Simulations of Historical and Future Trends in Snowfall and Groundwater Recharge for Basins Draining to Long Island Sound

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ABSTRACT: A regional watershed model was developed for watersheds contributing to Long Island Sound, including the Connecticut River basin. The study region covers approximately 40,900 km², extending from a moderate coastal climate zone in the south to a mountainous northern New England climate zone dominated by snowmelt in the north. The input data indicate that precipitation and temperature have been increasing for the last 46 years (1961–2006) across the region. Minimum temperature has increased more than maximum temperature over the same period (1961–2006). The model simulation indicates that there was an upward trend in groundwater recharge across most of the modeled region. However, trends in increasing precipitation and groundwater recharge are not significant at the 0.05 level if the drought of 1961–67 is removed from the time series. The trend in simulated snowfall is not significant across much of the region, although there is a significant downward trend in southeast Connecticut and in central Massachusetts. To simulate

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future trends, two input datasets, one assuming high carbon emissions and one assuming low carbon emissions, were developed from GCM forecasts. Under both of the carbon emission scenarios, simulations indicate that historical trends will continue, with increases in groundwater recharge over much of the region and substantial snowfall decreases across Massachusetts, Connecticut, southern Vermont, and southern New Hampshire. The increases in groundwater recharge and decreases in snowfall are most pronounced for the high emission scenario.

**KEYWORDS:** Watershed modeling; Climate change; Groundwater recharge; Snowfall trends; Groundwater recharge trends

1. Introduction

Knowledge of historical and future hydrologic trends is needed to develop and evaluate regional water-management strategies. In New England, the amount of groundwater discharge to streams and rivers (base flow), the amount of snow, and the timing of snowmelt are especially important factors for evaluating, managing, and planning for the health of in-stream habitat and recreation. This knowledge needs to include patterns of hydrologic change in both time and space. Regional spatially distributed hydrologic models are an effective tool to build this knowledge. Such models require strategies for parameterization and calibration that are efficient and scalable enough to allow the development of model applications of sufficient spatial resolution over large enough areas to be relevant for effective water management. Historical trends in hydrologic response can be developed by driving the model using field-based observations, but extrapolating these trends into the future requires the development of climatic forecasts, usually from general circulation models (GCMs).

Past station-based observations and GCM projections indicate that precipitation is increasing in the conterminous United States (Karl and Knight 1998; Groisman et al. 2005). These trends have also been noted in more localized studies in parts of New England (Miller et al. 2002) and New York State (Burns et al. 2007). Increases in precipitation occurring around 1970 also coincide with changes in the timing of snowmelt peaks in New England (Hodgkins et al. 2003). Key questions addressed by this paper are whether the recent increases in precipitation have also been accompanied by increases in evapotranspiration (ET) due to increasing temperatures (Fennessey and Kirshen 1994) and how the relative changes in precipitation and ET in the future will affect snowfall and groundwater recharge (Kirshen 2002). For example, if increases in evapotranspiration (due to increases in temperature) exceed increases in precipitation, the possibility exists that increases in ET could consume enough soil water to severely reduce groundwater recharge, which leads to the question, will climate change result in a net reduction in groundwater recharge, or will the opposite happen if evapotranspiration increases do not keep pace with increases in precipitation? This study does not consider changes in flow pathways that would result from land-cover change or other physical change in the hydrologic response unit (HRU). As such, the basic manner in which runoff occurs will not change even if the amount of precipitation $P$ or ET changes, and changes in surface runoff and groundwater recharge will largely remain proportional to the relative balance between $P$ and ET except in the case where the consumption of soil water by ET becomes limiting to recharge.
The Precipitation-Runoff Modeling System (PRMS) (Leavesley et al. 1983) is an effective tool for addressing questions about the interaction of hydrologic processes because it simulates surface, soil, subsurface, and groundwater storage flux; runoff; snow cover and snowmelt; and a large number of other hydrologic variables on a daily time step. PRMS can be used to understand spatial variation of the hydrologic responses over large areas as well. It can be parameterized at a wide range of scales with any discretization scheme for subdividing the modeled region. PRMS simulations are based on the spatial variation in measurable physical characteristics that can be quantified using geographical information systems (GIS) and widely available spatial datasets, including land cover, topography, soil, and geology.

There have been two recent large-area modeling efforts using PRMS completed by the U.S. Geological Survey (USGS), one in the Delaware River basin (Pennsylvania and New Jersey) and the other in the Flint River basin (Georgia). In the Flint River study, the basic PRMS model was used unmodified including the method to simulate solar radiation, evapotranspiration, groundwater recharge and discharge, snowfall and snowmelt, and subsurface flow routing. However, in the Flint River basin an additional process module was added to account for numerous small ponds used for water storage that were found to significantly affect routing of runoff and streamflow (Viger et al. 2010; Viger et al. 2011). In the Delaware River study, a newer version of PRMS, developed for use with the USGS modular groundwater flow model (MODFLOW) in a coupled groundwater surface water flow model was used. In this model, most of the algorithms used were the same as used in this study; however, this model was operated on an hourly time step and required additional routing procedures and included a transfer function between HRUs and a more complex shallow subsurface flow zone that accounts for a two-layer unsaturated zone above the saturated groundwater zone (Goode et al. 2010). In the study presented here, we are not as interested in subdaily routing of flows, and rely on the simpler Muskingum routing scheme used in the original PRMS.

This study is designed to assess the application of PRMS at a regional scale. The model is used to assess the spatial and temporal distribution of air temperature, precipitation, actual evapotranspiration, runoff and streamflow, snowfall, and groundwater recharge across southern and central New England within the Connecticut, Thames, and Housatonic River watersheds. Additionally, the model is used to simulate potential future changes in groundwater recharge and snowfall resulting from climate change using output from a GCM to adjust input weather records (Hay et al. 2011). The paper does not evaluate the quality of the future climate trends derived from the GCM, focusing on trends rather than absolute values of potential future changes in hydrologic quantities.

The intent of the paper is to look specifically at the changes in groundwater recharge and snowfall that could result from climate change. The model performance was assessed by looking at how well mean annual streamflow was simulated over a 46-yr period at 73 streamflow gauges in the modeled area. Over this time period, streamflow \( Q \) can be considered the residual of the water balance between precipitation \( P \) and \( ET \). Simulated groundwater and base flow were also assessed by comparison against observed trends; however, specific groundwater datasets that can be used for comparison are not available. Simulation of snowfall was assessed by comparison against Moderate Resolution Imaging Spectroradiometer (MODIS) snow-cover data, which is not a direct comparison but serves as an
indicator. The ET estimates are compared against general knowledge derived from previous studies.

Given that precipitation is an input to the model and streamflow can be compared to numerous stream gauges, the accuracy of the simulation of ET is assumed to be adequate if streamflow is adequately estimated. The accuracy of the groundwater recharge and snowfall simulations can only be evaluated on the limited comparative data that are available and on the general shape of the streamflow hydrograph. Comparison with other hydrologic abstractions such as subsurface flow, direct surface runoff, and snowmelt runoff, which the model simulates, are difficult to directly observe and there are no real observations for comparison on a study of this scale. Thus, the confidence that the model simulates these adequately comes from the streamflow characteristics and on the physical basis of the model algorithms (http://wwwbrr.cr.usgs.gov/projects/SW_MoWS/software/oui_and_mms_s/prms_page.shtml).

