


Figure 31: Monthly Time Series of Ottaquachee River at Gage 01151500

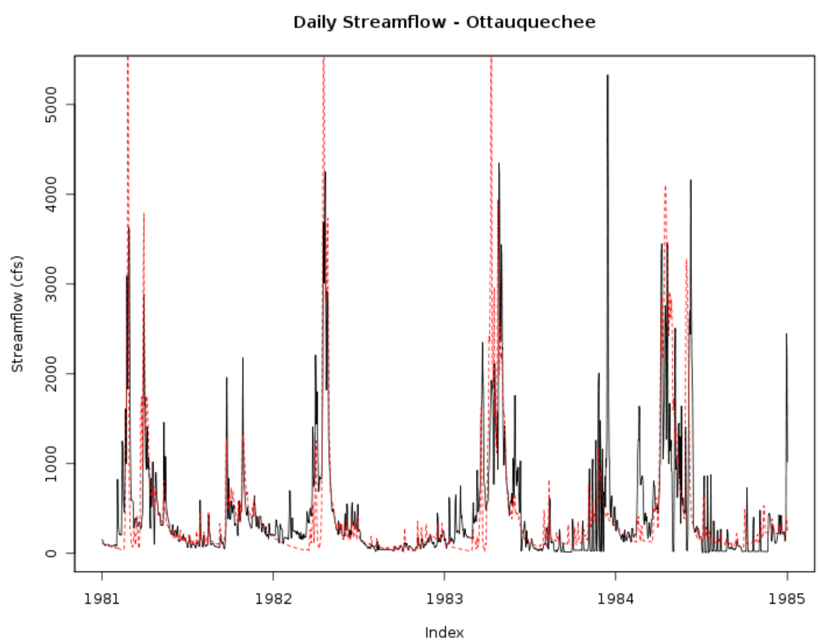


Figure 32: Daily Time Series of Ottaquachee River at Gage 01151500

Parameter Estimation and Verification Results - Lower Third Basins

West

The West River drains about 420 square miles of drainage area in southern Vermont. The 50 mile long river has its headwaters in the Green Mountains south of Rutland, VT and flows southeast to Brattleboro, VT where it enters the Connecticut River. The West River basin has a large elevation range, over 1100 meters, relative to the other sub-basins with a mean elevation of around 470 meters above sea level. Most of the basin consists of small mountains and hills covered with forest. There are 21 dams on the river including two Army Corps flood control dams at Ball Mountain and Townshend. A comparison between the SYE generated data and observed flows shows that the streamflow is regulated by these dams. There are only two USGS maintained flow gages with long monitoring records. One gage is located a couple miles downstream of Ball Mountain Dam (West River below Winhall River, Near Jamaica, VT- 01155349) monitors 179 square miles of drainage area. The second gage is located in Newfane, VT (West River at Newfane, VT-01156000), and monitors 308 square miles of drainage area. The record for this gage ends on September 30, 1989. Since the only other useful gage in the basin is located far upstream above one of the major dams, the gage at Newfane is used for calibration. Drainage area scaling was used to account for the streamflow below the gage. This is acceptable as the land cover and terrain for the area omitted from the gage data is similar to that of the rest of the basin.

Snow

In the West River basin, sufficient observed data is missing for the years 1986 through 1989. The model predicts two distinct peaks of mean and median snow depth at around 90cm. The first peak comes in mid-February while the second occurs in mid-March. The timing of peaks matches that of the observed snowpack, but the model predicts an added 30cm in depth (Figure 33). Over-prediction by the VIC model is fairly consistent throughout the calibration period for the grid cell over the meteorological station at Peru, VT. In this basin snow accumulation begins as early as November 1st and is completely melted before May 1st. In 1982 and 1983 the model appears to accurately capture the timing of snow depth variation with error in the depth only. Comparison in the remaining calibration years is difficult due to missing data, but it is evident that the over-prediction of depth continues. The prediction error results in a longer time period of modeled snow coverage throughout the basin. The error in modeled snowpack is a main source of streamflow prediction error discussed in the next section.

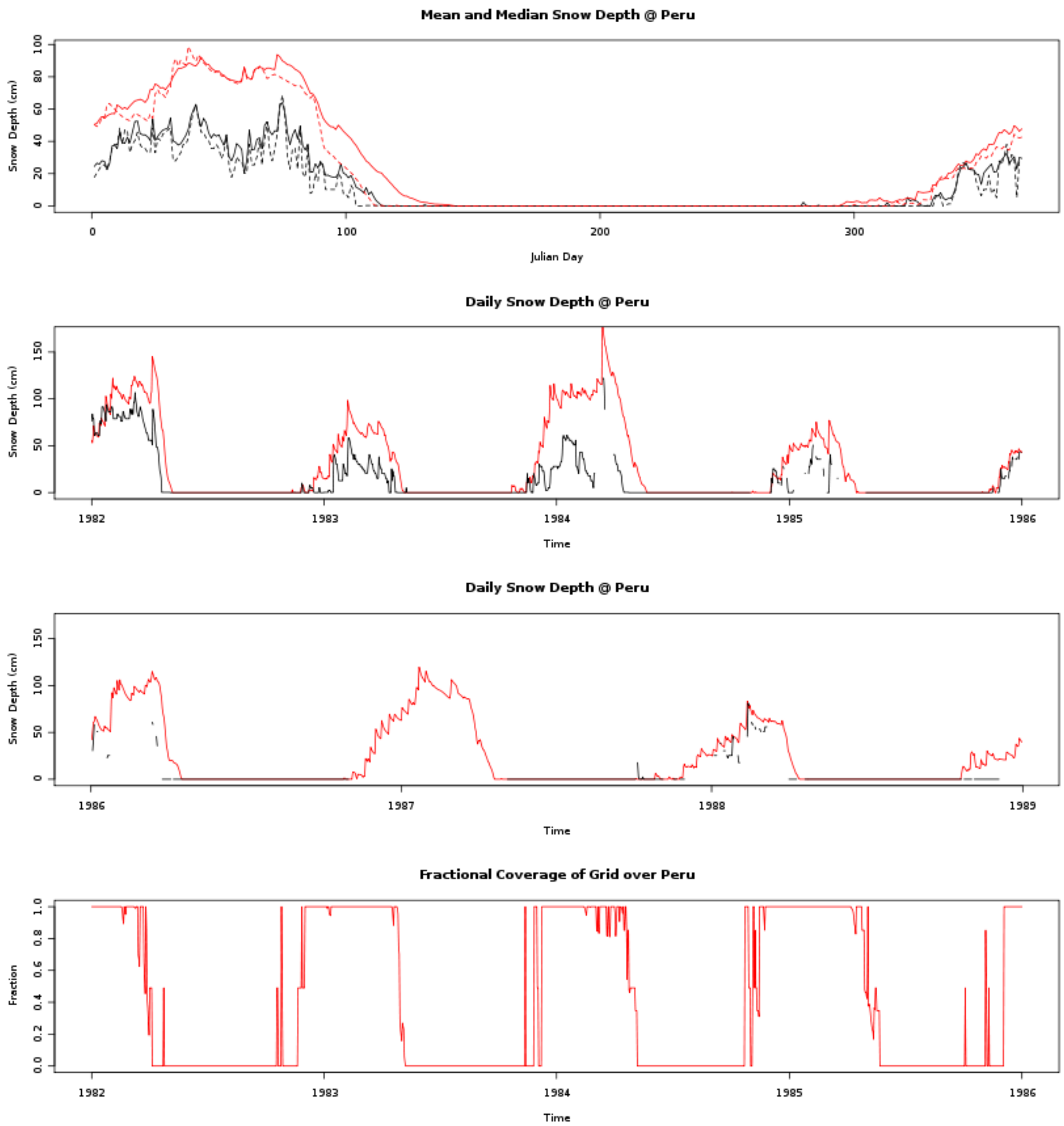


Figure 33: Snow Plots for the West River Basin at Peru, VT

Streamflow

The model skill and physical water balance of the West River basin is described in this section. Figure 34 presents the mean and median Julian day flows at gage 01156000 in Newfane, VT. In this basin, the model under-predicts winter flows and over-predicts the volume of spring flows. This error results from the over-prediction of snow depth which impacts the water balance. Precipitation that should contribute to winter flow is instead stored in snowpack and augments spring flow volume. The magnitude of mean and median peak streamflow is accurate, but extended longer into the season than the observed. The baseflow throughout the remaining part of the year is accurate, but the model consistently misses increased flow events.

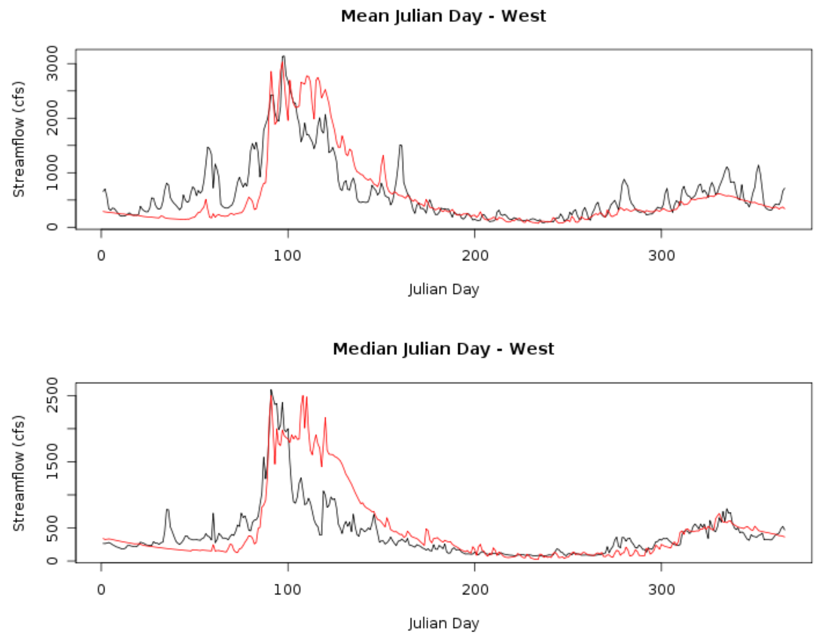


Figure 34: Mean and Median Julian Day Flow

Figure 35 shows the mean monthly streamflow at the Newfane gage on the West River. This plot provides a visual of the mass balance error in this basin of the model caused by inaccurate snow depth. The modeled monthly average winter flows are below the observed. This error contributes to the slight delay in spring flows from the observed. The average monthly spring peaks are captured well by the model. Figure 36 shows the daily streamflow at the Newfane gage. Timing of high flow events is somewhat accurate, but the magnitude is consistently off. Additionally, the model misses some events altogether. Missed events in the winter and spring can be attributed to the aforementioned error in snowpack prediction. Error in the summer may be due to the soil's response to precipitation events. The baseflow is accurate, but the flashiness of streamflow in drier months is not captured by the model indicating error in soil saturation thresholds.

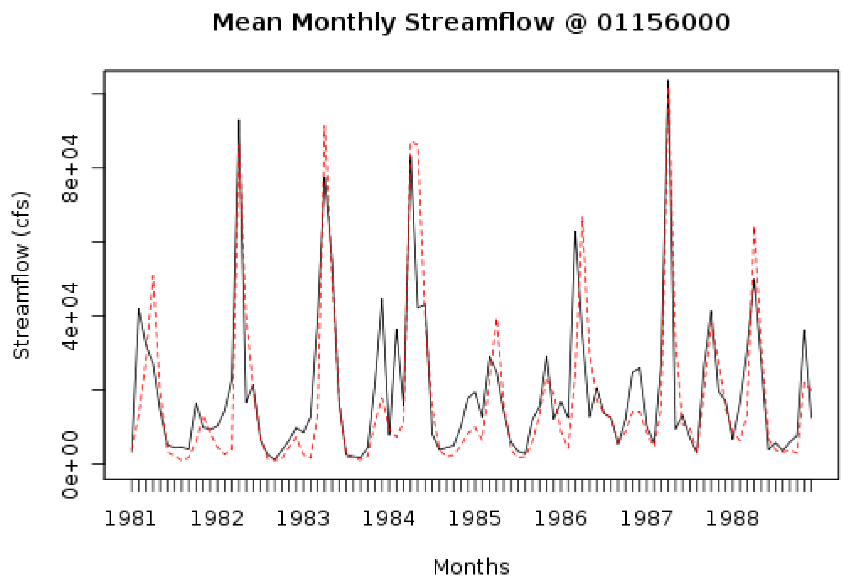


Figure 35: Time series of Mean Monthly Streamflow

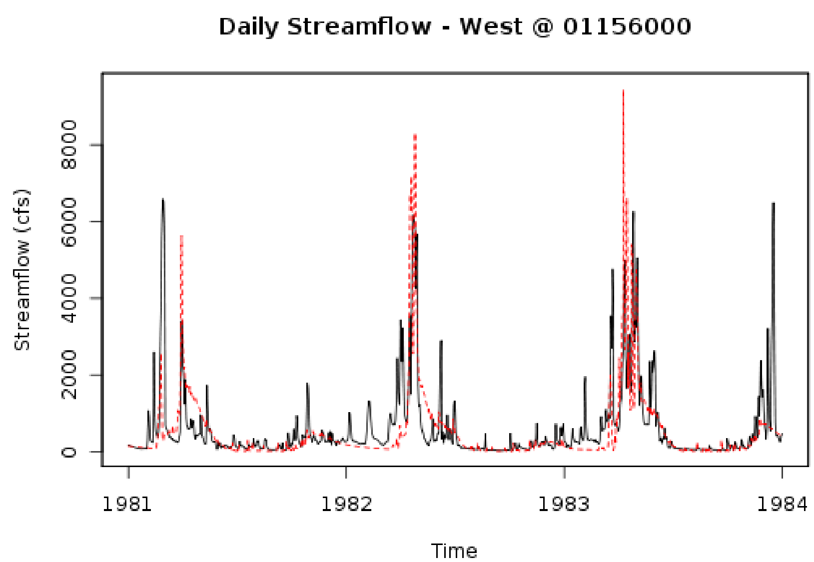


Figure 36: Time series of Daily Streamflow at

The daily and monthly Nash-Sutcliffe values for the calibration time period (1981 through 1988) are 0.29 and 0.65 respectively. The average annual flow has a positive bias of 10.26% and an RMSE value of 854 cfs. The low Nash-Sutcliffe values result from the inability to predict flows on a daily time step and the seasonal shift of the physical water balance in the model. Overall, the West is one of the least accurate sub-basins in the model. The inability to predict snow pack is the root of the error, which is reflected in the winter and spring streamflow. Error also exists in the ability of the model to replicate flashiness during drier months.

Ashuelot

The Ashuelot River is located in Southwestern New Hampshire, draining approximately 403 square miles. The northern headwaters begin on the southwestern slope of Mount Sunapee. Roughly 64 miles in length, the Ashuelot parallels the Connecticut River through the hills of southern New Hampshire, passing through the town of Keene. The river finally turns west in Winchester, NH and empties into the Connecticut roughly 3.5 miles north of the Massachusetts border. The basin has a mean elevation of 332 meters above sea level and an elevation difference of 900 meters from the highest to lowest elevation. There are several hydroelectric and flood control dams in the basin and many small run of river dams. Despite the total number of dams in the system exceeding 180, a comparison of SYE generated flows and observed flows shows the river running under almost natural, unimpeded conditions. The USGS maintains 3 gages in the basin with long monitoring records. The first gage is located a few miles upstream of Surry Mountain Lake (Ashuelot River near Gilsum, NH- 01157000) monitoring a 71.1 square miles of drainage area, the second gage is located just below the Surry Mountain Dam (Ashuelot River Below Surry Mountain Dam, Near Keene, NH- 01158000) monitoring a 101 square miles of drainage area, and the third gage is located at the mouth of the river (Ashuelot River at Hinesdale, NH- 01161000) monitoring the full 420 square miles of drainage area in the basin.

Snow

The VIC model for the Ashuelot River basin tends to under-predict snow during the peak periods, resulting in earlier complete melt for the pixel centered over the Keene, NH station. This under-prediction is evident in most years, particularly in 1987, although the large snow year of 1982 is modeled well. At the meteorological gage in Keene, NH, the mean and median snowpack occurs around the second week of February with approximately 25 cm of snow accumulation (Figure 37). Snow accumulation typically begins in the second week of November, reaching a maximum in early February, and melting completely by May 1st. The time series match well most years. Differences stem from large under-predictions of snowpack throughout the season. The results in the next section indicate that the effect of snowmelt on streamflow matches the observed conditions.

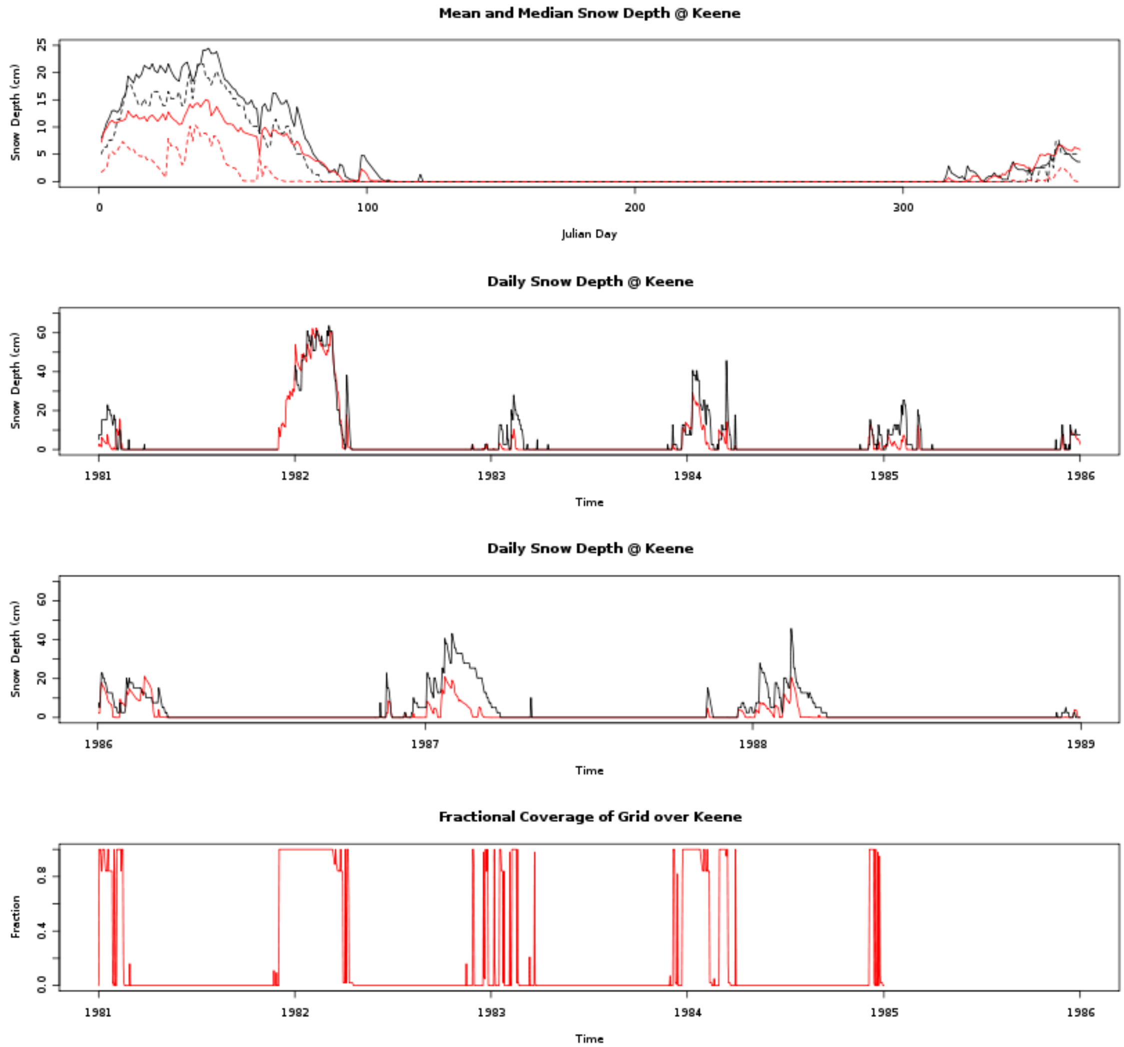


Figure 37: Snow Plots for the Ashuelot Basin at Keene

Streamflow

Three figures in this section describe the model skill and physical water balance in the Ashuelot basin. The mean and median Julian day flows at gage 01161000 in Hinesdale are displayed in Figure 38. Overall, the mean and median streamflow match the observed hydrograph well. The peak flows during the spring runoff and fall recharge periods are slightly under-predicted. Winter flows are also low during the same period of peak snowpack in early February. This issue is consistent with the discussion in the Ammonoosac basin section about winter flows. The model produces baseflow lower than the observed and the mass balance as described by mean annual flow shows a negative bias.

