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MODELING ANNUAL WATER BALANCE IN THE SEASONAL BUDYKO FRAMEWORK

by

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B.S. University of Tehran, 2010

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in the Department of Civil, Environmental, and Construction Engineering in the College of Engineering and Computer Sciences at the University of Central Florida Orlando, Florida

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ABSTRACT

In this thesis, the role of soil water storage change on the annual water balance is evaluated based on observations at a large number of watersheds located in a spectrum of climate regions, and an annual water balance model is developed at the seasonal scale based on Budyko hypothesis. The annual water storage change is quantified based on water balance closure given the available data of precipitation, runoff, and evaporation estimated from remote sensing data and meteorology reanalysis. The responses of annual runoff, evaporation, and storage change to the interannual variability of precipitation and potential evaporation are then analyzed. Both runoff and evaporation sensitivities to potential evaporation are higher under energy-limited conditions, but storage change seems to be more sensitive to potential evaporation under the conditions in which water and energy are balanced. Runoff sensitivity to precipitation is higher under energy-limited conditions; but both evaporation and storage change sensitivities to precipitation are higher under water-limited conditions. Therefore, under energy-limited conditions, most of precipitation variability is transferred to runoff variability; but under water-limited conditions, most of precipitation variability is transferred to storage change and some of precipitation variability is transferred to evaporation variability. The main finding of this part is that evaporation variability will be overestimated by assuming negligible storage change in annual water balance, particularly under water-limited conditions. Budyko framework which expresses partitioning of water supply at the mean annual scale, is adapted to be applicable in modeling water cycle in short terms i.e.,
seasonal and interannual scales. Seasonal aridity index is defined as the ratio of seasonal potential evaporation and the difference between precipitation and storage change. The seasonal water balance is modeled by using a Budyko-type curve with horizontal shifts which leads prediction of seasonal and annual storage changes and evaporation if precipitation, potential evaporation, and runoff data are available.
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CHAPTER 1: INTRODUCTION

1.1 Introduction

Water cycle consisting of precipitation, evaporation, runoff and storage change is one of the basic concepts in hydrology science. Although the concept of water cycle seems to be clear and simple enough to be understood, it has a convoluted and intricate nature due to unidentifiable bounds of the cycle per se and its components. Also the main global water cycle is an integration of many smaller cycles at different spatial scales which they have interactions as well. Integrated appreciation of hydrologic cycle at various temporal and spatial scales and understanding the controlling mechanisms on its components are one of the fundamental questions and research paths in the hydrology. The complicated processes affecting the partitioning of available water to runoff, evapotranspiration and storage make it difficult to estimate each component and predict the tempo-spatial related variability of them. In the past decades extensive efforts have been made to provide simple conceptual models which enable us to predict the hydrologic behavior of the catchment.

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1.2 Background and Problem Definition

Generally studies on watershed water balance are conducted on three different time resolutions ranging from long term studies to annual and then sub-annual scales mainly in monthly frames. Those studies working on long term trends of hydrologic cycle usually assume that the storage changes are negligible over the long period, thus focusing on partitioning of precipitation on evaporation and runoff only. Researches focusing on annual time scales, mostly make the rough assumption of neglecting storage changes to come up with quantification of runoff and evaporatranspiration or their interannual variability; nevertheless many of them consider the effects of storage changes in interpreting their results. This approach is mainly due to the fact that from the four main components of water cycle it is just precipitation and runoff that can be measured widely and with lower uncertainties, while the other two components are hard to be measured. Evapotranspiration is hard to be quantified because of the very continuous tempo-spatial nature of it. Evapotranspiration links the soil, vegetation and water so quantification becomes more difficult as one has to consider all the processes controlling vegetation transpirations as well as those which affect bare soil evaporation. Although there are several proposed methods for quantifying evaporation, they are all limited by their spatial and temporal scales validity as well as their bias. Storage change also is hard to be measured due to the complexity of the soil-related processes which store the water in the watershed. Also the spatially large volume of soil and water that storage is dealing with, makes any kind of direct measurement difficult and very costly. The third group of studies focusing on the balance of sub-annual water, have to consider the storage
changes; those methods used some rough estimations of storage changes such as using probe-wells or ground water levels. All these methods not only are costly to be implemented over large scales but also are biased because of uneven distribution of stored water over the entire catchment. So, the need to develop a systematic approach to quantitatively account the water storage change at the watershed scale seems to be of significant importance. More importantly, within this perspective, the responses of water cycle components to climate changes too would be reconsidered having in mind all those previous works that just dealt with evapotranspiration and runoff variability responses to climate changes.

At the next step the need to understand the main frame which controls the partitioning of available water into other components of hydrologic cycle is getting highlighted in mind. A parsimonious model which is capable of explaining the phenomena while maintaining the necessary simplicity of the model should be developed.

1.3 Research Questions

The followings are specific research objectives of this study:

I. Providing a systematic framework to evaluate storage changes at different timescales.

II. Evaluating the responses of water cycle component to climate changes at annual scale by considering the storage changes.
III. Studying the observed seasonal trend of water cycle components as a ground for a top-down approach to understand the behavior of water cycle components at catchment scale.

IV. Developing a parsimonious model capable of explaining the partitioning pattern of available water to other water cycle components at the seasonal scale and through which estimating annual storage change.

V. Model parameter identifiability in order to implement the model in ungagged watersheds.

1.4 Organization of the Thesis

The presented thesis has been broken down into 4 Chapters to represent the concepts of the indicated research tasks. Chapters 2 and 3 are intended to be two stand-alone journal papers and have been developed in a way to be ready to submit.

- Chapter 1: Introduction – This Chapter contains the Thesis abstract, identified gaps that motivated this research, the novel approach that this study adopts to address the identified gaps, and the overall objective and tasks defined and accomplished in this project.
- Chapter 2: Responses of annual runoff, evaporation and storage change to climate variability at the watershed scale – In this chapter the basic methodology and perspective to estimating storage changes as well as data sets are introduced. A general method for investigating the responses of water cycle components to
climate changes has been developed. Then a thorough discussion on the results is presented.

- Chapter 3: modeling annual water balance based on budyko hypothesis at the seasonal scale – This chapter is on extending the Budyko concept to seasonal scale. Then based on the observations, a parsimonious model developed in a modified Budyko framework which is capable to explain dynamic of water balance at seasonal scale and also is capable to model storage changes.