2. Previous studies

With the advent of GCMs and the increased availability of results from these models (e.g., Solomon et al. 2007), more scientists and natural resource managers are seeking to understand the impact of climate change on hydrology. Early works in this area focused on global- (e.g., Arnell 1999a) and continental- (e.g., Arnell 1999b; Lettenmaier et al. 1999) scale assessments of hydrologic impacts. To more effectively support natural resource managers, especially those relying on water resources, there is a need to establish scientifically defensible methods of integrating GCM forecasts with hydrologic models at more relevant spatial scales. As a result, a large number of regional hydrology model applications have been developed to evaluate the impact of climate change in North America, including studies in the American Midwest (e.g., Jha et al. 2004; Niemann and Eltahir 2005), the American West Coast (Gleick 1987), the North American Rockies (e.g., Jain and Eischeid 2008), the Pacific Northwest of the United States (e.g., Chang and Jung 2010; Salathe et al. 2007; Wenger et al. 2010), parts of Canada (e.g., Burn 1994; Choi et al. 2009), and the Connecticut River basin itself (e.g., Marshall and Randhir 2008). There are also many studies of European regions (e.g., Kleinn et al. 2005; Menzel et al. 2006; Shabalova et al. 2003). Some studies, such as Rowell (Rowell 2009), even compare impacts across continents.

As pointed out by Varis et al. (Varis et al. 2004) and by Xu et al. (Xu et al. 2005), the functional chain created by the definition of climate change scenarios, the selection and application of GCMs and hydrological models, and the downscaling techniques used to establish connections between these components is based on sometimes crude and rapidly evolving understanding. The uncertainty in the initial GCM forecasts and limitations of the downscaling technique to interpolate these forecasts to a finer resolution serve as an important constraint on the fineness of the spatial and temporal scales at which the hydrologic modeling can be carried out. As a result, many of the regional hydrologic studies of climate change impacts cited above and in general use lumped or semidistributed models and do not have an explicit routing scheme.

Gosling et al. (Gosling et al. 2010) point out that uncertainty of climate impacts on hydrology is least in regions where expected runoff changes are greatest, such as
in the high northern latitudes and central Asia. The implication is that regions
found in lower (i.e., mid) latitudes such as the conterminous United States and
southern Canada exhibit more variable hydrologic response to changes in tem-
perature. This could be in part due to the fact that in these geographic regions
important factors, such as minimum daily temperatures in winter, are at or near
important thresholds that control the balance of snow- and rain-dominated hy-
drological responses. Several authors have pointed out that various hydrologic
states in snow-dominated regions of the United States and Canada are sensitive to
climate change, including the phase of precipitation (e.g., Huntington et al. 2004;
Knowles et al. 2006), snowpack trends (Hamlet et al. 2005), frozen soil (Cherkauer
and Lettenmaier 2003), winter–spring streamflows (Hodgkins and Dudley 2006),
lake ice-out dates (Hodgkins et al. 2002), and ultimately water availability (e.g.,
Barnett et al. 2005).

These locales are important not only for the academic purpose of building
better scientific understanding but also because there are substantial human
populations in them. Further, the use of (regional) watersheds that are large
enough to encompass both snow- and rain-dominated areas can be used to un-
derstand geographic shifts in the transition between the areas within which these
processes dominate. The study uses this region to examine not only temporal
trends for the watershed as a whole but also spatiotemporal trends within the
watershed region.

3. Study area

The region modeled is approximately 40 900 km² in area and includes 20 eight-
digit hydrologic units (HUCs) (Seaber et al. 1994) [(http://www.ncgc.nrcs.usda.
gov/products/datasets/watershed/); accessed October 2008] (Figure 1). The HUCs
comprise the Connecticut River watershed, the Thames River watershed, the
Housatonic River watershed, and smaller coastal watersheds. The study region is
located entirely within the New England physiographic province, a plateau-like
upland that rises gradually from the sea but includes numerous mountain ranges
and individual peaks [(http://ct.water.usgs.gov/nawqa/description.htm); accessed
February 2010]. Elevations range from sea level in coastal Connecticut to 1914 m
above mean sea level (MSL) at the peak of Mount Washington in the White
Mountains of New Hampshire. Most of the area is within the New England upland
section of the province, where the topography is characterized by rolling hills and
low, rounded mountains interrupted by numerous, generally narrow valleys. The
Connecticut River valley forms a broad lowland that extends from a short distance
south of the Vermont–New Hampshire–Massachusetts border to Long Island
Sound. The relief is higher in the northern part of the study unit where the Green
Mountains of Vermont and the White Mountains of New Hampshire commonly
reach elevations between 600 and 1200 m MSL.

The climate varies considerably within the study unit but is generally temperate
and humid. Average annual temperature ranges from less than 4°C in the northern
mountainous areas to about 10°C in southwestern coastal Connecticut. Average
annual precipitation for the entire study unit is 109 cm but ranges locally from
about 86 cm at places in the Connecticut River valley to more than 165 cm in some
mountainous regions. Annual precipitation, however, commonly fluctuates as much
as 50 cm from these long-term averages ([http://ct.water.usgs.gov/nawqa/description.htm]; accessed February 2010).

The Connecticut River is the principal river in the study region, extending 616 miles from its source in the Connecticut Lakes of northern New Hampshire to its outlet at Long Island Sound. The river drains 30 100 km$^2$, or about 72% of the study unit. Flow of the Connecticut River near the Connecticut–Massachusetts border averages about 481 m$^3$ s$^{-1}$ (41 000 million L day$^{-1}$) ([http://wdr.water.usgs.gov/wy2008/pdfs/01184000.2008.pdf]/); accessed July 2010). Other major streams include the Housatonic and Thames Rivers, which together drain

Figure 1. Map of the study area showing major watersheds, weather stations used in the model to define precipitation and temperature distribution, and the stream routing network.
8860 km$^2$ or about 20% of the study unit. Numerous smaller streams and rivers that flow directly into Long Island Sound collectively drain 12 770 km$^2$ in coastal parts of the study unit.

Two general aquifer types underlie the study region: unconsolidated glacial sand and gravel (stratified drift) aquifers and aquifers composed of glacial till and/or fractured bedrock (crystalline and sedimentary rock) (Grady and Garabedian 1991). Aquifers composed of stratified glacial deposits are generally the most productive sources of groundwater in the study unit and typically occupy river valleys between bedrock hills and mountainous areas with variable thickness of deposits. Till and fractured bedrock and till aquifers underlie the entire study unit and are an important source of water for self-supplied domestic, commercial, and industrial users. Bedrock aquifers primarily store and transmit water through intersecting fractures in consolidated rock.