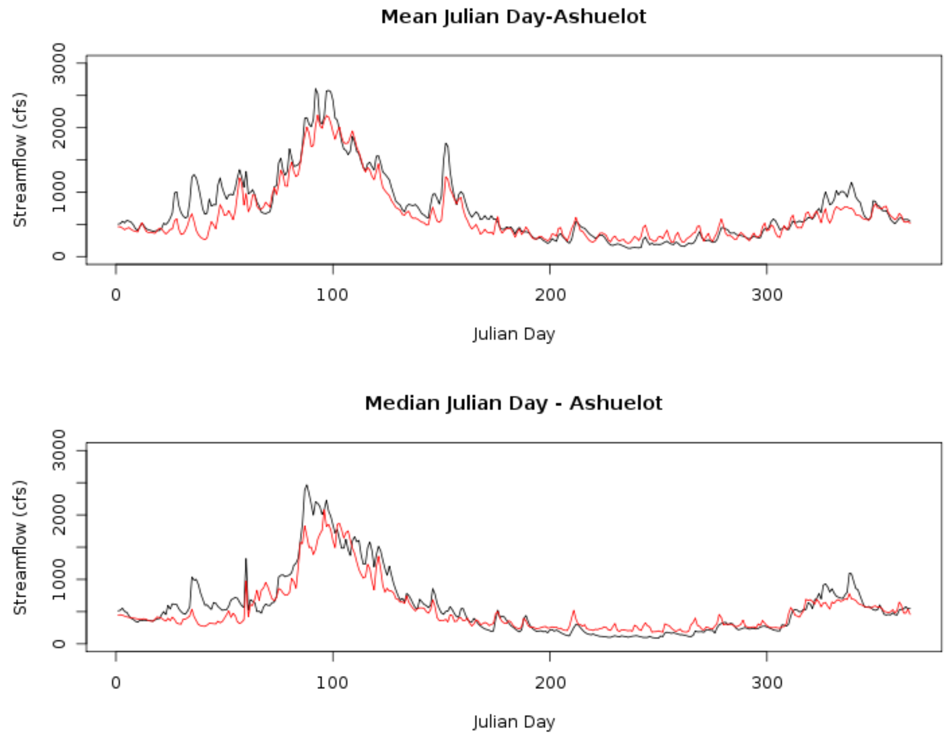


Figure 38: Mean and Median Julian Day Flow at 01161000

The mean monthly streamflow at the Hinesdale gage displayed in Figure 39 is well matched to the observed flows aside from a few under-predicted peaks between 1987 and 1988. There is a clear discrepancy between modeled and observed flows in early 1982. Figure 40 which shows the daily streamflow at the Hinesdale gage allows for the inspection of this difference and the determination that the model is clearly missing two increased flow events. It is suspected that this is a result of the occurrence of meteorological conditions that are difficult to capture in the model.

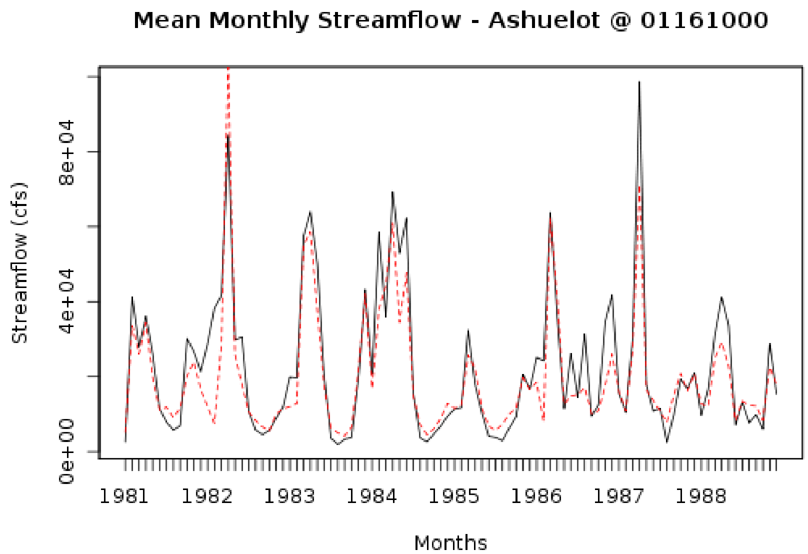


Figure 39: Mean Monthly Streamflow at 01161000

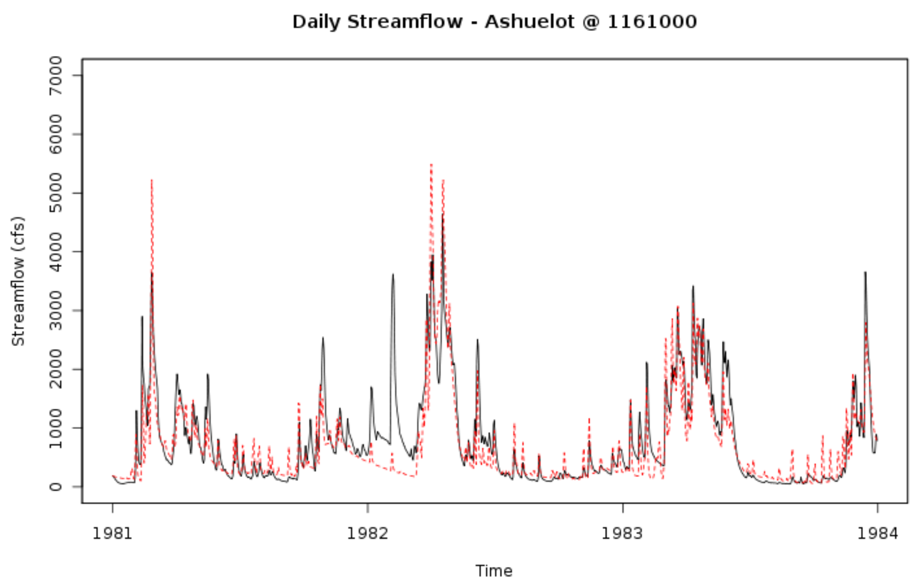


Figure 40: Daily Streamflow at 01161000

The daily and monthly Nash-Sutcliffe values for the calibration time period (1981 through 1988) are 0.73 and 0.81 respectively. The RMSE for the basin is 439.5 cfs which is roughly 60% of the average flow over the calibration time period. The average annual flow has a negative bias of 11.7% which can be used to explain the large RMSE value. The RMSE can also be attributed to the problems faced by the model in replicating the events in early 1982 where major peaks were missed altogether.

Deerfield

The Deerfield River Basin is located in southern Vermont and northwestern Massachusetts. The watershed covers a total drainage area of roughly 665 square miles (347 square miles in MA) and a mainstem length of 70.2 miles (44 miles in MA). The river begins at the base of Stratton Mountain in Southern Vermont and `south towards Greenfield Massachusetts. The landscape of the watershed is rugged and very typical of what can be found in most New England areas. Land surface altitudes range from 4,000 feet above sea level in the Vermont Mountains to 120 feet above sea level in the Connecticut River Valley. There are 10 hydroelectric dams along the river that use the river to power local communities. USGS maintains two gages along the Deerfield River. The first gage is located in Charlemont, MA (MA – 01168500) and has a drainage area of 361 square miles. The period of record for this gage is 1913 to the current year. The second gage is located in West Deerfield, MA (MA – 01170000) and has a drainage area of 557 square miles. The period of record for this gage is 1904 to the current year.

Streamflow

Given the regulation in the system, only two graphs are presented for verification purposes for the Deerfield. With the exception of the two larger reservoirs in the headwaters of the Deerfield Basin, majority of the hydropower facilities are run-of-river hydropower or pump storage facilities with minimal impact on overall flow through the system. Figure 41 plots the mean monthly streamflow over the period of 1981 to 1989. In general, there is good agreement between the observed and modeled flow at the West Deerfield gage. The model is negatively biased in the summer and fall months, which is likely due to higher flows in the river from dam releases. Figure 42 depicts this, as summer flows are highly variable due to the upper reservoirs releasing flow to match energy price signals. Overall the model replicates the annual and monthly water balances adequately, providing an unaltered inflow dataset for use in the optimization model.

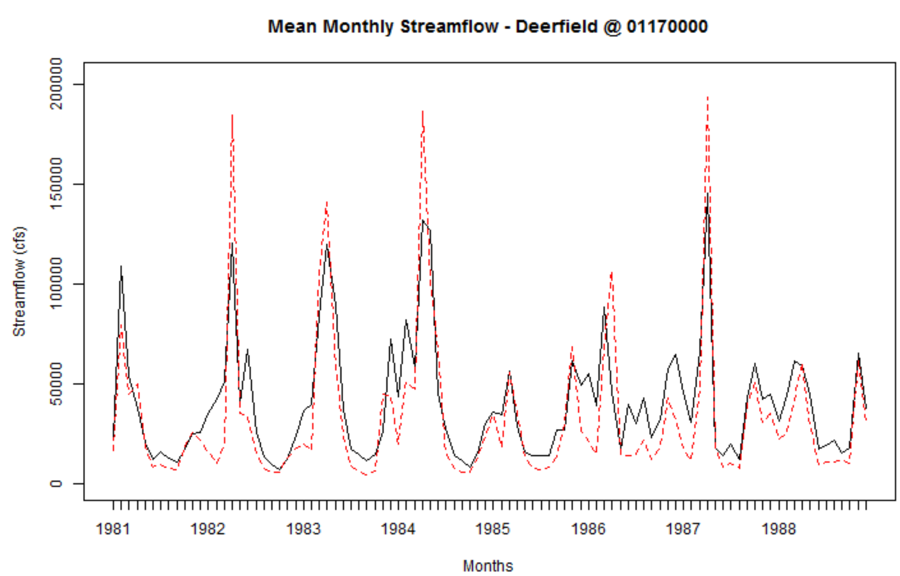


Figure 41: Mean Monthly Streamflow at 01170000

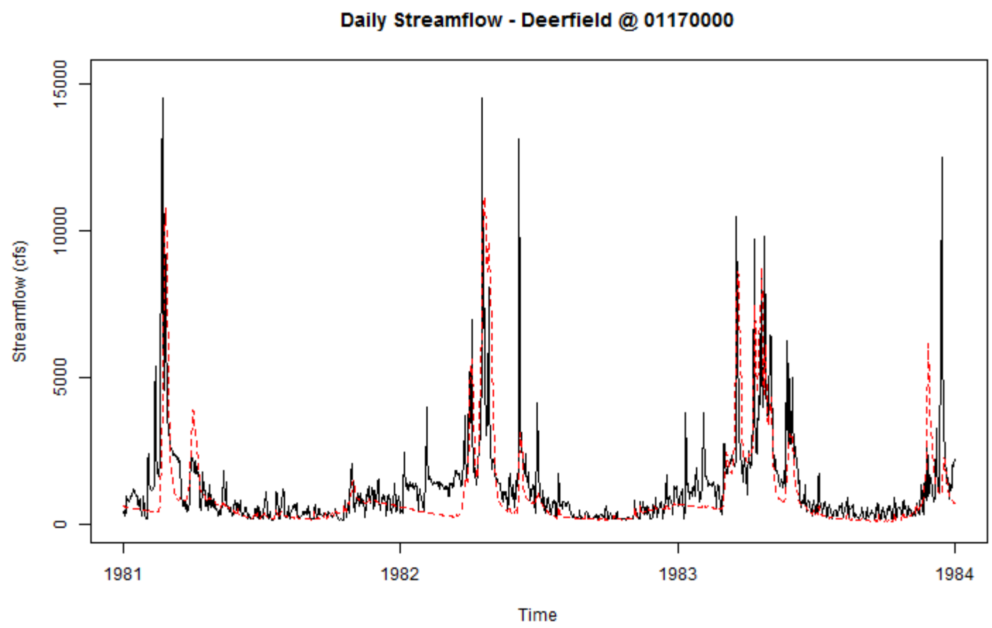


Figure 42: Daily Streamflow at 01170000

Westfield

The Westfield River is located in Western Massachusetts and drains about 517 square miles. From its headwaters located in northwest Massachusetts, the Westfield flows 78.1 miles south and east through the Berkshire Mountain Range to its confluence with the Connecticut River in Agawam, MA. The mean elevation of the basin is 350 meters. The range of elevation difference is 755 meters which is near the average for the basins in the Connecticut River watershed, and translates to an average slope of 7% for the basin. There are over 100 dams within the basin with varying functions including hydroelectricity, flood control, water supply, and recreation. A comparison between the SYE flows and the observed flows in the basin demonstrates the impact of regulation from these dams on the flows. It is also important to note that the Cobble Mountain Reservoir delivers water to areas outside of the basin which impacts the mass balance of the system. There are four USGS maintained stream gages with long data records in the basin. One gage a few miles upstream from the mouth (Westfield River near Westfield, MA- 01183500) monitors 497 square miles of drainage area, another gage located on the unregulated West Branch a mile above the confluence with the main stem(West Branch Westfield River at Huntington, MA- 01181000) monitors 94 square miles of the basin, a third gage at the mouth of the Middle Branch just below Littleville Dam (Middle Branch Westfield River at Goss Heights, MA- 01180500) monitors 52.7 square miles of drainage area, and the final gage just below Knightville Dam(Westfield River at Knightville, MA- 01179500) monitors 161 square miles of drainage area on the East Branch.

Snow

The historic snow depth record at Chester, MA is fragmented over the calibration period for the Westfield River. From the observed data that exists, it appears that the VIC model is over-predicting snowfall, mainly in late winter. This characteristic however, is only evident during 1982, which coincides with an under-prediction of streamflow during the same time period. The other two years with sufficient observed data, 1983 and 1985, are both well matched by the model. Because of the limited observed data it is difficult to evaluate how well the model predicts snow depth in the Westfield basin. Both modeled and observed snowpack exist from late November through early April. The mismatch between the modeled and observed snow depths in the mean and median plot may result from the lack of observed data. The depth comparison during 1982 does show the existence of model error. Results in the streamflow section may provide some insight into the accuracy of the snowpack prediction of the model.

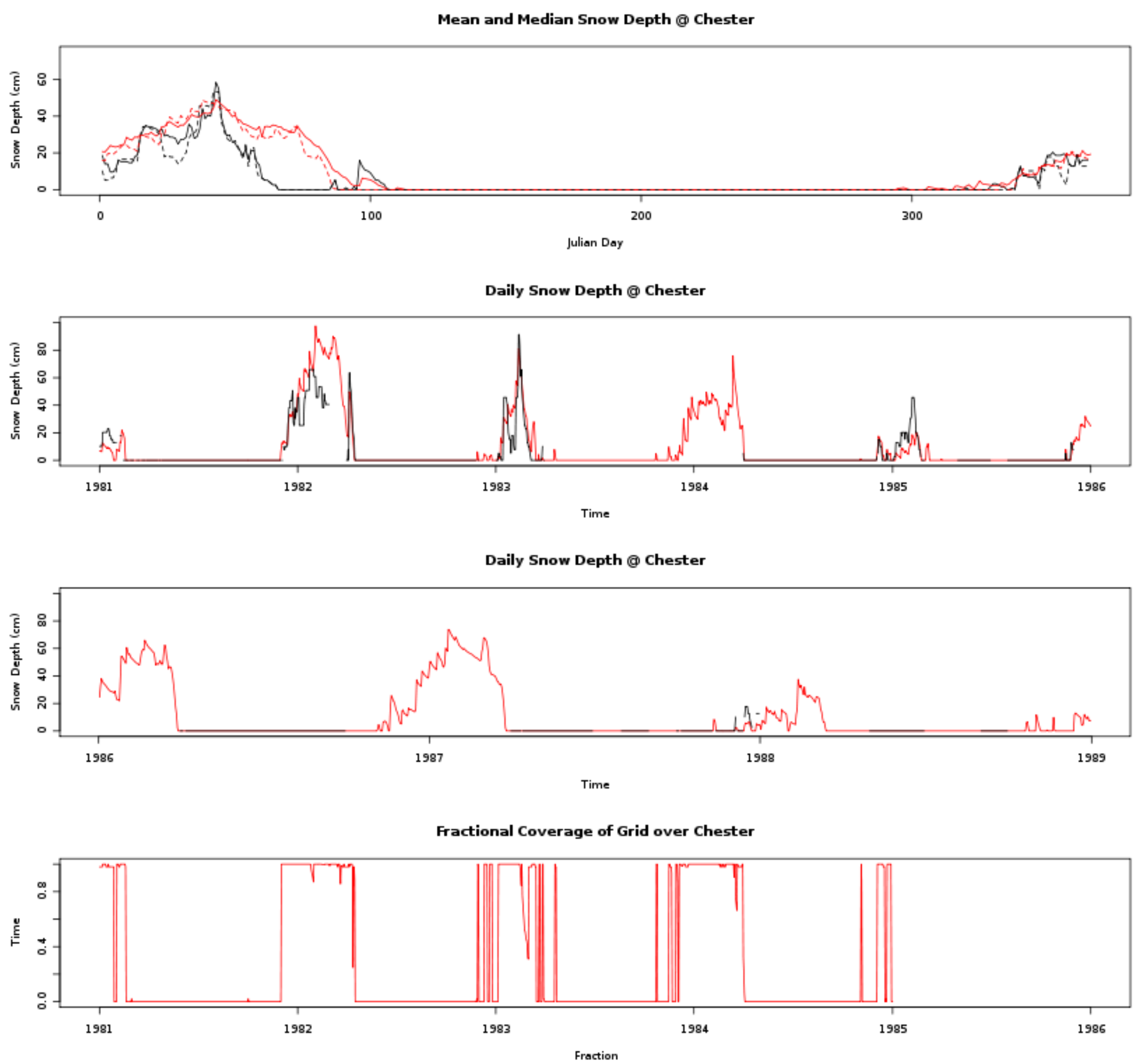


Figure 43: Snow Plots for the Westfield Basin at Chester

Streamflow

Three figures in this section are used to evaluate the streamflow output of the model. There are a few discrepancies to be addressed in these plots. The mean and median Julian day flows at gage 01161000 in Westfield are displayed in Figure 44. The early winter sees the model under-estimating flows which is a consistent problem in many of the basins. In late winter and early spring the flows are over-estimated, while later in the spring the flows are under-predicted. This difference is likely the result of missed snowpack during the winter resulting in a shift to earlier spring runoff in the model.

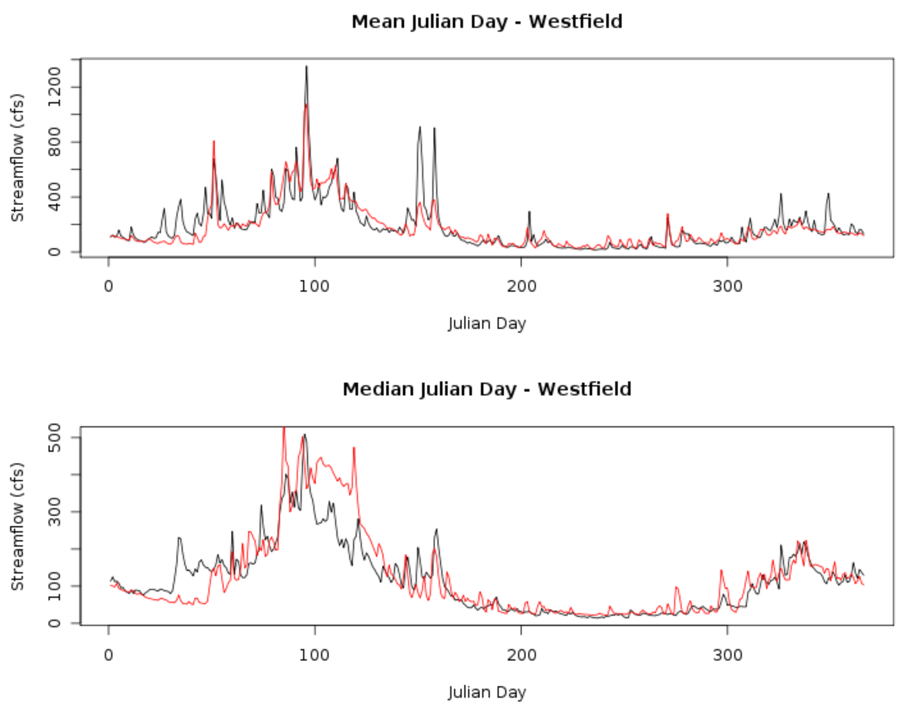


Figure 44: Mean and Median Julian Day Flow at

Figure 45 shows that the mean monthly streamflow at the Westfield gage is well matched to the observed flows. There is some notable under-prediction of some winter flows (1982, 1986, and 1987). The daily streamflow at the Westfield gage is shown in Figure 46. While the time series is modeled well here, there is consistent inaccuracy of peak flow prediction during the spring runoff period. Again this may be due to inaccurate snow prediction as well as the model's difficulties on the daily time step indicated by some of the statistics.

