IMPACTS OF CLIMATE CHANGE ON THE SURFACE WATER BALANCE OF THE CENTRAL UNITED STATES, 1984-2007

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IMPACTS OF CLIMATE CHANGE ON THE SURFACE WATER BALANCE
OF THE CENTRAL UNITED STATES, 1984-2007

by

Bo Dong

A THESIS

Presented to the Faculty of
The Graduate College at the University of Nebraska
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The climate system and the hydrologic cycle are strongly connected with each other. Understanding the interactions between these two systems is important, since variations in climate can trigger extensive changes in the hydrologic cycle, with significant impacts on agriculture, ecosystems, and society. Observations over the central U.S. in recent decades show numerous changes in climatic variables. This includes decreases in cloud cover and wind speed, increases in air temperature, and seasonal shifts in precipitation rate and rain/snow fraction. To assess the impacts of these variations in climate on the regional water cycle, a terrestrial ecosystem/land surface hydrologic model (Agro-IBIS) is employed in this study, forced by observed climatic inputs for the period 1984-2007. The results generally show an acceleration of the water cycle in the Upper Mississippi, Missouri, Ohio, and Great Lakes basins, but with significant seasonal and spatial complexity. Over the past 24 years, evapotranspiration has increased in most regions and most seasons, particularly during the fall, which is also a time of pronounced solar brightening. Trends in runoff are characterized by distinct spatial and seasonal variations. Since recent warming has led to a greater fraction of winter precipitation falling as rain rather than snow, spring runoff in some snow-dominated regions (such as the northern Great Lakes) has declined significantly since 1984. Other regions, however,
such as the northern Missouri basin, show large increases in runoff throughout all seasons, primarily as a result of increased precipitation. Sensitivity experiments show that the water balance is most linearly sensitive to solar radiation and relative humidity, followed by precipitation, air temperature and wind speed. Because of the interdependencies among the climate factors, the hydrological responses of climate change are highly non-linear. Seasonal hydrological responses are notably dependent on regional water and energy availability, and are affected by seasonal conditions of soil moisture and snow cover. Furthermore, precipitation is characterized as the predominant factor that affects the decadal scale hydroclimatic changes in the central U.S.
This thesis is dedicated to my parents

for their endless love, support and encouragement.
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Chapter 1

Introduction

1.1 Background

There is a scientific consensus that climate change has occurred in part as a fact of global warming since the 19th century. During the period of industrialization, a tremendous amount of greenhouse effects gases, especially CO$_2$ has been released due to fossil fuel combustion and deforestation. The increasing concentrations of greenhouse gases (GHGs) in the atmosphere have resulted in a rise of global mean temperature in the past century that is greater than any in human history – the Earth's average surface temperature has increased by about 0.8 °C (1.4 °F), with about two thirds of the increase occurring since 1980 (IPCC, 2007). Associated with the rapid warming, many changes in the hydrologic cycle have occurred, such as unprecedented recession of glaciers, melting of perpetual snow and increases in sea level. Increase in air temperature has also led to larger water holding capacity of the atmosphere, and thus more precipitation on the global scale. However, on the regional scale, changes in water availability are much more complicated, with some regions becoming increasingly wet, while some other regions experiencing drying.

During the same time period of global warming, the world population has risen up significantly, reached 7 billion in October 2011, and is expected to continue to grow
exponentially. In many regions of the world, water is in short supply under current
climate and population conditions. Although water is one of the most abundant resources
on the planet, less than 1 percent is fresh water available for agricultural, industrial, and
other consumptive uses. Such rapid population expansion with future climate changes
uncertain will intensely challenge the global water and food supply. The prospect of
global warming and changes in water supply imposes a broadened dimension for people
to understand the connections between the climate and water resource.

Many studies have indicated a strong connection between the climate system and
the hydrologic cycle (e.g., IPCC 2007; Thomson et al., 2005; Jones, 2011; Nijssen 2001;
Groisman 2004; Lu 2010; Christensen, 2004; Qian 2007). Understanding the interactions
between these two systems is important because variations in climate can trigger
extensive and even long term changes in the hydrologic cycle, with significant impacts on
agricultural production, natural disaster risk, environmental problems such as water
quality and biodiversity, public health, and other socio-economic issues. Therefore, due
to the changing climate, public concerns on the risk from the altering surface water
condition are growing.

The central United States is a unique area that is extraordinary sensitive to climate
variations, where the America’s No.1 field crop – corn, and one of the most important
cash crops – soybean are extensively cultivated. This domain includes the world’s largest
Corn Belt, which produces approximately 41% of the world’s corn supply and is
responsible for 48% of the total global exports (Ye, 2011). Currently, soybean production
in the central United States accounts for 33% of the world total and contributes 37% of
the global exports (American Soybean Association, 2012). In addition, high density
freshwater networks, including the U.S. major rivers such as the Missouri River, Mississippi River and Ohio River, as well as a group of world largest fresh water bodies – the Great Lakes are settled in this area.

In the context of climate change, especially like the rapid warming and obvious changes in other climatic factors since the 1980s (referred to chapter 2), regional water availability in the central United States could be strongly affected. Given that the importance of the river networks to agriculture and water resources, unprecedented changes in surface water condition in this area, particularly when associated with extreme weather events such as drought and flood, could be disruptive, or even disastrous to global agricultural economy and food security. In this case, there is a strong demand of such studies to assess the potential magnitude of the hydrological consequences of climate change in this region, and thus by which they can help human being improving the water resources management and policy-decision making processes.

Regional fresh water availability can be estimated by balancing the major components of the water cycle including precipitation, evapotranspiration, runoff, and water storage. However, due to the technical and practical limitations of the observations, the water availability could not be easily estimated based on in situ data, especially on the regional scale. Recent developments of a variety of hydrologic and land surface models provide a means to quantitatively assess the surface water balance on local, regional and global scales. In addition, with continual increasing spatial and temporal resolution, models’ performances are improved significantly and thus are able to contribute more accurate assessments.
1.2 Objective and Contents

The objective of this research is to assess the potential impacts of climate change on the surface water balance in the central United States by developing finer resolution climate scenarios, using smaller temporal and spatial analysis scale and provide results that will be useful for water resources management, field crop planning, policy-decision making and adaptation strategies. The corresponding scientific questions will be answered in this thesis:

1) What is the spatial and temporal variability of the regional water balance in the central United States (in terms of annual mean distribution, seasonal variability, and interannual variability)?

2) How have historical regional water balance changed in the past two decades? How are they associated with climate change?

3) What is the sensitivity of the water balance to imposed changes in climate? What is it implied for driving climate?

With these scientific questions, this thesis is composed of seven chapters. Chapter 1 (i.e., the current chapter) presents an introduction. Chapter 2 introduces the geographic and climate characteristics of the study area. The methodologies in this research are presented in Chapter 3. In chapter 4, a land surface model, namely Agro-IBIS, is employed to simulate surface water balance (referred to as the “control simulation”) in the study region. Scientific Question 1) will be answered in this chapter. In chapter 5, based on the high resolution meteorological dataset, trends analyses are done for precipitation, air temperature, diurnal temperature range, surface solar radiation, relative
humidity and wind speed for the period 1984-2007. Moreover, the physical causes of these changes are briefly discussed. Trends analyses are then applied to the water balance from the control simulation. This chapter is essentially a companion paper of Chapter 4, which will focus on the Question 2). In Chapter 6, a “future” climate scenario is created according to the historical climatic trends. Drawing on this climate scenario, a set of sensitivity experiments are conducted for individual atmospheric forcings, as well as some combinations of them (referred to as the “perturbed simulations”). In answer to scientific Question 3), the differences between the perturbed simulations and control simulation are then examined to see how sensitive the surface water balance is to various climate factors. Finally, conclusions and recommendations of this thesis are presented in Chapter 7.
Chapter 2

Study Area

2.1 Geographic Characteristics

The geographic region delineated for this study includes four hydrologic units: the Missouri River Basin, Upper Mississippi River Basin, Ohio River Basin and the Great Lakes Basin (Figure 2.1). The surface water resources are complex in these basins, knowing that the high dense river network and reservoir system, as well as the world largest fresh water bodies – the Great Lakes are located in this domain. The water system supports commercial navigation and a wide variety of ecosystems, including numerous wildlife refuges. Over 100 million residents in the region rely on these invaluable water resources for public and industrial supplies.

The topography in this domain is depicted in Figure 2.2 (the data for this map are described in chapter 3). Most of the area show flat topography. In the Missouri River Basin, the elevation increases as it goes west, where the Rocky Mountains are located. The substantial snow cover on the Rocky Mountains dominates the runoff in this basin. The southeast boundary of the Ohio River Basin is constrained by the Appalachian Mountains.

Land cover in this region is diverse, including agricultural lands, forest, shrubs, wetlands and prairies. Figure 2.3 depicts the spatial distribution of the natural vegetation
(the data for this map are described in chapter 3). Major natural vegetation covers include grassland, mixed forest, temperate deciduous forest and savanna. Temperate and boreal evergreen are growing in the Great Lakes Basin, northern part of the Upper Mississippi River Basin and Rocky Mountain region. Shrubs are found on the relative high elevated lands in western Missouri River Basin.

Maize and soybean are dominant crops in this region. The world largest Corn Belt is located in the central United States (Figure 2.4b). In the year 2011, the U.S. corn production reached up to 13 billion bushels, which ranked No.1 among all the crops in the country (USDA, 2011). Most of the corn fields are naturally rain-fed, while a small portion of them are irrigated fields, mostly located in eastern Nebraska. The fractional coverage of soybean is shown in Figure 2.4c. Over 3 billion bushels of soybean have been harvested in the year 2011 (USDA, 2011). The high transpiration demand for these crops strongly influences the hydroclimate in the Midwestern U.S. The spatial distribution of the dominant (>33% coverage) land cover types in this study are shown in Figure 2.5.

The soil texture features a great diversity (Figure 2.6). The dominant soil types in the study area are silt and loam. Sandy soils are shown in the Nebraska Sandhills, northern Michigan, Wisconsin and Minnesota. Distinct hydraulic conductivity and transmissivity in different soils have profound impacts on the surface hydrologic processes. For example, surface water from the precipitation easily infiltrates down through the sand texture, and recharges the underground aquifer. Thus, less soil moisture are available for the use of evapotranspiration in the sandy soils.
2.2 Climate Characteristics

The climate of this area is usually hot or warm in summer and cold in winter, which associates with considerable snowfall. The entire domain receives more precipitation in warm seasons and less precipitation in cold seasons, however, the water supply and demand varies significantly in different regions. Divided roughly along 95°-100°W, the eastern half has a humid climate while the western half has a semiarid climate. According to the Köppen climate classification, most of the eastern half is humid continental climate, while the southern part of the Ohio River Basin is humid subtropical climate. Based on the meteorological observations from 1984 to 2007, the annual and seasonal means of the climate factors including air temperature, diurnal temperature range, precipitation, relative humidity, solar radiation and wind speed are shown in Figure 2.7 and Figure 2.8-2.12 respectively.
Chapter 3

Methods

3.1 Agro-IBIS Description

The Integrated Biosphere Simulator (IBIS), first developed by Foley et al. in 1996, is a dynamic global vegetation model (DGVM) that integrates a variety of terrestrial ecosystem processes within a single, physically consistent framework. It performs assessments of biophysical, ecological, and hydrological processes on the local, regional and global scales (Foley et al., 1996; Kucharik et al., 2000). A more complete hierarchy of ecosystem phenomena that is represented in the IBIS includes land surface physics, canopy physiology, vegetation phenology, vegetation dynamics and terrestrial carbon balance.

The land surface module of IBIS is based on much of the basic structure from the land surface transfer scheme (LSX) of Pollard and Thompson (1995), which simulates the energy, water, carbon and momentum balance of the soil-vegetation-atmosphere system. The module represents two layers of natural vegetation with eight potential forest plant functional types (PFTs) in the upper canopy, and two grasses and two shrub PFTs in the lower canopy (Kucharik and Twine, 2007). A three-layer thermodynamic snow model is represented in the IBIS to simulate mass and energy balance of the snow surface. IBIS also includes a multi-layer soil model with eleven soil layers, which are parameterized with eleven soil textural categories. The thicknesses of different soil layers are varying
and the total soil depth is 2.5 meters. Richard’s equation and Darcy’s law are adapted to calculate and model vertical water flux between soil layers (Campbell and Norman 1998). IBIS does not include groundwater as a lower boundary condition, therefore free drainage is allowed in this model (Soylu et al., 2011).

In this study, an advanced agricultural version of the IBIS (referred to as Agro-IBIS; Kucharik, 2003) is used. The schematic structure of Agro-IBIS is illustrated as Figure 3.1, where crop phenology, crop management, belowground carbon/nitrogen cycling and solute transport modules are key additions to the original IBIS.

Agro-IBIS is processes based, capable of simulating both natural and managed ecosystems such as major crops (i.e., maize, soybean, and spring and winter wheat are added to this model as lower canopy PFTs) across the continental United States (Kucharik and Brye, 2003; Donner and Kucharik, 2003). Parameters and formulations of crop physiology, daily phenology, and carbon allocation (e.g., photosynthesis, stomatal conductance, and transpiration) vary according to generalized crop categories (e.g., C3 and C4), and thus govern the canopy exchange processes. The optimal planting dates for crops are determined based on algorithms of the 10-day running mean of maximum and minimum temperatures. Crop phenology stages (e.g., growth and development) are on the basis of growing degree days (GDDs). Agro-IBIS simulates the energy, water, carbon, and momentum exchanges between soils, vegetative canopies, and the atmosphere in each one-dimensional column (i.e., there is no lateral fluxes from one column to another). For a detailed description of the processes based approaches in the model, the reader is to refer to relevant publications (Donner and Kucharik, 2003; Kucharik, 2003; Kucharik and Brye, 2003; Kucharik et al., 2000).
Agro-IBIS has been well validated at multiple temporal and spatial scales. Kucharik and Twine (2007) evaluated the modeled carbon allocation, soil temperature and moisture, and energy fluxes by using the measurements at the AmeriFlux eddy covariance site in Mead, Nebraska. Twine and Kucharik (2008) validated the Agro-IBIS simulated vegetation phenology in comparison to the satellite information of greenness. In addition, on the regional scale, Agro-IBIS simulations of maize yield across the U.S. Corn Belt were found to be spatially consistent with observations (Kucharik, 2003).