In Connecticut, land-cover change between 1960 and 2006 (the span of the model calibration) has resulted from a mixture of conversion of agricultural land to forest (an ongoing trend since the early twentieth century) and conversion of forest to suburban development, with a net loss of 6% forest cover [Loomis et al. 2009; University of Connecticut Center for Land Use Education and Research (CLEAR); http://clear.uconn.edu]. During this same period, population in Connecticut increased by about 6%, which occurred primarily in suburban areas. The changes have been localized near the coast and urban centers. In Connecticut, the overall change in land cover and population (with its attendant change in water and land use) has been relatively small from a hydrologic perspective, as suggested by recent modeling in the Pomperaug River watershed in Connecticut (Bjerklie et al. 2010). Similar trends, but to a lesser degree, have occurred in western Massachusetts. Western New Hampshire and eastern Vermont have experienced greater percent population growth but remain more than 70% forested with low population densities. Areas of more significant land-cover change and population growth have occurred in eastern Massachusetts and New Hampshire; however, these are outside of the modeled region.

Bjerklie et al. (Bjerklie et al. 2010) indicate that effective impervious areas (those areas that contribute directly to streams) associated with development need to reach 10% or greater before important hydrologic effects are observed. Less than 5% of the HRUs in the model have impervious areas greater than 10%, and these are localized in the Hartford, Connecticut, and Springfield, Massachusetts, areas and along the coast. We assume that only a handful of HRUs may have experienced an increase in development and land-cover change over the course of the calibration period to a degree capable of altering hydrologic response. However, future population growth and development patterns and change of forest type (e.g., change from coniferous to deciduous trees) due to climate change may be important factors in future hydrologic change and have not been considered here.

4. Data and methods

The USGS GIS Weasel (Viger and Leavesley 2007), a GIS toolbox for the delineation and parameterization of spatial features used in modeling, was used to analyze digital elevation model (DEM) data; extract a drainage network; and define land surface units, which are referred to in PRMS as HRUs. The final output of the
GIS Weasel was an initial version of a PRMS parameter file that was populated with descriptions of the spatial features (drainage network segments and HRUs) delineated within the study region along with default values for many of the nonspatial PRMS parameters. Estimates of several other parameters pertaining to groundwater fluxes were made based on available GIS information, including soil characteristics, surficial geologic characteristics, and topographic slope for several watersheds in Connecticut (Bjerklie et al. 2010).

The PRMS model uses the Jensen–Haise (JH) method (Jensen and Haise 1963; Jensen et al. 1969) to estimate PET. This method derives coefficients directly from the temperature records and basin elevation. The method is semiempirical and uses mean near-surface air temperature and diurnal saturation vapor pressure differences (estimated from air temperature), elevation, and mean incident solar radiation. The mean incident solar radiation $R_s$ is calculated for latitude and day of the year and adjusted for atmospheric attenuation and reflectance (albedo) using a method based on precipitation and temperature ([http://wwwbrr.cr.usgs.gov/projects/SW_MoWS/software/oui_and_mms_s/prms.shtml]; accessed October 2008]. The variability of the daily evaporative demand is driven by the daily variation in average near-surface air temperature and the daily incident solar radiation. The model tracks soil water and ET, and recharge to the subsurface occurs only after the soil water/ET demand is satisfied.

The Jensen–Haise model is given by the following equations:

$$
\text{PET} = C_t(T_{\text{mean}} - T_x)R_s, \quad (1)
$$

$$
C_t = \frac{1}{\left(\frac{68}{278} - \frac{h}{278}\right) + \left(\frac{650}{e_{\text{max}} - e_{\text{min}}}\right)}, \quad \text{and (2)}
$$

$$
T_x = \left[-27.5 - 0.25(e_{\text{max}} - e_{\text{min}}) - \frac{h}{1000}\right], \quad (3)
$$

where PET and $R_s$ are in millimeters of potential water evaporation, $h$ is the mean elevation of the representative station (or in the case of the model, the HRU) in feet, temperature is in degrees Fahrenheit, and $(e_{\text{max}} - e_{\text{min}})$ is the saturation vapor pressure deficit between the daily maximum and the daily minimum temperatures for the warmest month of the year given in millibars. The vapor pressure function is an index for the actual vapor pressure deficit and is used in lieu of input data for relative humidity or dewpoint temperature. The coefficients of the Jensen–Haise model ($C_t$ and $T_x$) are empirical in nature and subject to calibration.

The part of precipitation that is not evapotranspired is either stored in the soil or subsurface reservoirs in the model or routed directly to runoff as overland flow. Details of the model operation can be found in Bjerklie et al. (Bjerklie et al. 2010). The surface runoff is generated on impervious surfaces and on variable source areas that are determined by the presence of low permeability soils and a defined riparian zone next to streams. The impervious surfaces are defined from land-cover data as a fraction (effective impervious) of urban land pixels (Bjerklie et al. 2010).
The direct runoff that is generated is routed to streams within the daily time step of the model minus depression storage, which is then subject to evaporation. Water is stored in the soil, and the soil has a set capacity based on the root zone depth that is a function of the vegetation type. This soil reservoir is subject to evapotranspiration. Below the soil, the model partitions infiltrated water to the subsurface zone and the groundwater zone. The subsurface zone is comparable to the zone of interflow or rapid groundwater flow. This region in the model is considered to be present in areas of secondary porosity, weathered rock and till, and perched high permeability layers exist above the groundwater (aquitard) zone. In some places, this zone is considered to be absent, and water flowing through the soil, once it is saturated, flows directly to the subsurface zone or, if this zone is absent, flows directly to the groundwater.

Water in the subsurface zone is routed either horizontally as rapid groundwater flow or continues to infiltrate vertically through the zone to the groundwater. Two routing parameters, vertical and horizontal routing coefficients, control the rate of this process. The groundwater reservoir in the model is conceived as a mixture of low permeability bedrock and higher permeability stratified glacial deposits above the aquifer. Water entering this zone either directly from the soil or vertically through the subsurface zone is considered groundwater recharge for the purpose of this study. This is the water considered to be generally available to the groundwater system that would be tapped by wells and would supply long-term base flow to streams. Flow out of the groundwater reservoir is modeled as a one-dimensional reservoir for each HRU and is routed via a reservoir routing coefficient. PRMS also computes snowfall, snow cover, and snowmelt based on accepted methods [(http://www.brr.cr.usgs.gov/projects/SW_MoWS/software/oui_and_mms_s/prms.shtml); accessed October, 2008]. The PRMS snow methodology has been used in numerous studies in the western United States (Clark et al. 2005).