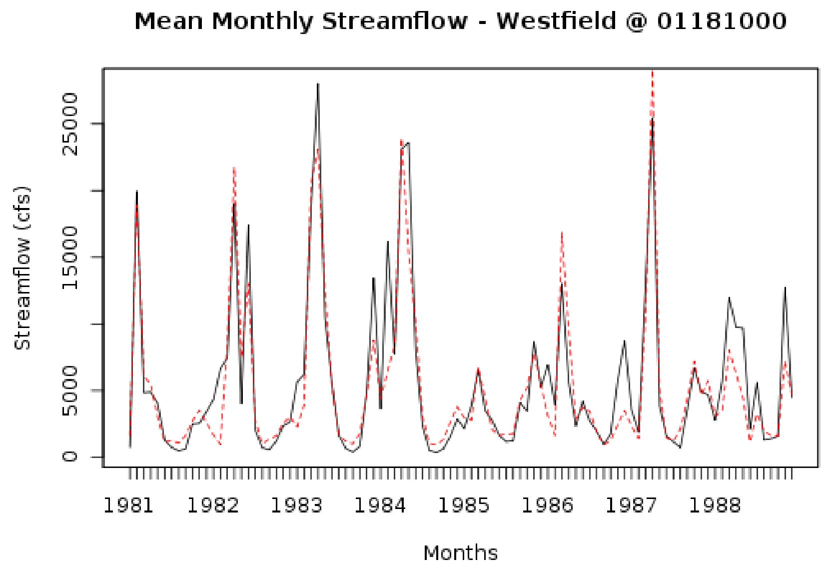


Figure 45: Mean Monthly Streamflow at Gage 01181000

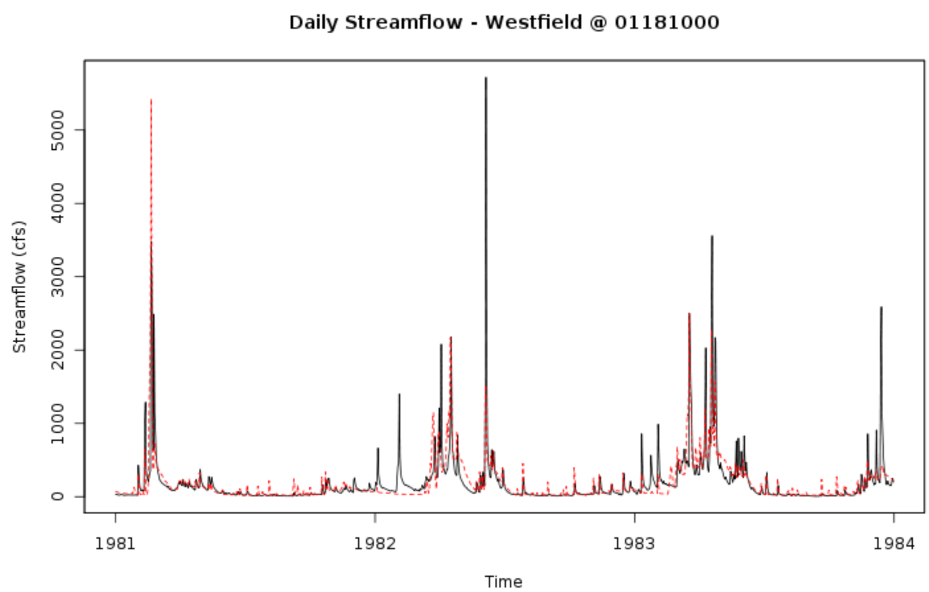


Figure 46: Daily Streamflow at 01181000

The daily and monthly Nash-Sutcliffe values for the calibration time period are 0.54 and 0.84 respectively. The difference in these values is an indication that the model does a better job of capturing flows averaged over time than it does on the daily time step. Peak flow estimates are frequently missed, but the overall mass balance of the system is represented well by the model. The modeled average annual flows have a 7.5% negative bias from the observed. The RMSE is 252 cfs, which is 135% of the average flow over the calibration period. This large value is likely due to the fact that RMSE is sensitive to under- and overestimation of peak flows, which is also indicated by the daily Nash-Sutcliffe value. The misrepresentation of spring runoff peaks is a common theme as evidenced by the daily streamflow comparison.

Millers

The Millers River is located in north central Massachusetts with the Quabbin Reservoir to the south and the New Hampshire border to the north. The river runs about 52 miles from east to west to its confluence with the Connecticut River just upstream of Greenfield, MA. The mostly forested terrain varies throughout the basin and includes many lakes, ponds, and wetlands in the northeast section of the watershed. The mean elevation is about 300 meters above sea level with a range of 520 meters between the highest and lowest points in the basin. There are over 140 dams on the Millers River, most of which are run of river. Two flood control dams, one at Tully Lake and one at Birch Hill, regulate the flow of the river. There are four USGS maintained gages in the watershed with long records. This includes the two gages on the tributaries, Priest Brook (Priest Brook near Winchendon, MA-01162500) and Tully River (East Branch Tully River near Athol, MA-01165000) which drain 19.4 square miles and 50.5 square miles respectively. Another gage far upstream on the Millers (Millers River near Winchendon, MA-0116200) drains 81.8 square miles of drainage area While the gage closest to the mouth (Millers River at Erving, MA-01166500) drains 372 square miles of the basin.

Snow

The model for the Millers River basin tends to over-predict snow throughout the winter. Altered runoff timing and later complete melt for the grid cell centered over Tully Lake are the direct result of this error. The largest over-predictions relative to actual depths occur in 1982 and 1986. Model prediction of snow depth over the remaining years during the calibration period is much closer to the observed. The peak modeled mean and median snowpack at the meteorological gage at Tully Lake occurs around the second week in February with snow depth around 40cm, about 10cm greater than the observed. Snow accumulation begins around the second week of November and is completely melted by mid-April. Modeled snowpack timing matches well over the calibration period. The main error occurs in modeled snowpack depth, which is reflected in streamflow predictions discussed in the next section.

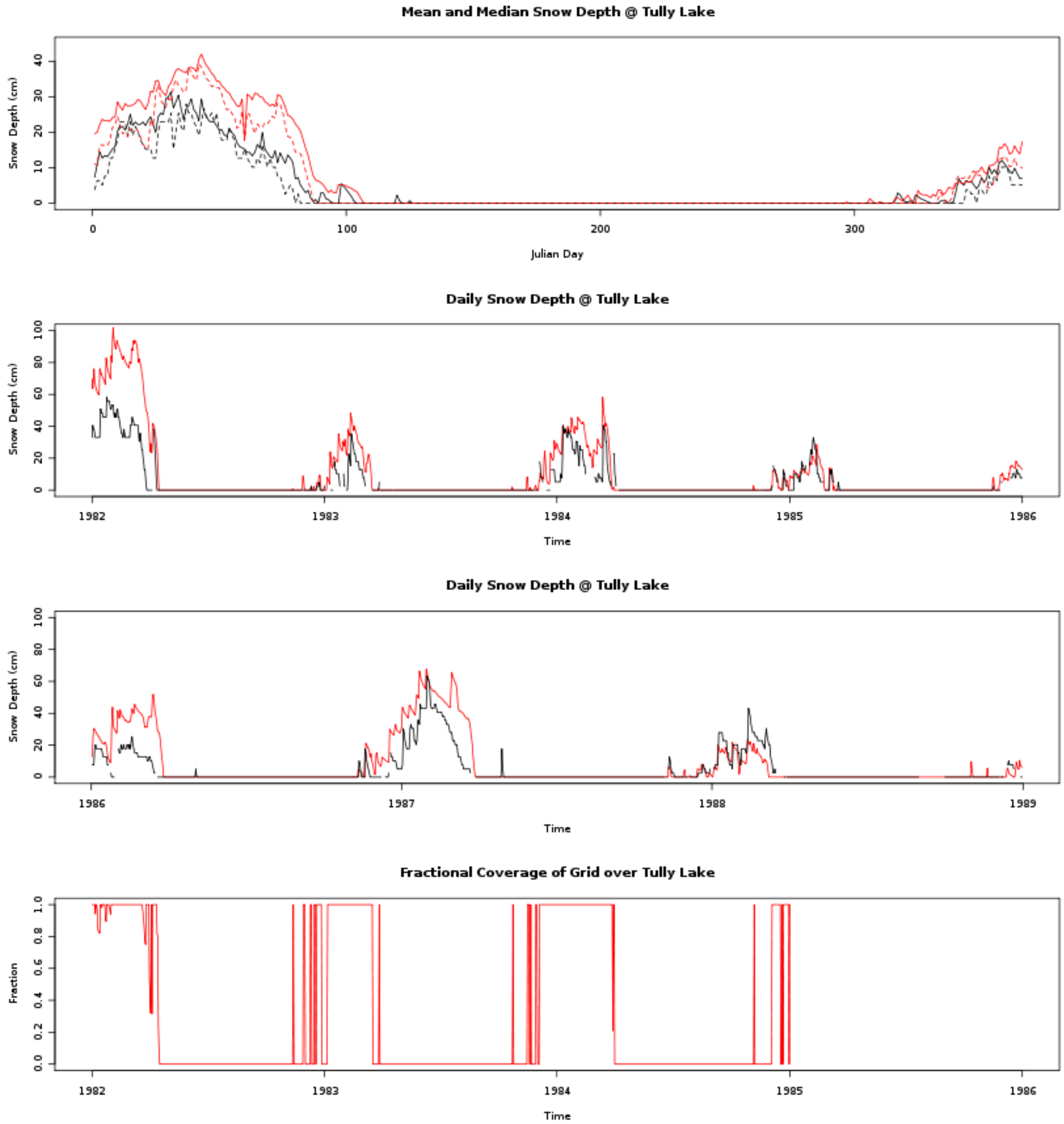


Figure 47: Snow Plots for the Millers Basin at Tully Lake

Streamflow

The model skill and physical water balance in the Millers basin is described by the three figures in this section. Figure 48 shows the mean and median Julian day flows at gage 01166500 in Erving, MA. The mean and median streamflow match the observed hydrograph fairly well with a couple of notable exceptions. The model over-predicts the peak spring flows. This is most likely due to the over-prediction of snowpack discussed in the previous section. The under-prediction of early winter flows also exists in this basin. This error, discussed in previous sections, is relatively consistent through the model. The mass balance, as described by mean annual flow, shows a negative bias mostly due to the lengthy periods of under-prediction during the winter.

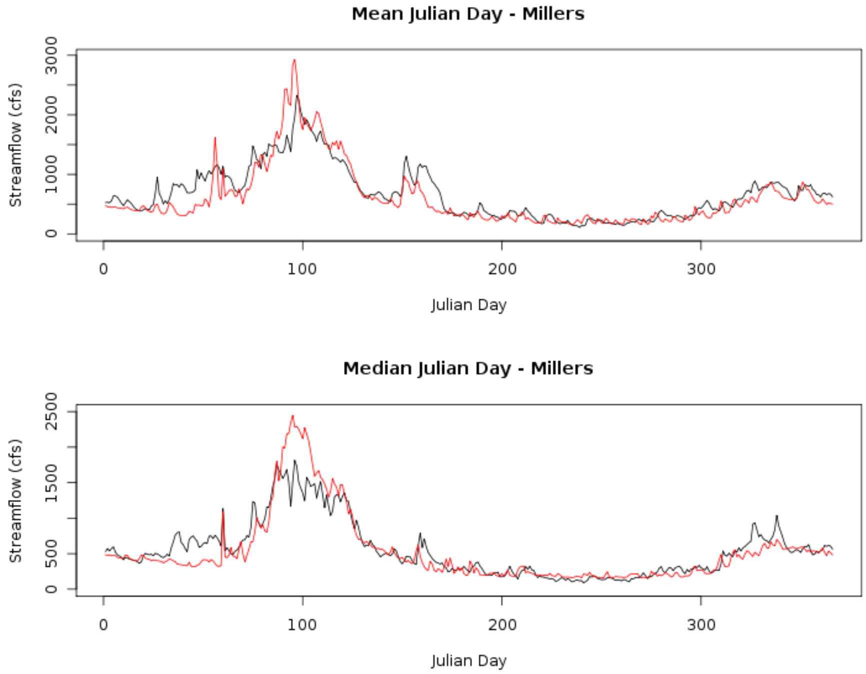


Figure 48: Mean and Median Julian Day Flow at Gage 01166500

The mean monthly streamflow at the Erving gage is displayed in Figure 49. When spring peaks are high (above 60,000cfs) the model tends to over-predict the flows, but when the spring peaks are lower (below 50,000cfs) the model tends to under-predict. Additionally, an over-prediction is usually paired with an under-prediction earlier in the season. This indicates that the error exists in the timing of the mass balance and supports the theory that the model experiences difficulty in classifying winter precipitation. The Millers is no exception to the clear discrepancy between modeled and observed flows in early 1982 that is consistent throughout the entire model.

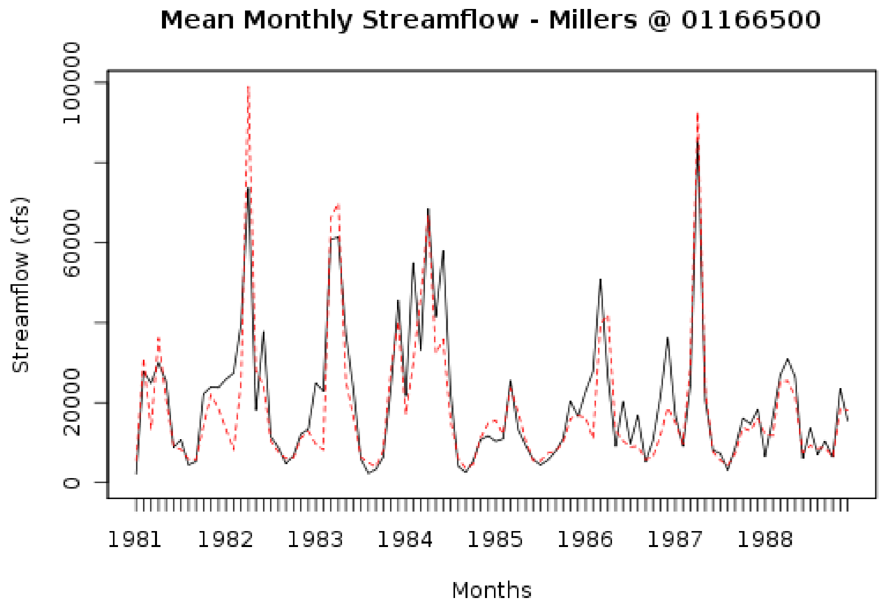


Figure 49: Mean Monthly Streamflow at Gage 01166500

Figure 50, which shows the daily streamflow over a three year period at the Erving gage, provides a closer look at the error mentioned above. Spring peak flows around the month of March are consistently over-predicted, while base and peak flows in January and February are under-predicted. Daily streamflow throughout the rest of the year is accurate.

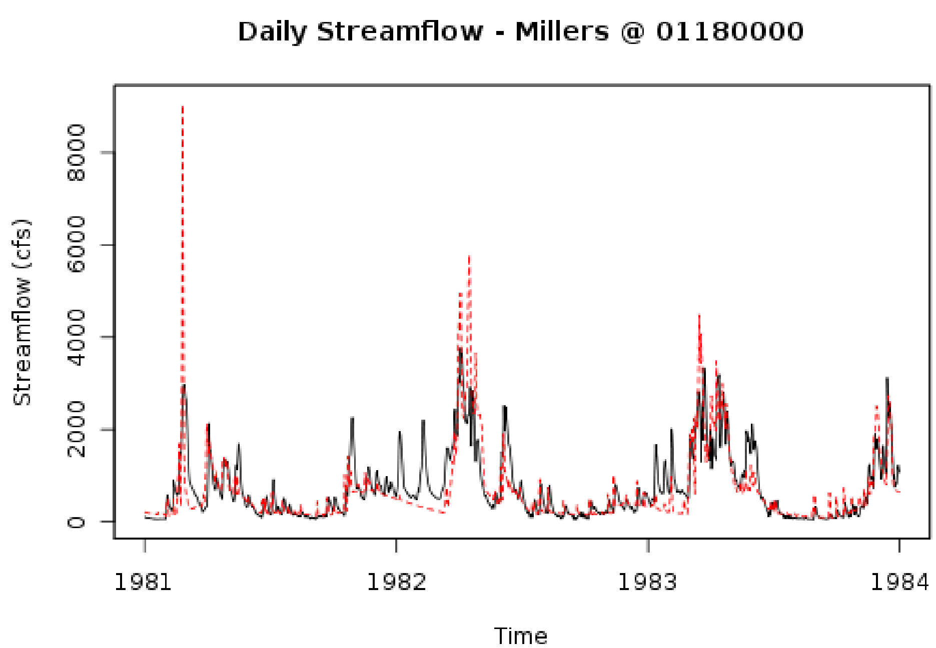


Figure 50: Daily Streamflow at Gage 01166500

The daily and monthly Nash-Sutcliffe values for the calibration time period (1981 through 1988) are 0.48 and 0.79 respectively. The RMSE for the basin is 506 cfs while the average annual flow has a negative bias of 9.90%. The low daily Nash-Sutcliffe value is due to the magnitude of the error in peak flow prediction. The negative bias for the average annual flow is due to the periods of under-prediction during the winter months.

Chicopee - Quaboag

The Chicopee River Basin is located in central Massachusetts in Franklin, Hampshire, Hampden, and Worcester counties. The basin covers a total watershed drainage area of 723 square miles and includes three major sub-basins (the Swift, Ware, and Quaboag systems) and also the Quabbin Reservoir. The landscape of the basin is characterized by rolling hills and alluvial plains that are divided by 136 rivers and 174 lakes. The topography of the basin rises over 1,500 feet above mean sea level and falls as low as only 40 feet above mean sea level in the Connecticut Valley lowlands. The river begins in Palmer, MA village of Three Rivers and flows into the Connecticut River in downtown Chicopee, MA. There are many hydroelectric power plants in the Chicopee River Basin, including the Chicopee Municipal Light Plant and the Ware River Power Plant. USGS currently maintains one gage along the Chicopee River and it is located at Indian Orchard, MA (MA – 01177000). This gage covers a drainage area of 689 square miles at an elevation of 125 feet above mean sea level. The gage has a period of record from August 1928 to the current year. The snow section is omitted for this basin as there were no immediate snow gages within the Quaboag River basin to compare modeled and historic precipitation.

Streamflow

The CRVIC model adequately represents stream flows in the Quaboag River. Figure 51 presents the mean and median Julian day streamflow. Both the mean and median flows are replicated well with the exception of flows during mid-winter months where a low bias occurs. The seasonality of the streamflow in the basin tracks the observed flow well during the spring, summer, and fall months. Annual variability of flows in the basin is modeled well (Figure 52) as are the monthly and annual totals for flow. The daily streamflow time series match well, with some over and under predictions of storm peaks (Figure 53). The largest 'misses' occur during the mid-winter months of January and February when warm rain-on-snow type events drive runoff. Some of the highest peak flows in the spring are under predicted as well. In general the model matches the observed record well, with NS-values of 0.83 and 0.67 for monthly and daily periods and an annual bias of approximately 0.02% over estimation of flows.

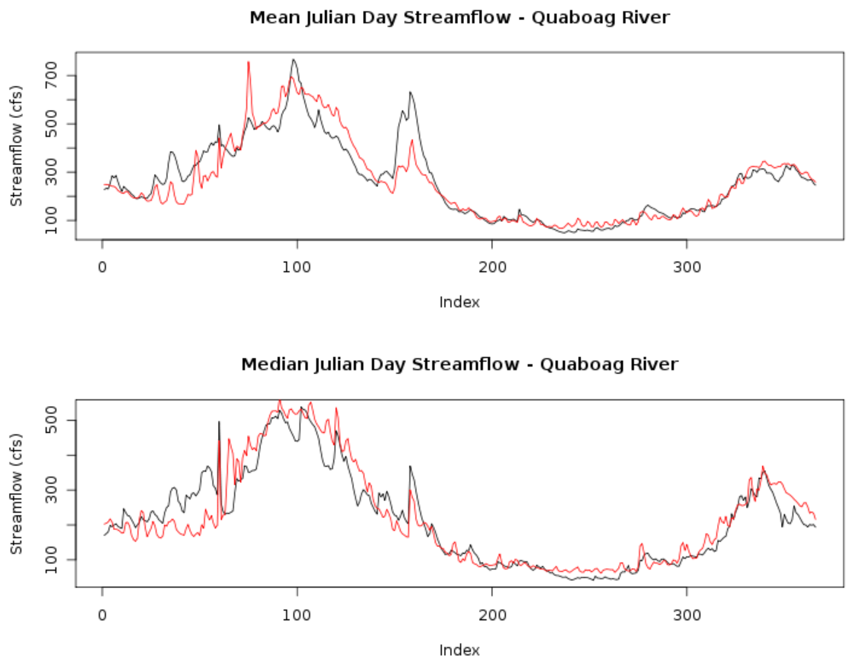


Figure 51: Mean and Median Julian Day Streamflow at Gage 01176000

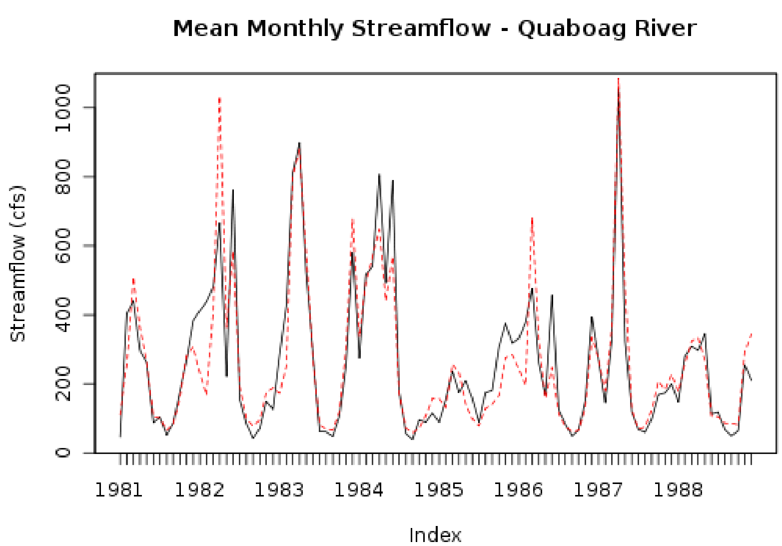


Figure 52: Mean Monthly Streamflow at 01176000

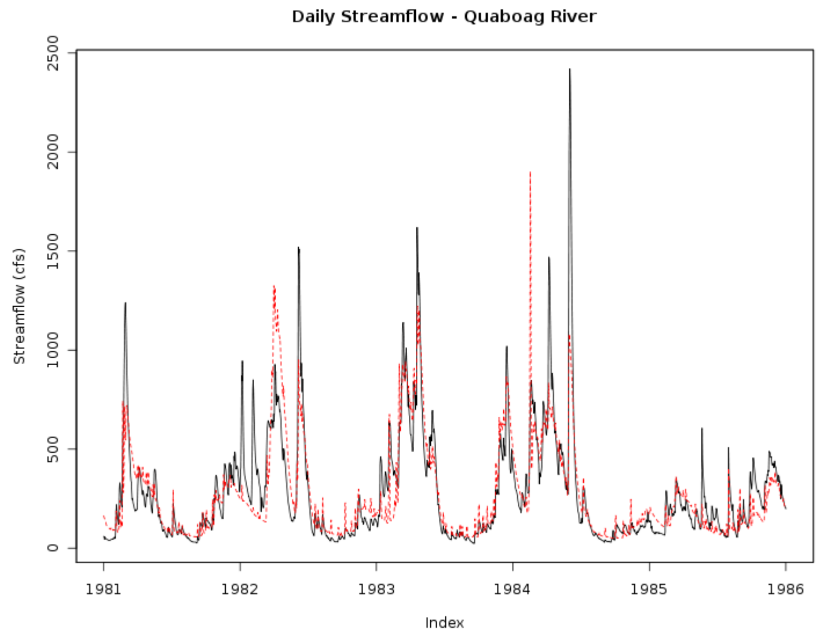


Figure 53: Daily Streamflow at 01176000

Farmington

The Farmington River Basin is located in southwest Massachusetts and northwest Connecticut. The Farmington River is the longest mainstem of any sub-basin of the Connecticut River with a length of 80.4 miles. The watershed covers a total drainage area of 609 square miles or 384,000 acres. The two main branches of the river start in southwestern Massachusetts and flow south until they meet in New Hartford, CT flowing on to meet the Connecticut River in Windsor, CT. There are 409 dams on the Connecticut portion of the Farmington River Watershed. Currently, USGS maintains one gage along the Farmington River in Massachusetts and one gage in Connecticut. The Massachusetts gage is located along the west branch of the river near New Boston, MA (MA – 01185500). The gage has a drainage area of 91.7 square miles at an elevation of 758 feet above mean sea level. The gage has a period of record from May 1913 to the current year. The Connecticut gage is located along the west branch of the Farmington River at Riverton, CT (CT – 01186000). The gage has a drainage area of 131 square miles at an elevation of 485 feet above mean sea level and has a period of record for daily discharge from October 1955 to the current year.