For the water balance components in Agro-IBIS, total evapotranspiration (ET) is calculated as the sum of three water vapor fluxes: 1) evaporation of water intercepted by vegetation, 2) evaporation from both dry and wet soil surfaces, and 3) plant transpiration. The rates of evaporation are calculated according to the standard mass transfer equation, which is a function of temperature, vapor pressure deficit, and air conductance (Campbell and Norman 1998). Transpiration rates are calculated independently for each PFT, and influenced by a series of plant physiological parameters such as the leaf area index (LAI) and stomatal resistance.

The total runoff in the model is a sum of surface runoff and drainage. When rainfall event happens, the rain (including melt water) is apportioned between surface runoff and puddle water/puddle ice based on an empirical coefficient. Some proportion of the puddle liquid goes to evaporation, while some proportion is transferred to infiltration until the soils are saturated. Any excess puddle liquid is finally turned into surface runoff.

Agro-IBIS and its predecessors have been evaluated extensively in the use of studying the surface water balance across a variety of spatial and temporal scales. On the
global scale, IBIS simulated runoff was shown to agree reasonably well with measurements (Kucharik et al., 2000). On the regional scale, IBIS simulations of seasonal and interannual variations in precipitation, evapotranspiration, runoff, soil moisture and snow depth show generally agreement with observations in the continental United States (Lenters et al., 2000). Donner et al. (2002) evaluated the modeled river discharge in the Mississippi River Basin against observed discharge, and validated the proper performance of the model. In addition, IBIS simulated water and energy cycling at daily to interannual timescales were also evaluated with reasonable accuracy in cold climates, such as the northern Wisconsin (Vano et al., 2006).

3.2 Data

3.2.1 Geographic Dataset

The soil texture dataset for Agro-IBIS was derived from the Pennsylvania State University Earth System Science Center’s CONUS dataset (Miller and White, 1998), which is on the basis of the USDA State Soil Geographic Database. It represents 12 types of soil texture in 11 soil layers, as a function of soil depth. The soil extends to a total depth of 2.5 m, with layer thickness of 5 cm (layers 1 and 2), 10 cm (layers 3-5), 20 cm (layers 6-8), and 50 cm (layers 9-11). The 1 km × 1 km resolution data set was aggregated to 5-minute resolution so as to be implemented in the Agro-IBIS. Since the CONUS dataset covers only the continental U.S., the soil texture in the Canadian part of the domain was extrapolated with a uniform loam type.
Natural vegetation cover was derived from the 1 km DISCover land cover dataset (Loveland and Belward, 1997). The original 94 types of land covers in the DISCover dataset were aggregated into 15 potential vegetation classes by Ramankutty and Foley (1999). Similar to the soil texture dataset, the 1 km × 1 km natural vegetation dataset was converted to a 5-minutes resolution one by selecting the dominant biome in each 0.083° × 0.083° grid.

Crop dataset is obtained from Monfreda et al. (2008) and Ramankutty et al. (2008; by merging two different satellite-derived products of Boston University’s MODIS-derived land cover product and the GLC2000 dataset). The data are presented at 5-minutes spatial resolution in latitude by longitude and it specifies the fractional coverage of 175 crops based on observations in the year 2000. In this study, maize and soybean are considered as the two types of land covers other than natural vegetation, since they are the top two major crops in this study area.

The Hydrologic Unit Code (HUC) runoff data obtained from the U.S. Geological Survey (USGS) are used to evaluate the hydrological performance of Agro-IBIS. The data are basin wide averaged, with monthly time step. They are computed based on historical observations from the USGS streamgages. The dataset is available for download at http://waterwatch.usgs.gov/index.php?id=romap3&sid=w__download.

### 3.2.2 Climate Inputs

The climate inputs for Agro-IBIS include maximum air temperature, minimum air temperature, precipitation, relative humidity, surface solar radiation and wind speed. The model is forced with daily meteorological data for the period 1984-2007. These data are
purchased from the ZedX Inc., gridded with 5-minutes spatial resolution (0.083° × 0.083°, approximately 8 km × 8 km) over the contiguous U.S. and southern Canada. Except the surface solar radiation, other variables are generated based on historical meteorological station observations.

3.2.3 Bias Correction of the Solar Radiation Dataset

Surface solar radiation data is basically a satellite product that is taken from the NASA/Global Energy and Water Cycle Experiment (GEWEX) Surface Radiation Budget (SRB) shortwave radiation dataset version 3.0 (obtained from the NASA Langley Research Center Atmospheric Science Data Center). The data is generated on the basis of the International Satellite Cloud Climatology Project (ISCCP) DX radiance and cloud parameters, using an updated version of the University of Maryland’s shortwave flux algorithm (Pinker and Laszlo, 1992). It has a spatial resolution of 1° × 1° and temporal resolution of 3-hours. In order to match the resolution of the other input variables, the SRB surface solar radiation data is interpolated to a 0.083° × 0.083° gridded dataset by adding the spatial anomalies of the North American Regional Reanalysis downward shortwave radiation flux (Mesinger et al., 2006). The technical details of this method are illustrated in Figure 3.2.

3.2.4 Model Implementation

Agro-IBIS uses a 60-minutes time step to simulate the energy, water, carbon, and momentum exchange processes in the soil-vegetation-atmosphere system. The hourly meteorological data is generated from daily data using a stochastic weather generator developed by Richardson and Wright (1984), Richardson (1981), and Geng et al. (1985).
The estimations of hourly air temperature and atmospheric humidity are based on Campbell and Norman’s (1998) method. Within each day, the starting/ending time and duration of a precipitation event are randomly generated. Therefore, extreme weather events, which are critical to surface water balance simulations, are able to be captured in the model. The hourly wind speed is estimated by using the equation from the Environmental Policy Integrated Climate (EPIC) weather generator (Williams 1995).

### 3.3 Trend Analysis

Trend analysis is an effective method to detect changes in climatic and hydrological variables (Yue and Wang, 2004). In this study, the Mann-Kendall trend test is used to analyze changes in surface water balance components.

#### 3.3.1 Mann-Kendall Trend Test

The Mann-Kendall (MK) test is a non-parametric test (Mann, 1945 and Kendall, 1975). Unlike parametric or semi-parametric tests, the MK test does not rely on the assumption of normality, linearity and independence of the data. In nature, the hydrological behaviors are mostly abnormally distributed (Yue and Pilon, 2004). Van Belle and Hughes (1984) pointed out that the power of non-parametric tests is higher for abnormally distributed data in comparison to their parametric counterparts. The Mann-Kendall test has been widely used to detect trends in hydrology and climatology (e.g. water quality, streamflow, and precipitation) and the results were satisfactory (Zhang et al., 2009; Shadmani et al., 2012; Zhang et al., 2010; Yue et al., 2002; Xu et al., 2010).
In the Mann-Kendall test, the statistic variable $S$ is first determined drawing on a data series \{x_1, x_2, \ldots, x_n\}, written as:

$$S = \sum_{i=1}^{n-1} \sum_{j=i+1}^{n} sgn(x_j - x_i)$$

where the $x_j$ and $x_i$ are the sequential data values, $n$ is the length of the data series, and

$$sgn(\theta) = \begin{cases} +1 & \theta > 0 \\ 0 & \theta = 0 \\ -1 & \theta < 0 \end{cases}$$

The null hypothesis $H_0$ is an assumption that the data is independent and identically distributed (i.e., no trend). The alternative hypothesis $H_a$ is that there is a trend in the data. The statistic $S$ is approximately normally distributed when $n \geq 8$, thus a standard normal statistic variable $Z$ is developed as:

$$Z = \begin{cases} \frac{S - 1}{\sqrt{Var(S)}} & S > 0 \\ 0 & S = 0 \\ \frac{S + 1}{\sqrt{Var(S)}} & S < 0 \end{cases}$$

where the variance $Var(S)$ is given by

$$Var(S) = \frac{1}{18} \left[ n(n - 1)(2n + 5) - \sum_{p=1}^{q} t_p(t_p - 1)(2t_p + 5) \right]$$

where $q$ is the number of tied groups, $t_p$ is the number of ties of extent $p$. 
For the sample size $n > 10$, a positive $Z$ value means an increasing trend, while a negative $Z$ value means a negative trend. In a two-tailed test, for a confidence level $\alpha$, if $-Z_{\alpha/2} < Z < Z_{\alpha/2}$ the null hypothesis $H_0$ is accepted, which means that no significant trend is detected in the dataset. Otherwise, the null hypothesis $H_0$ is rejected when a significant trend exists.

### 3.3.2 Theil-Sen Approach

To estimate the magnitude of the trend, the Thiel-Sen Approach (Thiel, 1950; Sen, 1968) is adopted in this study. Unlike the least square linear regression, this method has the advantage of limiting the influence of the outliers on the slope. The Theil-Sen’s slope is given by

$$
\beta = \text{Median} \left[ \frac{x_j - x_i}{j - i} \right] \text{ for all } i < j
$$
Chapter 4

Surface Water Balance of the Central U.S.: Long-term Mean and Seasonal Variability

4.1 Introduction

The hydrologic cycle, also known as the water cycle, consists of continuous movement of water among the subsystems (e.g. atmosphere, oceans, and land surfaces) of the Earth. The state of the water in the hydrologic cycle changes between liquid, solid and gaseous forms. Considering that the water has a high thermal capacity in nature, such phase changes and movements of water play important roles in transferring and transporting heat energy in the climate system, and hence they have strong influences on temperature changes. Physical processes such as evaporation, transpiration, condensation, precipitation, infiltration, runoff, and subsurface flow constitute the dynamic nature of the hydrological cycle. Moisture and energy exchanges are involved in these processes.

Land surface water is the largest source of fresh water in the earth system. Although the land surface water represents a very small proportion of the total water volume of the earth, taking up less than 3%, it plays a vital role in supporting human life and terrestrial ecosystem. It is closely related to water collecting as groundwater or atmospheric water, and thus an important subsystem in the hydrological cycle.
Precipitation is a major source to the land surface water. It is formed as the condensation of the atmospheric water vapor and then falls onto the land surface as rain and snow, as well as other forms such as hail and sleet. The quantity of available water in a given region is largely regulated by available water in the form of precipitation. Factors that affect regional precipitation will be discussed in chapter 5, section 5.2.3.

Evapotranspiration (ET) describes the sum of evaporation and transpiration from the earth surface. Evaporation from streams, rivers, lakes and moist soil are important sources to the atmospheric water. Transpiration from plants also supply moisture to the atmosphere. However, both evaporation and transpiration are major water losses with respect to the land surface water. Evapotranspiration is constrained by the surface water supply, available energy and the ability of the atmosphere to accommodate water vapor. It could be affected by a series of meteorological factors such as temperature, vapor pressure deficit, solar radiation and wind speed. Besides, influences of the plant physiological processes cannot be neglected.

Runoff is another form of water loss to the terrestrial water resources. Runoff describes the water flow from a drainage basin or watershed. Surface runoff occurs when excess water from the rain, melting water, or other sources flows over the land surface and through channels. Together with surface runoff, the subsurface runoff and groundwater runoff constitute the total runoff. They route into the stream, and eventually flow into the ocean. Runoff is highly sensitive to changes in precipitation and evapotranspiration. Extremes of runoff could result in flood or drought, thus it has significant impacts on the agriculture and economic societies.
The storage of land surface water primarily exists as a form of soil moisture, groundwater and snow cover. Soil moisture is critical to the local climate and agriculture, since plants depend on it to carry out critical chemical and biological processes. When the gravitational forces of the soil water exceed the field capacity, the water flows down and becomes part of the groundwater. Groundwater is a long-term reservoir of the water cycle, so it acts as a natural storage that can buffer against shortages of surface water during drought period. As a result, groundwater has a vital function in the context of long term surface water balance. Snow cover is an additional form of water storage, especially in the regions where wet and drying seasons are distinct. The duration of snow cover is of great climatological importance, as it impacts the timing and magnitude of runoff. Snow also alters the surface albedo, and hence has profound implication for the surface energy balance.

The surface water balance is a conceptual structure supporting a quantitative assessment of water supply and demand at the land-atmosphere interface (Shelton, 2009). The simplified water balance over the land can be written as

\[
dW/dt = P - E - R
\]

where \(dW/dt\) is the time rate of change of total land water storage (which include soil liquid water, soil ice, snow and groundwater), \(P\) is the precipitation, \(E\) is the evapotranspiration, \(R\) is the runoff.

On the regional scale, \(R\) and \(dW/dt\) are considered to be the primary indicators of the water availability. Consequently, equation (1) could be reformatted as

\[
R + dW/dt = P - E
\]
4.2 Agro-IBIS Surface Water Balance Validation

4.2.1 Model Simulations

In this study, Agro-IBIS is run over the central United States for a 24-year period 1984-2007, with historical observations as climate inputs (referred to as the “control run”) after 5 years spin up. Agro-IBIS is totally executed 3 times. The first simulation is carried out using the natural vegetation land cover, and the other two simulations are performed with maize and soybean as land covers respectively. When running with crop land covers, Agro-IBIS assumes that the crops are growing everywhere with 100% fractional coverage. The purpose of these 3 simulations is to establish a more realistic climate scenario, that is, the “actual” land cover of each grid cell is made up of some portion of natural vegetation and the rest portion of crops. Then the outputs are weighted according to the fraction cover of natural vegetation, maize and soybean contained in each grid cell.

Considering that the Agro-IBIS does not represent the water table dynamics, irrigation is not performed in these simulations, because the irrigated water will disturb the surface water balance without groundwater interactions. Outputs of the simulations include total evapotranspiration (ET), total runoff, volumetric water content (VWC), rainfall and snowfall (in terms of snow water equivalent). They are monthly values that are averaged from daily outputs. The time rate change of water storage ($dW/dt$) is calculated based on the surface water balance equation (equation (1) in section 4.1).

4.2.2 Validation of Simulated Runoff

For evaluating the performance of Agro-IBIS in simulating surface water balance, the simulated runoff is compared against the USGS observations. Basin-wide averaged
runoff are used in this set of evaluations. Three model evaluation criteria are examined in this section: 1) Are the seasonal timing and magnitude of water flows captured by the model? 2) Does Agro-IBIS reasonably simulate the annual average water balance and interannual variability? 3) Are the spatial variations in surface water balance captured by the Agro-IBIS?