The modeled region was discretized into 511 HRUs (Figure 1) that are delineated based on the HUCs at the 12-digit level of resolution (Seaber et al. 1994) [(http://www.ncgc.nrcs.usda.gov/products/datasets/watershed/); accessed October, 2008]. Although PRMS does not require HRUs to be delineated on watershed boundaries, this model uses the 12-digit HUCs as HRUs to ensure that each HRU has a single outlet through which all streamflow passes. GIS characteristics of each HRU, including the coordinates of the HRU centroid, topography [slope, aspect, and elevation; (http://ned.usgs.gov/); accessed October 2008], soils [(http://soils.usda.gov/survey/geography/ssurgo/); accessed October, 2008], land cover (Vogelmann et al. 2001) [(http://landcover.usgs.gov/); accessed October, 2008], surficial geology, and hydrography [(http://nhd.usgs.gov/); accessed October, 2008], are determined from the national datasets. A daily water and energy balance is simulated for each of the HRUs based on precipitation and temperature input data.

The input time series of daily precipitation and minimum and maximum temperature used in the model application were derived from 69 National Weather Service (NWS) Cooperative Observer Program (COOP) weather stations in the region (Figure 1). The weather stations range in elevation from nearly 2000 m at the top of Mount Washington in New Hampshire to near sea level and are distributed from near the Canadian border to Long Island Sound and from the western edge of Vermont to the eastern edge of New Hampshire (Figure 1). The 46-yr period of daily record (1961–2006) was chosen for the analysis so that recent
climate trends could be evaluated over a relatively wide range of weather variability (Miller et al. 2002) and to incorporate this variability into the future climate datasets. The weather stations were selected to provide geographic coverage of the model domain, to represent varying elevations and environmental settings within the domain, and to have records of sufficient quality.

We recognize the difficulty and inherent issues related to distributing input weather data. In this study, we assume that the distribution pattern of the gauge records is representative, even considering issues of undercatch or overcatch at stations and observer bias (Daly et al. 2007; Hay et al. 2000; Lougeay 1976; Michelson 2004). More specifically, we assume that the measurement error around the mean is random, recognizing that there may be a consistent bias one way or the other (up or down) relative to the “true” (and unknown) mean value due to systematic measurement problems, as pointed out above. If the error about the mean is random, a consistent bias would not change the difference between time series (even though it would change the absolute magnitude of the time series), thus minimizing the implication for trend and change analysis.

All weather stations are subject to undercatch (and in some cases overcatch), depending on the specific conditions of where they are located. Obstructions, wind, and temperature (an index to evaporation) affect both the undercatch and overcatch qualities of rain gauges. Additionally, observer bias can be a factor. The method for distributing precipitation to the HRUs used in the PRMS model is the x–y–z method (Hay and Clark 2000; Hay and Clark 2003), which applies a consistent 1% correction for undercatch. Of particular concern is the measurement of snowfall (typical COOP stations measure this by taking a core of snow from the open area and determining the water equivalent). Developing and applying specific corrections for each station is beyond the scope of this study; thus, we must qualify the results to the extent that the station network inherently includes uncertainty and potential bias.

The x–y–z methodology develops a relation between the x, y, and z coordinates of each station and then applies this relation to derive adjusted values of temperature and precipitation specific to each HRU. The methodology preserves the statistical characteristics of the input dataset. This method will tend to spread precipitation more evenly across the region than might actually be the case, especially with regard to isolated thunderstorms. However, days of no precipitation at a group of nearby stations will also be reflected in no precipitation in those HRUs represented by those stations (Hay and McCabe 2002; Hay et al. 2000; Hay and Clark 2000).

In the Connecticut basin, the method provides precipitation distribution similar to other analyses for the region with the exception that it overpredicts relative to some estimates in the region north of the White Mountains in New Hampshire and in portions of the Connecticut River valley in New Hampshire and Vermont by 1–2 in. yr\(^{-1}\) (on the order of 5%) and underpredicts in the coastal region of Connecticut by the same amount. These precipitation distribution issues must be considered in the interpretation of the results.

Measurements from 73 streamflow gauges are used for calibration and comparison, along with independent estimates of evapotranspiration where available. The 73 selected streamflow gauges all have watershed areas larger than 65 km\(^2\) and a period of record of 46 years or more [(http://waterdata.usgs.gov/nwis)]; accessed
October 2009]. The streamflow gauge with the largest contributing watershed is the Connecticut River at Thompsonville (USGS station 01184000; watershed area 25,000 km$^2$). This streamflow gauge measures runoff from approximately 61% of the modeled region. On average, approximately 6.5 m$^3$ s$^{-1}$ is diverted out of the Connecticut River watershed upstream from the Thompsonville gauge to the Quabbin Reservoir, which serves the Boston metropolitan area ([http://www.mwra.state.ma.us/04water/html/hist1.htm]; accessed February 2010). The consumptive water use was estimated by assuming 15% of the mean annual water use is consumptive (Solley et al. 1998). The estimated water use was derived from population density for each HRU assuming a water use of 560 L day$^{-1}$ per person (Solley et al. 1998), which includes domestic, agricultural, industrial, and commercial uses on a per capita basis for New England.

Future climate datasets were generated by using GCM output to modify the historical 46-yr record (1961–2006) from all 69 NWS COOP weather stations used in the model (Hay et al. 2011). The GCM output were obtained from the World Climate Research Program’s Coupled Model Intercomparison Project phase 3 multimodel dataset archive, which was referenced in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report Special Report on Emission Scenarios (SRES) (Solomon et al. 2007). Four datasets were derived from the Bjerknes Centre for Climate Research Bergen Climate Model (BCC-BCM2.0) developed in Norway. This model uses a 128 × 64 square mile grid, which is approximately 2°, and was chosen as a typical model based on comparison with five other GCMs. The BCC-BCM2.0 model simulates changes in temperature and precipitation and results in PRMS streamflow simulations that tend toward the average of other GCM models (Hay et al. 2011) based on a limited sensitivity analysis of the GCM input from these models run through the PRMS simulations. Future land use, land cover, and water use were not considered in the future climate scenario.

Climate change fields for daily precipitation and daily maximum and minimum air temperatures were derived by adjusting the observed record by the change simulated by the GCM for a 12-yr moving average window (Hay et al. 2011). Two periods of downscaled data, 2001–12 and 2088–99, representing the current/near-future condition and the end of century, were used to calculate the difference between the simulated current/near-future conditions and the simulated end of century conditions under each of the two emission scenarios. Although the downscaled data are based on the 12-yr moving window, the average adjustments are applied to the historical 46-yr record of input data, thus providing a dataset that is comparable in its variability with the original input data. The modified PRMS input files incorporate the detail of the station records as well as the change factor derived from the closest GCM grid node. The only difference between the original and modified PRMS input files are in the mean, maximum, and minimum values. The temporal variability and sequencing remains unchanged, limiting this method to studies examining changes in mean climatic conditions. The basic assumption behind the downscaling method is that the variation of the recent past climate (weather) is similar to what it will be in the near future and that only the mean values (of precipitation and air temperature) will change.

Two carbon emission scenarios were chosen for each of the two time periods (thus four datasets): one is the most optimistic with regards to future greenhouse
gas emissions (SRES scenario B1) and the other is less optimistic (SRES scenario A2). The smallest projected changes in precipitation and temperature are associated with the SRES scenario B1, and the largest projected changes, as well as the largest uncertainties, are associated with SRES scenario A2. In general, the largest variability and uncertainty in the GCM projections is with precipitation.