Streamflow

Model verification for the Farmington Basin is challenging because the reservoir systems in the basin operate for hydropower, flood control, water supply, and the reservoirs are large enough to alter flow significantly. Figure 54 plots the mean monthly observed and modeled streamflow at gage 0119000. Over the period 1981-1986, model agreement with the observed record near the mouth of the Farmington has some correlation over time, though many of the observed peaks are shifted or muted.

Figure 55 provides an explanation for the discrepancy between the two, as flood peaks are stored (winter/spring of 1984) and released later in the year over longer periods or diverted for water supply. In general, the daily time series tend to mimic one-another for most events, though the sporadic releases during the summer months are not replicated by the hydrology model as these releases are hydropower influenced.

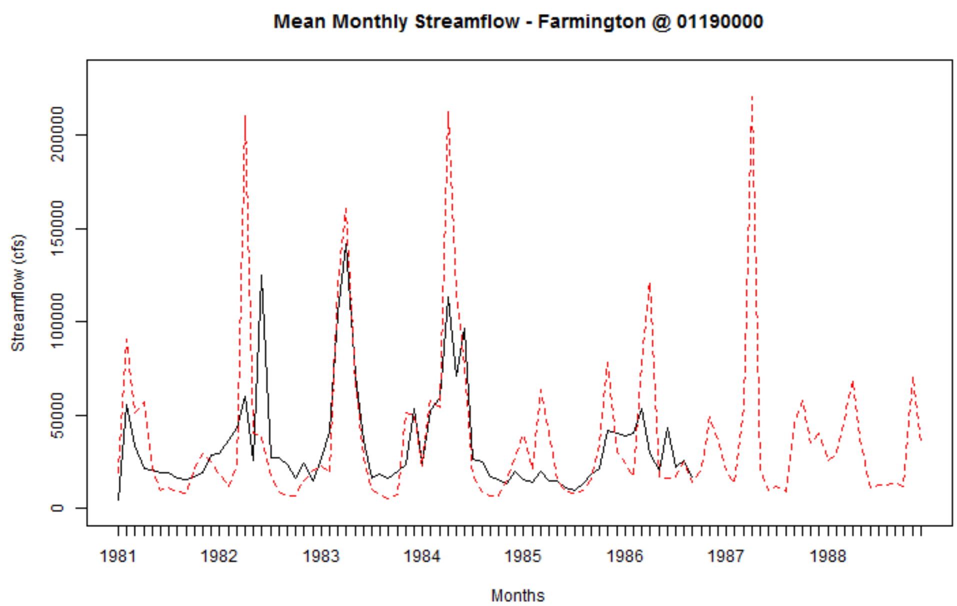


Figure 54: Mean Monthly Streamflow at Gage 01190000

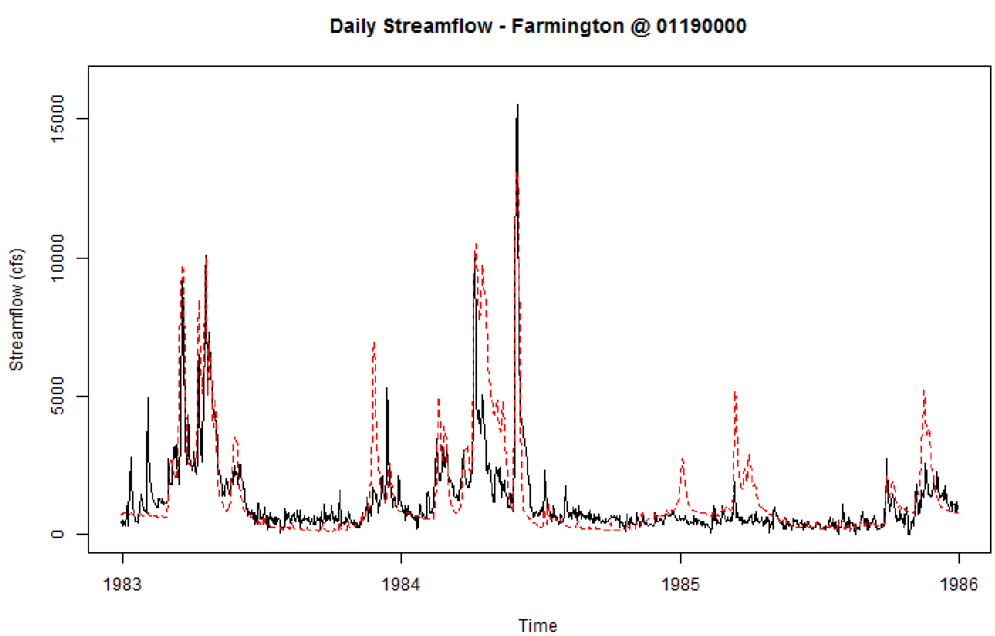


Figure 55: Daily Streamflow at Gage 01190000

Salmon

The Salmon River Watershed is located in Connecticut and covers a total drainage area of 96,000 acres across ten towns in the state. The river begins at the confluence of the Blackledge and Jeremy Rivers about a mile west of North Westchester, CT and runs for 10.4 miles before flowing into the Connecticut River in East Haddam, CT. The Salmon River has a significant drop in elevation over the course of its flow which provided substantial water power to the old textile mills located along the river. Currently, USGS maintains one gage along the Salmon River. This gage is located near East Hampton, CT (CT – 01193500). It has a drainage area of 100 square miles at an elevation of 64 feet above mean sea level and has a period of record for daily discharge from October 1928 to the current year.

Snow

Snow time series are compared with data from the meteorological gage at Middletown, CT. The mean and median daily snowpack occurs around mid-February, with approximately 20 cm of snow accumulation (Figure 56). Snow starts accumulating in the middle to end of December, increasing throughout February and then steadily decreasing until the end of March, with little to no snowpack by April. The model overestimates snowpack during the melt period in March but in general follows the seasonality and amount of snow well over the calibration period. The daily time series match well for most years when snow data are available, with no clear bias. In general, the effect of snow melt on runoff timing replicates observed conditions well as evidenced by the next section.

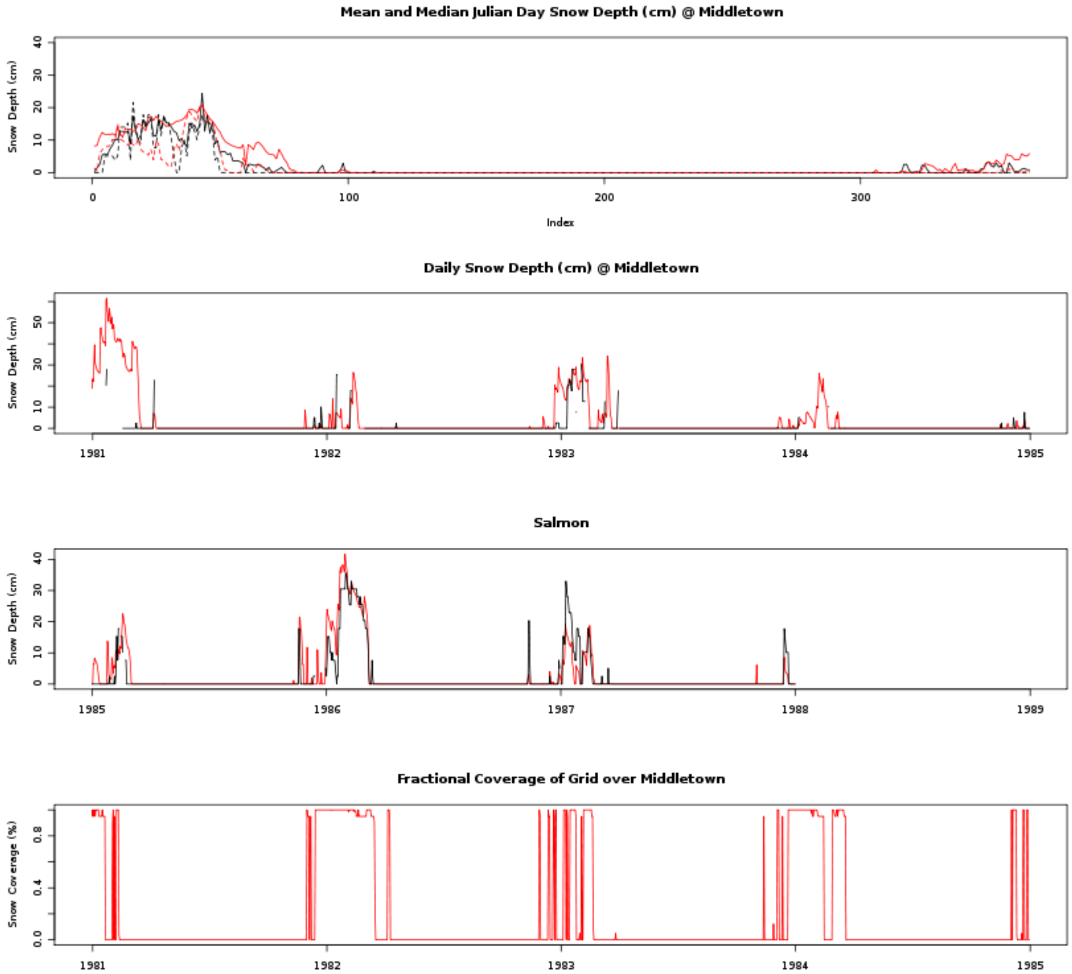


Figure 56: Snow Plots for Salmon River Basin at Middletown

Streamflow

As with many of the lower elevation and latitude basins, the CRVIC model performs well in representing streamflows in the Salmon River. Figure 57 presents the mean and median Julian day streamflow. Both the mean and median flows are modeled exceptionally well over the entire season. The seasonality of the streamflow in the basin tracks the observed flow well during the spring, summer, fall and winter months. The annual variability of flows in the basin is modeled well as indicated by Figure 58 and the high NS-value of 0.84. The daily streamflow time series match well, with some over and under predictions of storm peaks (Figure 59). Some of the highest peak flows in the spring are slightly under predicted. In general the model matches the observed record well, with NS-values of 0.84 and 0.63 for monthly and daily periods and an annual bias of approximately 1.0% over estimation of flows.

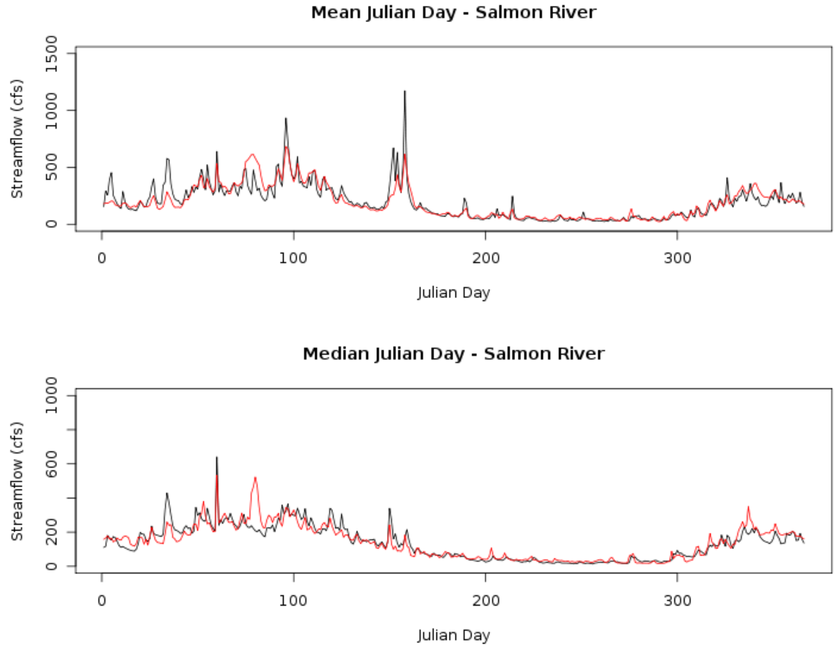


Figure 57: Mean and Median Julian Day Streamflow at 01193500

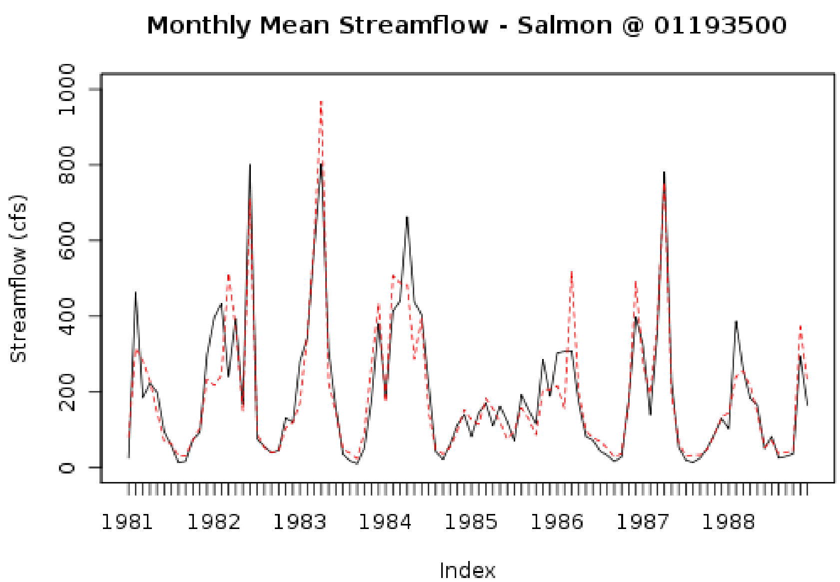


Figure 58: Mean Monthly Streamflow at 01193500

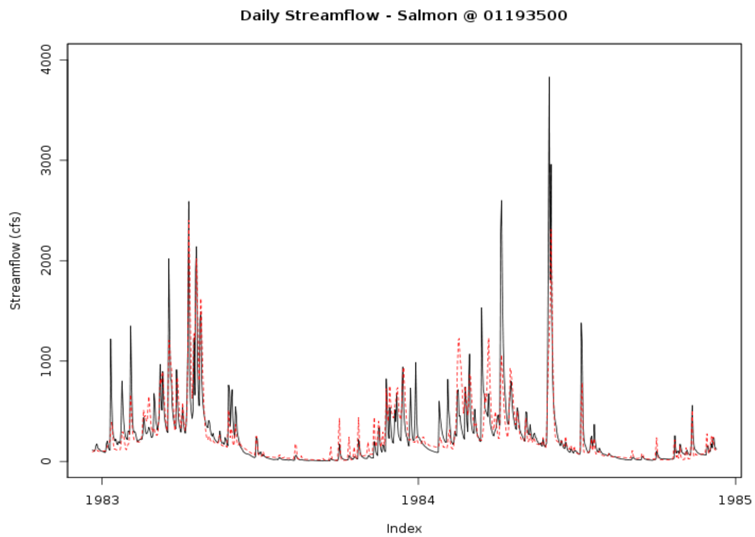


Figure 59: Daily Streamflow at 01193500

Climate Change Analysis

In this section the CRVIC model is used to investigate the impacts of climate change on streamflow and other hydrologic variables in the basin. Climate change has been shown to impact streamflow in regions throughout the world. The Northeast is expected to experience increased temperatures and precipitation as a result of climate change. Alteration of streamflow patterns can be expected throughout the watershed. The goal of this analysis is to obtain climate impacted streamflow for the Connecticut River Basin. These results will be used to inform management of reservoir systems within the watershed. The analysis focuses on the White, Ashuelot, Deerfield, and Chicopee sub-basins as well as the Connecticut River Basin in its entirety. These locations are chosen based on their representation of the basin and the importance of understanding climate change impacts on the water resource systems of the sub-basins.

A total of 112 downscaled climate projections are used to perform the climate change analysis with the CRVIC model. The projections are in the form of meteorological sequences consisting of daily minimum and maximum air temperature, wind speed, and precipitation amount. The sequences are based on the output produced by running 16 different General Circulation Models (GCMs) with three different climate change emissions scenarios. The three emissions scenarios represent varying degrees of future greenhouse gas emissions and are described in depth in the IPCC's Special Report Emissions Scenarios (Nakicenovic et al 2000). In general, the 37 B1 scenarios represent the lowest level of projected anthropogenic carbon emissions, the 39 A1B scenarios represent the greatest, and the 36 A2 scenarios represent the mid-range emissions level.

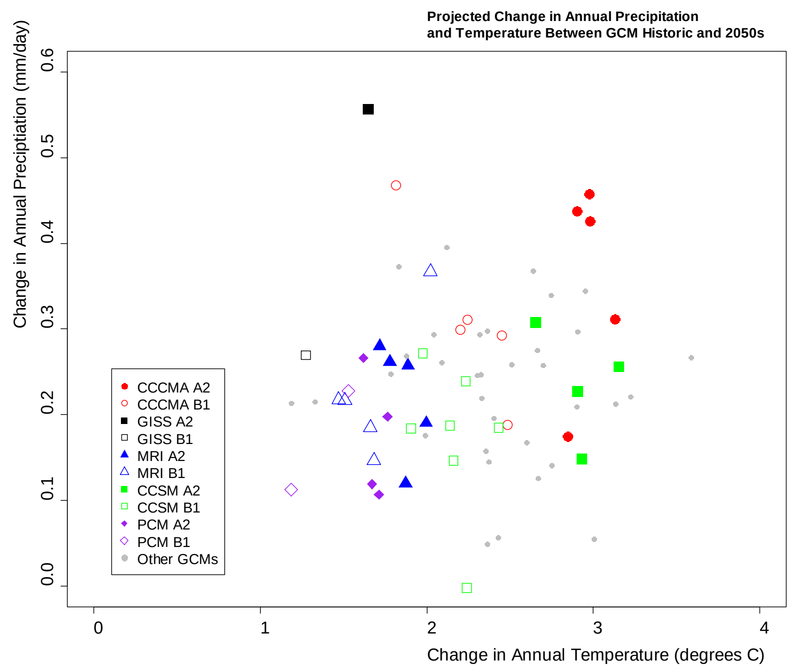


Figure 60: Projected Change in Annual Precipitation and Temperature between GCM Historic and 2080s

The downscaled forcing data produced by the GCMs are used as input into VIC in place of the gridded observed data used to force the calibration runs. A summary of the average changes in precipitation and temperature for each IPCC scenario into the future is shown in Figure 60 and Table 1 and Table 2. Running the 112 datasets through the CRVIC model produces a range of climate impacted streamflows for the Connecticut River basin from the time period of 1950 through 2099. In this analysis, three 30-year time periods centered around 2025, 2050, and 2075 are used to describe changes in the future. The time period of 1950 to 2000 is referred to as the historic time period. It should be noted that this historic period is modeled results which differs from actual observed historic streamflow. The results referring to these time periods use annual averages of the results over these 30-year periods of data.

Table 1: Average precipitation change (mm) from baseline for the Connecticut River Basin

	B1	A1B	A2
2025	42.2	42.7	39.4
2050	59.3	64.4	64.4
2075	72.1	88.6	82.2

Table 2: Average temperature change (^{o}C) from baseline for the Connecticut River Basin

	B1	A1B	A2
2025	1.06	1.14	1.03
2050	1.55	1.96	1.85
2075	2.00	2.67	2.88

The impact of climate change on hydrology throughout the basin is assessed by analyzing the changes in individual hydrologic variables. Figure 61 and Figure 62 contain plots of the mean model response for 5 GCM runs for different hydrologic variables. These scenarios project changes in hydrologic factors across the CRB. Annual precipitation is expected to rise across the watershed with the magnitude of change smallest in the valley and largest near the coast and in the northern mountains, specifically the Ammonoosuc Basin in the White Mountains. The expected increase in annual runoff and soil baseflow follows the spatial trend of precipitation change across the basin. The predicted increase in evaporation is likely a result of the presence of higher temperatures and more available water to evapotranspire. The reduced evaporation around the lower third section of the Connecticut River mainstem is possibly due to a shift in seasonal runoff timing and less water available during seasons where evaporation takes place. Higher future temperatures result a diminishing snow depth and snow water equivalent (SWE) most prevalent in the northern and mountainous regions of the basin (Figure 63). This change is the cause of shifted spring streamflow further explored in the streamflow analysis of each basin.