First, the seasonal cycle of 1984-2007 mean monthly runoff values for each basin are plotted against the USGS observations (Figure 4.1). Comparisons show that the seasonal cycles are well captured in all these basins. For the Missouri River Basin and Upper Mississippi River Basin, the Agro-IBIS simulated winter time runoff agree well with the observations, but the warm season (spring and summer) runoff values are higher than the observed ones, with highest biases of nearly 0.5 mm day\(^{-1}\) in the early summer. One of the reasons for this warm season runoff biases is that the irrigation is not implemented in the model simulations (as discussed in section 4.2.1). Although most of the corn and soybean fields in the study area are naturally rain-fed, a small proportion of irrigated lands do exist in the Missouri River Basin and Upper Mississippi River Basin. During the growing seasons, a large amount of groundwater is pumped up to the ground surface for crops use. The irrigated water is finally evaporated or transpired by crops, consequently contributing to the total ET. Because Agro-IBIS calculates runoff as the precipitation subtracted by ET and water storage, the underestimation of actual ET can result in the overestimation of runoff.

Another reason for such warm season runoff biases is that not all the crops are considered in the model simulations. Although maize and soybean are major crops in these basins, over 100 other types of crops are cultivated here and they could account for
as much as 80% of the total land vegetation coverage in some specific area (Figure 2.4d; Ramankutty et al. 2008). For example, there are some spring wheat growing in the northern part of the Missouri River Basin and some winter wheat growing around the southern edge of the basin. The water demand for wheat during the growing season is much higher than the potential vegetation (i.e., the actual ET rates in these areas are higher). As a result, the ET is underestimated by the Agro-IBIS and the high biases of runoff appear in the simulations.

In addition, a certain portion of the runoff biases could be attributed to the errors when calculating basin-wide averaged values. Agro-IBIS does not simulate ET from the inland water bodies (i.e., all the large open water area are masked out before simulations), thus the simulated spatial averaged “actual” total ET for the basin is not the actual value. Considering that the evaporation rate for open water in summer is higher than that over land, the actual basin-wide averaged ET is underestimated by the model. Consequently, on account of the Agro-IBIS surface water balance algorithm, the modeled runoff is underestimated.

Figure 4.2 depicts the comparisons of the interannual variability of modeled runoff and observed runoff for each basin. These time series suggest that for the most part, Agro-IBIS captures spatial variations in the annual mean runoff, as well as interannual variations. For the Missouri River Basin and Upper Mississippi River Basin, although high biases of the annual mean runoff are found, the interannual variability show well agreement with observations. The high biases are primarily caused by the overestimation of summer time runoff, as discussed above.
4.3 Regional Surface Water Balance

Basin-wide averaged monthly climatological mean and interannual variability (1984–2007) for water balance components are depicted in Figure 4.3. Distinct spatial variations in surface water balance are found among these four basins. Seasonal cycle of surface water balance for the Missouri River Basin is similar to that for the Upper Mississippi River Basin but with smaller amplitudes. Both basins have maximum precipitation in summer and minimum precipitation in winter. Summertime water storage exhibits an evident deficit due to large amounts of ET during that season. The timing of peak runoff appears in the late spring or early summer, associated with rapid increasing precipitation and a great portion of snowmelt water. Thereafter, the magnitude of runoff starts to decrease due to the high rate of ET in summer. For the Upper Mississippi River Basin, the summer precipitation shows a bimodal distribution, with an early summer maxima in June, a mid-summer “dry” in July and a secondary maxima in August. The bimodal pattern of precipitation is a unique climatic characteristic in the U.S. Midwest region, and the physical causes of the phenomena are referred to chapter 5, section 5.2.3.

The seasonality of precipitation in the Ohio River Basin and Great Lakes Basin is not that obvious compared to that in the other two basins. In winter, substantial amount of precipitation are also received in these two basins. Wintertime runoff generally follows the precipitation. The peak runoff occurs in spring, and after that the runoff gradually decreases due to water consumptions of warm season ET. For the snow dominated Great Lakes Basin, the step change of runoff in spring is attributed to a large amount of melting snow.
The monthly based interannual variabilities for each water balance component are represented by the corresponding errorbars. In general, precipitation has greatest interannual variability, especially in summer. Interannual variations of ET are relatively small or even negligible, which indicates that water supply is not the only factor that affects ET, but other factors such as the available energy also play an important role in controlling the latent heat fluxes. Large anomalies in precipitation seem to have strong influences on the water storage (e.g., soil moisture), because relative significant interannual variations in \( \frac{dW}{dt} \) are found along with variations in precipitation.

Figure 4.4 shows the basin-wide averaged monthly climatology (November - April) of snow depth and associated interannual variability. Snow depth is the product of snow heights and fractional cover, which has profound implications for the seasonal timing of runoff, especially in snow-dominated regions. For the seasonal variation, greatest amount of snow appears in February for all of these basins. The Great Lakes Basin receives a greater amount of snow during cold seasons among all the basins, because there is abundant moisture evaporated from the lakes. Largest interannual variability is also shown in the Great Lakes Basin, most of which are contributed by the lake effect snow events. Shallowest snow depth is found in the Ohio River Basin, on account of the relatively warmer climate in winter.

Spatial patterns of annual mean and seasonal mean surface water balance are shown in Figure 4.5-4.9. For each mean surface water balance result, precipitation, -ET, runoff, and \( \frac{dW}{dt} \) are plotted as a form of geospatial maps, where the blue color means “more water” and red means “less water” (i.e., blue precipitation means more precipitation, blue runoff means more runoff, blue \( \frac{dW}{dt} \) means more water being
stored, and blue -ET means less water being taken away from the land surface). In total, according to equation (2), one would simply add $P$ and $-ET$ to match $R + \frac{dW}{dt}$ (i.e., the total water availability).

The distribution of annual precipitation shows a general southwest-northwest gradient pattern (Figure 4.5a), with the highest value in the southern Ohio River Basin and the lowest value in the western Missouri River Basin. This gradient pattern is associated with atmospheric moisture fluxes (c.f. chapter 5, section 5.2.3). The maritime tropical air masses originate from the southern Atlantic Ocean, Caribbean Sea and the Gulf of Mexico and transport warm moist water vapor northwestward into the central United States. Similar gradient patterns are also found for annual ET and runoff. In Figure 4.5d, the annual water storage is almost zero over the entire domain. This suggests a good performance of Agro-IBIS in balancing the surface water budget.

In winter (DJF), both of the rates of ET and runoff are very low, while most of the precipitation are stored as a form of soil water, soil ice and snow cover (Figure 4.6). As it gets warmer in spring (MAM), ET and runoff increases, while the water storage ($\frac{dW}{dt}$) is reduced significantly, especially for the Ohio River Basin and Great Lakes Basin (Figure 4.7). Large quantities of melting snow lead to increasing amounts of runoff, and thus result in losses in water storage. Higher rates in ET in the southern edge of the Ohio River Basin also account for the decreases in water storage. In summer, ET exceeds precipitation in most parts of the area, as a result of excessive amounts of available energy (Figure 4.8). Consequently, runoff decreases evidently as compared to that in spring, and extensive soil water deficit appears over the entire domain. As the precipitation and the available energy at the land surface are reducing in fall (SON), the
rate of ET decreases dramatically, and most of the precipitation are to replenish the soil, or stored as a form of snow (Figure 4.9). Runoff in this season shows a slightly reduction compared to that in summer.

Wintertime rainfall directly controls the runoff of the time, while snowfall tends to have lag effects, primarily constrain the spring runoff. The fractions of precipitation falling as rain against snow (in terms of snow water equivalent) for wintertime (DJF) and snow season (November - April) are shown in Figure 4.10. Snowfall dominates the wintertime precipitation in most parts of the domain except the Ohio River Basin and small parts of the southern Upper Mississippi and Missouri River Basin.

The spatial distributions of snow depth are shown in Figure 4.11. The seasonal mean snow depth is calculated based on accumulative monthly snow depth value. Reasonable north-south gradient pattern are shown in the domain. In fall (SON), high snow depth in the western Missouri River Basin is attributed to the topography of the Rocky Mountains (i.e., high elevated mountain snow falls earlier than that on the plains). Seasonal runoff in the Missouri River Basin largely depends on the snow accumulation in this region.
Chapter 5

Hydroclimatic Trends in the Central U.S. from 1984-2007

5.1 Introduction

Over the twentieth century, evident changes in both climate and water cycle have occurred in the United States (IPCC, 2007). Since terrestrial water balances are strongly controlled by climate factors such as temperature and precipitation, the response of water resources to climate change can be significant but not constant through time. Numerous studies have indicated the influence of climate change on the hydrological cycle. For example, Portmann et al. (2009) found a clear connection between regional changes in daily maximum air temperatures and the climatological mean precipitation through much of the year in the southern United States. Huntington et al. (2009) suggested that increased temperature in the last few decades have brought about earlier spring streamflow peaks, decreased summertime evapotranspiration, reduced summer streamflow, and declined winter ice cover on lakes and streams in the eastern United States. Based on observed temperature and precipitation, Dai et al. (2004) used the Palmer Drought Severity Index to estimate soil moisture and found that since the middle 1950s, soil had been losing water over much of Eurasia, northern Africa, Canada and Alaska.
The partitioning of precipitation into runoff and evapotranspiration (ET) are key elements of the surface water balance study. Unlike runoff, which can be observed through streamgages measurements and usually have long historical records over 100 years, there are very limited direct observation of ET over land. Near-real time observations of water exchange have been established in recent years in selected locations, such as the AmeriFlux, the Atmospheric Radiation Measurement/Cloud and Radiation Testbeds (ARM/CART), the Oklahoma Mesonet, and some short-term field experiments from eddy covariance flux towers in Asia and Europe (Baldocchi et al., 2001; Qian et al., 2007). Satellite measurements have increased the availability of monitoring ET at large scales (Kustas and Norman, 1996), but the relative short records period is not ideal to detect either long term trend or decadal variability.

Controversy continues over ET trends (e.g., Ohmura and Wild 2002; Peterson et al., 1995; Szilagyi et al., 2001; Wild et al., 2004). The controversy mainly arises from indirect estimation of ET owing to incomplete observations of the water and energy budget (Qian et al., 2007). Nevertheless, many studies support an upward trend in ET in the United States over the last century. For instance, Brutsaert and Parlange (1998) presented a complementary relationship that the actual ET is negatively correlated with pan evaporation, and thus explained that in response to enhanced landscape’s ET, the rising surrounding humidity could account for observed weakening in pan evaporation. Further studies (Lawrimore and Peterson, 2000; Golubev et al., 2001) also observed this inverse relationship and suggested increasing trends in ET over the United States. Milly and Dunne (2001) found an upward trend in ET (1949-1997) by synthetically analyzing precipitation, net radiation data and human disturbance over the Mississippi River Basin.
On the basis of the water balance approach, Walter et al. (2004) analyzed published precipitation and streamflow data across the United States and concluded that ET has increased over the past 50 years.

Aside from these relatively simple and indirect approaches, a large number of land surface models (LSMs) have been developed for surface energy and water balance research since the late 1980s. Land surface models consider complicated physical processes in terms of sophisticated parameterizations of vegetation and soil, and thus are able to provide a more comprehensive view of the hydrological impacts of climate change. Alkama et al. (2010) used Organizing Carbon and Hydrology in Dynamic Ecosystems (ORCHIDEE) land surface model to simulate global hydrological processes and found that climate change largely drives the twentieth century runoff increase. Qian et al. (2007) studied hydroclimatic trends in the Mississippi River Basin from 1948 to 2004 by using the Community Land Model version 3 (CLM3) and found that the upward ET trend is primarily due to increases in precipitation in the basin, while changes in temperature and solar radiation have only small effects.

Most of the previous studies have focused on climatic and hydrological trends on the century or half-century long scale. However, climate change is not usually linear over a long time period, and the decadal-scale variation might differ from the long term signal. Evidence shows that the global air temperature has increased more rapidly since the 1980s and the rapid warming is predicted to continue through the 21st century (IPCC, 2007). Huntington et al. (2009) found that the most pronounced changes in global snow and hydrologic regimes associated with global warming have been observed since 1970s. McCabe and Wolock (2002) analyzed 59 years (1941-1999) of streamflow for 400 sites
in the conterminous United States and their results indicated a step increase of annual streamflow around 1970 rather than a gradual trend; furthermore, they suggested that the hydrological cycle has shifted to a new regime and the new gradual trend will behave relatively constant until another new step change occurs. Since the 1980s, changes in climatic variables such as temperature, solar radiation, precipitation, wind speed are found to be significant over the central United States and in contrast with century long trends (c.f. section 5.2). Therefore, it is worthy of analyzing decadal-scale variations in surface water balance, considering that the century long trends might mask the full magnitude of climatic and hydrological changes since the 1980s.

Water management and regulation decisions are usually made on the state size scale or even smaller local scale, while the diverse local consumptions of water in turn react on the water resources in large water basins. Many previous studies have only paid attention to national-wide trends or large basin-wide trends of the water balance (Walter et al., 2004; Qian et al., 2007; Milly and Dunne, 2001; Mackay et al., 2003), probably due to the limitations of their study method, data or coarse resolution of the modeling. Nevertheless, since hydrological components usually appear significant spatial heterogeneous within a large basin, these results are too broad so that they actually provide very limited sense for the policy makers. In this case, there is a strong demand of such information of changes in water resources on both large basin scale and local scale (e.g., subbasin, small watershed and county scale).

Moreover, consumptive uses of water for agriculture, industry and residence are subject to vary by seasons. Berbery et al. (2003) and Huntington et al. (2009) pointed out that the water cycle has diverse land surface-atmosphere interactions at seasonal and
monthly timescales. Hence the hydrological trends at annual time step cannot meet all the needs for decision making processes. Therefore, it is necessary to study changes in surface water balance not only at annual but also monthly or seasonal time steps, which requires high accuracy of climatic and hydrological data and better performance of LSMs.

Although contemporary LSMs are capable of simulating hydrological changes on both global and regional scales, their representations of land surface conditions are not realistic and fine enough. In order to simulate surface energy and water balance more accurately, a land surface model must also meet three important criteria.