The model provides spatial distribution of input variables (temperature and precipitation) as well as various hydrologic output variables. The recent trends in the data inputs and selected output variables therefore are assessed spatially in this study. The most severe drought in the study area since the 1930s occurred during 1961–67 (Miller et al. 2002). To assess the effects of the drought on the climate dataset, trends in the data were assessed for the period including the drought (1961–2006) and for the period excluding the drought (1967–2006). The Kendall tau rank correlation coefficient (Kendall 1938) was used to evaluate the nature of the trends over the two periods of record (1961–2006 and 1967–2006).

The Kendall tau rank correlation coefficient is a nonparametric statistic used to measure the degree of correspondence between two rankings (time series) and assessing the significance of this correspondence. In this case, the first time series is the date (ranked as a time step, starting with 1 to the end of the series $n$) and the second time series is the sequential hydrologic value of interest. The date, which is always increasing between time steps (ranks), is compared with the direction of change in the variable time series, either increasing or decreasing. If the hydrologic variable increases between time steps, the paired observations (ranked time and hydrologic variable) are called concordant, meaning they change in the same direction and, if the variable change decreases relative to the increasing time step, the paired observations are called discordant. If the agreement between the two rankings is perfect (the two rankings are the same), the coefficient has value 1. If the disagreement between the two rankings is perfect (one ranking is the reverse of the other), the coefficient has value $-1$. For all other arrangements, the value lies between $[-1, 1]$, and increasing values imply increasing agreement between the rankings. If the rankings are completely independent, the coefficient has value 0 on average. Kendall’s tau is a measure of correlation and so measures the strength of the relation between two variables. The Kendall’s tau coefficient is defined as

$$\tau = \frac{n_c - n_d}{\frac{1}{2}n(n-1)},$$

where $n$ is the total number of pairs, $n_c$ is the number of concordant pairs, and $n_d$ is the number of discordant pairs in the dataset.

5. Calibration

The model was calibrated to the discharge at 73 USGS streamflow gauges from the period 1961–2006 by adjusting the Jensen–Haise coefficients to correct for latitudinal differences in evapotranspiration. The Quabbin Reservoir diversion, as well as estimates of consumptive water use in the watershed, was compensated for in the PRMS calibration process by subtracting from the simulated streamflow. The size of the HRUs used in this application and the smoothing of the $x$–$y$–$z$
methodology limit the temporal and spatial resolution and thus limit the interpretation of the data. The model simulations were evaluated on the monthly and annual time step, and no inferences are made about the hydrologic process on the daily time step or for streams and areas smaller than about 30 km$^2$.

A range of JH coefficients was tested to evaluate their importance in determining the simulation results of snowfall and groundwater recharge. The Jensen–Haise coefficient is not important in the simulation of snowfall but is very important in the simulation of groundwater recharge because it determines the balance between change in precipitation and change in the actual evapotranspiration (AET). The distribution pattern of average annual AET after calibration is in general agreement with contoured estimates of average annual AET derived by Randall (Randall 1996) and estimated by Church et al. (Church et al. 1995), with areas of higher and lower AET generally located in the same regions; however, the magnitude is somewhat different. The input data used for the PRMS model showed somewhat higher precipitation than that mapped by both Randall (Randall 1996) and Church et al. (Church et al. 1995). As a consequence, the PRMS-simulated AET is also somewhat higher than the values reported in these previous studies. However, more recent studies (Hayhoe et al. 2006; Hayhoe et al. 2007) report precipitation quantities and patterns similar to that distributed by the PRMS model based on the observations at the 69 weather stations and indicate average AET similar to this study.

6. Results

The simulated mean annual water balance for the entire modeled area is illustrated in Figure 2. The figure shows the mean monthly values, in centimeters per month, for precipitation minus evapotranspiration (P-ET), total runoff (surface plus groundwater plus subsurface runoff), groundwater runoff, and soil moisture for the period 1960–2006. The monthly variation is a function of season and storage effects. Over the annual cycle P-ET equals the total runoff. The simulated mean annual streamflow for the historical 46-yr period at the 73 gauges used for comparison are shown in Figure 3. The largest errors in the simulation occur in streams with the smaller mean annual streamflows (smaller watershed area). This is likely due, at least in part, to the importance of local complexity not captured in the model in the smaller watersheds relative to the larger ones. Corrected for consumptive use and out-of-basin diversions, the mean annual daily streamflow error for all 73 stations is 4.4%, with a maximum error of approximately ±30% and a standard deviation of errors of 13%. The mean annual (Figure 4) and mean monthly (Figures 5, 6) simulated and measured streamflows for the Connecticut River at Thompsonville have $R^2$ of 0.84 and 0.77, respectively, with an overall mean error of 6% and a standard deviation of the error of 9% for annual flow and a standard deviation of 46% for the monthly error. The simulated data for the Connecticut River at Thompsonville tend to overestimate the low monthly flows (Figures 5, 6). This error could be due in part to flow management and storage upstream from the Thompsonville gauge but could also be caused by a general overestimation of the groundwater recharge and base flow and an underestimation of rapid subsurface flow. However, not all of the simulated flows exhibit this tendency because some show an underestimation of seasonal base flow (and groundwater discharge) and others show the overestimation of seasonal base flow.
The simulation error varied across the modeled area. Table 1 lists a sampling of 15 of the 73 streamflow gauges with simulated and measured annual-mean streamflow statistics, coefficient of determination $R^2$ between the simulated and measured annual means for the 46-yr record, and percent simulation error statistics for the 46-yr annual-mean simulated record. The simulated mean monthly streamflows ($N = 564$ months) have a larger range of error, with the general pattern indicating the largest percent errors in monthly simulations occurring during periods of lower flow and during the winter months, when the error is frequently negative (simulated discharge less than the measured discharge). The data in Table 1 show that the standard deviation of the simulated flows tend to be somewhat larger but follow the same trend as the observed flows, indicating that the model captures the year-to-year spatial variability of flow. This is also the case with monthly flows.

The cumulative errors tend to average out as watersheds and time spans over which data are averaged become larger. In large part, the errors associated with smaller spatial and temporal scales within the model are due to the effects of the coarse scale of the model tending to reduce the importance of local physical variability in weather, geology, elevation and relief, and land cover. The potentially wide range of error for individual months in specific watersheds that comprise a part of the whole indicates a key calibration issue regarding parameter optimization for large-area models: minimizing mean overall error can mask larger errors in specific regions of time and space within the larger model.