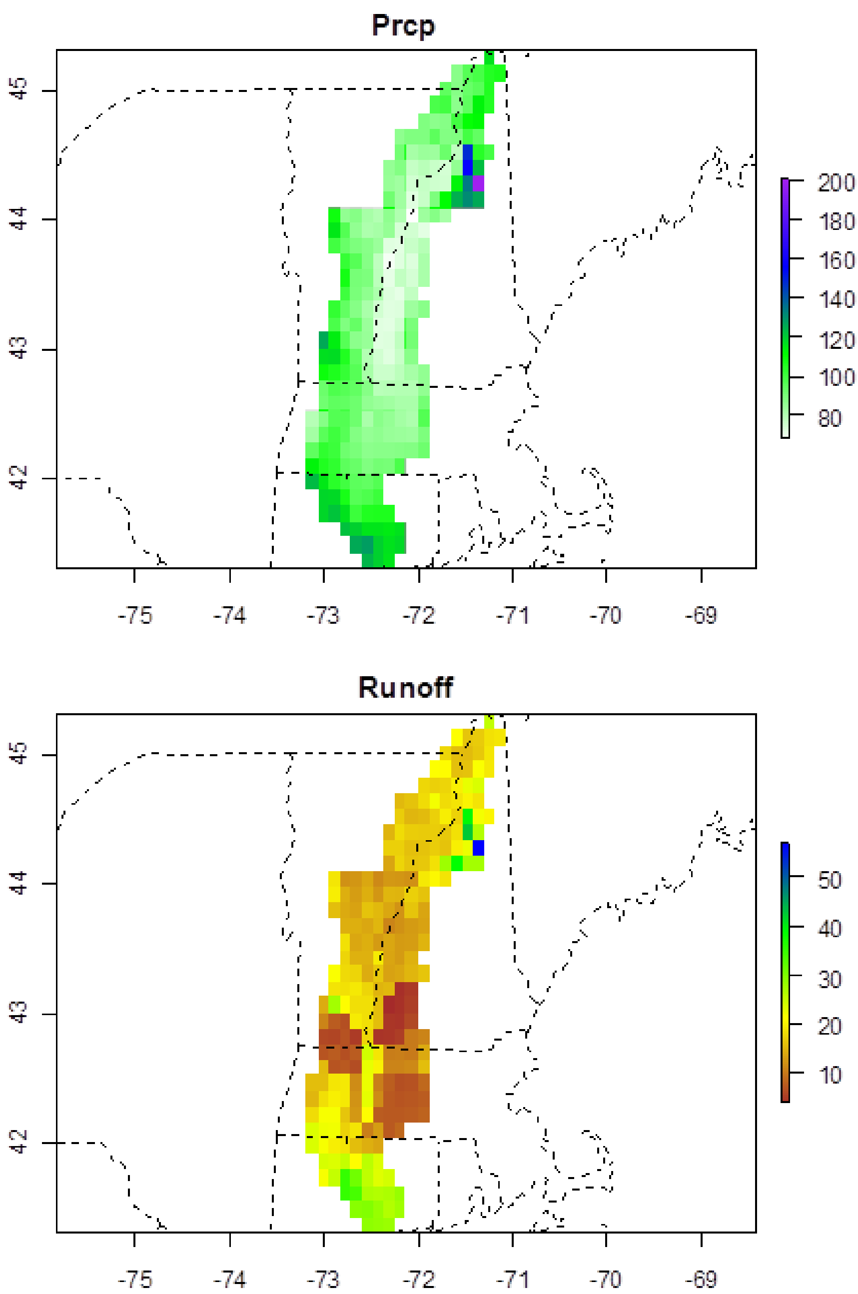


Figure 61: Average change (mm of water) in precipitation and runoff in the basin between 2080s and baseline

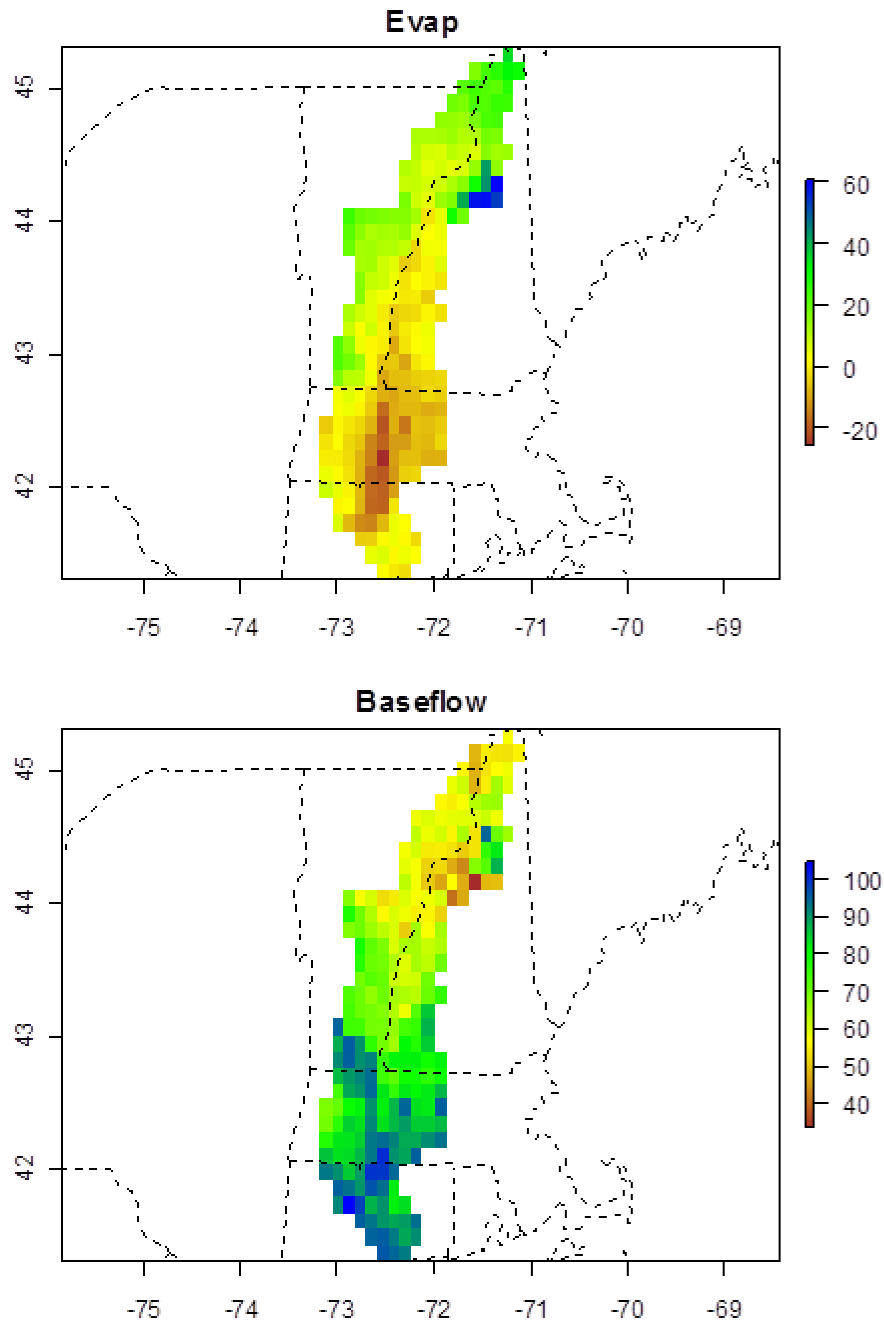


Figure 62: Average change (mm of water) in evaporation and baseflow in the basin between 2080s and baseline

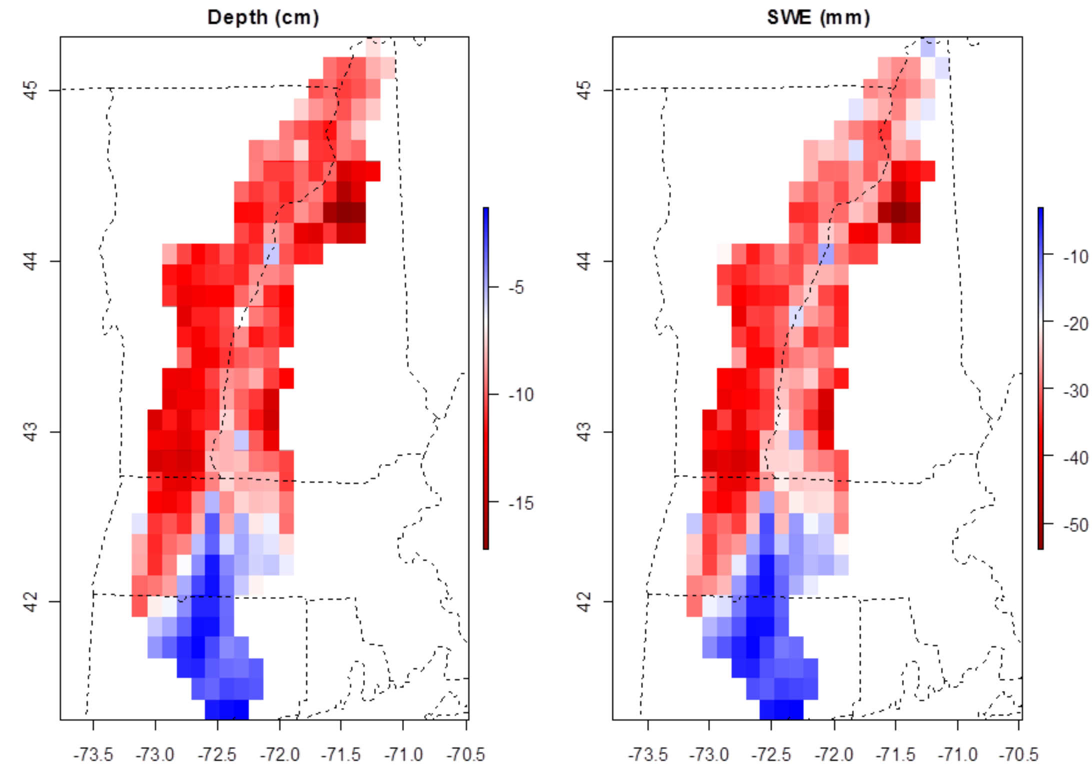


Figure 63: Average change snow depth and snow water equivalent (SWE) in the basin between 2080s and baseline

Streamflow is evaluated at nine designated points throughout the basin that capture the diverse characteristics of the watershed. The White and Ammonoosuc sub-basins as well as a point on the mainstem at the mouth of the Wells River, VT represent the upper third of the basin. The Black and Ashuelot sub-basins and a mainstem gage at Montague, MA represent the middle third of the basin. The Westfield and Salmon sub-basins and the gage near the mouth of the mainstem at Thompsonville, CT are representative of sections on the lower third of the basin. Figure 64 compares the average weekly streamflow from the gridded observed data to the average of the climate impacted streamflow under the B1 emissions scenarios over the historical period of 1950-2000. The range of streamflow generated from the GCM forcing data is also represented. Over the fifty year historic period the flows generated from the GCM forcing closely resemble those of the gridded observed data. The comparison of generated streamflow validates the ability of downscaled GCM datasets to reproduce the gridded observed data.

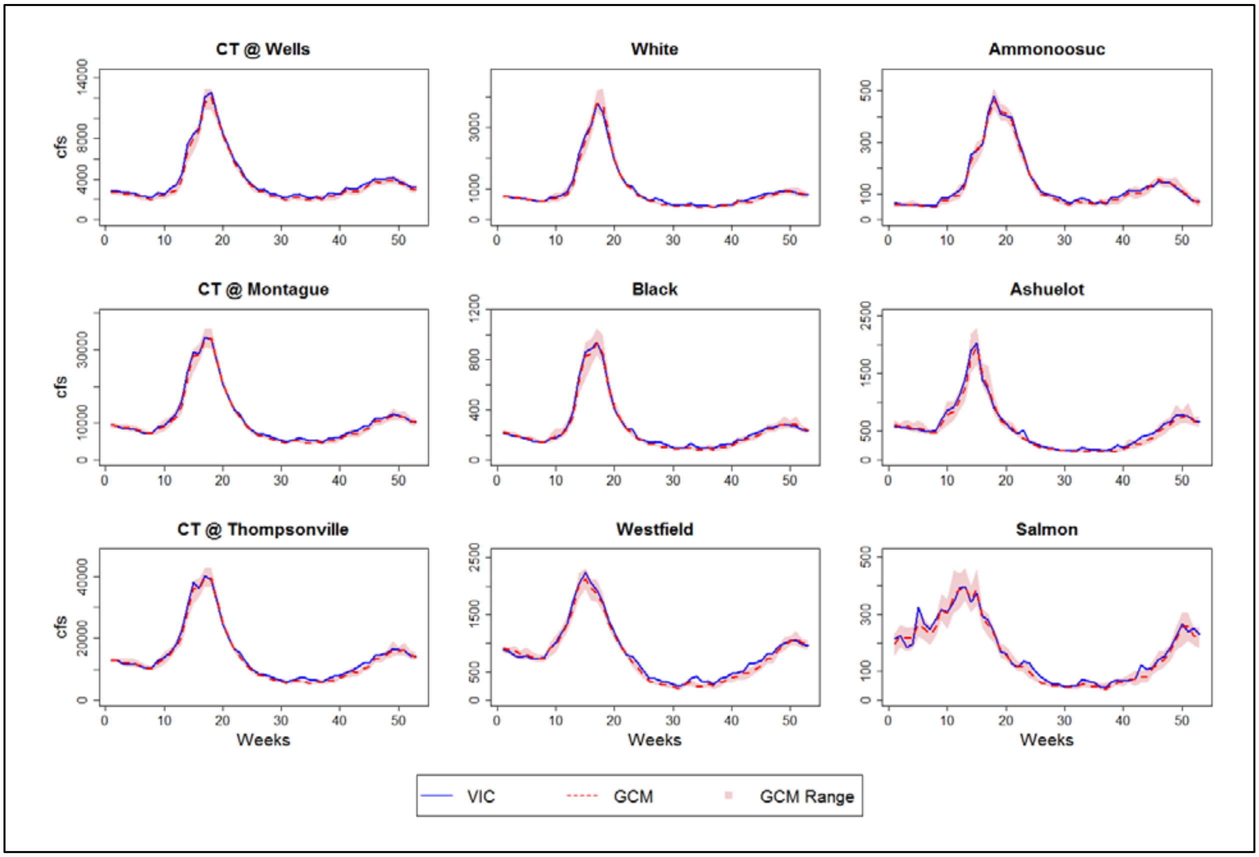


Figure 64: Comparison of average weekly streamflow between the streamflow generated from the gridded observed data and the average of the 33 B1 scenarios at select locations throughout the basin over the historical period (1950-2000)

Mainstem at Thompsonville, CT

In this study, the USGS gage on the Connecticut mainstem at Thompsonville, CT is the location used to measure streamflow impacts throughout the entire basin. Modeled streamflow projections for the basin as a whole are represented below (Figure 65). The CRVIC model predicts increasing streamflow in late fall and winter seasons for all three of the SRES emission scenarios. This is expected based on the snow depth and SWE predictions shown in the previous section. Precipitation that historically fell as snow and was stored in the snowpack is predicted to fall as rain and runoff immediately. This effect is amplified as temperature continues to increase into the future (Table 2). Another impact of the increase in temperature is the change in timing of the spring flow peak, which occurs about 2-3 weeks earlier than historic peaks. This is consistent across all scenarios and time periods. The consistency of this timing shift into the future may indicate that the basin is nearing a threshold where an increase in annual temperature from 0°C to 1°C will have a much greater impact on spring peak flow timing than an increase from 1°C to 2°C. The magnitude of the spring peak however, is not predicted to deviate from historic values. The model predicts minimal change from mid-spring through mid-fall in the streamflow across all scenarios.

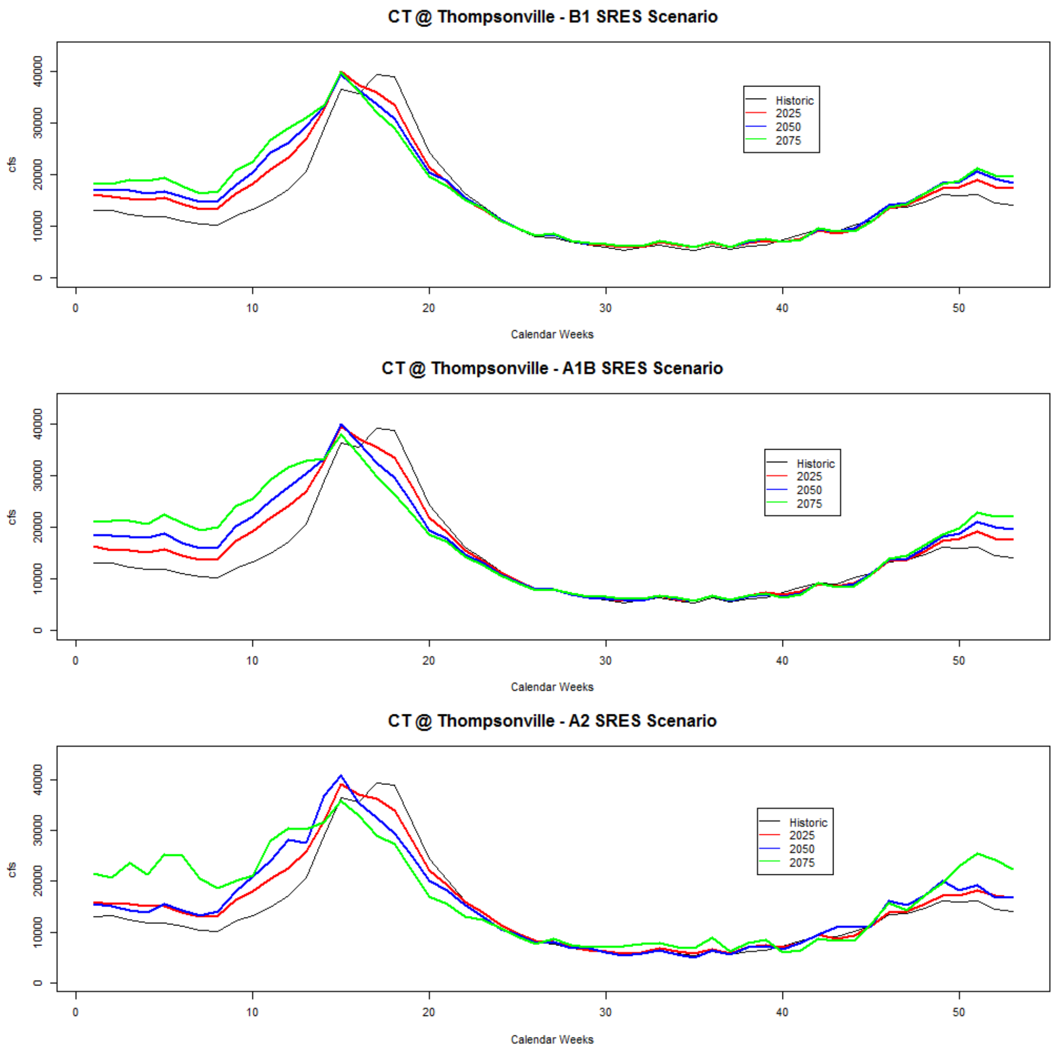


Figure 65: Future mean weekly streamflow projections for each SRES emissions scenario at the Thompsonville, CT gage

The range of streamflow predictions from all of the SRES emissions scenarios is shown below (Figure 66). It is important to note that while the mean streamflow is changing through time, it is not until the 2050 time period that the range of GCM predicted streamflow departs from the historic mean. In the 2075 time period the difference is more significant and all of the climate change results during the winter and early spring months show an increase in streamflow.

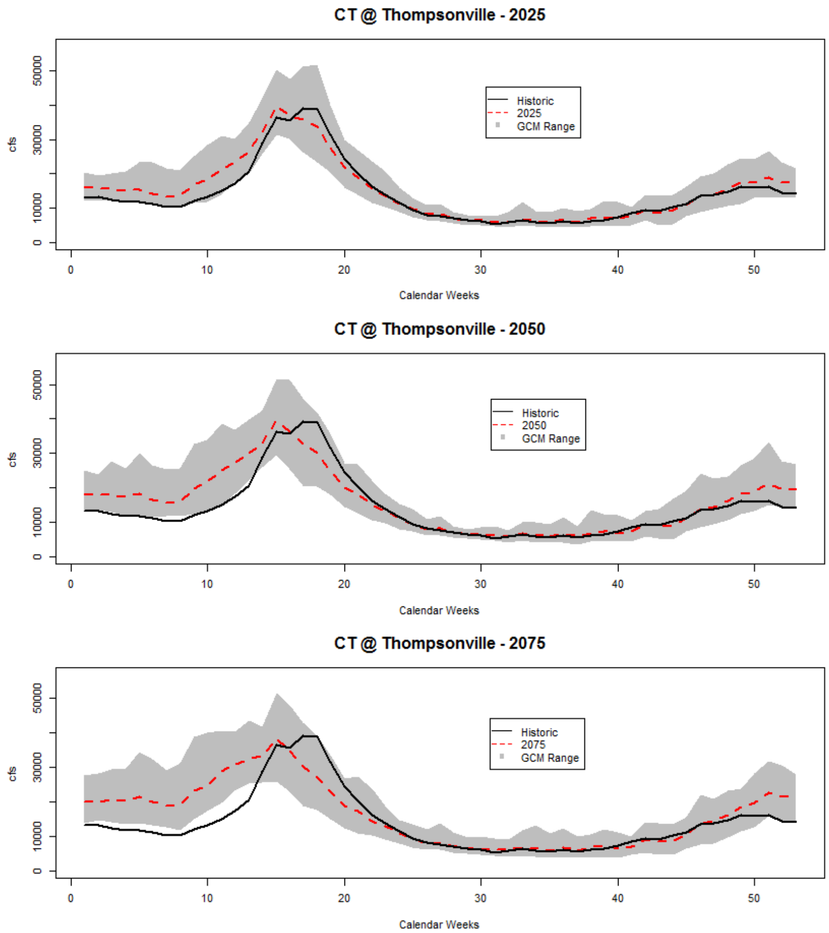


Figure 66: Mean annual streamflow and range on the Connecticut River mainstem at Thompsonville, CT of all 112 GCM runs

The differences between streamflow predictions under the B1, A2, and A1B SRES scenarios are minimal (Figure 67). The B1 scenario differs slightly from the A1B and the A2 scenarios further into the future. This trend holds true for all of the other basins examined in this section. For this reason, different SRES emissions scenarios will not be plotted against each other beyond this section.