First, major crop types over land must be explicitly represented in the model. The phenology (e.g., planting date, growth of leaves, stems, roots and grain, senescence) and physiology (e.g., photosynthesis, leaf respiration, stomatal conductance) of each type of seasonal crops are unique and cannot be alternatively surrogated by other natural vegetation. Twine et al. (2004) studied the effects of land cover changes on surface energy and water balance of the Mississippi River Basin by using the Integrated Biosphere Simulator (IBIS) and found that when forest are converted to croplands, the runoff has increased while the ET has decreased significantly because of the reduced leaf area index (LAI), particularly during summer; the net radiation that is driven mainly by land surface albedo has decreased with forest conversion to crop fields; opposite changes have occurred when crop covers are converted from grasslands. Based on simulations with the variable infiltration capacity (VIC) model, Mishra et al. (2010) found that the reduction of net radiation and sensible heat flux from forest-to-cropland conversion are more prominent in winter and spring due to changes in snow albedo, while the summertime latent heat flux has increased resulting from the increased snow water
equivalent in the preceding seasons. Therefore, the absence of major crop fields in a land surface model may lead to considerable biases in simulating energy and water fluxes.

Second, realistic near-surface forcing data are necessary to drive the land surface model. Numerous studies have used General Circulation Models (GCMs) or coupled LSMs to modeling the hydroclimate on the global or regional scales (e.g., Marks et al., 1993; Nijssen et al., 2001; Bukovsky and Karoly, 2010; Music and Caya, 2007; Mishra et al., 2010; Winter and Eltalir, 2012a,b). Although coupled modelings provide a means to predict future changes, the simulated atmospheric forcing (i.e., temperature and precipitation) as input for the LSMs usually do not agree well with ground measurements, or do not agree with each other (Lettenmaier et al., 1999; Kirshen and Fennessey, 1995; Wolock and McCabe, 1999). These inaccurate forcing data can potentially bring about large uncertainties in land surface modeling. On the contrary, when using more reliable in situ data to drive the LSMs (i.e., offline studies), there are advantages for better performance regarding the land surface schemes.

Third, the resolution of a model should be as fine as possible. For the regional scale study, the spatial distribution of soil texture, land cover and topography have prominent influence on the dynamical and physical processes of the models. Inaccurate subgrid parameterizations of coarse-resolution climate models have little regional scale predictive ability (Duffy et al., 2003). Elguindi et al. (2011) and Rummukainen (2010) pointed out that model performances improve significantly with higher resolution. Therefore, finer resolution simulations are needed in order to better assess the magnitude and spatial variation of the regional hydrological effects of climate change.
In this study, a sophisticated land surface model, namely Agro-IBIS, is employed to simulate impacts of climate change on the surface water balance of the central United States for the period 1984-2007. Agro-IBIS is the first dynamic vegetation model that incorporates major crop types of the United States. Detailed descriptions of the model are referred to chapter 3, section 3.1. High resolution meteorological datasets (5-minutes spatial resolution, about 8 km × 8 km) are used to drive the land surface processes over three subbasins of the Mississippi River Basin and the Great Lakes Basin. Monthly to seasonal analysis scale are used so that it is able to provide useful information for crop planning and water resources management.

5.2 Climate Change in the Central U.S.

5.2.1 Temperature

Over the twentieth century, the average annual air temperature has risen nearly 0.6 °C in the United States (Karl and Knight, 1998). There is also mounting evidence indicating increased annual air temperatures in the Midwestern United States, accompanied by shorter winters (Sinha and Cherkauer, 2010; O’Neal et al., 2005). This regional temperature change might “catch up” to the global warming, attributable to observed increases in anthropogenic greenhouse gases (GHGs) concentrations (IPCC, 2007). The increases in GHGs cause a rise in the global average air temperature by increasing the amount of long wave radiation that is trapped in the atmosphere.
Analyses based on historical observations from 1984 to 2007 (detailed description of data is in chapter 3) show that the annual temperature has increased at a rapid rate of 0.3 - 1°C decade⁻¹ in the central United States (Figure 5.1a; Since the outline of the study region is stippled on the map, discussions of the trends focus on the patterns only within this region). While breaking the annual trend into seasonal time steps (Figure 5.2), it is clear to see that the fall and winter temperature trends have primarily contributed to the annual signal, which also indicates a shift toward warmer winters. The warming patterns are especially obvious in the Great Lakes basin, the northern Upper Mississippi River basin, and the southwestern Missouri River basin. A spring cooling pattern, denoted as a “warming hole” is found in the northern part of Missouri River basin, which is a clear math-up with the surface solar radiation (Figure 5.2b). In addition to the response to the GHGs, Kunkel et al. (2006) found that the seasonal and decadal variations of temperature in this region are influenced by external forcings, and to more extent, associated with internal dynamic variability, such as variations of sea surface temperatures in the Central Equatorial Pacific and North Atlantic. On the regional scale modeling, Pan et al. (2004) found that local land surface feedback may be a potential cause for the lack of warming in warm seasons.

5.2.2 Diurnal Temperature Range

The diurnal temperature range (DTR) is the difference between daily maximum and minimum air temperatures. There is evidence of narrowing in DTR from 1950 to the 1990s over the United States because maximum temperatures have remained constant or increased only slightly, whereas there have been greater increases in minimum temperatures (Easterling et al, 1997). Factors that affect DTR include large-scale climatic
effects such as changes in cloud cover, precipitation, soil moisture, solar heating, greenhouse gases and aerosols (Kanamitsu et al., 2002; Trenberth and Shea, 2005; Dai et al., 1997; Dai et al., 2006; Peterson et al., 1995; Easterling et al., 1997); Local effects can also affect the DTR, such as urban growth, irrigation, desertification and variations in local land use (Zhou et al., 2004; Dai et al., 1999). In the United States, Peterson et al. (1995) found a negative correlation between DTR and cloud cover, while Gallo et al. (1996) suggested that changes in land use/land cover are main drivers in DTR trends.

In the central United States, both maximum and minimum air temperatures have increased in the past two decades (Appendix A.1 and A.2). However, the minimum air temperature (Tmin) has increased at a faster rate than the maximum air temperature (Tmax), which results in a decreasing trend in DTR (Figure 5.1b). In winter, spring and summer, there was extensive narrowing in DTR, while in fall, such pattern disappeared, or even turned into an increasing trend instead in some regions. These increasing trends in DTR during fall season might be associated with those trends in surface solar radiation (Figure 5.3d). Considering that solar brightening is usually associated with decreases in cloud cover and aerosol, Tmax increases rapidly during daytime due to stronger solar heating at the surface; whereas in the night, less long wave radiation is trapped in the atmosphere, therefore Tmin does not increase as fast as Tmax. Besides cloudiness, various factors can result in variations in DTR and the primary cause for nighttime warming has been the subject of much debate (Walters et al., 2007).
5.2.3 Precipitation

Over the past century, precipitation is observed to have increased in most parts of the North America (IPCC, 2007). Frontal precipitation, associated with cyclonic activities, are dominant weather regime in the central plains and Upper Midwest of the United States (Rudd, 1961; Trewartha, 1981), and are strongly influenced by shifts in westerlies and upper-level jet streams (Trenberth and Guillemot 1996; Kunkel and Liang, 2001). Previous studies have indicated that changes in mid latitude upper-level jet streams are forced by anomalies in sea surface temperature and atmospheric pressure (e.g., El Nino and Southern Oscillation, North Atlantic/Artic Oscillation), as well as anomalies in continental surface conditions, such as snow and soil moisture over Eurasia and North America (Bell and Janowiak 1995; Trenberth and Guillemot 1996; Liang et al., 1997). Interdecadal variations in precipitation are documented to be connected to large-scale oscillation patterns, such as the Pacific Decadal Oscillation and Atlantic Muti-decadal Oscillation (Birk et al., 2010; Hu et al., 1998). On the regional scale, the Great Plains low-level jet (LLJ) is a major contributor to the precipitation by transporting moisture from the Gulf of Mexico into the central United States (Higgins et al., 1997). During warmer seasons, water recycling through land surface processes, such as deep convection, contributes a lot to the formation of the rain, and it is the main cause for the nocturnal maximum precipitation. (Lee et al., 2008; Bukovsky and Karoly, 2011; Liang et al., 2006; Gutowski Jr. et al., 2003). In the late summer, northward penetration of tropical cyclones with advection of highly unstable air masses inland could also bring precipitation in this region (Bryson and Hare, 1974).
Trend analyses based on observations in the past two decades indicate that in general, annual precipitation did not show significant changes in the central United States, except some discontiguous small areas where significant upward trends are found, such as Ohio, Indiana, North Dakota and Oklahoma. (Figure 5.1c). Nosier patterns show up on the seasonal maps (Figure 5.4a-d). Precipitation has generally increased during cold seasons (DJF, MAM) but decreased in warm season (SON). On smaller regional scales, drier conditions are found to have strong connections to the solar brightening (Figure 5.4d and Figure 5.6d), which might be associated with decreases in cloud cover. Observations also indicate the increases in both intensity and frequency of heavy precipitation events in this region, which mainly cause the wetter conditions (Karl and Knight, 1998; Groisman et al., 2005). On the diurnal time scale, Dai (1999) pointed out that the nighttime precipitation has increased in terms of both frequency and amount in summer and fall.

5.2.4 Relative Humidity

Relative humidity (RH) is a meteorological term used to describe the amount of water vapor in a mixture of air and water vapor. It is defined as the ratio of the amount of water vapor in the air at a specific temperature to the maximum amount that the air could hold at that temperature; it can also be expressed by the ratio of partial vapor pressure to the saturated vapor pressure. The vapor pressure deficit (VPD), denoted as the difference between the saturated vapor pressure and actual vapor pressure, is usually considered as a measure of the drying power of the air. In the context of global warming, changes in relative humidity have important implications for evaporative demand, because the atmospheric water-holding capacity increases at a rate of about 7% per °C (IPCC, 2007).
As atmospheric humidity is highly dependent on temperature, the atmosphere is expected to become more humid as the air temperature increases, therefore altering the characteristics of precipitation in terms of amount, frequency, intensity, duration and type and consequently affects the hydrologic cycle (Trenberth et al., 2003).

Large RH increases (0.5%–2.0% decade^{-1}) have occurred over the central United States over the past 30 years (Dai, 2006). For the period of 1984 to 2007, the annual trend in RH is found to have extensively increased over the central United States (Figure 5.1d), and the increasing patterns are more significant than the precipitation trends (Figure 5.1c). Figure 5.5 show that the Great Lakes basin was getting wetter all through the seasons. Trends in RH in the Missouri River, Upper Mississippi River and Ohio River basin are found to have increased during spring and summer, but kept almost constant in fall and winter. Atmospheric humidity is not only controlled by air temperature; besides, surface water availability is an important factor that governs the water that evaporates into the atmosphere.

5.2.5 Solar Radiation

Surface solar radiation (SSR) refers to the solar incident at the Earth’s surface. It is a major component of the surface energy balance, which not only governs the temperature, but also provides energy to the biosphere that is closely linked with agricultural productivity and human life. Surface solar radiation takes an important role in the water cycle, as it drives a number of hydrological processes, such as evapotranspiration and snowmelt (Roderick and Farquhar, 2002). Such energy input to the Earth’s surface is not constant over the years but undergoes substantial decadal
variations on both regional and global scales (Pinker et al., 2005; Wild, 2009). The widespread decreases and increases in SSR are usually referred as the “dimming” and “brightening” respectively. Observation shows a decline in SSR from 1960s to 1990s over the United States, at a rate of -6 W m$^{-2}$ decade$^{-1}$ (Liepert, 2002). In the same region, Long et al. (2009) found a recovery phenomenon from 1995 to 2007, with an 8 W m$^{-2}$ decade$^{-1}$ rise in SSR. Although the physical causes of the decadal variations in SSR are controversial, there is a consensus that these phenomena are not triggered by external forcings (e.g., variations of solar constant), but as a result of internal variations, such as the changing cloud and aerosol characteristics (Roderick and Farquhar, 2002; Wild, 2009). Liepert (2002) found a shift from cloud-free to overcast sky conditions over the United States from 1961 to 1990, with increased optical thickness of both cloudy and clear skies, and explained that the “dimming” is dominantly resulted from changes in overcast frequency and associated optical thickness, while the reduction of the clear sky optical properties has secondary effect.

Satellite observations (data descriptions are referred to chapter 3) indicate that the central United States was experiencing a “brightening” period from 1984 to 2007 (Figure 5.1e). Seasonal patterns suggest that the upward annual trend is primarily as a result of significant solar brightening in summer and fall seasons. The trends are found to be strongest in the Great Lakes basin, while lack of significance in the Missouri River basin. In spring, the trend is decreasing in the northern part of Missouri River and Upper Mississippi basin, which is linked to the pattern on the temperature map (Figure 5.6b and 5.2b). The major factors that affect SSR trend in this region are not clear by reason of limited observations of aerosol loading and optical properties.
5.2.6 Wind Speed

Near-surface wind plays a vital role in the surface water and energy balance, as it transports heat and moisture between the land surface and the atmosphere. Wind speed is a major factor that affects the rate of evaporation, and thus has profound implications for the hydrological cycle (Roderick et al., 2007). Recent observations of near-surface wind speed have shown prevalent declining trends (“stilling”) over the last 30 to 50 years in the mid-latitude regions of North America (Hobbins, 2004; Klink, 1999; Pryor et al., 2009). In the past 24 years (1984-2007), overwhelming declining trends in near-surface wind speed are found in the Upper Mississippi River, Ohio River and Great Lakes basins (Figure 5.1f). Because the prevalent trends exist in each season (Figure 5.7), the annual trend exhibit similar pattern to the seasonal ones.

Atmospheric pressure gradient is one of the factors that govern the wind speed. Because pressure is closely related to temperature, it is possible that changes in surface temperature may produce systematic changes in surface winds (Klink, 1999). Global warming has particularly enhanced temperature increases at higher latitudes compared to the tropics (Easterling et al. 1997). The difference in latitudinal warming could reduce the pressure gradient over the United States and consequently reduce the wind speeds. Klink (1999) also pointed out that increases in minimum air temperatures may weaken the nighttime inversion, reduce the near surface wind speed gradient and the turbulence transports, and thus result in the decline in nighttime wind speed.

Friction force is another factor that affects surface wind speed. For this reason, changes in land use/land cover will alter the surface roughness, and consequently affect
near surface wind speed. Studies noted that some of the wind speed declines are owing to local effects such as urbanization (e.g., building constructions or other obstacles obstructing the air flow) and growing trees (Roderick et al., 2007; McVicar et al., 2008).