- Chapter 4: Conclusions and Future Work – General conclusion of the thesis as well as the gaps and shortcomings is explained. Future steps for further development of the presented framework are described.
CHAPTER 2: RESPONSES OF ANNUAL RUNOFF, EVAPORATION AND STORAGE CHANGE TO CLIMATE VARIABILITY AT THE WATERSHED SCALE

2.1 Introduction

In the long-term mean annual water balance at the watershed scale, mean annual change of water storage (\(\Delta S\)) is negligible and mean annual precipitation (\(\bar{P}\)) is partitioned into mean annual runoff (\(\bar{Q}\)) and evaporation (\(\bar{E}\)). Budyko [1958] postulated that the partitioning of precipitation, to first order, was determined by the competition between available water (\(\bar{P}\)) and available energy measured by potential evaporation (\(\bar{E}_P\)). Based on datasets from a large number of watersheds and the work of Schreiber [1904] and Ol’dekop [1911], Budyko [1974] proposed a relationship between mean annual evaporation ratio (\(\bar{E}/\bar{P}\)) and mean annual potential evaporation ratio or climate dryness index (\(\bar{E}_P/\bar{P}\)). Other functional forms of Budyko-type curves have been developed for assessing the long-term water balance [e.g., Pike, 1964; Fu, 1981; Choudhury, 1999; Zhang et al., 2001; Porporato et al., 2004; Yang et al., 2008; Gerrits et al., 2009]. Besides the climate dryness index, the effects of other variables on the mean annual water balance have been studied to explain the observed deviation from the Budyko curve, e.g., the competing effects of climate fluctuations and watershed storage capacity [Milly, 1994a and 1994b], rainfall seasonality and soil moisture capacity [Potter et al., 2005; Hickel and Zhang, 2006; Zhang et al., 2008], the relative infiltration capacity, relative soil water storage, and the watershed average slope [Yang et al., 2007],
climate seasonality, soil properties and topography [Yokoo et al., 2008], vegetation type [Zhang et al., 2001; Oudin et al., 2008], and vegetation dynamics [Donohue et al., 2007]. Particularly for water storage change, at the seasonal scale, the storage carryover has an impact on the mean annual water balance [Milly, 1994b; Zhang et al., 2008; Jothityangkoon and Sivapalan, 2009; Donohue et al., 2010]; at the mean annual scale, steady-state of water balance, i.e., $\overline{\Delta S} = 0$, is assumed in the Budyko framework [Donohue et al., 2007].

Water balance at the annual scale has also been studied in the literature. Ignoring the groundwater inflow and outflow, the annual water balance at the watershed scale is represented as:

$$P_i = Q_i + E_i + \Delta S_i$$

where $P_i$, $Q_i$, $E_i$, and $\Delta S_i$ are annual precipitation, runoff, evaporation, and water storage change during year $i$, respectively. The Budyko-type functions have been extended to study the relationship between annual evaporation ratio ($E_i/P_i$) and annual potential evaporation ratio ($E_{pi}/P_i$) [e.g., Yang et al., 2007; Zhang et al., 2008]. Potter and Zhang [2009] tested the relationship of $E_i/P_i$ and $E_{pi}/P_i$ with six functional forms of Budyko-type curves and one linear model, and found that rainfall seasonality was important in determining the functional forms. Jothityangkoon and Sivapalan [2009] examined the effects of intra-annual variability of rainfall (e.g., storminess and seasonality) on interannual variability of water balance through the simulation of annual runoff in three semi-arid watersheds. Due to the limitation of data availability on evaporation and storage change, the annual evaporation is usually computed by $E_i = P_i - Q_i$ assuming
steady-state conditions [e.g., Koster and Suarez, 1999; Potter and Zhang, 2009]. However, in several recent studies, the interannual water storage carryover has been found to be significant in some studied watersheds [Flerchinger and Cooley, 2000; Tomasella et al., 2008]. Milly and Dunne [2002] accounted for the interannual storage changes for 175 large basins worldwide, and found that the annual storage change effect was important in some basins. Zhang et al. [2008] found that Fu’s equation, one functional form of Budyko-type curves, performed poorly on estimating annual streamflow in some watersheds in Australia, explaining that it might be due to the impact of water storage changes which could not be neglected on the annual scale. Ohta et al. [2008] studied the water balance of the Siberian forest from 1998 to 2006 and found that the interannual variation of storage was even more significant than precipitation for this particular watershed. Donohue et al. [2010] studied the annual water balance in 221 watersheds in Australia, and found that the effect of non-steady state conditions were an important source of variation at the annual scale and needed to be accounted for.

One of the important questions in studying the annual water balance that should be addressed is the responses of components of water balance to climate variability. The sensitivity of runoff to climate has been discussed in many studies [e.g., Schaake, 1990; Fu et al., 2007; Harman et al., 2011]. Runoff sensitivity has been represented by precipitation elasticity of streamflow which is defined as the ratio of percentage change of runoff to percentage change of precipitation [Schaake, 1990]. Sankarasubramanian et al. [2001] proposed a nonparametric estimation of sensitivity of streamflow to rainfall directly from historical data, and Sankarasubramanian and Vogel [2003] documented the
precipitation elasticity of streamflow for 1337 basins in the United States (US). Some hydrologic climate sensitivity studies involve calibrating a conceptual hydrologic model, and then varying the model’s atmospheric inputs, to observe the resulting changes in streamflow [e.g., Vogel et al., 1999; Chiew, 2006]. Budyko framework has been successfully used as an equilibrium interpretation of the climate elasticity of streamflow [e.g., Dooge, 1992; Dooge et al., 1999; Arora, 2002; Yang and Yang, 2011]. Recently, Roderick and Farquhar [2011] developed an analytical framework for determining runoff sensitivities to precipitation, potential evaporation, and catchment properties based on Budyko curves, and the framework has been successfully applied to the Murray Darling Basin in Australia with an emphasis on the spatial variation [Donohue et al., 2011].

In order to represent the response of interannual evaporation to precipitation variability, Koster and Suarez [1999] proposed the evaporation deviation ratio denoted as $\sigma_E/\sigma_P$ which is defined as the ratio of standard deviations between evaporation and precipitation. Following this, Sankarasubramanian and Vogel [2002] defined the runoff deviation ($\sigma_Q/\sigma_P$) as the ratio of standard deviations between runoff and precipitation. Koster and Suarez [1999] derived a powerful equation, which is a function of $\bar{E}_P/\bar{P}$, to estimate the evaporation deviation ratio by assuming (1) Budyko-type curves can be applicable for interannual water balance; (2) interannual changes in water storage are much smaller than the annual precipitation, evaporation and runoff; and (3) interannual variability on potential evaporation is negligible.

In this study, the role of annual water storage carryover in annual water balance is investigated directly in a systematic framework based on the data availability. The
interannual variability of water storage is quantified by utilizing the long-term observations of precipitation, runoff, and evaporation estimation from remotely sensed data. The purposes of this research are to test whether interannual storage change is negligible and to explore the behavior of annual storage carryover from energy-limited to water-limited conditions. Meanwhile, sensitivities of annual runoff, evaporation, and storage change to interannual climate variability including precipitation and potential evaporation will be quantified.