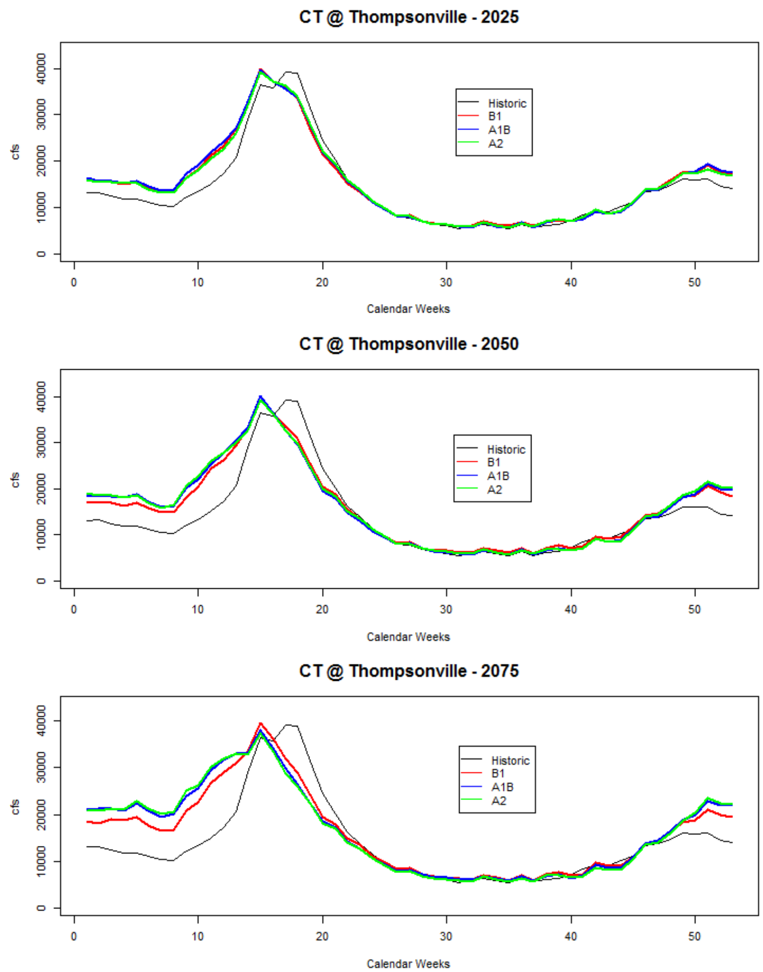


Figure 67: Future mean weekly streamflow projections at each time period for the 3 SRES emissions scenarios at the Thompsonville, CT gage

A multivariate regression model is used to explore the impact of precipitation and temperature change on streamflow in the basin. Mean annual precipitation and temperature change are simultaneously regressed against mean annual streamflow change. The R^2 and p-values are provided in Table 3 below. Figure 68 shows plots of precipitation and temperature change separately versus streamflow change. It is evident from these figures that precipitation proves to be the dominant factor on streamflow change in the basin. A 10% increase in precipitation results in a consistent 18% increase in streamflow through all future projections time periods. The impact of temperature change is insignificant but is found from the multivariate regression to increase in significance into the future.

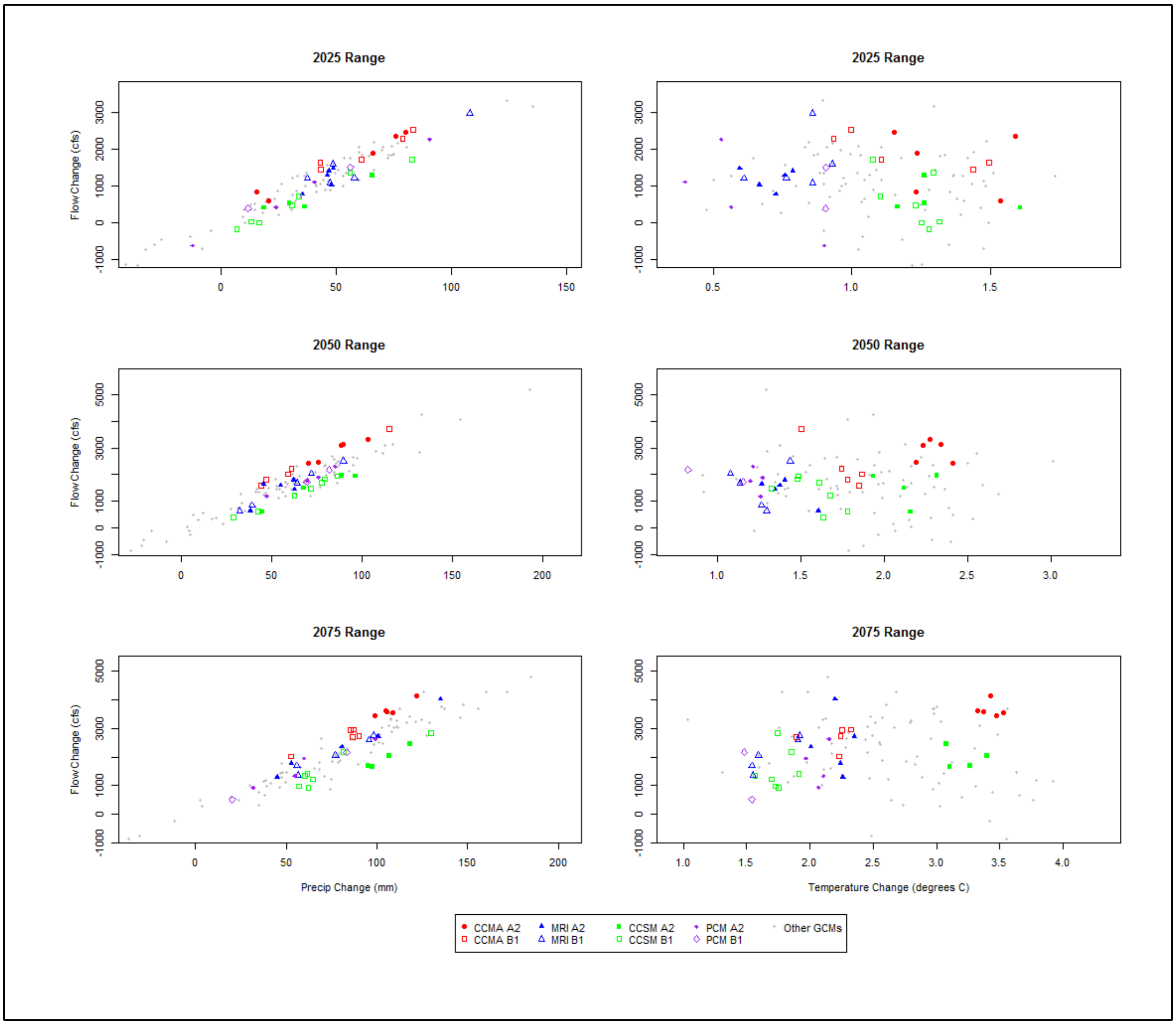


Figure 68: Change in mean annual streamflow for mean annual precipitation and temperature changes over all the grid cells in the CRB for the 112 GCMs

Table 3: Slopes of the precipitation vs flow change plots, (Multivariate regression R^2 -values), and the P-values significance codes for temperature only: 0 '*' 0.001 '**' 0.01 '*' 0.05 '.' 0.1 '' 1. The p-values for precipitation are consistent at 2e-16**

Basin	2025	2050	2075
CT River	27.18 (0.917)	27.25 (0.896)	26.56 (0.867)
White	1.715 (0.935)*	1.711 (0.923)*	1.671 (0.904)
Ashuelot	0.977 (0.877)	0.971 (0.849)	0.124 (0.802)
Deerfield	1.378 (0.900)	1.367 (0.888)	1.315 (0.852)
Chicopee	1.546 (0.879)	1.556 (0.853)*	1.496 (0.809)*

The effect of climate change on hydrology has a wide range of implications in the Connecticut River Basin. The projected impacts vary across the entire basin and different sub-basins are faced with very different problems. Public safety, energy production, resource availability, and ecological impacts highlight the major concerns in the CRB. The sub-basins chosen in this section attempt to illustrate the range of issues that may need to be addressed under a changing climate.

White

The analysis on the White River basin provides an example of the impacts of climate change on streamflow in an unregulated watershed. Modeled streamflow projections for the White are represented below (Figure 69). The predicted impacts follow trends similar to those discussed for the entire basin in the previous section. The B1 and A1B scenarios both show a consistent decrease of about 500 cfs in the mean annual streamflow peak with an increasing flow from late fall and winter. The A2 scenario shows greater fluctuation of streamflow into the future but still follows the same trend as the other two scenarios. The 2075 range of this scenario projection shows the greatest differentiation from the historic, consistent with Table 3. This is consistent across all of the sub-basins analyzed in this section. All scenario projections predict that by 2075 the spring flow peak in the White will arrive about 2 weeks earlier in the season than it has historically. Snowmelt plays a significant role in the hydrology of the northern and mountainous White River basin. Model predictions suggest a large decrease in snow depth and SWE in this sub-basin relative to the rest of the basin. The shift from snow to rain, and the consequent swing in seasonal runoff timing, is the cause of the hydrologic changes seen in the White River.

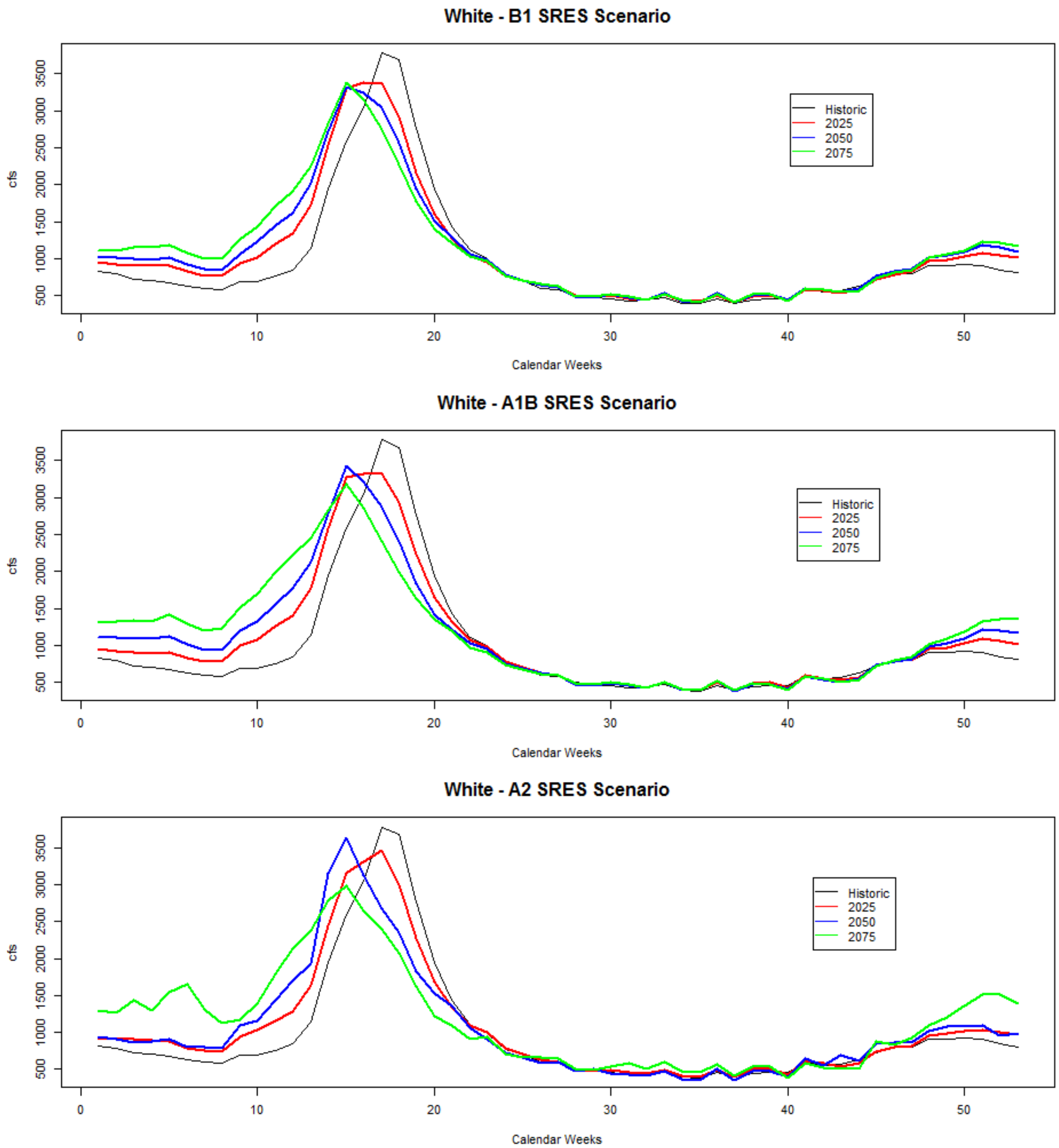


Figure 69: Future mean weekly streamflow projections for each SRES emissions scenario at the mouth of the White River.

Figure 70 shows the range of streamflow predictions from all of the SRES emissions scenarios in the White River. In 2025 historic streamflow for winter and early spring is on the lower bound of the range of GCM predictions. By the 2075 the range of GCM predictions during these seasons departs from the historic, which implies a consensus among GCM predictions that there will be a shift in winter and spring streamflow in the future.

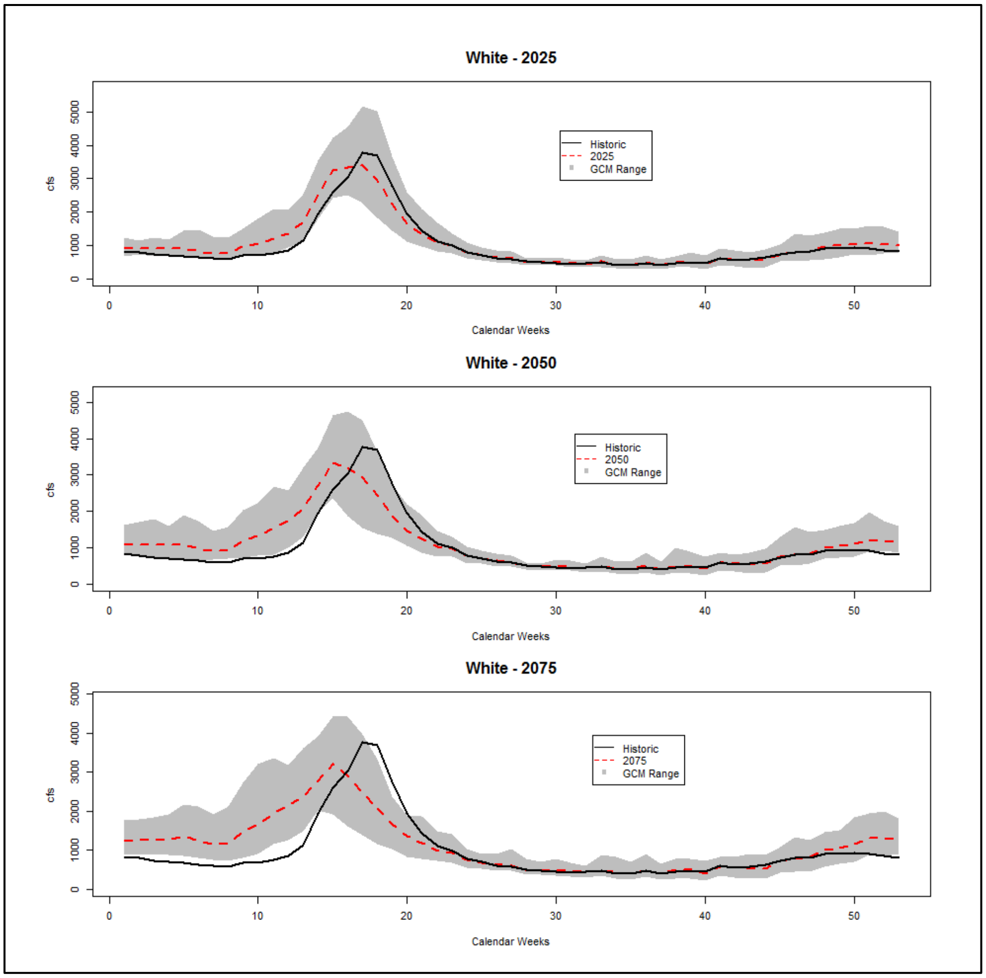


Figure 70: Mean annual streamflow and range at the mouth of the White River of all 112 GCM runs

Based on the multivariate regression of precipitation and temperature change, The White River basin shows the strongest correlation of the basins analyzed. The R^2 values are reported in Table 3 in the previous section. Figure 71 shows the spread of GCM projections for the impact of temperature and precipitation changes on streamflow. The clear correlation between precipitation and streamflow, as well as the lack of correlation between temperature change and streamflow is visually evident. In 2025 a 10% increase yields an 18% increase in streamflow. This flow increase drops to 17% for a 10% precipitation increase by 2075. The multi-variate regression also revealed that the significance of temperature change with respect to streamflow diminishes further into the future. This result is unique to the White River among the sub-basins analyzed in this section.

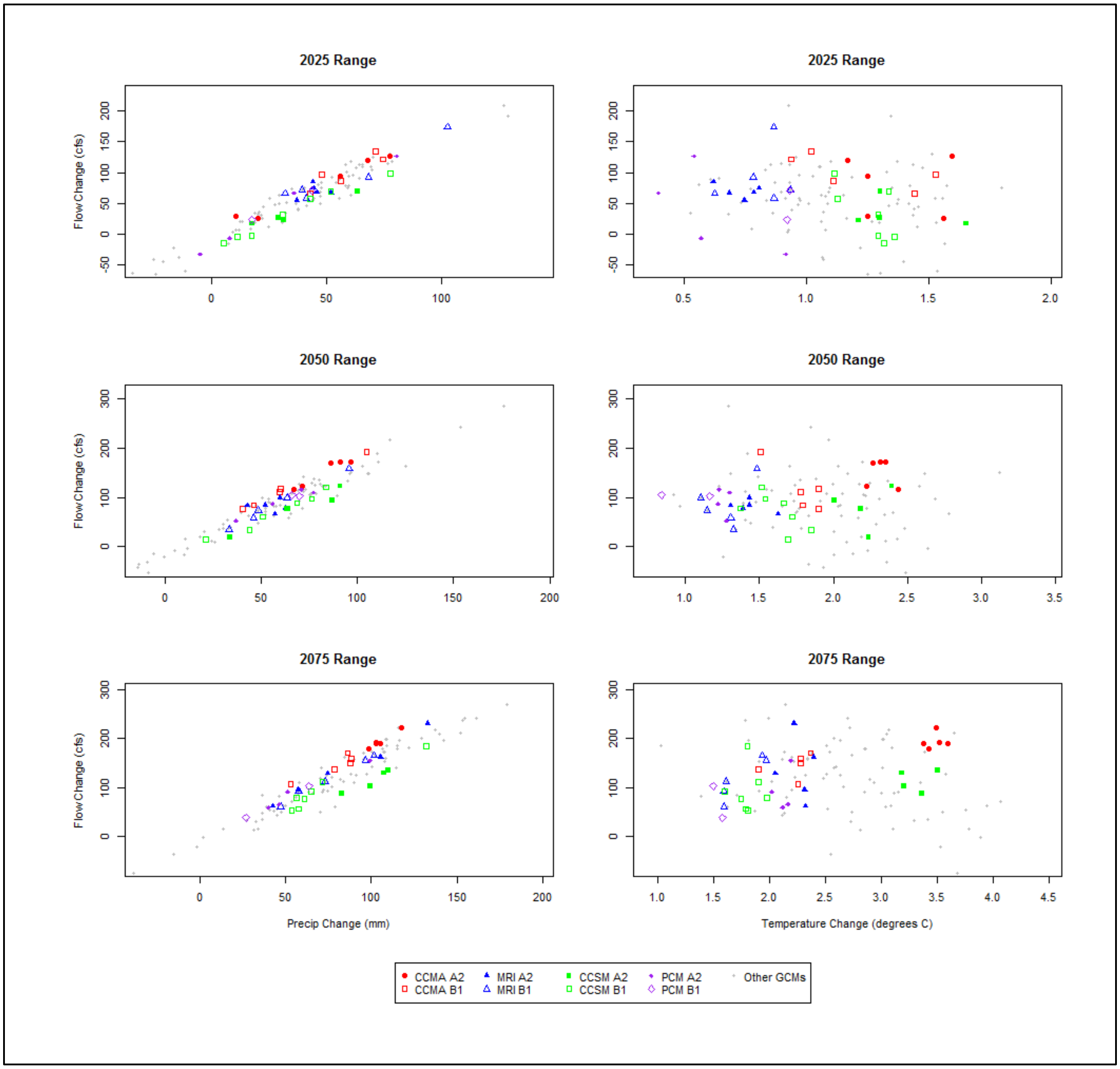


Figure 71: Change in streamflow resulting from GCM projected precipitation and temperature change in the White River Basin

The hydrologic impacts due to climate change discussed in this section have some important implications for the White River system. Reservoir management is not a concern because the basin is unregulated. The concern in this sub-basin lies primarily with the environment to which aquatic organisms have grown accustomed. A seasonal shift in flow timing and quantity can disrupt the delicate balance that fish and other aquatic organisms have with the riparian environment. Stream temperatures, sediment transport, and vegetation are a few of the characteristics that can change with a shift in hydrology. Any shift in hydrology will change the environment and possibly disrupt the established lifecycles of aquatic organisms. Environmental impacts of climate change are the principal concern for unregulated sub-basins in the region.