<table>
<thead>
<tr>
<th>Table 1: Simulation Error Statistics</th>
</tr>
</thead>
<tbody>
<tr>
<td>Streamflow Gauge</td>
</tr>
<tr>
<td>-------------------</td>
</tr>
<tr>
<td>1</td>
</tr>
<tr>
<td>2</td>
</tr>
<tr>
<td>3</td>
</tr>
</tbody>
</table>

![Figure 2](image.png)

**Figure 2.** Regionally averaged monthly values of key water balance variables for the period 1960–2006. The variables include precipitation minus evapotranspiration, total runoff, groundwater runoff, and soil moisture. The values are reported in centimeters per month and represent the spatially averaged value for the period of calibration for the entire modeled area.
large-area model calibrated as a whole therefore need to be used with some caution when interpreting smaller spatial and temporal scales. However, the use of a single calibration provides advantages as well, particularly when comparing relative change between different spatial and temporal regions, because the comparative calibration is consistent. If different calibration methods and criteria were used for different HRUs or for different regions in the model, then the comparability of the model simulations between HRUs and regions would be put into question, because the differences observed could be more a function of the calibration than the physical characteristics of the HRUs.

The PRMS model simulates total snowfall and snow-cover percentage for each HRU. The mean annual snowfall (in centimeters of water equivalent) simulated by the model for the entire modeled area for the 46-yr period is plotted in Figure 7 with a locally weighted scatterplot smoothing (LOWESS) curve also shown. The simulated snowfall shows an inflexion change around 1970, which is similar to changes in melt timing as reported by Hodgkins et al. (Hodgkins et al. 2003) for rivers in New England, suggesting a link between a reduction in snowfall and earlier snowmelt peaks in streamflow. The simulated percentage of snow-covered land surface can be compared to statistics of snow-cover area (SCA) maps derived from the MODIS data [(http://modis-snow-ice.gsfc.nasa.gov/); accessed February 2010] (Riggs et al. 2006).

The per-HRU estimates of SCA from MODIS and PRMS, averaged for the 7-yr simulation period that overlaps with the MODIS observations (2000–06), are
shown in Figure 8. The position on the x axis, shown as HRU numbers, on this plot indicates the north–south coordinate (in the Albers projection) of each HRU centroid, with more northerly values positioned closer to the origin. The average daily snow-cover percentage for the entire simulation region as observed by MODIS is 32% (meaning that on average over the course of the period of record, on any given day of the year, the study region is 32% snow covered) and for the PRMS simulation is also 32% (note that these percentages would be higher if only the winter months were used in the average). Additionally, the correlation coefficient between the mean annual MODIS and PRMS snow-cover percentages for each HRU is 0.92 ($R^2$ of 0.84), indicating that PRMS and MODIS show similar patterns of snow cover. However, as can be seen in Figure 7, there is a high bias for the SCA values simulated for HRUs in the northern end of the watershed (small HRU number on the left of the x axis) and a low bias for HRUs in the southern end of the watershed. The low bias can be explained, at least in part, by the tendency for MODIS to misclassify bright surfaces that are not snow (such as clouds) as snow (Riggs et al. 2006) and the high bias due to the inability of PRMS to account for uneven melting (patchiness) within the HRU, snow blown off of conifer forest cover (predominant at higher latitude and elevation HRUs), and blown snow leaving bare ground in parts of the HRU. Given that the overall mean value for the MODIS-observed and PRMS-simulated SCA are in agreement and the close association between simulated and observed snow-cover variation between HRUs, it is concluded that the PRMS snow-cover estimates are reasonable for comparison across the region and for estimated future conditions.
The model simulation results for precipitation, temperature, AET, runoff and streamflow, groundwater recharge, and snowfall are shown in Figures 9–15. Each figure shows a different variable. Within each figure, three things are shown: (i) the per-HRU mean for the variable during the period 1961–2006, (ii) the per-HRU Kendall tau rank correlation trend statistic (Kendall 1938) for 1961–2006 of that variable, and (iii) the per-HRU Kendall’s tau rank correlation trend statistic (Kendall 1938) for 1967–2006. The first time period includes the drought of 1961–67, which is the most severe drought in the region since the 1930s (Miller et al. 2002). The inclusion of this period of time affects the statistics of the Kendall tau analysis, and for this reason the second time period excludes the drought.

The spatial distribution of the Kendall’s tau coefficient for each HRU can be interpreted as a trend surface such that the size of the correlation indicates a more pronounced temporal trend. The maps show the variation of the strength of any trend. If the trend is not significant at the 0.05 probability level or lower, then the HRU has no color. If a higher significance level were mapped, for example the 0.1 level (significant at the 90% level), then the trends would appear to be more prevalent.

The trend surfaces for the 0.05 level show that, over the recent past (1961–2006), precipitation is increasing most notably in the eastern and western highlands region of Connecticut and Massachusetts (Figure 9b). However, for the period 1967–2006, the increase is not significant anywhere in the region. The increase in the maximum temperature (Figure 10b) is weaker than the increase in minimum
temperature (Figure 11b) in all HRUs, indicating that the minimum temperatures are rising faster than maximum temperatures, and thus the difference between the maximum and minimum temperature is narrowing. Similar changes in temperature have been measured in New York State (Burns et al. 2007). In addition, the rise in minimum temperature is more widespread across the region.

AET shows the lowest values at higher elevations, which is consistent with expectations (Figure 12a) and is fairly uniform over much of the rest of the region. AET is also increasing across the study area, with the most pronounced increases in the western highlands of Massachusetts, Connecticut, and southern Vermont (Figures 12b,c). This trend is evident for both time periods, 1961–2006 and 1967–2006. Runoff is generated primarily in the upland (higher elevation) areas as is expected (Figure 13a), gradually accumulating in the stream network (Figure 13b). Streamflow trends are increasing within most of the stream network, particularly in the western highlands and in the northern parts of the study area (Figure 13c), where precipitation is increasing faster than AET. However, for the time period 1967–2006, excluding the drought of 1961–67, the streamflow trends are not significant (figure not shown).

Groundwater recharge is also increasing over most of the region, particularly in the northern part (Figure 14b); however, if the drought is excluded (Figure 12c), the trend is not significant except in the extreme northern New Hampshire. Additionally, in some HRUs the trend is negative for the time period 1967–2006 (with the drought removed from the time series), although these negative trends are not significant (and not shown). PRMS does not simulate recharge to groundwater from losing streams, which could be an important source in many areas, particularly mountainous areas. Consequently, the effects of changes in streamflow on
Table 1. Mean annual simulated and measured flow statistics.