2.2 Methodology

2.2.1 Study Watersheds and Datasets

The study watersheds are obtained from the international Model Parameter Estimation Experiment (MOPEX) dataset which is described by Duan et al. [2006] and is downloaded from ftp://hydrology.nws.noaa.gov/. The dataset includes mean areal precipitation, climatologic potential evaporation, streamflow, and maximum and minimum air temperature for 432 watersheds with an adequate number of precipitation gages. Several recent studies have been based on the MOPEX watersheds [e.g., Sivapalan et al., 2011; Voepel et al., 2011; Wang and Hejazi, 2011]. The available precipitation and streamflow data during the period of 1948-2003 will be used in this study. In this paper, 277 watersheds, in which there is no missing data in any single day, are selected for analysis, and the daily variables are aggregated into annual values. Figure 1 shows the spatial distribution of the selected watersheds and the land use/land cover (LULC) within them. The LULC includes forest land, woodland, shrub land,
grassland, cropland, and urban and built-up. In the North Eastern US watersheds, forest is dominant; in the southern watersheds, grassland and shrub land are dominant; in the Mid-West, cropland is dominant. Summarized over the total area of all the selected watersheds, 26% is covered by forest, 29% by croplands, 33% by woodlands and wooded grasslands, 9% by grassland, 2% by shrub land, and less than 1% by bare ground and urban. The area of the study watersheds ranges from 67 km\(^2\) to 10329 km\(^2\) and the watersheds cover a wide spectrum of climate regions with \(\frac{\bar{E}_D}{\bar{E}_P}\) ranging from 0.24 to 3.84.

Figure 2.1. Spatial distribution of the 277 study watersheds and their associated land use/land cover

The actual daily evaporation, which is estimated from remote sensing data and meteorology reanalysis, is obtained from the University of Montana (UM) [Zhang et al.,
The evaporation data is grid-based with a spatial resolution of 8 km and is available from 1983 to 2006. In this dataset, the canopy transpiration, soil evaporation and open water evaporation are quantified using a modified Penman-Monteith approach coupled with a normalized difference vegetation index (NDVI) based biome-specific canopy conductance model [Zhang et al., 2010]. This dataset has been used to detect the trend of global land evaporation [Jung et al., 2010].

Monthly potential evaporation data are also provided by the University of Montana estimated using Priestley-Taylor method with the same extent and resolution as the actual evaporation data [Zhang et al., 2010]. The MOPEX dataset also provides mean annual potential evaporation estimation which is consistent with rainfall and runoff data. Since the remote sensing monthly potential evaporation data provides the interannual variability of potential evaporation, the monthly values are scaled by comparing the computed mean annual potential evaporation from the two datasets for each watershed. Then the scaled monthly potential evaporation is aggregated to annual potential evaporation which is used for the analysis.

In this study, the annual water balance analysis is conducted during 1983-2003 when precipitation, runoff, evaporation, and potential evaporation data are available. Figure 2 shows $\frac{E}{P}$ versus $\frac{EP}{P}$ of the watersheds in the Budyko framework where $\bar{E}$ is computed based on the estimated evaporation from remote sensing data. Generally, the data points follow the Budyko curve with some extent of scatters.
2.2.2 Estimation of Interannual Water Storage Change

It is a challenge to provide techniques and methods to quantify integrated storages at the watershed scale [Beven, 2006]. The terrestrial water storage change estimates from the Gravity Recovery and Climate Experiment (GRACE) satellite are spatially averaged over regions having areas of 1,000,000 km$^2$ and greater [Swenson et al., 2006]. However, the observational data of watershed water storage or dynamic storage change are usually not available at the watershed scale. In some studies, storage changes are estimated from localized measurements of piezometer wells and soil moisture probe [Wang, 2012], but
the estimation is strongly dependent on spatial heterogeneity of subsurface properties [Kirchner, 2009]. The dynamic storage change can also be estimated as the residual by water balance [Sayama et al., 2011], but the method is constrained by the data availability and uncertainty on observations or estimations especially evaporation.

In this paper, storage changes are estimated as the residual of water balance closure. The daily precipitation and runoff from MOPEX dataset, the daily evaporation from Zhang et al. [2010], and the scaled monthly potential evaporation are aggregated into annual values for each watershed. Given the annual precipitation, runoff and evaporation, the interannual water storage change can be estimated using equation (1), i.e., \( \Delta S_i = P_i - Q_i - E_i \).

2.2.3 Indicators for Interannual Variability of Water Balance

For a given year, the departure of an annual quantity from its mean annual value is called the annual anomaly. For example, \( \bar{P}_i = P_i - \bar{P} \) is the precipitation anomaly at year \( i \). Similarly, annual potential evaporation anomaly, runoff anomaly, and evaporation anomaly are denoted as \( \bar{E}_{Pi} \), \( \bar{Q}_i \), and \( \bar{E}_i \), respectively. Positive anomaly represents the value in a particular year is higher than that in a normal year. It is reasonable to assume negligible mean annual storage change (\( \bar{\Delta S} \)) since the number of years is large enough (i.e., 21 year in this study) and if there is no significant trend in groundwater table in the case study watersheds. One can obtain,

\[
\bar{P}_i = \bar{Q}_i + \bar{E}_i + \Delta S_i \tag{2}
\]
which can be interpreted as that precipitation anomaly is partitioned into runoff anomaly, evaporation anomaly, and storage change which are watershed responses to interannual climate variability. This partitioning is controlled by watershed properties (such as soil and vegetation) and human activities.

The standard deviations for annual precipitation, runoff, evaporation, and storage changes are computed for each watershed based on the 21-year data. Following Koster and Suarez [1999], the storage change deviation ratio is defined as $\sigma_{\Delta S}/\sigma_P$. Three standard deviations ratios ($\sigma_Q/\sigma_P$, $\sigma_E/\sigma_P$, and $\sigma_{\Delta S}/\sigma_P$) and their total values are then computed.