Ashuelot

Flood control, small hydropower operations, and anadromous fish interests make the Ashuelot River an important basin in the region. Streamflow impacts of climate change on the Ashuelot are focused on the same seasons as the White and the mainstem. The changes follow the same general trend of decrease in spring peak streamflow and increase in late fall and winter streamflow (Figure 72). The timing shift of annual peak streamflow shows a clearer progression into the future with peak timing remaining stationary through 2025 but coming roughly a month earlier in the season by 2075. Additionally, high spring flow is predicted to last longer but at a reduced magnitude from historic levels, while winter streamflow is expected to increase. This prominent “flattening” of the hydrograph is characteristic of a snowmelt driven system shifting toward a rainfall driven system. The A2 SRES scenario projections show this shift well with large streamflow peaks becoming more prevalent during the cold months. The streamflow shift is consistent with the snow depth and SWE projections in the Ashuelot River basin. Summer streamflow, consistent with much of the CRB, remains relatively stationary.

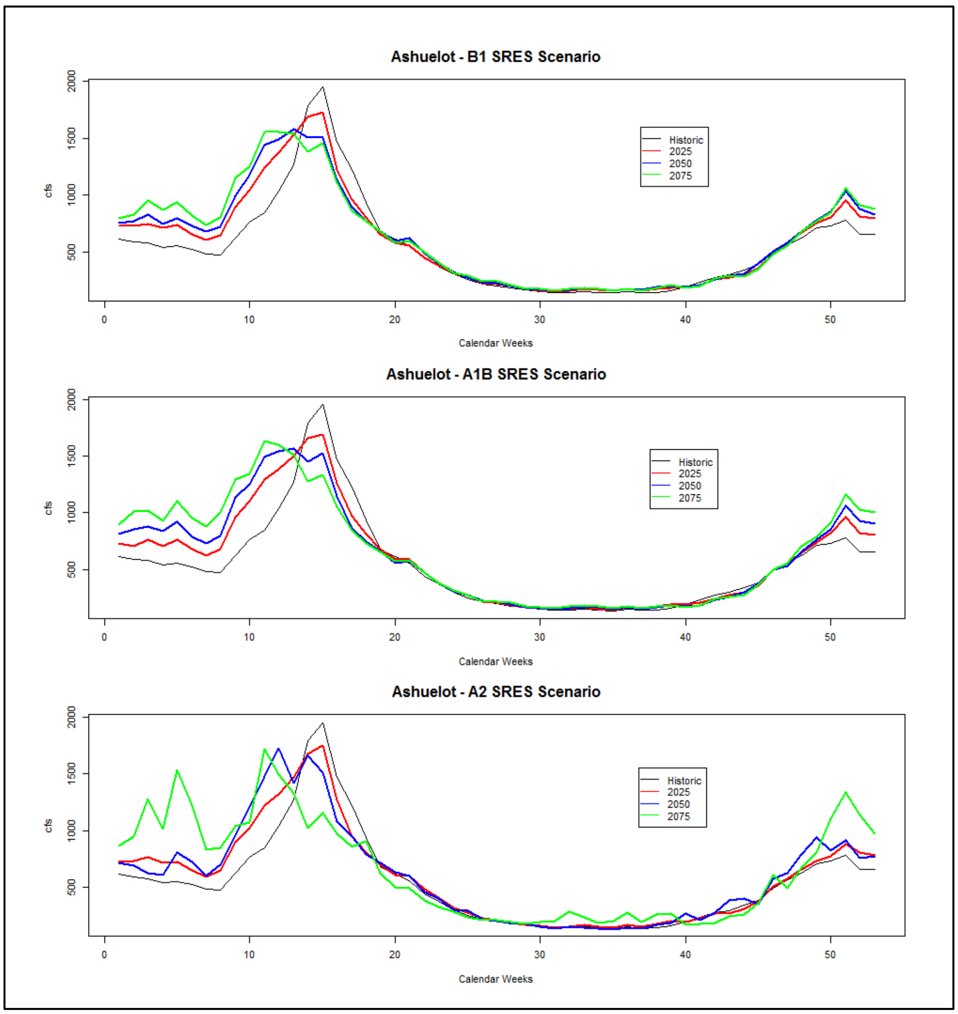


Figure 72: Future mean weekly streamflow projections for each SRES emissions scenario at the mouth of the Ashuelot River

A greater portion of the historical streamflow remains within the range of GCM predictions in this sub-basin than in the White as evident in Figure 73. The climate change projections agree that winter streamflow will increase in the Ashuelot but there is less of a consensus as to the timing and magnitude of the spring peak.

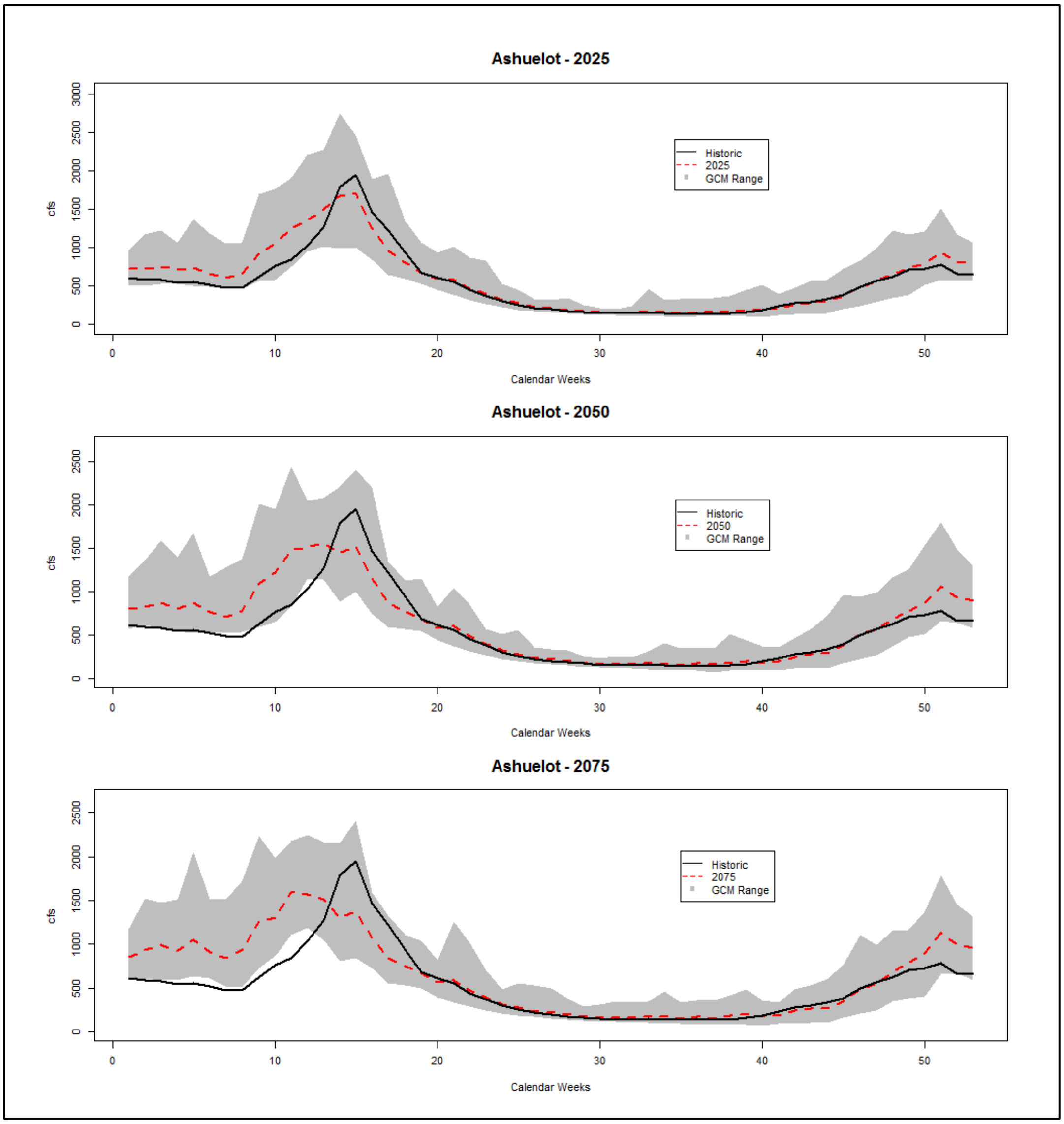


Figure 73: Mean annual streamflow and range at the mouth of the Ashuelot River of all 112 GCM runs

In Ashuelot basin, the multivariate regression again shows precipitation as the dominating factor in streamflow change (Figure 74). A 10% precipitation increase results in a 19% streamflow increase in the basin. This flow increase jumps to 20% for the same precipitation increase by 2050. Temperature remains statistically insignificant, as in all of the other basins, but does increase in relative significance through time.

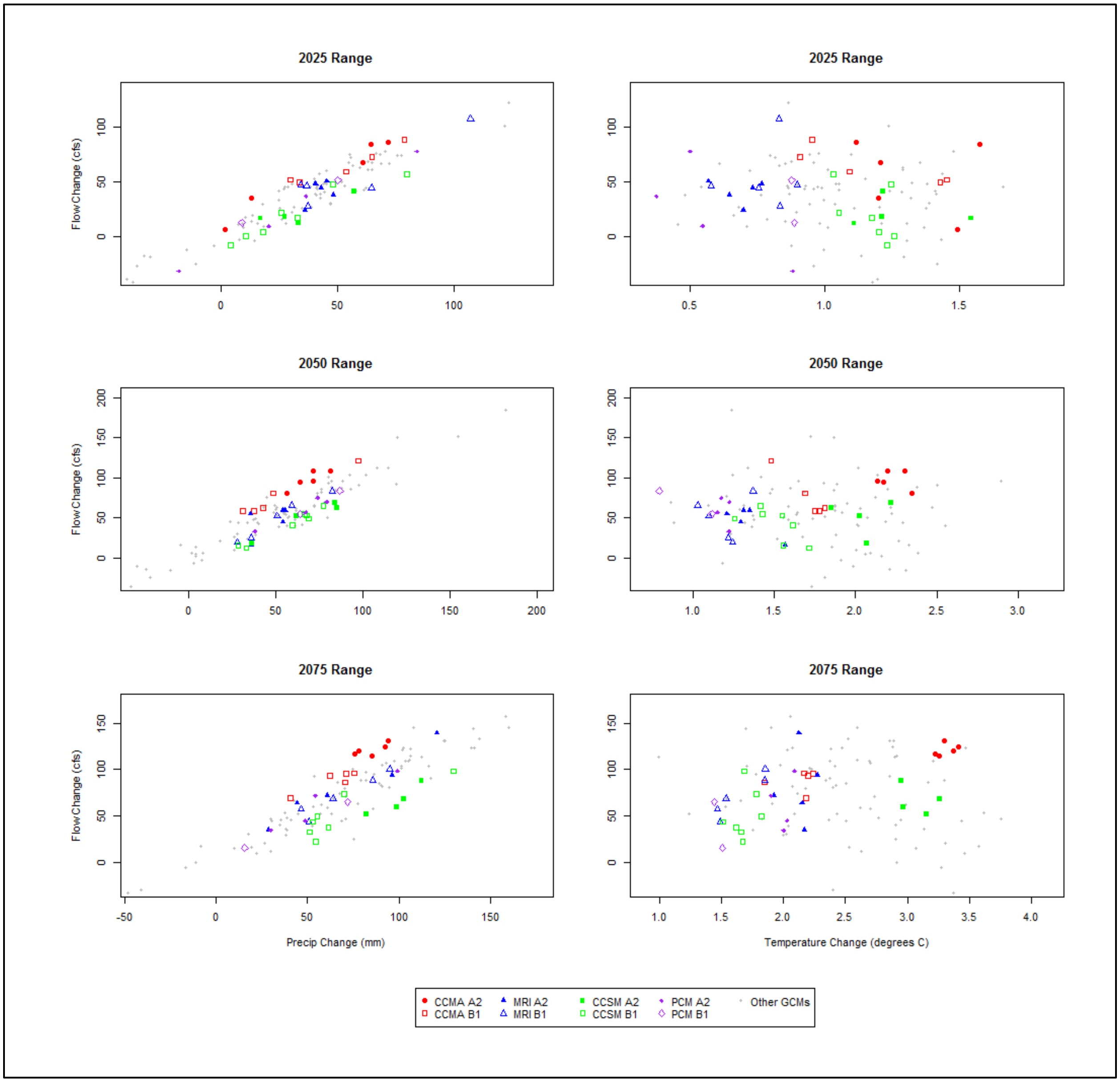


Figure 74: Change in streamflow resulting from GCM projected precipitation and temperature change in the Ashuelot River Basin

The Ashuelot River is shared by many different stakeholders with different purposes, all of which will be impacted if the hydrology of the basin is altered. Flood control operators are mainly interested in controlling peak flows. The model predicts an increase in seasonal range of peak streamflow, lengthening the time that reservoirs need to be prepared to catch a flood. Reservoir operating procedures may therefore need to be adjusted under changing hydrologic conditions, ultimately allowing for greater flexibility during seasons of uncertainty. Deviation from the historic streamflow will also impact the low head, run of river hydropower facilities on the Ashuelot River which have limited storage ability. Power production of these facilities hinges on the amount of streamflow in the river and is therefore directly linked to any changes within the system. Finally, as with the White River, there are environmental implications to changing hydrology of a system. Efforts in dam removal on the Ashuelot have been made in recent years with the purpose of restoring natural fish habitat. A shift from the historic hydrology of a river system will impact its suitability as fish habitat.

Deerfield

The analysis in this section serves as an example of the impacts of climate change on a heavily used and regulated river system. Consistent with projections throughout the entire CRB, mean winter streamflow in the Deerfield is predicted to increase while flows during the warmer half of the year are not expected to change significantly. Spring peaks are expected to decrease, especially as average annual temperature increases by more than 2^{o}C, as it does in the A1B and A2 scenarios in 2075. The shift in timing of the mean annual peak streamflow and flattening of the hydrograph is not expected be as drastic as some of the other sub-basins and the hydrograph. Peak spring streamflow is predicted to come about 2-3 weeks earlier than the historic high flow. The less drastic shift is possibly due to the fact that the streamflow is not as strongly dependent on snowmelt as is a basin such as the White River.

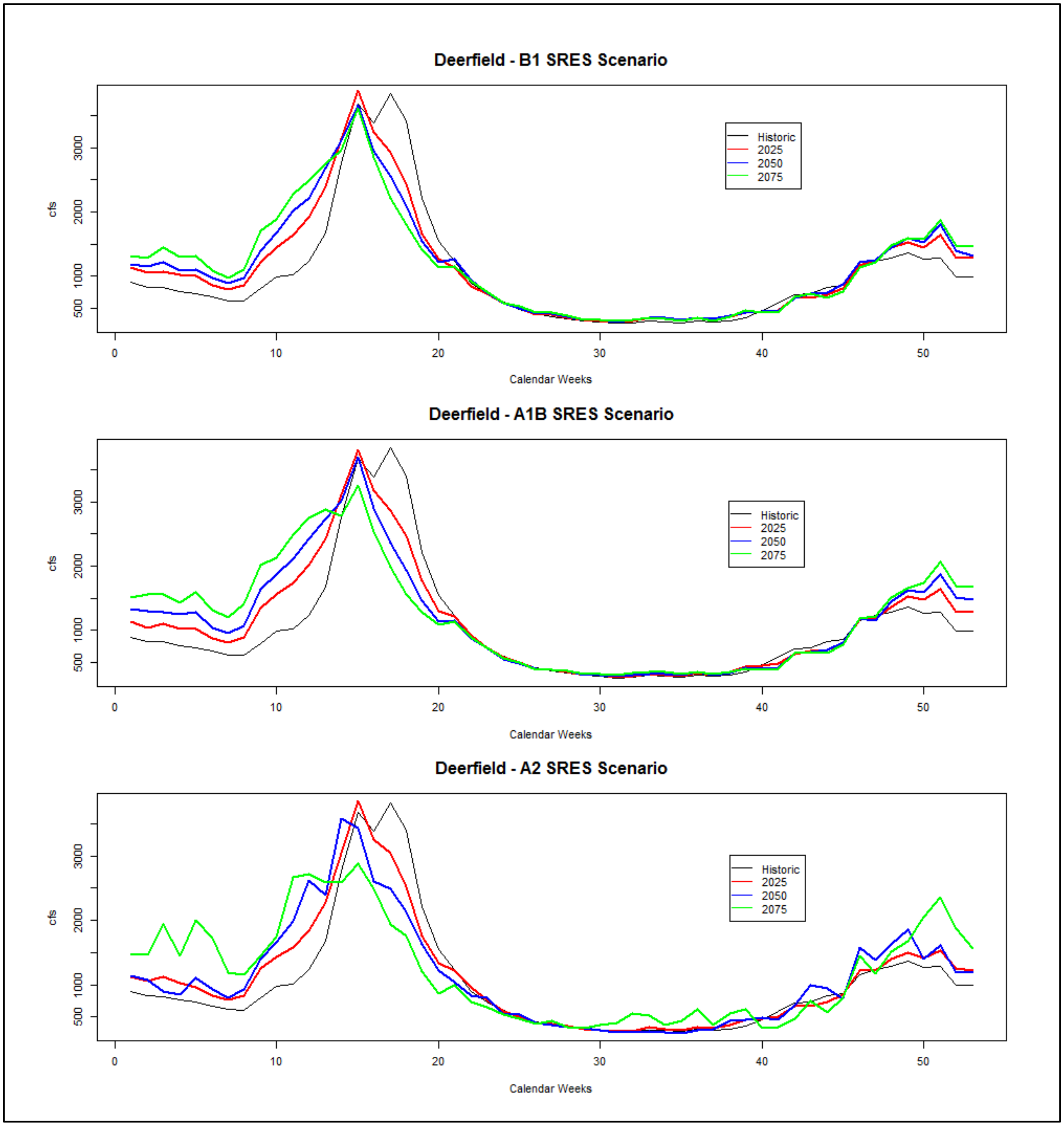


Figure 75: Future mean weekly streamflow projections for each SRES emissions scenario at the mouth of the Deerfield River

The spread of GCM projections follows a similar pattern as other sub-basins. The historic mean annual streamflow is within the GCM range in 2025, but departs from it by 2075 during the seasons most impacted by climate change.

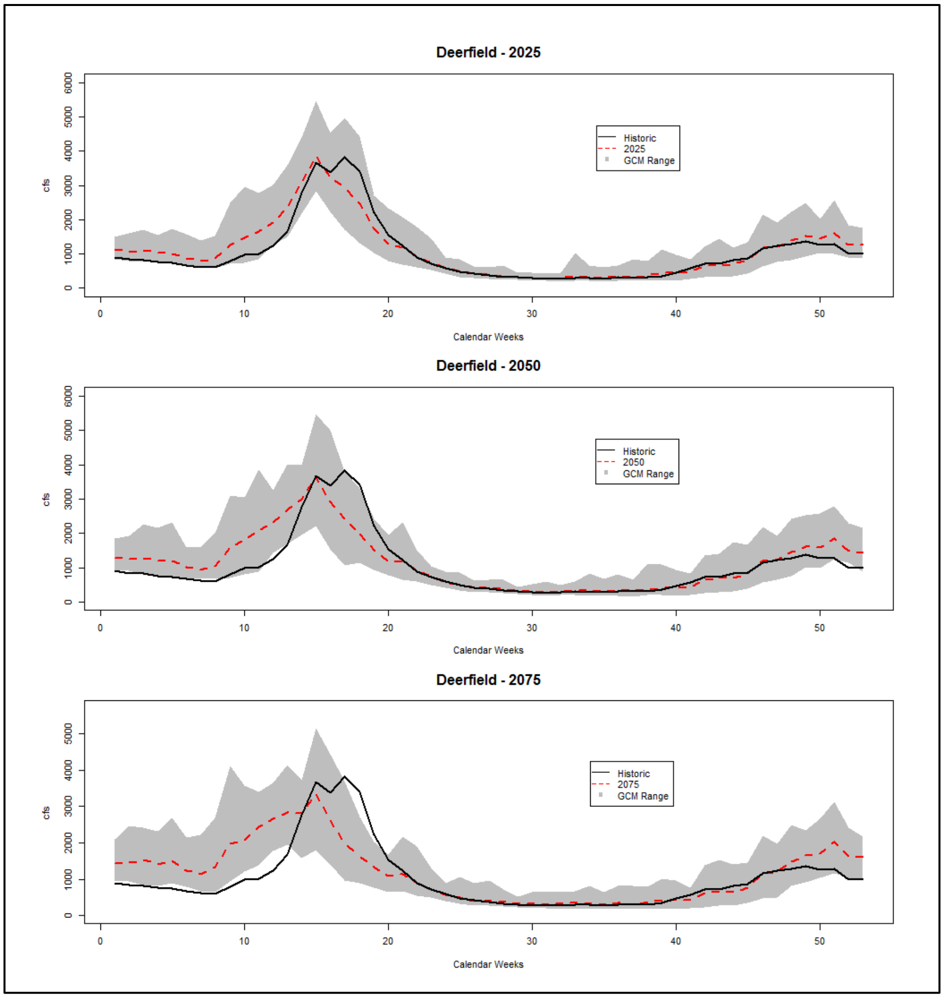


Figure 76: Mean annual streamflow and range at the mouth of the Deerfield River of all 112 GCM runs

The multivariate regression shows that precipitation is once again dominant over temperature with regard to streamflow change. A 10% precipitation increase results in a 16% increase in streamflow in the Deerfield basin. Temperature is predicted to increase in significance into the future, but still remains insignificant when compared with the impact of precipitation change on streamflow.