<table>
<thead>
<tr>
<th>Streamflow Gauge No.</th>
<th>Station name/period of record</th>
<th>Lat</th>
<th>Lon</th>
<th>Watershed area (km²)</th>
<th>Mean simulated (m³ s⁻¹)</th>
<th>Mean observed (m³ s⁻¹)</th>
<th>Std dev simulated (m³ s⁻¹)</th>
<th>Std dev observed (m³ s⁻¹)</th>
<th>Coef of determination $R^2$</th>
<th>Mean of the error</th>
<th>Std dev of the error</th>
</tr>
</thead>
<tbody>
<tr>
<td>01134500</td>
<td>Moose River at Victory, VT/1947–present</td>
<td>44.511 723 1</td>
<td>-71.837 314 3</td>
<td>202</td>
<td>5.1</td>
<td>4.3</td>
<td>0.96</td>
<td>0.91</td>
<td>0.50</td>
<td>0.21</td>
<td>0.19</td>
</tr>
<tr>
<td>01139000</td>
<td>Wells River at Wells River, VT/1940–present</td>
<td>44.150 341 3</td>
<td>-72.065 091 5</td>
<td>259</td>
<td>5.7</td>
<td>4.3</td>
<td>1.10</td>
<td>1.22</td>
<td>0.61</td>
<td>0.36</td>
<td>0.21</td>
</tr>
<tr>
<td>01137500</td>
<td>Ammonoosuc River at Bethlehem Junction, NH/1939–present</td>
<td>44.268 674 1</td>
<td>-71.630 361 7</td>
<td>365</td>
<td>6.4</td>
<td>5.9</td>
<td>0.96</td>
<td>1.22</td>
<td>0.60</td>
<td>0.11</td>
<td>0.13</td>
</tr>
<tr>
<td>01204000</td>
<td>Pomperaug River at Southbury, CT/1932–present</td>
<td>41.480 651 4</td>
<td>-73.224 558 6</td>
<td>199</td>
<td>3.6</td>
<td>3.7</td>
<td>0.74</td>
<td>1.13</td>
<td>0.67</td>
<td>-0.01</td>
<td>0.18</td>
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<tr>
<td>01186000</td>
<td>West branch Farmington River at Riverton, CT/1955–present</td>
<td>41.962 872 8</td>
<td>-73.017 606 1</td>
<td>332</td>
<td>8.0</td>
<td>7.2</td>
<td>1.67</td>
<td>1.78</td>
<td>0.66</td>
<td>0.14</td>
<td>0.18</td>
</tr>
<tr>
<td>01151500</td>
<td>Ottauquechee River at North Hartland, VT/1930–present</td>
<td>43.602 570 8</td>
<td>-72.354 258 2</td>
<td>570</td>
<td>12.0</td>
<td>11.7</td>
<td>2.41</td>
<td>3.09</td>
<td>0.52</td>
<td>0.05</td>
<td>0.18</td>
</tr>
<tr>
<td>01152500</td>
<td>Sugar River at West Claremont, NH/1928–present</td>
<td>43.387 573 3</td>
<td>-72.362 034 1</td>
<td>715</td>
<td>14.2</td>
<td>12.2</td>
<td>2.66</td>
<td>3.48</td>
<td>0.72</td>
<td>0.21</td>
<td>0.20</td>
</tr>
<tr>
<td>01197500</td>
<td>Housatonic River near Great Barrington, MA/1913–present</td>
<td>42.232 033 1</td>
<td>-73.354 832 8</td>
<td>730</td>
<td>15.6</td>
<td>15.0</td>
<td>2.97</td>
<td>4.11</td>
<td>0.69</td>
<td>0.07</td>
<td>0.17</td>
</tr>
<tr>
<td>01168500</td>
<td>Deerfield River at Charlemont, MA/1913–present</td>
<td>42.625 917 5</td>
<td>-72.855 095 6</td>
<td>989</td>
<td>25.6</td>
<td>26.0</td>
<td>4.56</td>
<td>5.35</td>
<td>0.77</td>
<td>-0.01</td>
<td>0.10</td>
</tr>
<tr>
<td>01161000</td>
<td>Ashuelot River at Hinsdale, NH/1907–present</td>
<td>42.785 915 8</td>
<td>-72.486 199 7</td>
<td>1096</td>
<td>20.6</td>
<td>20.5</td>
<td>3.99</td>
<td>5.72</td>
<td>0.73</td>
<td>0.05</td>
<td>0.18</td>
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<tr>
<td>01183500</td>
<td>Westfield River near Westfield, MA/1914–present</td>
<td>42.106 760 3</td>
<td>-72.698 981 1</td>
<td>1267</td>
<td>25.5</td>
<td>27.0</td>
<td>5.10</td>
<td>7.30</td>
<td>0.73</td>
<td>-0.03</td>
<td>0.15</td>
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<tr>
<td>01170000</td>
<td>Deerfield River near West Deerfield, MA/1904–present</td>
<td>42.535 919 4</td>
<td>-72.653 423 3</td>
<td>1484</td>
<td>35.5</td>
<td>38.0</td>
<td>6.51</td>
<td>7.96</td>
<td>0.78</td>
<td>-0.06</td>
<td>0.10</td>
</tr>
<tr>
<td>01122500</td>
<td>Shetucket River near Willimantic, CT/1928–present</td>
<td>41.700 376 4</td>
<td>-72.182 022 2</td>
<td>1085</td>
<td>18.5</td>
<td>21.1</td>
<td>3.65</td>
<td>5.72</td>
<td>0.68</td>
<td>-0.09</td>
<td>0.14</td>
</tr>
<tr>
<td>01144000</td>
<td>White River at West Hartford, VT/1915–present</td>
<td>43.714 236 1</td>
<td>-72.418 148 9</td>
<td>1829</td>
<td>37.9</td>
<td>34.8</td>
<td>7.67</td>
<td>8.86</td>
<td>0.69</td>
<td>0.11</td>
<td>0.15</td>
</tr>
<tr>
<td>01184000</td>
<td>Connecticut River at Thompsonville, CT/1928–present</td>
<td>41.987 318 6</td>
<td>-72.605 366 9</td>
<td>25 019</td>
<td>514.1</td>
<td>491.2</td>
<td>95.64</td>
<td>108.44</td>
<td>0.84</td>
<td>0.06</td>
<td>0.09</td>
</tr>
<tr>
<td>Avg</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>49.9</td>
<td>48.2</td>
<td>9.37</td>
<td>11.07</td>
<td>0.68</td>
<td>0.08</td>
<td>0.16</td>
</tr>
</tbody>
</table>
recharge are not considered. There are few published reports that document recent changes in groundwater levels in New England; however, one report (Dudley and Hodgkins 2010) indicates that summer base flow in streams has been increasing throughout New England. Base flow is generally considered to be a reflection of groundwater contribution to streams and thus would also indicate that groundwater levels and recharge are increasing during the summer as well. Snowfall for both time periods is decreasing significantly most notably in southeastern Connecticut and central Massachusetts (Figures 15b,c).

The difference between snowfall and groundwater recharge variables simulated for the current time period (2001–12) and the future time period (2088–99) under the two climate change emission scenarios (designated as A and B) are shown in Figures 16 and 17. These figures show the spatial distribution of the projected change in mean annual snowfall (in centimeters of snow-water equivalent) and mean annual groundwater recharge (in centimeters) between the current and future climate. For the low emission scenario (SRES scenario B1; Figure 16a), the maps show that snow will be decreasing over most of the region except in the northernmost area (White Mountains) but the decrease is not large.