### 2.2.4 Sensitivity of Annual Runoff to Climate Variability

Following Roderick and Farquhar [2011] and ignoring changes of watershed properties, the streamflow variability responding to rainfall and potential evaporation changes can be expressed as

$$dQ = \frac{\partial Q}{\partial E_P} dE_P + \frac{\partial Q}{\partial P} dP$$

where $\frac{\partial Q}{\partial E_P}$ and $\frac{\partial Q}{\partial P}$ are sensitivity coefficients of runoff to potential evaporation and precipitation, respectively. The sensitivity coefficients can be estimated through annual anomaly values:

$$\tilde{Q}_l = a\tilde{E}_{Pl} + b\tilde{P}_l$$

(3)

Where $a = \frac{\partial Q}{\partial E_P}$ and $b = \frac{\partial Q}{\partial P}$. The values of $a$ and $b$ are estimated by linear regression:

$$\frac{Q_l}{P_l} = a\frac{E_{Pl}}{P_l} + b$$

(4)

Since equation (4) is applicable for individual year, by taking average of equation (4) over all the years, one can obtain the equation for mean annual values, i.e.,

$$\frac{\bar{Q}}{\bar{P}} = a\frac{\bar{E}}{\bar{P}} + b.$$
Subtracting its mean annual form from equation (4), one obtains equation (3). Therefore, the values of $a$ and $b$ are calculated for each watershed by linear regression on annual data of $\frac{Q_i}{P_i}$ and $\frac{E_{Pl}}{P_i}$ during 1983-2003. The reasons that the values of $a$ and $b$ are estimated by linear regression through equation (4) instead of multilinear regression through equation (3) are discussed in the next section.

Figure 3 shows the runoff deviation ratio computed for the selected watersheds used in this study. The mean potential evaporation ratio is a major control on the runoff deviation ratio. Other factors, such as soil storage capacity, can also affect the runoff variability. Generally the data points in Figure 3 match the theoretical line derived by Koster and Suarez [1999] and the data cloud is similar with the watersheds presented by Sankarasubramanian and Vogel [2002], i.e., $\sigma_Q/\sigma_p$ decreases with the increase of $E_p/\bar{P}$. Some data points are above 1 which is the upper bound of the theoretical line.
2.2.5 Sensitivity of Annual Evaporation to Climate Variability

The interannual variability of water balance is important for understanding the response of evaporation to a changing environment (e.g., precipitation and land use changes). Similar to runoff sensitivity, the sensitivity of annual evaporation to interannual variability of potential evaporation and precipitation is expressed as

\[ dE = \frac{\partial E}{\partial E_p} dE_p + \frac{\partial E}{\partial P} dP \]

where \( \frac{\partial E}{\partial E_p} \) and \( \frac{\partial E}{\partial P} \) are sensitivity coefficients of annual evaporation to potential evaporation and precipitation, respectively. The sensitivity coefficients can be estimated through annual anomaly values:

\[ \bar{E}_i = \alpha \bar{E}_{p_i} + \beta \bar{P}_i \]  

(5)

Where \( \alpha = \frac{\partial E}{\partial E_p} \) and \( \beta = \frac{\partial E}{\partial P} \). Similarly to runoff sensitivity, the values of \( \alpha \) and \( \beta \) are estimated by linear regression:

\[ \frac{E_l}{P_l} = \alpha \frac{E_{p_l}}{P_l} + \beta \]  

(6)

Since this linear function holds for individual year \( i \), it is also applicable to the mean annual condition, i.e., \( \frac{\bar{E}}{\bar{P}} = \alpha \frac{\bar{E}_p}{\bar{P}} + \beta \). Combining the specific year \( i \) and the year with mean values, we can obtain equation (5). The coefficients \( \alpha \) and \( \beta \) are estimated by linear regression through the ratio model represented in equation (6) following the work by Cheng et al. [2011] who analyzed the relationship between \( \frac{E_l}{P_l} \) and \( \frac{E_{p_l}}{P_l} \) over 500 watersheds in US and found that a strong linear relationship exists and discussed the controlling factors on the linear relationship such as climate, soil water storage, vegetation, and human activities. The ratios in equation (6) can usually cancel out some
covariant factors between components and represent a more generalized relationship, and the linearity of component model (equation (5)) is not as strong as the ratio model [Cheng et al., 2011]. For consistency, the ratio model of equation (4) is also used for estimating the coefficients of runoff sensitivities.

2.2.6 Sensitivity of Annual Storage Change to Climate Variability

In this study, the sensitivity of interannual storage change to the climate variability is also investigated following equations (3) and (5). The sensitivity of annual storage change to potential evaporation and precipitation is expressed as:

\[ \Delta S_i = \gamma \hat{E}_{pl} + \varphi \hat{P}_i \quad (7) \]

where \( \gamma \) and \( \varphi \) are sensitivity coefficients of annual storage change to potential evaporation anomaly and precipitation anomaly, respectively. Substituting equations (3) and (5) into equation (2) and comparing it with equation (7), one can obtain the relationship among the sensitivity coefficients of runoff, evaporation, and storage change to potential evaporation and precipitation anomalies:

\[ \begin{align*}
\gamma &= -(a + \alpha) \\
\varphi &= 1 - b - \beta 
\end{align*} \quad (8) \]

If annual storage change is negligible, the sensitivity coefficients of runoff and evaporation to potential evaporation are opposite \((a = -\alpha)\) and the sensitivity coefficients of runoff and evaporation to precipitation are complementary \((b + \beta = 1)\) [Roderick and Farquhar, 2011]. The values of \( \gamma \) and \( \varphi \) for the study watersheds are computed by equation (8) since the values of \( a, b, \alpha, \) and \( \beta \) have been calculated. The linear regression of \( \frac{\Delta S_i}{P_t} = \hat{\gamma} \frac{E_{pl}}{P_t} + \hat{\varphi} \) can be conducted and the obtained values of \( \hat{\gamma} \) and \( \hat{\varphi} \)
\( \hat{\phi} \) are the same as values of corresponding \( \gamma \) and \( \varphi \) obtained by equation (8), i.e., \( \gamma = \hat{\gamma} \) and \( \varphi = \hat{\varphi} \).

In the following section, the six sensitivity coefficients (i.e., \( a, b, \alpha, \beta, \gamma \) and \( \varphi \)) for the study watersheds are presented and the climate control on the sensitivity coefficients is discussed.

2.3 Results and Discussions

2.3.1 Interannual Variability of Runoff

Figure 4 shows the values of runoff sensitivity coefficients (\( a \) and \( b \)) as a function of mean annual climate dryness index. Streamflow is positively related to precipitation (\( b > 0 \)) but negatively related to potential evaporation (\( a < 0 \)). The trends of \( a \) and \( b \) shown in Figure 4 are consistent with the findings by Milly and Dunne [2002] who reported that as evaporation ratio increases from 0 to 1, the runoff sensitivity to surface net radiation increases from -1 to 0, and that the runoff sensitivity to precipitation decreases from 1 to 0. With the increase of climate dryness index, both values of \( a \) and \( b \) approach zero. As discussed in Donohue et al. [2011], under energy-limited conditions, the sensitivity of runoff to precipitation and potential evaporation are both high. In Figure 4a, there are some outliers where the values of \( a \) are positive, and it is found that these watersheds are snow/glacier-melt dominated systems in the state of Washington. Runoff is positively correlated with temperature and this induces positive sensitivity coefficient of annual runoff to potential evaporation. Even though the values of \( a \) and \( b \) are fixed for a specified watershed, the climate elasticity of streamflow is not fixed and is
dependent on the ratio of potential evaporation anomaly and precipitation anomaly [Donohue et al., 2011]. From Figure 4b, the values of b match the runoff deviation ratio shown in Figure 3 because the variation of potential evaporation is much smaller than that of precipitation variation [Koster and Suarez, 1999]. The interannual runoff variability is mainly driven by the precipitation variability.