Other possible causes of the “stilling” phenomena include changes in cyclone and anticyclone frequency and instrumentation/observation biases (Klink, 1999; Kysel´y and Huth 2006). The declining trends in surface wind speed were also found to have been associated with increases in precipitation, because less available energy was partitioned into the sensible heat flux and consequently the mechanically produced turbulence was reduced (Ozdogan et al., 2006). Some climate models have captured that these phenomena are associated with poleward expansions of the Hadley cell (Lu et al., 2007).

5.3 Validation of Agro-IBIS Simulated Hydrological Trends

In addition to the comparisons of Agro-IBIS simulated runoff and the USGS computed runoff (referred to chapter 4, section 4.2.2), further steps are taken to evaluate how well the simulated seasonal and annual trends in runoff agree with the trends from observations. Modeled trends (MT) and observed trends (OT) in runoff, associated with their p-values, are listed in Table 5.1 and Table 5.2 respectively. The best agreement between MT and OT are found in the Great Lakes Basin, through all seasons. However, for the Missouri River Basin and Upper Mississippi River Basin, relationships between MT and OT are not consistent all the time. To make the evaluation more straightforward, a scatter plot is presented (Figure 5.8), which include MT and OT for all basins and all seasons. The correlation shows a high bias in MT, although the regression is not that
strong. This indicates that Agro-IBIS is generally able to capture the trend pattern, but overestimates the magnitude of the runoff trends for about 0.05 mm day\(^{-1}\) decade\(^{-1}\) on average. Many factors can affect the basin-wide averaged value, such as the soil texture, land cover, topography and water distribution. In this case, in order to more accurately quantify the hydrological changes, trends in surface water balance are calculated for each grid cell respectively.

5.4 Trends in Surface Water Balance

For quantitatively assessing changes in surface water balance, the Mann-Kendall trend test and Thiel-Sen Approach are applied to the Agro-IBIS outputs. Grid cell based trends are calculated at both seasonal and annual time scale. Results show that in the past 24 years, evapotranspiration (ET) has increased extensively over the study area (Figure 5.9b). Unlike ET, trends in runoff are not uniformly distributed across the basins (Figure 5.9c). Significant increasing trends are found in the Ohio River Basin and northern Missouri River Basin; however, in the northern Great Lakes Basin and southern Missouri River Basin, decreasing trends dominate. The patterns are mostly in accordance with precipitation trends. Hence, it seems that changes in runoff are largely controlled by the changes in precipitation. Water storage has remained almost constant (Figure 5.9d). All these trends together indicate an intensification of the water cycle in the central United States.

Seasonal trends in surface water balance show contrasting patterns compared to the annual trends. In winter, the water cycle has a tendency to weaken. Although it has been
getting wetter in some areas (Figure 5.10a), broad decreases in ET are shown in most parts of the Missouri River Basin, Upper Mississippi River Basin and Great Lakes Basin (Figure 5.10b). Runoff is also found to have decreased except in the northern Missouri and Ohio River Basins (Figure 5.10c). Significant increases in $dW/dt$ (Figure 5.10d) suggest that during the past two decades, winter precipitation is more likely to be reserved at the land surface rather than consumed as forms of ET and runoff.

Hydrological trends become inconspicuous as spring neared (Figure 5.11a-d). ET remained almost unchanged in all the four basins. The northern Great Lakes Basin has a wetter trend in precipitation. Nevertheless, more precipitated water did not lead to more runoff at the same time. On the contrary, the amount of runoff has reduced in the context of a wetter climate (Figure 5.11c). In this case, the declines in runoff seem to be strongly associated with $dW/dt$ trends (Figure 5.11d), since most of the springtime runoff is attributed to the abundant melt water in this season. Rather than concluding that there is a significant increase in the rate of water storage, in fact, the upward trends in $dW/dt$ (Figure 5.11d) indicate a significant decrease in the rate of water loss (i.e., snow melt), which result in declines in runoff.

In summer, an intensifying water cycle is found in the study area with uniformly upward trends in precipitation, ET and runoff, except in the northern Great Lakes Basin (Figure 5.12a-d). In contrast, all the water balance components are found to have downward trends in the northern Great Lakes Basin, which indicates a weakening water cycle in that region.
In the fall season, evident decreasing patterns dominate the precipitation trends, particularly in the Upper Mississippi River Basin, southern Great Lakes Basin and southwestern Missouri River Basin (Figure 5.13a). The drier changes are associated with obvious solar brightening in this season, as discussed in section 5.2.5. Together with extensive significant increasing trends in ET (Figure 5.13b), these changes result in a more severe water deficit (Figure 5.13d). The runoff trends in this season exhibit a similarity to the annual pattern (Figure 5.13c and 5.9c).

For the cold season snow depth, a significant decline pattern is shown in the northern Great Lakes Basin (Figure 5.14a-d). This pattern is consistent with the trends in spring runoff (Figure 5.11c), since the runoff in this area is snow-dominated. Widespread decline in snow depth is also found in fall (Figure 5.14c), which indicates a lag tendency of snow onset season as a result of regional warming.

Figure 5.15a-d show that wintertime rainfall has increased extensively for all these basins, while the snow fall has largely remained unchanged or even declined in some areas. These changes indicate that a greater fraction of the winter precipitation has fallen as rain rather than snow. It is also an indicator of a warming climate besides observations of increased temperature.

5.5 Discussion and Conclusions

In this chapter, changes in surface water balance (1984-2007) over central United States are studied by using the Agro-IBIS. Model evaluations suggest that Agro-IBIS is capable of simulating seasonal and annual trends in runoff on the regional scale. Results
of trend analysis demonstrate extensive enhanced ET and thus a general intensification of
the water cycle over the Upper Mississippi, Missouri, Ohio River basins and the Great
Lakes Basin. Furthermore, changes in water balance components (ET, runoff, \(dW/dt\))
are highly variable by basins and seasons.

Hydrological changes for the period 1984-2007 in some of the study basins are
found to differ from the century long trends or larger regional scale trends. On the four-
basins regional scale, there are scarcely any significant increases in the precipitation over
the past 24 years. However, previous studies had found upward trends in precipitation of
7\% per century (1908-2002) for the United States (Groisman et al., 2004), and significant
increasing trends of 176 mm century\(^{-1}\) (1950-2000) for the Mississippi River Basin
(Walter et al., 2004). Karl and Knight (1998) reported that for the period 1910-1996, the
Midwest region had become increasingly wet in autumn but dryer in winter, and that the
Great Lakes Basin had received more precipitation in summer, all of which are shown
opposite patterns in comparison to the results in this study.

Runoff trend patterns largely match the changes in precipitation in the past two
decades, which appear no significant changes or even slightly decreases over the domain
of the central United States. These patterns are also found somewhat inconsistent with the
long term trends. Groisman et al. (2004) pointed out that the United States runoff had
increased for the period 1939-2002, at a rate of 26\% per century. Besides, an increasing
runoff rate of 65 mm century\(^{-1}\) (1950-2000) over the Mississippi River Basin is
documented (Walter et al., 2004). For the Upper Mississippi River Basin, Zhang and
Schilling (2006) found an annual runoff increase of 28-320 mm century\(^{-1}\) from 1940 to
2003.
The relatively short sample length (24 years) is in part not well capable of revealing significant trends, and thus potentially results in statistically insignificant basin-wide trends in water balance. However, results do detect significant trends for different seasonal time steps, and smaller regional scale or even local scale. In addition, these trends evidently show contrasting patterns from those century long trends. All of these suggest that the water resources management and decision making processes should be highly dependent on specific objectives, such as the planning time period, target seasons and geographic regions, particularly small watersheds. In other words, policy maker cannot make decisions simply based on long term or large regional scale trends.

Changes in the timing and amount of snowmelt are important indicators of hydrological changes for the Missouri River Basin, Upper Mississippi River Basin and Great Lakes Basin, because snowmelt constitutes a significant part to the runoff and thus strongly impacts the timing and magnitude of maximum streamflow of the year. Agro-IBIS simulations indicate that proportionally more of the winter precipitation had fallen as rain rather than snow during the past 24 years, along with extensively decreases in snow depth, which is consistent with observations and previous studies (Groisman et al., 2001; Regonda et al., 2005; Thomson et al., 2005). Groisman et al. (2001) found a significant retreat of snow (1950-1998) in terms of both early spring snow cover extent and the mean date of last snow on the ground over the central U.S., especially for the Missouri River Basin; therefore, the snowmelt runoff had gradually shifted to earlier dates but with lower peak flow. Such changes in runoff are not shown in this study except for the region north of the Lake Superior. This is mainly because the seasonal time step analyses are not fine enough to capture the shift in seasonal streamflow. On the other
hand, runoff from the more precipitation in spring and early summer has largely compensated the weakening runoff from the reduced snowmelt.

Besides changes in surface water balance components (ET, runoff, $dW/dt$), it is also worth noting how soil moisture (SM) has changed correspondingly, since SM modifies the energy balance and the rate of water cycling between land surface and atmosphere. SM is also a good mediator of droughts and floods. Although the index of dryness (the ratio of annual potential evapotranspiration to precipitation; Budyko, 1974) was widely used as a surrogate of aridity that affects the water balance (Milly, 1994), SM is more appropriate to represent the land surface dryness because it directly support the transpiration processes for plants, particularly for crops. Changes in SM in the past two decades could be mediately roughly inferred from the trends in $dW/dt$ and snow cover. For a more intuitive understanding, the annual and seasonal trends in SM (in terms of volumetric water content) are shown in Figure 5.16. The Agro-IBIS modeled volumetric water content show general improved SM conditions during growing seasons (MAM and JJA) in the Missouri, Upper Mississippi and Ohio River Basins. By using the Keetch-Byram Drought Index (Keetch and Byram, 1968), Groisman et al. (2004) found similar SM trends in these areas over the past century.

To have a better understanding of the hydrological impacts of climate change, limitations in this study must be identified. One of the limitations is the uncertainties of the Agro-IBIS forcing data. Observations of precipitation are likely to have been underestimated because of rain gauge evaporation and the undercatch due to winds (Qian et al., 2007); topography is also a factor that induces biases in rain gauge records. Measurements of minimum air temperature and near surface wind speed are always
coupled with signals induced by changes in land uses and land cover (e.g. urbanization),
as discussed in section 5.2.2 and 5.2.6. Surface solar radiation from satellite
measurements is of relatively good quality, yet it is recorded only every 3 hours. The 3-
hourly temporal resolution is not fine enough to capture short lifetime clouds, thus
introduces short term biases to the daily averaged value. In addition to the biases induced
from measurement, some uncertainties may potentially arise from the interpolation when
data are gridded from the heterogeneous distributed station records.

Agro-IBIS is forced by hourly meteorological data that are generated by the
stochastic weather generator from daily averaged inputs. The weather generator is not
"intelligent" enough to adapt to changing climate, though it is able to generate the data of
extreme weather events. In the past several decades, precipitation events are documented
to have increased in terms of both intensity and frequency (Karl and Knight, 1998;
Kunkel et al., 1999). These heavy precipitation events, particularly when they occurred
around noon, could significantly affect plant physiology such as photosynthesis, and
further affect the rate of ET. There is also evidence of increasing nighttime precipitation
events (with greater amount and higher intensity) along with regional warming over the
central U.S. (Dai, 1999). Because a very small portion of precipitated water turns into ET
in the night, the upward trend of nocturnal rainfall mostly result in increasing runoff or
water storage. Therefore the model is not adept at capturing these types of changes and
consequently can possibly underestimate runoff to some extent.

However, streamflow (1940-1999) is documented to have increased only in the low
to moderate range (i.e., more baseflow rather than stormflow) over the central U.S. (Lins
and Slack). Pielke and Downton (1999) pointed out an apparent paradox that the
unchanged peak streamflow does not coincide with the increased extreme rainfall for the same period. Zhang and Schilling (2006) found that the increasing baseflow in the Mississippi River Basin is mainly as a result of land use changes and accompanying agricultural activities; the expansion of conversion tillage increases water infiltration since runoff is slowed or captured by row croplands. If one were to combine Zhang and Schilling’s (2006) points with previous documented precipitation and runoff trends, the paradox is addressed. Since changes in land cover are not considered in Agro-IBIS, the modeled surface runoff is likely to be overestimated over time, which compensates the potential underestimates of runoff from the increasing heavy precipitation.

The absence of anthropogenic dimensions in Agro-IBIS is another limitation in this study. Groundwater pumping is one of the many important human activities that directly affect surface energy and water balance. Most of the groundwater withdrawals in the Great Plains and Midwest region are used for agricultural cultivation (Brozović et al., 2010; McGuire, 2009). The overdraft of water consumption yields larger ET because pumped water is more susceptible to ET when redistributed on crop lands (Kustu et al., 2010; Walter et al., 2004). Mahmood and Hubbard (2004) reported that the annual total ET was 34% higher for irrigated corn than that for rain-fed corn. Milly and Dunne (2001) estimated that over the Mississippi River Basin, the rate of groundwater depletion (1949-1997) is about 3 mm century\(^{-1}\), and approximately 27.4% of the upward trend of ET is attributed to irrigation and other human activities (e.g. water management). The out of consideration of consumptive water uses in Agro-IBIS largely explains the high biases of modeled runoff trends.
Changes in agricultural land use and land cover are also considerable aspects that affect terrestrial water cycle. Over the Mississippi River Basin, there is evidence of rapid expansion of soybean cultivation that was converted from natural grassland and forests during the second half of the twentieth century (Donner, 2003; Zhang and Schilling 2006). During this time, corn-soybean rotation cultivation was widely used. The intensive agricultural land management did not only affect average runoff, but also altered ET losses (Renner and Bernhofer, 2011). As the average transpiration rate of seasonal crops is lower than that of perennial vegetation (FAO, 1998), Zhang and Schilling (2006) pointed out that the conversion of perennial vegetation to row crops may have reduced ET in the basin. Furthermore, unlike natural vegetation, seasonal crops do not transpire throughout the unfreezing seasons but until mid-growing season (Dinners, 2004). As a result, such changes in land uses and land cover may bring uncertainties to the interannual and intraseasonal variability in ET and runoff that are modeled in Agro-IBIS.