Groundwater recharge is projected to increase over most of Connecticut and central Massachusetts but remains the same or decreases over the northern part of the region. The potential for decreasing groundwater recharge in the northern part of the region results from the increasing rate of evapotranspiration relative to the increasing rate of precipitation. This trend is also reflected in a decrease of total

Figure 7. Simulated mean annual snowfall for the study area in centimeters of water equivalent for the 46-yr simulation period. The solid line shows a LOESS smoothed curve through the data, indicating changes in the trend of the annual time series.
basin streamflow between the periods 2001–12 and 2088–99. For the high emission scenario (SRES scenario A2; Figure 17b), snow is simulated to decrease over the entire region and groundwater recharge is projected to increase across the entire region. The increase in groundwater recharge indicates that, for this scenario, precipitation increases outstrip increases in evapotranspiration, which is also reflected in an increase in streamflow. The largest increases in recharge are projected to occur in the central Connecticut River valley.

7. Discussion

The trends detected over the recent 46-yr period (1961–2006) in the observed record (as distributed by the model) indicate general increases in temperature and precipitation, and the future simulations indicate that these trends are expected to continue. Recent trends in snowfall have not been strong; however, this may not change until temperatures consistently exceed the freezing mark threshold, turning snow into rain. The low emission scenario shows that groundwater recharge in much of the northern part of the study region will not increase significantly or might decrease. However, for the high emission scenario, groundwater recharge increases throughout the region, indicating that precipitation increases will exceed increases in AET. Part of this may be due to soil moisture storage limitations. Groundwater recharge is very sensitive to the estimate of AET with the potential for different projected outcomes for the water balance and thus different outcomes.

Figure 8. Plot comparing simulated against remotely sensed measurements of snow-covered area based on MODIS data for each HRU numbered along the x axis.
Figure 9. Maps of (a) the mean daily precipitation (cm) for 1961–2006 and the Kendall tau correlation surface for (b) 1961–2006 and (c) 1967–2006.
Figure 10. Maps of (a) the mean maximum daily temperature (°C) for 1961–2006 and the Kendall tau correlation surface for (b) 1961–2006 and (c) 1967–2006.
Figure 11. Maps of (a) the mean minimum daily temperature (°C) for 1961–2006 and the Kendall tau correlation surface for (b) 1961–2006 and (c) 1967–2006.
Figure 12. Maps of (a) the mean daily actual evapotranspiration (°C day⁻¹) for 1961–2006 and the Kendall tau correlation surface for (b) 1961–2006 and (c) 1967–2006.
Figure 13. (a) Map of the mean annual runoff for each HRU (°C day⁻¹) for 1961–2006. (b) Map of the mean annual streamflow (m³ s⁻¹) accumulated in the stream network for 1961–2006. (c) The Kendall tau correlation surface for the stream network for 1961–2006. Note that the streams shown as red show no significant trend (p < 0.05). There were no significant trends for any stream reach for the period 1967–2006.
Figure 14. Maps of (a) the mean daily groundwater recharge (cm day$^{-1}$) for 1961–2006 and the Kendall tau correlation surface for (b) 1961–2006 and (c) 1967–2006.
Figure 15. Maps of (a) the mean daily snowfall (annual total divided by number of days in the year; cm day\(^{-1}\)) of snow-water equivalent for 1961–2006 and the Kendall tau correlation surface for (b) 1961–2006 and (c) 1967–2006.
for groundwater recharge and streamflow (Kingston et al. 2009), depending on the method.

As previously discussed, the coarse scale of the model presented here tends to reduce the importance of local physical variability in weather, geology, elevation, relief, and land cover. This is a key calibration issue for large-area models that use a single calibration, and the results need to be used with some caution when interpreting smaller spatial and temporal scales. However, a single calibration is useful for comparing differences in time and space in the model, because the calibration parameters were all optimized in a similar way everywhere in the model, eliminating the potential for comparative discontinuities resulting from different calibration strategies. As indicated by MODIS snow cover, it is important to understand how remotely sensed observations relate to the comparable simulated variable so that remote data can be appropriately used for calibration and evaluation of the hydrologic model.

8. Errors and limitations

Model simplifications, the degree to which input climate and parameter data are representative, the exclusion of land-cover change, and the method for spatial averaging the input climate (weather) variables all contribute to model errors.
Potential errors due to the input data have not been assessed. Weather station data were assumed to be representative of the region and were not put through any quality-assurance procedures or systematic corrections. The degree to which the GCM data are representative of the current and future scenarios has also not been assessed, and the method for downscaling and updating the existing record is another source of error. For these reasons, the data trends rather than absolute values of the projected change in hydrologic variables have been the focus of this study. Additionally, the analysis has been restricted to annual and monthly averages to smooth out temporally local noise in the GCM output used to modify the climate input to the model. Inclusion of land-use and land-cover change in time, as well as water-use change, in the model would also improve the accuracy of the simulations relative to historical streamflow, which inherently includes the effects of these changes.

9. Summary and conclusions

The 46-yr record of measurements from the point weather stations that were spatially distributed across the modeled area showed that temperature and precipitation have increased throughout the region; however, if the drought is excluded from the time series, precipitation trends are not significant at the 0.05 probability
level. The simulation also indicates AET and streamflow are increasing throughout the region; however, streamflow—similar to precipitation—is not increasing significantly if the drought is removed from the data. Similarly, groundwater recharge has increased generally across the region, most notably in the northern part of the study area, but, if the drought is removed, the increases are significant only in a small area in the extreme north of the study area. Snowfall has decreased most notably in southeast Connecticut and central Massachusetts. Hydrologic simulations based on future climate scenarios indicate that the trend in groundwater recharge is dependent on the emission scenario and the method used to simulate evapotranspiration (Kingston et al. 2009). Assuming the method used to estimate PET is sufficiently accurate, the low emission scenario indicated that in some areas groundwater recharge will decrease because of evapotranspiration increasing faster than precipitation. For the high emission scenario, groundwater recharge will generally increase throughout the region. The trend in snowfall (decreases in the southernmost and easternmost areas) measured in the past record is projected to expand, with larger areas of Connecticut and Massachusetts and parts of southern Vermont and New Hampshire experiencing lower snowfall as the temperature threshold for snow to become rain is exceeded more generally across the region.

The results here suggest that streams, depending on the scenario and location in the basin, may experience increasing or decreasing groundwater discharge. This is an important consideration in in-stream habitat, because groundwater generally provides cooler water in the summer and maintains flows at sufficient levels to support healthy aquatic habitat. Similarly, reductions in snowmelt and likely earlier snowmelt freshet dates will also have an effect on not only fish and aquatic habitat, particularly spawning and migration, but also seasonal in-stream temperature. The results also indicate that recreational activities dependent on snow, such as skiing, may be impacted over the course of this century, with lower snowfalls and shorter snow seasons.

These results do not address potential land-cover changes. Extensive urban growth and increasing demands for off-stream water use, combined with the effects of climate change, will potentially affect future flow regimes in the region. Future studies should address the combined effects of climate and land-cover change, as well as by incorporating improved algorithms for distributing precipitation data and simulating evapotranspiration and including other GCM simulation results for comparison. For more local sensitivity within a regional modeling application, the results of this study suggest the need for the creation of subregions for distributing the precipitation and temperature input data to assess the daily variability in precipitation.

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**References**