Figure 2.4. The estimated sensitivity coefficients of runoff to (a) potential evaporation and (b) precipitation as a function of mean annual climate dryness index
2.3.2 Interannual Variability of Evaporation

Since the interannual variability of evaporation in this study depends on the annual evaporation estimation from satellite remote sensing-based algorithm, the uncertainty or bias of the evaporation dataset is discussed first. The evaporation dataset includes two potential uncertainty sources which are measurements of tower eddy flux and satellite-based NDVI [Zhang et al., 2010]. Comprehensive validation of the dataset has been conducted by Zhang et al. [2010] who used flux tower evaporation measurements at the daily and monthly time scales. Zhang et al. [2010] also verified the dataset at 261 major global basins by comparing it with evaporation estimation inferred from the long-term water balance and found that the two evaporation estimations are similar with root mean square error (RMSE) of 186.3 mm/year. Particularly, the relative difference between the two evaporation estimations in most regions of US is within ±10%. The detailed uncertainty analysis on the daily and monthly evaporation estimation is referred to Zhang et al. [2010].

The uncertainty of remote sensing-based evaporation estimation for the MOPEX watersheds is assessed in this study. The average annual evaporation for each watershed is computed by aggregating the daily evaporation estimation from remote sensing, and denoted as $\overline{E}_{RS}$. Based on the daily precipitation and runoff data obtained from the MOPEX dataset, the average annual evaporation is estimated by water balance assuming negligible storage change, $\overline{E}_{WB} = \overline{P} - \overline{Q}$. Figure 5a is the scatter plot of the relationship between $\overline{E}_{WB}$ and $\overline{E}_{RS}$. The RMSE and $R^2$ are computed as 91.7 mm/year and 0.65, respectively. The average difference of $\overline{E}_{RS}$ and $\overline{E}_{WB}$ over the study watersheds is
approximately -30.3 mm/year. Figure 5b shows the percent difference between $\bar{E}_{RS}$ and $\bar{E}_{WB}$ which is defined as $\frac{\bar{E}_{RS} - \bar{E}_{WB}}{\bar{E}_{WB}} \times 100$. Among the 277 study watersheds, the percent differences of 243 watersheds (i.e., 88% of the study watersheds) are within the range of ±20% and the percent differences of 158 watersheds (i.e., 57% of the study watersheds) are within the range of ±10%. It should be noted that the uncertainty of precipitation and runoff data can contribute the uncertainty of $\bar{E}_{WB}$.

Figure 2.5. Comparison of mean annual evaporation computed by (a) water balance ($\bar{E}_{WB}$) and (b) estimated from remote sensing-based method ($\bar{E}_{RS}$)
The generally favorable agreement between the two mean annual evaporation estimations in the most study watersheds provides some supports for the accuracy of the remotely-sensed evaporation estimation and a basis for further interannual variability analysis using remote-sensing based daily evaporation estimation.

To explore the interannual variability of evaporation at each watershed, the evaporation deviation ratio is computed for two cases: (1) annual evaporation is computed by the difference between annual precipitation and runoff, i.e., \( E_l = P_l - Q_l \); (2) annual evaporation is from the remote sensing data. Evaporation deviation ratios from the two cases are shown in Figure 6a. The theoretical line derived by Koster and Suarez [1999], who presented the evaporation deviation ratio from a 20-yr GCM simulation at the spatial scale of 4° x 5°, is located at the lower envelope of the data points (grey dots) at case (1). When evaporation is estimated from remote sensing independently, the evaporation deviation ratio (red circle) decreases significantly compared with that in case (1). For watersheds with large climate dryness index, the data points are much further below the theoretical line with assumptions discussed earlier. Therefore, the interannual storage carryover in these watersheds mitigates the evaporation variability to climate variability. The interannual storage change from GCM may be much smaller due to the large spatial scale (i.e., ~400 km x 500 km) in the study of Koster and Suarez [1999]. To assess the impact of evaporation estimation uncertainty, Figure 6b plots evaporation deviation ratios for the 158 watersheds with percent difference within ±10%, and the same trend can be observed.
Figure 2.6. Evaporation deviation ratio versus mean annual climate dryness index: (a) all the 277 watersheds; (b) 158 watersheds with the percent differences within ±10%.
Therefore, if interannual storage carryover was ignored, evaporation deviation ratio would be overestimated, especially for watersheds in water-limited regions. It has been documented in the literature that interannual variability of evaporation in undisturbed watersheds is reduced by vegetation responses to climate variability [Jones, 2011]. For example, although drought in temperate deciduous forests decreases transpiration rates of many species, total evaporation is often reported to exhibit less interannual variability than precipitation [Wullschleger and Hanson, 2006]. Oishi et al. [2010] studied the transpiration of a mature oak-hickory forest in North Carolina and found that despite the large interannual variation in precipitation (ranging from 934 to 1346 mm), annual evaporation varied much less (610–668 mm). Therefore, interannual evaporation variability can be mitigated by vegetation responses through decreasing storage during drought period.

In semi-arid ecosystems, trees have developed adaptive mechanisms that buffer themselves from the year-to-year variations in precipitation and maintained the evaporation level [Raz-Yaseef et al., 2010]. For a grassland watershed in the Mediterranean climate zone of California, Ryu et al. [2008] reported that annual evaporation ranged a little despite a two-fold range in precipitation. As they found, in water-limited seasons, most evaporation was regulated by stomatal closure; in wet season high rainfall did not lead to high evaporation because of the marginal available energy. Therefore annual evaporation is not sensitive to annual precipitation. In agricultural watersheds particularly water-limited regions, human interferences can contribute to the smaller variability of evaporation. Cheng et al. [2011] found that watersheds with higher
agricultural land coverage generally have a stronger linear relationship between annual evaporation ratio and potential evaporation ratio. The sensitivity coefficients of evaporation to potential evaporation ($\alpha$) and precipitation ($\beta$) in equation (5) are estimated for each watershed. Figure 7 shows the histograms of ($a+\alpha$) and ($b+\beta$). The values of ($a+\alpha$) are located between 0 and 1 and the average value is 0.34; the values of ($b+\beta$) are located between 0.1 and 1 and the average value is 0.65. Therefore, $a = -\alpha$ and $b + \beta = 1$ are not held statistically and annual storage change may mitigate evaporation variability.