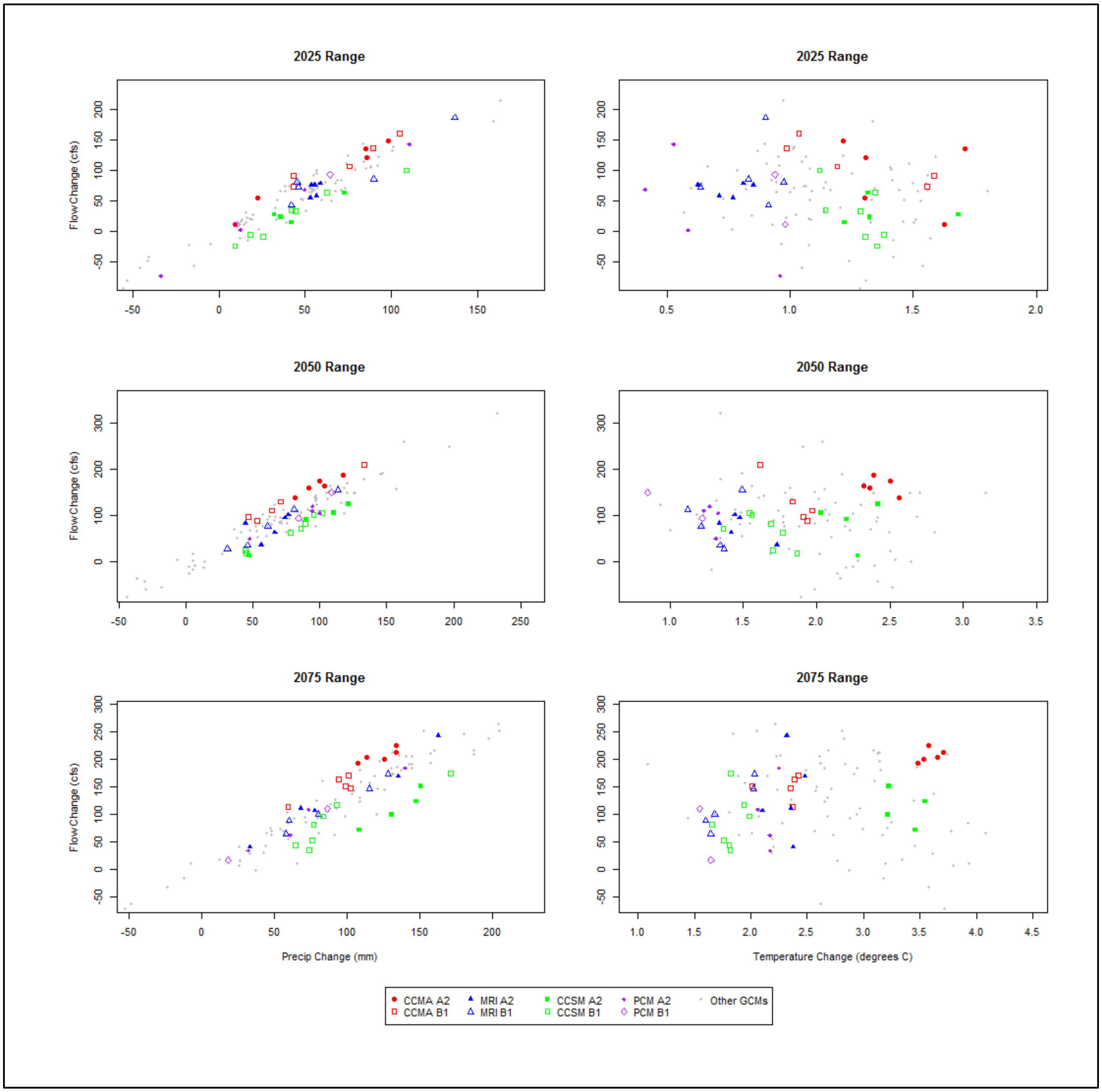


Figure 77: Change in streamflow resulting from GCM projected precipitation and temperature change in the Deerfield River Basin

The Deerfield River is used heavily for hydropower production, utilizing storage, pumped storage, and low head run of river hydroelectric facilities. Hydrologic changes resulting from climate change will be felt by these facilities. The impacts on run of river facilities are addressed in the section about the Ashuelot River. Impacts on storage facilities will differ such that the streamflow availability on which the operational procedures are based will change. Storage targets will likely need to be modified during the seasons where flow is predicted to be altered. The possibility also exists of new regulations regarding in-stream flow requirements for aquatic organisms may be implemented under changing hydrologic conditions. These restrictions will certainly impact hydropower operations on the Deerfield. All of the concerns mentioned in previous sections pertain to the Deerfield River, but the most prominent is hydropower production.

Chicopee

The Chicopee River differs from the other basins examined in previous sections. The mean annual streamflow peak is more rounded and the streamflow is steadier than the other basins. Snowpack and SWE, which play a limited role in streamflow, are not predicted to change significantly relative to changes in other regions of the CRB. The streamflow predictions in this section are unique in that there is no projected decrease in volume, only an increase from late fall into the early spring. All scenarios are in agreement on this and only differ slightly on the quantity of streamflow increase. Of the sub-basins analyzed for impacts of climate change, the Chicopee River appears to be the most robust system.

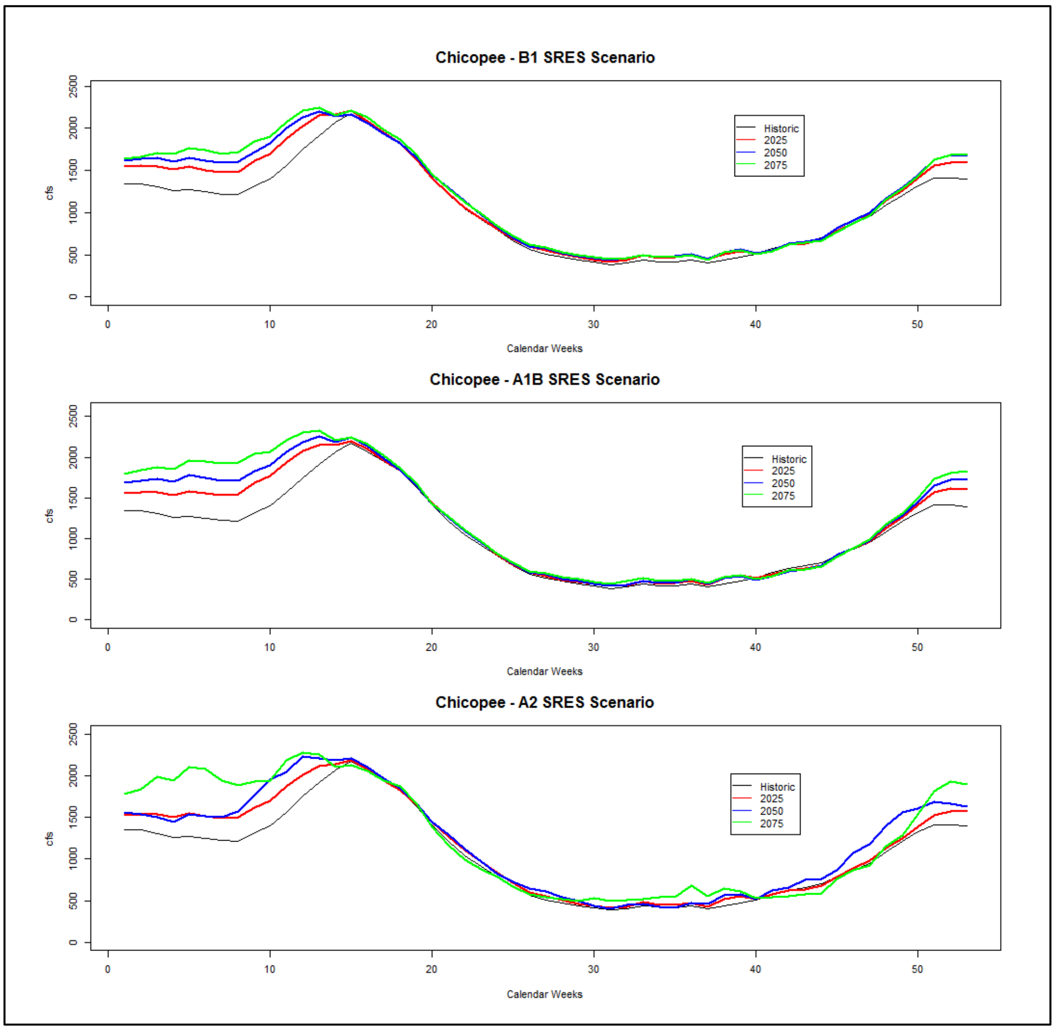


Figure 78: Future mean weekly streamflow projections for each SRES emissions scenario at the mouth of the Chicopee River

The historic modeled streamflow for the Chicopee falls into the range of GCM projections for much of the year. The GCMs are only unanimous in predicting an increase from the historic streamflow for a few weeks during the winter months. The mean GCM generated streamflow projections follows the historically modeled flow for roughly two-thirds of the year.

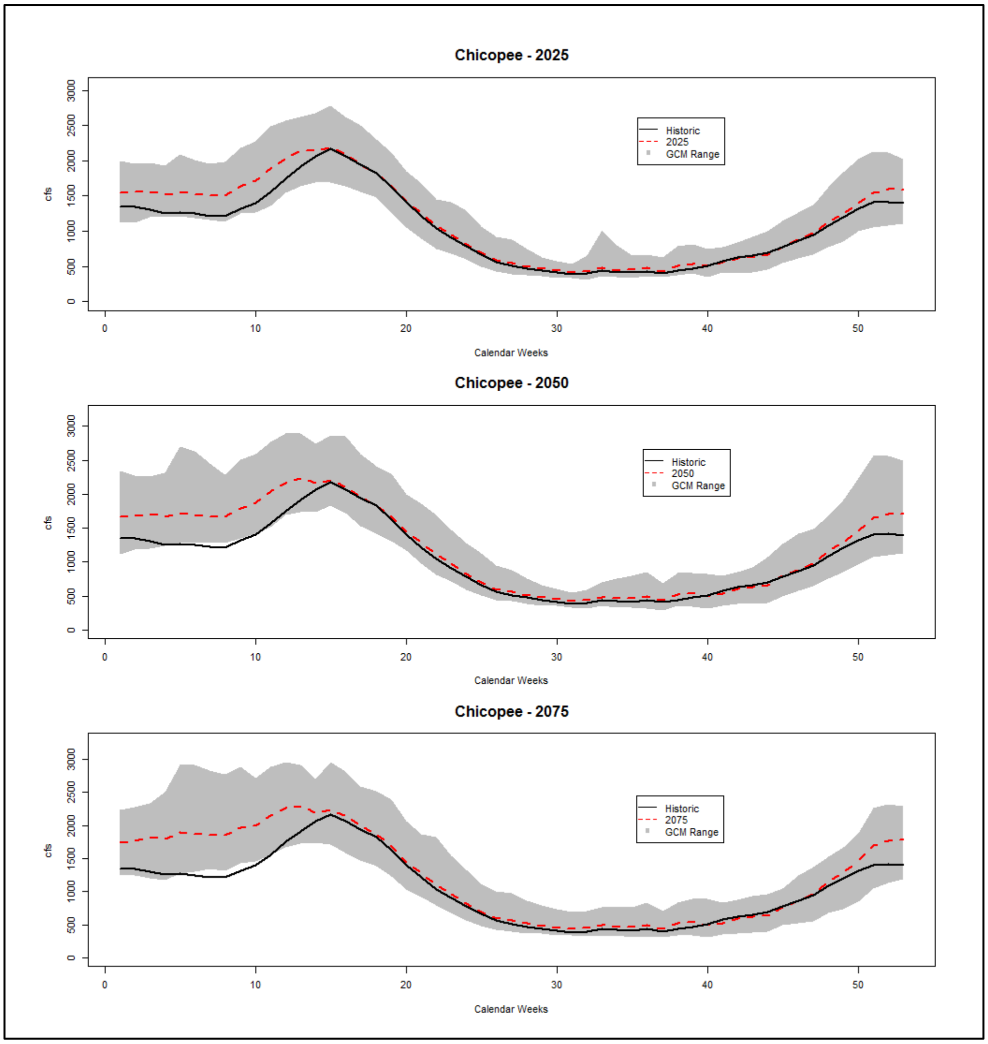


Figure 79: Mean annual streamflow and range at the mouth of the Chicopee River of all 112 GCM runs

Change in precipitation is once again revealed by the multivariate regression to be dominant over temperature change in impacting mean annual streamflow. In 2025 a 10% increase in average annual precipitation results in an 18% streamflow increase, which jumps to 19% by 2050. As in most of the other basins, temperature is expected to increase in significance in future projections, but remain ultimately insignificant when compared with precipitation.

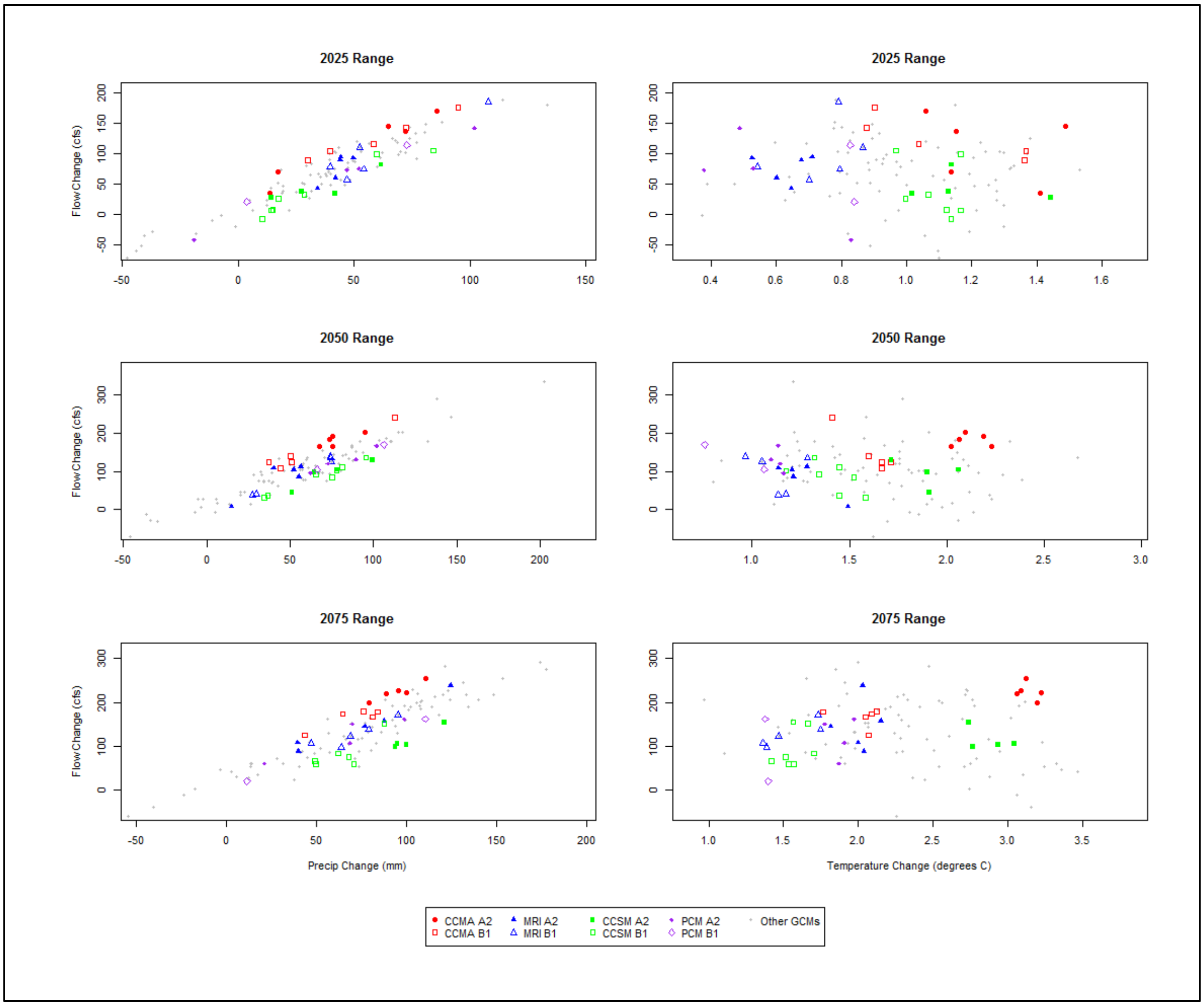


Figure 80: Change in streamflow resulting from GCM projected precipitation and temperature change in the Chicopee River Basin

The impact of climate altered hydrology on water supply is the main concern in the Chicopee basin. The Quabbin Reservoir, which supplies water to the city of Boston, is located in this sub-basin. Fortunately for the city of Boston, model predictions suggest an overall increase in volume with no seasonal decrease which occurs in the other sub-basins. Hydrologic impacts in the Chicopee may actually benefit the system when viewed from a water supply standpoint. Increased length of high flow periods may impact flood control in the system, but the water supply operation in the sub-basin appears poised to benefit from the projected streamflow changes.

Climate Impacted Optimization Model Inflows

The CRVIC model is used to generate climate impacted streamflow that may be input into the Connecticut River Environmental Assessment Model (CREAM). CREAM is a reservoir operations optimization model used to examine hydropower production, flood control, water supply, recreation metrics, and ecological targets on the Connecticut River. CREAM requires input of streamflow at each of the 54 reservoir locations and 27 ecological target locations in the basin. These locations differ from the flow locations generated by the CRVIC model, usually at the mouth of each sub-basin and a few gaged points along the Connecticut mainstem. The inconsistency in flow locations requires a reliable method for transposing the CRVIC generated flows into CREAM.

A simple drainage area scaling method is chosen as the remedy to incongruity of flow locations. Inflows at each location in the CREAM are obtained by scaling the flows from the VIC locations by respective drainage areas as shown in Figure 81. This scaling is achieved by multiplying ratio of CREAM location drainage (A) area to CRVIC drainage area (B) by the CRVIC generated flow (C). The result is a streamflow value at the CREAM location that has been scaled from the CRVIC location.

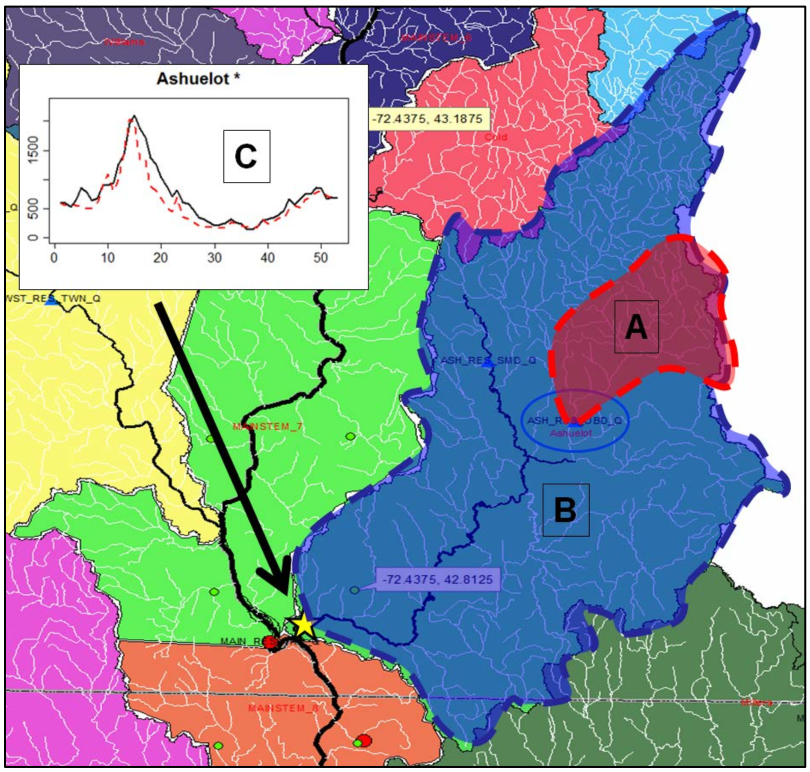


Figure 81: Visual of drainage scaling method for the Ashuelot River basin. The red shaded region represents the drainage area of the CREAM point (A), the blue shaded region represents the drainage area of the CRVIC point (B), and the hydrograph shows the streamflow at the CRVIC point (C)

This method was determined to be the most sensible because of its simplicity. The process does not involve parameter calibration and new points can be added on demand. It is worth noting that although the optimization flow points may fall within a smaller VIC area, about 1 or 2 grid-cells in the example above, the streamflow was scaled from the mouth of the sub-basins. The reason for this is that the flows at the mouth have been calibrated to a USGS gage. There is greater confidence in the accuracy of the streamflow at the gaged locations than the streamflow routed from each individual grid-cell. Overall the drainage scaling method is chosen for its simplicity and minimal time requirement.

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