In addition to changes in rural land uses, rapid urbanization during the past century played an influential role in altering the surface hydrological processes. Many studies indicated that runoff has increased along with the development of urban area (Sala and Inbar, 1992; Tang et al., 2005; Palmer et al., 2004). Rainfall is more likely to be collected by city drainage system and then rapidly routed into streams. More remarkably, snow cover in the urban area melts faster than in the rural areas; therefore most of the winter precipitation are partitioned into runoff rather than evaporation. This explains the seasonal runoff biases in this study. Considering that the Ohio River Basin and Great Lakes Basin include densely populated areas with numerous highly developed cities and towns, more wintertime runoff occurs in these regions rather than the sparsely populated
Missouri and Upper Mississippi River Basins. Since Agro-IBIS does not take urban areas into account, low biases in wintertime runoff are found with respect to observations. In addition, with the continuous growth in population and wealth, the trends in runoff are likely to be underestimated over time. For the consideration of flood mitigation and water management, it is worth noting that high level of urbanization increases the societal vulnerability to potential climate changes (Palmer et al., 2004).

Some uncertainties also arise from the internal algorithms in Agro-IBIS. Most of the algorithms in energy, water and carbon fluxes in the model are purely empirical and therefore the accuracy of simulations usually could not satisfy all locations. For example, although the basin-wide downward longwave radiation is well simulated by the model (c.f. Appendix B), some unsymmetrical biases are found in specific locations such as the Republican River site in Nebraska (Mykleby, 2012). For the snow thermodynamics, Agro-IBIS assumes a constant snow density, and the snowpack density is the same as snowfall density (Vano et al., 2006). Also, snow compaction and refreezing are ignored in the model, and fractional snow cover is assumed to vary linearly with snow height (Lenters et al., 2000). Therefore, wintertime water and energy flux are potentially influenced. Moreover, Twine and Kucharik (2008) evaluated the model with satellite information of greenness and found biases in leaf onset and growing season LAI, which indicates some uncertainties in vegetation phenology and physiology. Summing up the above, continuing improvements in land surface models are necessary for better understands of the regional-scale land surface hydrological/terrestrial ecosystem processes.
Chapter 6

Sensitivity of Modeled Hydroclimatic Trends to Individual and Combined Forcing Factors

6.1 Introduction

Historical changes in hydrological cycle are suggested being associated with climate change and variability (Portmann et al., 2009; Huntington et al., 2009). Because intricate feedbacks constitute the coupled nature of these two systems, the interactions between climate variables (e.g., air temperature, precipitation, wind speed, atmospheric humidity, solar radiation) and hydrological variables (e.g., ET, runoff, soil moisture, snow cover, groundwater table depth) are highly complex. For this reason, it is not rational to simply relate changes in climatic and hydrological variables based only on observations. Considering that changes in climate and water cycle have far-reaching implications for water resource planning and management, it is necessary to understand the sensitivity of surface water balance to changes in climate factors.

One of the methods to assess the hydrological impacts of climate change is theoretical or analytical derivative equation based. These approaches usually consider the climate variables as inputs to the formula while the outputs are indices that represent hydrological characteristics. For example, the Keetch-Byram Drought Index (KBDI;
Keetch and Byram, 1968) is a cumulative algorithm for estimating soil moisture deficiency, which is dependent on meteorological input parameters including daily maximum air temperature, daily precipitation, and mean annual precipitation. Using the KBDI and future scenarios of temperature and precipitation for the period 2070–2100, Liu et al. (2010) found that the drought potential increased significantly in the United States, and this change was mainly caused by climate warming in the 21st century.

Climate elasticities are also important indicators for evaluating the effects of changing climate on hydrological elements. Yang and Yang (2011) derived the elasticities of runoff to precipitation, net radiation, air temperature, wind speed, and relative humidity in order to separate the contributions of different climatic factors; they found that in the Futuo River basin of China, decreases in precipitation was mainly responsible for declined runoff, and the reduced wind speed had the second greatest effect.

Sensitivity experiments provide a means to quantitatively assess the effects of climate change on surface water balance. Such approach usually requires semi-empirical formula as the basis in order to examine how the experiment output depends on the varying input. For example, based on the Penman-Montieth equation, Irmak et al. (2006) analyzed the sensitivity of reference evapotranspiration (ET_{o}, the potential ET over the well-watered, full-cover grass surface) to diverse climate variables in 7 different locations of the United States. Meteorological variables were prescribed to increase and decrease by certain unit intervals so as to quantify the corresponding changes in ET_{o}. They found that the reference ET was most sensitive to vapor pressure deficit than other variables including wind speed, surface solar radiation, minimum and maximum air temperatures. In addition, sensitivity of ET_{o} to the same climate variable showed significant variation
by locations and seasons. Singh and Xu (1997) investigated sensitivity of evaporation to vapor pressure, wind speed and temperature on the basis of seven generalized mass transfer-based evaporation equations. ±5%, ±10% and ±20% errors of the daily and monthly input parameters are employed on these equations to examine the corresponding errors of the estimated evaporation. They found that evaporation was particularly sensitive to vapor pressure, less sensitive to wind speed and least sensitive to temperature. Since such methods are highly dependent on semi-empirical equations, these experiments are usually taken on specific sites and the results do not apply to broader areas.

Land surface models (LSMs) are useful tools for sensitivity studies on the regional scale. LSMs implicitly treat the effect of the vegetation canopy and soil on hydrological processes and thus are able to investigate the corresponding changes in water cycle more comprehensively. By using a rule-based hydrology-vegetation dynamic model, Poiani et al. (1995) studied the sensitivity of a semi-permanent prairie wetland in North Dakota to temperature and precipitation changes. The model was run with increased temperatures of 2°C combined with a 10 percent increase or decrease in precipitation for spring, summer and fall seasons respectively. They found that increases in spring precipitation played an important role in maintaining an extensive open water area, but increases in precipitation for other seasons did not mitigate the water loss in the context of a warming climate. Lynch et al. (2001) applied a set of perturbations to the atmospheric forcing of the Alaska Arctic area to the NCAR LSM and found that the summer soil water content was strongly controlled by precipitation and downwelling longwave radiation rather than other factors including shortwave radiation, air temperature, atmospheric humidity and wind speed. In the lower Mississippi River Basin, river discharge was found to be sensitive to soil
moisture according to the Community Land Model 3.5 simulations (De Lannoy et al., 2011).

Some LSMs based sensitivity experiments generally assume a certain amount of perturbations of meteorological variables or land surface parameters for everywhere in the study area (e.g., increase the precipitation by 5% everywhere) and for any time of year. However, such kind of climate scenarios might be unrealistic for some locations and seasons. For example, during the past two decades, the annual precipitation has increased in North Dakota and Minnesota, but decreased significantly in southwestern United States (referred to Figure 5.1c in chapter 5). In addition, the seasonal precipitation in the western United States has increased significantly in winter but decreased in summer. In this case, although such idealized perturbations do help us understand how the hydrological elements depend on the varying climatic variables, they do not make any practical significance.

In recent decades, land surface models are widely coupled to general circulation models (GCMs) and regional or mesoscale atmospheric models (e.g., Winter and Eltahir, 2012; Bonan et al., 2012; Gedney et al., 2000). Unlike LSMs, the atmospheric forcings of the land surface scheme are internally obtained from the coupled atmospheric models. Therefore, these models are able to generate relatively realistic perturbed climate scenario in comparison to those perturbations for LSMs. Coupled land-atmosphere models have advantages that they do not require numerous weather data as inputs, and thus offer an effective method to investigate the sensitivities of land surface processes to potential future climate scenarios. Although GCMs are able to simulate large scale features of global climate with some skill, their regional scale features or downscaled outputs
sometimes show large biases in comparison to observations, and their future projections
do not usually agree with each other (Thomson et al., 2005; Lettenmaier et al., 1999;
Kirshen and Fennessey, 1995; Wolock and McCabe, 1999). In addition, the GCMs
usually operate at relatively coarse spatial resolutions (typically half to several degrees
latitude by longitude) with imprecise parameterizations (Nijssen et al., 2001). As a result,
their projected meteorological variables do not satisfy the high resolution demand of the
sophisticated land surface models. Therefore, despite that the coupled land-atmosphere
models are capable of predicting future scenarios of hydrological changes, results from
their simulations have relatively limited confidence for regional and local scale water
resources planning and management.

For the time period from 1984 to 2007, evident changes in climate were detected
over the central United States (referred to chapter 5, section 5.2). In the meanwhile,
significant changes in water cycle were also found (referred to chapter 5, section 5.4) in
the region. For assessing the sensitivity of water balance to changes in climate factors
over the past two decades, a hypothetical-empirical based approach is presented in the
current study. A process-based land surface model, namely Agro-IBIS, is applied for
carrying out the sensitivity experiments. Both annual and seasonal responses of
hydrological components are investigated through a set of simulations under perturbed
climate scenarios. Furthermore, factors other than climatic variables that affect the
surface water balance are discussed in the latter part of the study.
6.2 Agro-IBIS Simulations

A control simulation (referred to as “CTL”) has been completed in chapter 4 that Agro-IBIS is run over the entire study domain for the period 1984-2007 with historical climate drivers. Another group of simulations that are done in the current study are similar to the CTL except that the model is forced with perturbed climate drivers with respect to historical climate inputs (referred to as “PTB”). Then deviations of each PTB output from the CTL output are examined respectively. The significance of difference for the deviations are examined by the Student’s t-test at 90% confidence level.

In order to generate new climate scenarios, perturbations are applied to each climate input including minimum air temperature (TMIN), maximum air temperature (TMAX), surface solar radiation (RADS), precipitation (PREC), relative humidity (RH) and wind speed (WSPD). Unlike some other LSM based sensitivity studies, perturbations applied in this study are not uniform over space and time. Instead, each location has a specific perturbed scenario that is based on historical climate change from 1984-2007 (i.e., climate change scenario that simply “added” the historical trend to the original input data). Moreover, these climate change scenarios are created at seasonal time steps (i.e., historical trends are applied across each season of the 1984-2007 period). Technically speaking, the historical climate trends are likely extrapolated to another 24 years period (think of these “future scenarios” as being like the period 2008-2031, with the “predictions” being based not on GCM scenarios, but rather on the trends of the past 24 years). What this method basically implies is to test “what the next 24-year period (compared to 1984-2007) would be like” if it were to continue to trend in the same way as the past 24 years have.
The algorithm below explicitly illustrates how a new 24-year dataset is created for being used as the perturbed climate scenario,

\[ Var_{PTB} = Var_{CTL} \times \frac{Sen_{end}}{Sen_{start}} \]

where, \( Var \) could be any climate variable of PREC, WSPD, RH and RADS. \( Var_{CTL} \) is a climate variable in the control simulation, while \( Var_{PTB} \) is the corresponding variable in the perturbed scenario. \( Sen_{start} \) means the value of the starting point on the Sen’s slope, while \( Sen_{end} \) is the value of the ending point on the Sen’s slope. On a seasonal basis, this algorithm is applied to each daily data for the period 1984-2007 and consequently to generate a “future” daily dataset for the next 24-year period. The purpose to use Sen’s ratio in representing percentage change is to avoid potential negative values in the new dataset. For TMAX and TMIN, the following algorithm is applied.

\[ Var_{PTB} = Var_{CTL} + (Sen_{end} - Sen_{start}) \]

In the current study, totally 6 sensitivity climate scenarios are developed. For the first 5 scenarios, perturbations are imposed to a certain climate factor, while other factors are kept isolated. In addition, a combined sensitivity experiment is undertaken in order to examine the effects of co-varying changes in climate drivers.

These climate scenarios are listed in Table 6.1. The notation “d” means “simulation with imposed perturbation”, while the acronym right after it means the corresponding variable. For example, dRADS means that surface solar radiation are imposed by perturbations while the other factors remain as what they were. These climate
scenarios are useful to evaluate how sensitive and how linear the results are in response of changes in different climate factors.

6.3 Results

The deviations of each PTB water balance (in terms of the 24-year annual and seasonal means) from the CTL’s are shown in 8-panel maps (panel (a), (b), (c), (e), (g) and (h) in Figure 6.1-6.20). To keep the color scheme consistent with that in the previous chapters (blue means “more water” and red means “less water”), deviations of -ET are plotted in Figure 6.1-6.5 rather than ET. Deviations of runoff and \( \frac{dW}{dt} \) are presented in Figure 6.6-6.10 and 6.11-6.15 respectively. Differences in VWC are also shown in Figure 6.16-6.20 in order to distinguish changes in soil moisture and other forms of water storage such as snow cover. Changes in climatic inputs are referred to Figure 5.1-5.7.

For dPREC simulations, on the annual time scale, changes in precipitation result in consistent changes in ET and runoff (i.e., \( \Delta \) ET and \( \Delta \) runoff are positively scaled with \( \Delta \) precipitation; Figure 6.1a). The perturbed precipitation is mostly reflected in changes in runoff, followed by ET, while the difference in \( \frac{dW}{dt} \) is not distinctive (Figure 6.11a). However, on the seasonal scale, the spatial pattern of such consistence is disturbed. For example, although there is a drying condition in the southern Upper Mississippi River basin (Figure 5.4d), the ET in this area continues increasing (Figure 6.5a). The undiminished ET is attributed to substantial available water in the soil according to the \( \Delta \) VWC map in Figure 6.19a. Results also show that with respect to the changing precipitation, seasonal runoff and \( \frac{dW}{dt} \) are more sensitive than ET.
Unlike dPREC runs, results in dRADS simulations show more spatial and temporal complexity in the connections between solar radiation and surface water balance. The annual mean deviation in ET (Figure 6.1b) shows that ET declines in response of solar brightening (Figure 5.1e). As simply relate \( \Delta \) ET, \( \Delta \) runoff and \( \Delta dW/dt \) to solar radiation trends, no distinct matching patterns are found. However, when take \( \Delta \) VWC into consideration, it is found that seasonal \( \Delta \) ET is more strongly affected by soil water condition in the previous season rather than solar brightening or dimming in its corresponding season. For instance, although the solar brightening in SON is very significant in the northwestern Ohio River basin, southern Upper Mississippi River basin and eastern Missouri River basin (Figure 5.6d), ET doesn’t increase but in turn decreases in this area (Figure 6.5b). This is attributed to the lack of available water for plant transpiration because the soil has lost a large amount of water in JJA (Figure 6.19b). Since precipitation in dRADS runs doesn’t change in reference to the baseline scenario (i.e., the same as that in the CTL run), patterns of \( \Delta \) runoff and \( \Delta dW/dt \) show a clear complementary relationship (i.e., \( + \Delta \) runoff is related with \( - \Delta dW/dt \), and vice versa).