![Histograms of (a) ($a+\alpha$) and (b) ($b+\beta$)](image)

Figure 2.7. Histograms of (a) ($a+\alpha$) and (b) ($b+\beta$)
The sensitivity coefficients are plotted versus mean annual climate dryness index. As shown in Figure 8a, in humid regions where energy is limited, evaporation anomaly is highly sensitive to the potential evaporation anomaly and the value of $\alpha$ approaches to 0.8; but in arid regions where energy supply ($\bar{E}_p$) is larger than water supply ($\bar{P}$), potential evaporation anomaly is not the controlling factor on the interannual evaporation variability and the value of $\alpha$ approaches to 0.1. The sensitivity of evaporation anomaly to precipitation anomaly is complex due to the correlation between precipitation and potential evaporation (Figure 8b). In humid regions where $\bar{E}_p<\bar{P}$, higher precipitation induces lower potential evaporation, therefore the evaporation anomaly can be negative in wet years, i.e., $\beta<0$; in arid regions, $\beta$ is positive since precipitation is limited. The values of $\beta$ in most watersheds are bounded between 0 in humid regions and 0.15 in arid regions and do not approach -1 or 1. This is consistent with the discussion above that interannual variability of evaporation is much lower than that of precipitation. The sensitivity coefficients obtained have the same trend as found by Cheng et al. [2011]
2.3.3 Interannual Variability of Water Storage Change

Annual storage change ($\Delta S_t$) in each year is computed by equation (1), and the standard deviation of annual storage change is calculated over the 21-year data. It should be noted that uncertainty of estimated annual storage change depends on the uncertainties...
of precipitation, runoff, and evaporation estimates. Figure 9 shows the storage change deviation ratio \( \sigma_{\Delta S}/\sigma_P \) versus climate dryness index. The values of storage change deviation ratio are bounded between 0.2 and 1. The scattering of \( \sigma_{\Delta S}/\sigma_P \) is significant in humid regions with \( \bar{E}_P/\bar{P} \) between 0.5 and 1.0, but generally the storage change deviation ratio increases with climate dryness index. For watersheds in humid regions, the soil is wet and the interannual soil water storage carryover is less sensitive to the precipitation anomaly compared with the watersheds in arid regions. The soil moisture and groundwater table fluctuation in arid regions is more sensitive to the precipitation variability at the annual scale. Comparing Figure 9 and Figure 6, in energy-limited watersheds, the deviation ratio of storage change is about twice that of evaporation, and in water-limited watersheds storage variability can be 3 times that of evaporation variability. Therefore, the interannual variability of storage change is more significant than that of evaporation.

![Figure 2.9. Storage change deviation ratio versus climate dryness index](image)

Figure 2.9. Storage change deviation ratio versus climate dryness index
The sensitivity coefficients of interannual storage change to potential evaporation ($\gamma$) are plotted versus climate dryness index (Figure 10). Figure 10a plots the values of $\gamma$ for all the 277 watersheds. For most watersheds, the value of $\gamma$ is between 0 and -0.6. The negative value of $\gamma$ represents water storage deceases with the increase of $E_p$. In humid regions, $\gamma$ approaches to 0 with decreasing $E_p/P$; while in the arid regions, $\gamma$ approaches to -0.1 with increasing $E_p/P$. To explore the impact of uncertainty in the evaporation dataset, Figure 10b only plots the values of $\gamma$ for the 158 watersheds with percent difference within $\pm 10\%$ shown in Figure 5b. It seems that the trend of the data clouds in Figure 10 is not monotonic, i.e., $\gamma$ decreases then increase with climate dryness index.

The sensitivity coefficients of interannual storage change to precipitation ($\phi$) are plotted in Figure 11 under different climate regions. For most watersheds, the value of $\phi$ ranges from 0 to 0.85 and increases with climate dryness index. The positive value of $\phi$ represents water storage increases with precipitation. The water storage change in the humid regions is less sensitive to precipitation than that in the arid regions. In arid regions, the value of $\phi$ approaches to 0.85, and most of the precipitation variability is transferred to storage variability.
Figure 2.10. The estimated sensitivity coefficients of storage change to potential evaporation anomaly ($\gamma$): (a) all the 277 watersheds; (b) watersheds with the percent differences of mean annual evaporation within $\pm 10\%$. 
2.3.4 Partitioning of Precipitation into Runoff, Evaporation, and Storage Change

Table 1 summarizes the sensitivity coefficients for runoff, evaporation, and storage change to potential evaporation and precipitation under energy-limited and water-limited conditions. With the increase of climate dryness index, the sensitivity of runoff and evaporation to potential evaporation increases from -0.8 to 0 (\(\alpha\) as shown in Figure 4a), decreases from 0.8 to 0.1 (\(\alpha\) as shown in Figure 8a), respectively. According to equation (8), the value of \(\gamma\) should be 0 under extreme humid conditions and -0.1 under the extreme arid conditions. Therefore, considering the data points in Figure 10, it seems that the minimum value of \(\gamma\) occurs when energy and water are balanced, i.e., \(E_p/P = 1\). When \(E_p/P <1\), with the decrease of \(E_p/P\) soil becomes wetter and soil water storage approaches to the soil water capacity, therefore the storage change sensitivity to potential evaporation approach 0 when \(E_p/P\) approaches to 0; When \(E_p/P <1\), energy is limited and the storage change sensitivity to potential evaporation approaches 0 when \(E_p/P\) is large. Both runoff and evaporation is more sensitive to potential evaporation under energy-limited conditions, but storage change is more sensitive to potential evaporation under the conditions where water and energy are balanced.

With the increase of climate dryness index, the sensitivity of runoff and evaporation to precipitation decreases from 1 to 0 (Figure 4b) and increases from 0 to 0.15 (Figure 8b), respectively.
Table 2.1. The sensitivity coefficients at the lower and upper bounds of climate dryness index in the study watersheds

<table>
<thead>
<tr>
<th>Variables</th>
<th>$\Delta E_{pi}$</th>
<th>$\Delta P$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Sensitivity coefficient</td>
<td>Humid $\rightarrow$ Arid</td>
</tr>
<tr>
<td>$\Delta Q_i$</td>
<td>$a$</td>
<td>$-0.8 \rightarrow 0$</td>
</tr>
<tr>
<td>$\Delta E_i$</td>
<td>$\alpha$</td>
<td>$0.8 \rightarrow 0.1$</td>
</tr>
<tr>
<td>$\Delta S_i$</td>
<td>$\gamma$</td>
<td>$-1 \rightarrow 0$</td>
</tr>
</tbody>
</table>

Sensitivity of storage change to precipitation ($\phi$) increases from 0 to 0.85 as shown in Figure 11.

Figure 2.11. The sensitivity coefficients of storage change to precipitation anomaly ($\phi$) versus mean annual climate dryness index.
Therefore, runoff is more sensitive to precipitation under energy-limited conditions; but both evaporation and storage change is more sensitive to precipitation under water-limited conditions. Under energy-limited conditions, most of precipitation anomaly is transferred to runoff anomaly; but under water-limited conditions, most of precipitation anomaly is transferred to storage change and some of precipitation anomaly is transferred to evaporation anomaly.

The total deviation ratio of runoff, evaporation, and storage change are plotted for all the watersheds (Figure 12a) and the 158 watersheds with percent differences of $\bar{E}$ within $\pm 10\%$ (Figure 12b). Considering interannual storage change, the variance of rainfall can be written as $\sigma_p^2 = \sigma_Q^2 + \sigma_E^2 + \sigma_{\Delta S}^2 + 2\rho_{Q,E}\sigma_Q\sigma_E + 2\rho_{Q,\Delta S}\sigma_Q\sigma_{\Delta S} + 2\rho_{E,\Delta S}\sigma_E\sigma_{\Delta S}$ where $\rho_{Q,E}$, $\rho_{Q,\Delta S}$, and $\rho_{E,\Delta S}$ are the correlation coefficients between the corresponding variables. Since $\rho_{Q,E}$, $\rho_{Q,\Delta S}$, and $\rho_{E,\Delta S}$ are less than 1, $\sigma_p^2 < (\sigma_Q + \sigma_E + \sigma_{\Delta S})^2$. Therefore, the total deviation ratio is greater than one, i.e., $$(\sigma_Q + \sigma_E + \sigma_{\Delta S})/\sigma_p > 1,$$ as shown in Figure 12. The trend and upper bound of total deviation ratio in Figure 12a depends on the correlation coefficients between the variables. There is a clear decreasing trend in the total deviation ratio when climate dryness index is larger than 1. When climate dryness index is less than 1, it seems that there is increasing trend which may be due the uncertainty of evaporation and other data as shown in Figure 12b. From humid to arid regions, $\sigma_Q/\sigma_p$ decreases (Figure 3), $\sigma_E/\sigma_p$ increases (Figure 6), and $\sigma_{\Delta S}/\sigma_p$ (Figure 9) increases. The increasing trend of total deviation ratio in energy-limited regions is dominated by the increasing trend of evaporation and storage change.
deviation ratios. The decreasing trend of total deviation ratio in water-limited regions is dominated by the decreasing trend of the runoff deviation ratio.

Figure 2.12. The total deviation ratio of runoff, evaporation, and storage change: (a) all the 277 watersheds; (b) 158 watersheds with the percent differences within ±10%.
2.4 Summary and Conclusions

The annual soil water storage change is usually assumed to be negligible in interannual and mean annual water balance at the watershed scale. In this study, the role of interannual variability of soil water storage change in annual water balance is assessed for 277 watersheds located in a spectrum of climate regions. The annual water storage change is estimated based on water balance closure given the available data of precipitation, runoff, and evaporation estimated from remote sensing data and meteorology reanalysis. The partitioning of annual precipitation anomaly to runoff anomaly, evaporation anomaly, and storage change is studied. The interannual storage carryover in the study watersheds mitigates the evaporation variability caused by climate variability. The main finding is that evaporation variability is overestimated by assuming negligible storage change, and that storage change is the most sensitive component to precipitation under water-limited conditions.

The sensitivity coefficients of runoff, evaporation, and annual soil water storage change, to interannual variability of potential evaporation and precipitation are computed for the study watersheds, respectively. Both runoff and evaporation are more sensitive to potential evaporation under energy-limited conditions, but storage change is more sensitive to potential evaporation under the conditions where water and energy are balanced. Runoff is more sensitive to precipitation under energy-limited conditions; but both evaporation and storage change are more sensitive to precipitation under water-limited conditions. Under energy-limited conditions, most of precipitation anomaly is transferred to runoff anomaly; but under water-limited conditions, most of precipitation
anomaly is transferred to storage change and some of precipitation anomaly is transferred to evaporation anomaly.

The scattering of the sensitivity coefficients can be due to uncertainties of the datasets and other controlling factors such as seasonal distribution of precipitation, soil water storage capacity and vegetation coverage and types, etc. Also, since the evaporation data is obtained from a different source with the precipitation and runoff data, they may have some discrepancy. Further research is needed to verify the interannual variability of evaporation and storage change to climate variations using alternative datasets, and to explain the scattering of the sensitivity coefficients by investigating the roles of other controlling factors besides annual climate.
2.5 References


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CHAPTER 3: MODELING ANNUAL WATER BALANCE BASED ON BUDYKO HYPOTHESIS AT THE SEASONAL SCALE

3.1 Introduction

The physical controls of climate, vegetation, soil, and topography on the water balance at the mean annual and interannual scales have been an important research question in watershed hydrology [Budyko, 1974; Milly, 1994; Dooge et al., 1999; Koster and Suarez, 1999; Zhang et al., 2001; Milly and Dunne, 2002; Sankarasubramanian and Vogel, 2002; Potter et al., 2005; Yang et al., 2007; Yokoo et al., 2008; Zhang et al., 2008; Wang et al., 2009; Donohue et al., 2011; Roderick and Farquhar, 2011; Cheng et al., 2011; Zanardo et al., 2012; Wang, 2012]. A comprehensive understanding of water balance is challenging partly due to the fact that among the components of water balance, only precipitation and runoff may have reliable measurements available at the watershed scale. Evaporation and soil water storage data are usually not available. Evaporation at the watershed scale is controlled by complex factors such as atmospheric condition, vegetation, and water availability and the associated spatial variability. With the advancement of measurement technology, evaporation can be estimated by remote sensed data [e.g., Mu et al. 2007; Zhang et al. 2010]. Remote sensing observations can be utilized to study the variability of evaporation with high-temporal and spatial resolutions [Miralles et al. 2011].
The difficulties involved in measurement of water storage changes are due to the spatial variability of soil moisture and groundwater storage [Donohue et al., 2009; Peter and Aulenbach, 2011; McNamara et al., 2011]. Gravity Recovery and Climate Experiment (GRACE) mission is to monitor the terrestrial water storage (TSW) changes through the changes in Earth’s gravity field [Tapley et al. 2004; Strassberg et al. 2009]. However, the spatial coverage is as large as 400 km and the temporal resolution is on monthly basis. Therefore, GRACE measurements are useful for large scale soil water storage monitoring, but are not suitable for watershed scale studies. Water storage changes can also be estimated based on point-based observations of groundwater level and soil moisture, or water balance closure. Sayama et al. [2011] quantified storage changes of 17 watersheds in California based on water balance closure at the hourly scale. Billah and Goodall [2011] applied a water balance approach to analyze the interannual water storage variations during a drought period in South Carolina. Comparing storage change values with groundwater levels they highlighted water balance approach as a valuable method to study water cycle variability. Wang and Alimohammadi [2012] estimated water storage change as water balance residual using remote sensing-based evaporation estimations and investigated the interannual variability of water storage changes for 277 watersheds in the United States. Peters and Aulenbach [2011] analyzed water storage in a 41-ha watershed located in Georgia, and estimated annual storage changes using a streamflow recession relation coupled with a water balance closure. Wang [2012] estimated the interannual water storage changes of a watershed in Illinois using long-term observations of soil moisture and groundwater
levels. *Istanbulluglue et al.* [2012] estimated the soil water storage changes of four watersheds in Nebraska by groundwater level observations and examined the role of soil texture and baseflow contributions to water storage changes.