For dTEMP simulations, changes in ET are positively scaled with changes in air temperature (i.e. \( +\Delta \) temperature give rise to \( +\Delta \) ET and vice versa; Figure 6.1c) except in some certain cases such as the western Missouri River basin in DJF and JJA. Similar to the situations in dRADS runs, such inconsistency could be explained by the constraint of soil moisture availability. On the annual time scale, \( \Delta \) ET and \( \Delta \) runoff show a nice complementary relationship, and therefore \( dW/dt \) seems not sensitive to changes in temperature. However, significant patterns of \( \Delta dW/dt \) show up on the seasonal
deviations maps since the seasonal $\Delta$ ET and $\Delta$ runoff does not complement each other as nicely as they do in the annual results.

Results in $d$RH simulations are very similar to the results in $d$RADS simulations. Exceptions of such similarities appear on $\Delta$-ET in the Great Lakes basin in JJA (Figure 6.5b and 6.5g) and SON (Figure 6.5b and 6.5g). The ET increases significantly in the $d$RADS runs but decreases in the $d$RH runs in contrast. This indicates that rather than water, energy is the limited climatic factor in controlling the water balance in the corresponding regions and seasons.

$\Delta$ ET and $\Delta$ runoff in $d$WSPD simulations show clear matchup with changes in wind speed on both the annual and seasonal scales. Changes in ET are negatively scaled with changes in wind speed. In addition, $\Delta$ ET and $\Delta$ runoff are complementary throughout the seasons. Responses of $dW/dt$ to changes in wind speed are not conspicuous.

The purpose of these sets of sensitivity experiments is to examine how linear the hydrological responses are to different individual perturbed factors. Comparing among the results of all single perturbed climatic variable simulations (i.e., compare panels (a), (b), (c), (e), (g) and (h) in Figure 6.1-6.20 respectively), it seems that the surface water balance is most sensitive to surface solar radiation and relative humidity, less sensitive to precipitation, followed by air temperature and wind speed. Panel (d) in each figure shows the composite of map (a), (b) and (c). Comparing among the top 4 maps in each figure respectively, it’s clearly to see that the sensitivity of surface water balance to surface
solar radiation dominates the other two (i.e., sensitivities to precipitation and temperature).

Panel (e) in each figure is a companion map to the panel (d), which shows the deviations of the surface water balance from the multiple perturbed climatic variables simulations \( d(\text{PREC}+\text{RADS}+\text{TEMP}) \) in reference to the results of the CTL simulations. By comparing map (d) and (e), one can tell whether or not the response to combined climate factors is as linear as the responses to individual climate factors. Results show that there is hardly a consistence pattern between the two simulations, which indicates that the hydrological impacts of each climatic variable are not independent, but exhibit highly nonlinear behaviors.

Panel (f) in each figure shows the 24-year changes in water balance that are generated on the basis of historical water balance trends of the control simulation (referred to Figure 5.9-5.13 in chapter 5). These changes are considered as the surrogates of the results from the simulations with “all perturbed” climatic inputs. By comparing the results of the “individual perturbed” simulations (map (a), (b), (c), (g) and (h) in each figure) with the results of the “all perturbed” simulation, it tells what the most likely predominant climate factor is that controls the hydroclimatic trends in the study domain. Results indicate that precipitation is the dominant climate factor that affects changes in ET, runoff and soil moisture (VWC) on both the annual and seasonal scales. For changes in annual \( \frac{dW}{dt} \), it seems that none of the “individual” perturbed patterns show similarity to the “all perturbed” pattern. This might be in part because that the annual \( \frac{dW}{dt} \) trend is not significant, so that the potential matching pattern doesn’t show up on
the map. However, on the seasonal scale, it is obvious that changes in $\frac{dW}{dt}$ are dominated by precipitation.

### 6.4 Discussion and Conclusions

A series of sensitivity experiments are conducted in this study in order to investigate the sensitivity of modeled hydroclimatic trends to individual and combined forcings, as well as the contribution of key meteorological variables to the variations of ET, runoff and water storage from 1984 to 2007 in the central U.S.. In addition to the control simulation that is done in chapter 4, Agro-IBIS is applied for another 6 simulations with perturbed climatic inputs in reference to the baseline climate scenario. Results show that the sensitivities of water balance to the imposed changes in climate are not constant over time but exhibit distinct seasonal variations, and in some cases, the seasonal responses present contrasting patterns to the annual ones. Besides the variations in climate drivers throughout the year, the seasonal differences of the sensitivities can be largely attributed to the seasonal variations in soil moisture, and snow packs in the areas where snow constitutes the majority of the precipitation.

Results also show that the water balance in this domain is most sensitive to solar radiation and relative humidity, followed by precipitation, air temperature and wind speed. Although the sensitivities can be characterized by the linear responses of the surface water balance to the independent changes in climatic factors, the linearity of the sensitivities disappears when relating the changes in water balance to the co-varying perturbed climate scenarios.
By comparing the results of sensitivity experiments with independent perturbed climate forcings to the historical linear trends, it appears that among all the climate variables, precipitation is the predominant factor that affects decadal scale hydroclimatic changes in the study domain. This implies that rather than available energy, the land surface water balance is restrained by available water. Therefore, on the annual to decadal time scale, it seems that the central U.S. is actually a “water-limited” region.

The surprisingly similar results of dRADS and dRH simulations draw attention to the interdependency between solar radiation and relative humidity. According to the historical trend of the two climate factors (Figure 5.1d, 5.1e, 5.5 and 5.6), it is not obvious to find any direct connections between each other. In spite of this, the sensitivity patterns uncover the strongly coupled nature of solar radiation and relative humidity on their role of governing the surface water balance.

Although isolating one factor and fixing other factors provide a useful means to examine the linear effect of different climate factors on surface water balance, climate scenarios with “separate” forcings are sometimes unrealistic to investigate the real effects. Because in nature, these climate factors are always coupled to one another, such idealistic sensitivity experiments can result in plausible issue in land-atmosphere interactions. For example, results of dRH simulations show a declining ET with increasing RH as the atmospheric forcing. It shows contrasting pattern against the historical hydroclimatic trends and contradicts the facts that high relative humidity is always associated with ample amount of ET.
In nature, changes in relative humidity are closely connected with changes in air temperature, precipitation and evapotranspiration. For $d$RH simulation in this study, relative humidity is the only perturbed meteorological variable but the air temperature and precipitation are not. For this reason, as the relative humidity increases, the vapor pressure deficit (VPD), a function of temperature, decreases. VPD describes the capacity of the atmosphere to take up water, which strongly controls the rate of ET. The larger the VPD, the more the water will evaporate or be transpired into the atmosphere by plants, assuming that the terrestrial water supply are sufficient. Since Agro-IBIS calculates ET using a standard mass transfer method, changes in relative humidity directly impacts the variations in ET. This explains why ET is inverse related with RH in the sensitivity experiment. Therefore, attentions should be paid to the effects of interdependent climate variables on the surface water balance in order to reasonably explain the results of the sensitivity experiments.

Other than the interdependencies of climate variables, the analysis time scale in sensitivity experiments is another factor that affects the results. For example, although the solar brightening over the study domain is significant (Figure 5.1e), the expected increases in annual ET do not show up in the dRADS results (Figure 6.1b). On the contrary, decreases in ET are shown on the maps (Figure 6.1b). For dRADS simulations, the water related climate forcings (e.g., precipitation and relative humidity) do not co-vary with perturbed solar radiation. In this case, although the evaporative demand increases due to more available energy, the available water does not increase accordingly. For a short time period (e.g., daily to seasonal time scale), ET increases in line with solar brightening, but as time goes by, the soil is becoming drying, and the climate system is
losing water. Thereby a negative feedback occurs to suppress the ET: reduced uptake of water from drying soil results in stomatal closure of the plants to restrict further water loss, so that the rate of ET is restrained. Such negative feedback is magnified over time due to the decoupled water related forcings in the simulations. Consequently, for the long run (e.g., annual to decadal time scale), ET is more dependent on the land surface water condition rather than solar radiation, and hence exhibits extensively declining pattern in Figure 6.1b.

Geographical variance of hydroclimate (i.e., regional dry and wet conditions) is another factor that impacts the sensitivities of surface water balance to climate variables. For instance, in the Missouri River Basin (a water-limited region), ET is more sensitive to changes in precipitation in the fall season (Figure 5.4d and 6.5a), while in the Ohio River Basin (an energy-limited region), ET is more sensitive to changes in solar radiation (Figure 5.6d and 6.5b) and temperature (Figure 5.2d and 6.5c). Similar geographical differences in hydrological sensitivities were also reported in other studies. For example, Irmak (2006) found that the effects of climate factors on reference ET are characterized by distinct differences in arid, semi-arid and humid regions. These results indicate that the seasonal hydrological responses to climate change are highly dependent on the regional availability of water and energy.

In the similar way, the predominant factor that governs the surface water balance varies by locations. For example, Zuo et al. (2012) found that relative humidity was the most sensitive variable for potential ET in the Wei River Basin, followed by wind speed, air temperature and solar radiation. Hence for the policy making purpose, though
precipitation is generally the dominant factor in the central U.S., scrutinizes on smaller spatial scales (e.g., subbasins, or even local scale) are highly recommended.

It is worth noting that the hydrological sensitivity to climate change varies according to what the estimation method is applied. Using eight alternative approaches (e.g., Thornthwaite, Blaney-Criddle, Hargreaves, Samani-Hargreaves, Jensen-Haise, Priestley-Taylor, Penman, and Penman-Monteith methods), McKenney and Rosenberg (1993) estimated the sensitivity of potential evapotranspiration to climate change and found that their sensitivities to climate drivers differ from one another. Sometimes even if the estimation approaches consider the same climate factors, their sensitivities to climate can still be different. Singh and Xu (1997) compared the performance of 20 mass transfer based evaporation equations and found that they differ in both their estimations and sensitivities to climate variables.

Agro-IBIS calculates ET using a standard mass transfer equation, so it is not surprising to see that the ET is more sensitive to relative humidity than precipitation. However, because the runoff is directly associated with changes in precipitation, it shows less sensitivity to relative humidity than precipitation. In addition, because the runoff is calculated as the residual of the precipitation, ET and $dW/dt$ in the model, some complementary patterns are found among these water balance components.

The influences of temporal resolution of the climate inputs on the sensitivity experiments are necessary to be aware of. Singh and Xu (1997) reported a much more sensitive effect on evaporation estimates in the monthly cases than in the daily case. Agro-IBIS is capable of handling sensitivity experiment with a relative precision, since it
simulates water balance using hourly data that are generated by a stochastic weather
generator. Nevertheless, similar as the discussion in chapter 5, the model errors occur
with the increasing extreme weather events that are companied with global warming. For
example, based on the precipitation data from the National Climate Data Center (NCDC)
for the period 1970-1999, Groisman et al. (2004) found increases in heavy precipitation
(upper 5% precipitation event; 4.6 mm decade\(^{-1}\))
, very heavy precipitation (upper 1%
precipitation event; 7.2 mm decade\(^{-1}\))
and extreme precipitation (upper 0.1% precipitation
event; 14.1 mm decade\(^{-1}\)) over the contiguous United States, especially significant for the
Upper Mississippi, Ohio River Basins and the Great Lakes Basin. Because more and
more heavy precipitation can occur at minute to hour time scale and the timing of the
events are shifting during the day, the hydrological sensitivity is subject to change. Such
effects are particular important for mitigation strategy making and future projection
considerations.

Apart from the climate forcings, other environmental factors that are not tested in
this study play unnegligible roles in determining the surface water balance as well.
Groundwater, the largest form of belowground water storage in the water cycle, behaves
as the moderator in the role of water balance. Variations in groundwater, in terms of
changes in water table depth, can be induced by either climate variations or human
activities (e.g., pumping), or both. Many studies have documented that while
groundwater conditions are altering, the ET-groundwater interactions are subject to
change, and thus so are the sensitivities (e.g., Maxwell and Miller, 2005; Kollet and
Maxwell, 2008; Yeh and Eltahir, 2005; Soylu, 2011; Nosetto et al., 2009; Kustu et al.,
2010; Walter et al., 2004; Mahmood and Hubbard, 2004; Milly and Dunne, 2001).
Natural land cover and anthropogenic land use changes can alter the sensitivity of surface water balance through modifying the ET processes. For example, Twine et al. (2004) analyzed the effects of land cover changes on the energy and water balance of the Mississippi River basin using the IBIS model, and found that the ET rates decrease (up to ~0.75 mm day\(^{-1}\); 20%) over summer crops (corn and soybean) converted from forest and increase (up to ~0.4 mm day\(^{-1}\); 45%) over summer crops converted from grassland. Other forms of land changes, such as the invasion of non-native vegetation are also documented to have important effects on ET, runoff and soil moisture in many regions of the central U.S (Hooper et al., 2005; Soylu, 2011).

Studies suggested that the differences in soil texture and soil utilization may strongly modify the hydrological changes as well (e.g., Wang et al., 2009; Zhang and Schilling, 2006). In addition, Gedney et al. (2006) reported that the CO\(_2\)-induced stomatal closure may result in the suppression of plant transpiration. In the 21\(^{st}\) century, the carbon emission will continue increasing and so do the human induced changes in land uses. Although Piao et al. (2007) pointed out that changes in climate and land use have a larger direct impact than rising CO\(_2\) on the runoff trends, practically and more accurately, for decadal or century-long scale strategic planning and environmental mitigation purpose, all of the factors affecting surface balance that are discussed above are not recommended to be neglected.
Chapter 7

Summary and Conclusions

The climate system and the hydrologic cycle are strongly connected with each other. Observations over the central U.S. in recent decades show numerous changes in hydrologically significant climatic variables. This includes increases in air temperature and solar radiation (the solar brightening suggest decreases in cloud cover as well), decreases wind speed and seasonal shifts in precipitation rate and rain/snow fraction. Understanding the interactions between these two systems is important, since variations in climate can trigger extensive changes in the hydrologic cycle, with significant impacts on agriculture, ecosystems, and society. The sufficiency of water supply to meet the changing demands is a challenging issue, and the study of the surface water balance provides a good means to assess the water availability on the regional scale.

The overall goals of this study are to assess the impacts of climate change on the surface water balance of the central United States for the period 1984-2007, and present useful information for managing the water resources, planning the field crop, and making pertinent policy and adaptation strategies. In answer to the three scientific questions in this research, this thesis was correspondingly written in three main chapters:

- Surface Water Balance of the Central U.S.: Long-term Mean and Seasonal Variability (Chapter 4)
- Hydroclimatic Trends in the Central U.S. from 1984-2007 (Chapter 5)
Sensitivity of Modeled Hydroclimatic Trends to Individual and Combined Forcing Factors (Chapter 6)

In these studies, a terrestrial ecosystem/land surface hydrologic model, namely Agro-IBIS, was employed to simulate the surface water balance and hydrological trends, and conduct sensitivity experiments. A higher spatial resolution (5-minutes) and finer temporal resolution (daily) meteorological dataset was used for representing the input climate scenarios in order to improve the accuracy of the regional scale modeling. Analyses of the model outputs were considered on not only annual time scale, but also seasonal scale.

In chapter 4, The Agro-IBIS was forced by observed climatic inputs for the period 1984-2007 (denoted as the “control simulation”). Basin averaged results showed that the seasonal cycle of the surface water balance and the snow depth are characterized by distinct variations among the Upper Mississippi, Missouri, Ohio, and Great Lakes basins. Also, the snow depths are characterized by more interannual variability than the water balance components, which implied the importance of snow cover in the role of affecting the interannual variations of the water balance. Results of the spatial distributions of the snow depth and the wintertime rain/snow fraction indicated that the northern areas of the study domain are hydrologically snow-dominated regions, where received more snow but less precipitation in winter than the southern areas of the domain. In addition, the long term annual and seasonal means of each water balance component were analyzed for each location in the study domain, which were presented in reference of interpreting the results in the following chapters.
In chapter 5, in addition to calculating the long term means, trend analyses were accordingly applied to the outputs of the control simulation. The results generally showed an acceleration of the water cycle in the Upper Mississippi, Missouri, Ohio, and Great Lakes basins, but with significant seasonal and spatial complexity. Over the past 24 years, evapotranspiration (ET) has increased in most regions and most seasons, particularly during the fall, which is also a time of pronounced solar brightening. Trends in runoff are characterized by distinct spatial and seasonal variations. Since recent warming has led to a greater fraction of winter precipitation falling as rain rather than snow, spring runoff in some snow-dominated regions (such as the northern Great Lakes) has declined significantly since 1984. Other regions, however, such as the northern Missouri basin, showed large increases in runoff throughout all seasons, primarily as a result of increased precipitation.

Finally, in chapter 6, a group of sensitivity experiments were conducted in order to investigate the sensitivity of the surface water balance to imposed climate change, as well as the contribution of key meteorological variables to the variations of ET, runoff and water storage from 1984 to 2007 in the central U.S.. In reference to the baseline climate scenario, 6 sets of perturbed climate scenarios (with perturbed air temperature, solar radiation, precipitation, relative humidity and wind speed respectively, and a combination of perturbed air temperature, precipitation and solar radiation) were created based on the historical climatic trends in the past 24 years. Agro-IBIS was then run with these climate inputs respectively (denoted as the “perturbed simulations”). Result showed that the sensitivities of water balance to the imposed changes in climate are not constant over time but exhibit distinct seasonal variations, and in some cases, the seasonal responses
present contrasting patterns to the annual ones. Besides the seasonal variations in the climate forcings, the seasonal differences among the sensitivities can be largely attributed to the seasonal variations in soil moisture, and in the snow-dominated region, the snow packs.

It was found that generally the water balance in this domain is most sensitive to solar radiation and relative humidity, followed by precipitation, air temperature and wind speed. Although the sensitivities of surface water balance to the independent changes in climatic factors were characterized by the linear responses, in reality the effects were highly non-linear and exhibited more complexity. This is because of the interdependencies of the climate factors that they are always coupled to each other in nature. In addition, sensitivity patterns uncovered the strongly coupled nature of solar radiation and relative humidity on their role of governing the surface water balance.

Results also indicated that the seasonal hydrological responses to climate change are highly dependent on the regional availability of water and energy. In the western half of the central U.S., for example (a water-limited region), ET was more sensitive to changes in precipitation, while in the energy-limited eastern half, ET was more sensitive to changes in solar radiation and temperature. Comparisons of each annual result of the perturbed simulations with the historical linear trends in the surface water balance indicated that among all the climate variables, precipitation is the predominant factor that affects decadal scale hydroclimatic changes in the study domain. This implied that rather than available energy, the land surface water balance is restraint by the available water, and thus on the annual to decadal time scale, the central U.S. is actually more like a “water-limited” region.
In conclusion, this research provides insight into the impacts of climate change on the surface water balance in the central U.S. for the period 1984-2007. Although the hydrologic cycle was found to have been accelerating, the changes in each water balance component are certain to be complex, affected by various climate factors, and varied by time and location. Such a complexity challenges the policy making process, which means that the decision cannot be made simply based on the results from large basin-wide or long term trend analyses. In turn, analyses of the surface water balance on more comprehensive spatial (i.e., not only the state scale, but also the basin-wide and local scales) and temporal (i.e., both the annual and seasonal scales) scales are recommended to be taken into account. In addition, it is highly suggested to consider the combined effects of climate factors on their roles of affecting the surface water balance, though precipitation was found to be the predominant factor that governs the hydrologic change in the central U.S. in recent decades.

Though being beyond the scope of this research, the impacts of human activities on the surface water balance would be interesting for future studies. As Milly (2008) stated “Stationarity is dead”, the hydrologic cycle is strongly influenced by anthropogenic disturbance, such as the groundwater pumping, changes in land use/land cover and the reservoir management. Therefore, a more sophisticated land surface model is in demand to explicitly represent both climate and human dimension. In addition, continuing improvement of the model resolution and the input data are preferable in order to build a higher confidence on the accuracy of the modeling results.
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Figures

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Figure 5.3. Same as Figure 5.2 but showing seasonal trends in diurnal temperature range.
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Figure 5.9. Maps showing spatial distribution of annual trends in observed and Agro-IBIS simulated water balance components for the period 1984-2007, including (a) observed precipitation, (b) ET, (c) runoff, and (d) dW/dt. Trends that are statistically significant at 90% confidence level are hatched. Each basin is highlighted with the dark line as the boundary.
Figure 5.10. Maps showing spatial distribution of seasonal (DJF) trends in observed and Agro-IBIS simulated water balance components for the period 1984-2007, including (a) observed precipitation, (b) -ET, (c) runoff, and (d) dW/dt.
Figure 5.11. Same as Figure 5.10 but for the season of MAM.
Figure 5.12. Same as Figure 5.10 but for the season of JJA.
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Figure 6.2. Same as Figure 6.1 but for the DJF deviations.
Figure 6.3. Same as Figure 6.1 but for the MAM deviations.
Figure 6.4. Same as Figure 6.1 but for the JJA deviations.
Figure 6.5. Same as Figure 6.1 but for the SON deviations.
Figure 6.6. Deviations of annual mean (1984-2007) runoff in the Agro-IBIS perturbed simulations from those in the control simulation. Climate scenarios in the perturbed simulations are the same as those in the control run except that (a) precipitation, (b) solar radiation, (c) temperature, (e) precipitation, solar radiation and temperature, (g) relative humidity, and (h) wind speed are modified. Map (d) showing a composite pattern by adding (a), (b) and (c) together. Map (f) showing 24-year changes in ET based on historical (1984-2007) linear trend.
Figure 6.7. Same as Figure 6.6 but for the DJF deviations.
Figure 6.8. Same as Figure 6.6 but for the MAM deviations.
Figure 6.9. Same as Figure 6.6 but for the JJA deviations.
Figure 6.10. Same as Figure 6.6 but for the SON deviations.
Figure 6.11. Deviations of annual mean (1984-2007) $\Delta dW/dt$ in the Agro-IBIS perturbed simulations from those in the control simulation. Climate scenarios in the perturbed simulations are the same as those in the control run except that (a) precipitation, (b) solar radiation, (c) temperature, (e) precipitation, solar radiation and temperature, (g) relative humidity, and (h) wind speed are modified. Map (d) showing a composite pattern by adding (a), (b) and (c) together. Map (f) showing 24-year changes in ET based on historical (1984-2007) linear trend.
Figure 6.12. Same as Figure 6.11 but for the DJF deviations.
Figure 6.13. Same as Figure 6.11 but for the MAM deviations.
Figure 6.14. Same as Figure 6.11 but for the JJA deviations.
Figure 6.15. Same as Figure 6.11 but for the SON deviations.
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Figure 6.17. Same as Figure 6.16 but for the DJF deviations.
Figure 6.18. Same as Figure 6.16 but for the MAM deviations.
Figure 6.19. Same as Figure 6.16 but for the JJA deviations.
Figure 6.20. Same as Figure 6.16 but for the SON deviations.
Tables

Table 5.1. Basin integrated linear trends in runoff that are simulated from Agro-IBIS control run. Trends that are calculated based on seasonal and annual time step are included in the table. Units are % decade$^{-1}$. P-values for each trend are shown in parentheses.

<table>
<thead>
<tr>
<th>Seasons RUNOFF</th>
<th>Missouri River Basin</th>
<th>Upper Mississippi River Basin</th>
<th>Ohio River Basin</th>
<th>Great Lakes Basin</th>
</tr>
</thead>
<tbody>
<tr>
<td>DJF</td>
<td>-12.1 (0.50)</td>
<td>-20.4 (0.09)</td>
<td>8.1 (0.44)</td>
<td>4.9 (0.82)</td>
</tr>
<tr>
<td>MAM</td>
<td>-2.6 (0.90)</td>
<td>4.9 (0.75)</td>
<td>-1.4 (0.86)</td>
<td>-1.1 (0.79)</td>
</tr>
<tr>
<td>JJA</td>
<td>5.9 (0.64)</td>
<td>5.2 (0.60)</td>
<td>13.5 (0.36)</td>
<td>0.0 (0.98)</td>
</tr>
<tr>
<td>SON</td>
<td>0.0 (0.90)</td>
<td>-4.1 (0.71)</td>
<td>11.4 (0.39)</td>
<td>-18.1 (0.22)</td>
</tr>
<tr>
<td>JAS</td>
<td>8.5 (0.60)</td>
<td>7.7 (0.36)</td>
<td>14.7 (0.41)</td>
<td>-1.7 (0.79)</td>
</tr>
<tr>
<td>Annual</td>
<td>-0.0 (0.98)</td>
<td>1.2 (0.98)</td>
<td>4.0 (0.64)</td>
<td>-1.9 (0.54)</td>
</tr>
</tbody>
</table>
Table 5.2. Same as Table 5.1. but for trends that are calculated based on USGS computed runoff data.

<table>
<thead>
<tr>
<th>Seasons</th>
<th>Missouri River Basin</th>
<th>Upper Mississippi River Basin</th>
<th>Ohio River Basin</th>
<th>Great Lakes Basin</th>
</tr>
</thead>
<tbody>
<tr>
<td>DJF</td>
<td>-23.3 (0.13)</td>
<td>-16.6 (0.06)</td>
<td>-1.9 (0.86)</td>
<td>-1.7 (0.82)</td>
</tr>
<tr>
<td>MAM</td>
<td>-19.8 (0.12)</td>
<td>-6.8 (0.54)</td>
<td>-0.6 (0.94)</td>
<td>-0.3 (0.94)</td>
</tr>
<tr>
<td>JJA</td>
<td>-10.9 (0.32)</td>
<td>2.4 (0.86)</td>
<td>2.8 (0.75)</td>
<td>0.3 (0.98)</td>
</tr>
<tr>
<td>SON</td>
<td>-24.4 (0.04)</td>
<td>-16.5 (0.14)</td>
<td>-0.4 (0.98)</td>
<td>-10.7 (0.17)</td>
</tr>
<tr>
<td>JAS</td>
<td>-13.7 (0.16)</td>
<td>-5.1 (0.60)</td>
<td>1.3 (0.94)</td>
<td>-2.9 (0.78)</td>
</tr>
<tr>
<td>Annual</td>
<td>-21.3 (0.07)</td>
<td>-10.0 (0.31)</td>
<td>0.0 (0.98)</td>
<td>-2.4 (0.64)</td>
</tr>
</tbody>
</table>
Table 6.1. List of perturbed simulations. The notation "d" means "simulation with imposed perturbation", while the acronym right after it means corresponding variable.

<table>
<thead>
<tr>
<th>Order Number</th>
<th>Simulation</th>
<th>Perturbed Climate Variable</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>dPREC</td>
<td>Precipitation</td>
</tr>
<tr>
<td>2</td>
<td>dRADS</td>
<td>Solar radiation</td>
</tr>
<tr>
<td>3</td>
<td>dTEMP</td>
<td>Air temperature</td>
</tr>
<tr>
<td>4</td>
<td>dRH</td>
<td>Relative humidity</td>
</tr>
<tr>
<td>5</td>
<td>dWSPD</td>
<td>Wind speed</td>
</tr>
<tr>
<td>6</td>
<td>d(RADS+PREC+TEMP)</td>
<td>Solar radiation &amp; precipitation &amp; air temperature</td>
</tr>
</tbody>
</table>
Appendix A. Trends in maximum and minimum air temperatures from 1984-2007

Figure A.1. Spatial distributions of annual and seasonal maximum air temperature trends from 1984 to 2007 for (a) annual, (b) winter, (c) spring, (d) summer, and (e) autumn. Trends that are statistically significant at 90% confidence level are hatched. The stippled area is the study area with the dark line as the boundary.
Figure A.2. Spatial distributions of annual and seasonal minimum air temperature trends from 1984 to 2007 for (a) annual, (b) winter, (c) spring, (d) summer, and (e) autumn. Trends that are statistically significant at 90% confidence level are hatched. The stippled area is the study area with the dark line as the boundary.
Appendix B. Validation of the Agro-IBIS modeled longwave radiation.

Figure B.1. Scatter plots showing the Agro-IBIS modeled 24 years (1984-2007) longwave radiation against satellite observations (NASA/GEWEX Surface Radiation Budget, v.3.1). Each black dot represents a daily value at a specific location. The red dots are longwave radiation values in Nebraska.