In many watersheds, the annual storage changes are found to be significant compared with other components (i.e., precipitation, evaporation, and runoff) thus having an important role in annual water balance [*Ohta et al.* 2008; *Donohue et al.* 2010]. The role of this component is even more significant in the seasonal water dynamics [*Feng et al.*, 2012]. Uncertainties and difficulties involved in measurements of water storage lead hydrologists to develop parsimonious but reliable models to estimate watershed storage changes at the interannual and intra-annual scales, which is useful for the purpose of water resources planning and management. A top-down analysis [*Sivapalan et al.*, 2003] seems to be a viable approach following the limit concept proposed by *Budyko* [1974]. For example, *Zhang et al.* [2008] extended Budyko framework from interannual to seasonal and monthly, and also included water storage capacity in their model to reflect the dynamics of water balance at shorter timescales. *Donohue et al.* [2007] suggested that vegetation dynamics and rooting depth needs to be incorporated into water balance models for timescales less than 1 year. *Yokoo et al.* [2008], in addition to aridity index, incorporated storage capacity index and drainability index to model water balance at the seasonal scale.

The roles of storage dynamics in water balance at the annual and seasonal scales are worth-while to be further investigated for watersheds located in different climate conditions. In this study, Budyko framework is extended to seasonal water balance, but
the climate aridity index and evaporation ratio are redefined at the seasonal scale following the concept of taking effective precipitation as water supply [Wang, 2012]. Budyko-type functions are then extended to modeling the partitioning of precipitation into runoff, evaporation and storage changes. Then the performance of the models is evaluated based on a large number of study watersheds. Given observations of precipitation, runoff, and potential evaporation, the extended Budyko-type models can be used to estimate seasonal evaporation and storage change which can be further aggregated into annual storage dynamics. The extended Budyko-type model at the seasonal scale is introduced in the following section.

3.2 Methodology

3.2.1 Data Sets

This study is based on the Model Parameter Estimation Experiment (MOPEX) watersheds which are highlighted as watersheds with low human impacts [Duan et al., 2006]. For the watersheds of this study (Figure 1) the values of daily precipitation, climatic potential evaporation, and runoff are available from 1948 to 2003. Data can be downloaded from ftp://hydrology.nws.noaa.gov/. For all watersheds the daily values of evaporation and monthly potential evaporation for 24 years starting 1983 to 2006 are obtained from the data set provided by University of Montana [Zhang et al, 2010]. The gridded remote sensing evaporation data sets are globally produced with a resolution of 8 km. The daily evaporation and monthly potential evaporation is spatially averaged to watershed scale values. The detailed description of data sets is available in Wang and
Alimohammadi [2012]. The study is conducted on the overlapped time span of the two data sets, i.e., from 1983 to 2003. A group of 277 watersheds has been selected for these studies which are characterized as watersheds with no missing data for the entire period of 21 years and these watersheds span over different climate regions.

Figure 3.1. 277 study watersheds in contiguous United States, and the groups of watersheds based on numbers of months in dry season.

Figure 2 plots the long-term mean evaporation ratio, which is computed as the ratio of remote sensing-based evaporation to precipitation, as a function of aridity index (ratio of potential evaporation to precipitation) for the watersheds in the Budyko frame work. Mostly, the study watersheds especially those located in wet regions follow the Budyko curve.
3.2.2 Seasonal Aridity Index

Annual soil water storage dynamics are controlled by climate, watershed properties, and human activities. The model developed in this paper follows Budyko hypothesis which expresses the partitioning of precipitation into evaporation and runoff based on the available energy and water supply at the mean annual scale. In the original Budyko framework, available energy is represented by potential evaporation and available water is represented by precipitation since long-term mean storage change is negligible. At the seasonal scale, water storage dynamics is significant as discussed earlier and needs to be considered for accounting available water. The available water supply in dry seasons includes not only precipitation but also the depletion of stored
water in the watershed, while in wet seasons watershed storage is replenished by infiltrated rainfall and increased storage needs to be subtracted from precipitation. Following Wang [2012], water availability, defined as effective precipitation, is computed as the difference between seasonal precipitation and storage change.

To account for the effect of storage change on water availability, the water balance analysis at the intra-annual scale is conducted based on the ratio of energy supply and effective precipitation. Given the data of monthly precipitation ($P_m$), potential evaporation($E_{pm}$), runoff ($Q_m$) and evaporation ($E_m$), the long-term mean monthly values of precipitation ($\bar{P}_m$), potential evaporation ($\bar{E}_{pm}$), runoff ($\bar{Q}_m$), and evaporation ($\bar{E}_m$), are computed for each month $m (=1, 2, \ldots, 12)$. Then the long-term mean monthly storage changes are estimated as a residual of the water balance closure:

$$\Delta S_m = \bar{P}_m - \bar{E}_m - \bar{Q}_m$$

(9)

The “observed” storage changes for individual month can be estimated based on the available data on precipitation, runoff, and evaporation, i.e., $\Delta S_m = P_m - E_m - Q_m$. Wet and dry months are identified based on long-term mean monthly aridity index which is defined as the ratio of monthly energy supply to monthly effective precipitation:

$$\bar{A}_m = \frac{\bar{E}_{pm}}{\bar{P}_m - \Delta S_m}$$

(10)

Months with $\bar{A}_m \geq 1$ are identified as dry months, and months with $\bar{A}_m < 1$ are indentified as wet months. Wet months are grouped into wet seasons and dry months are
grouped into dry seasons. Therefore, wet and dry seasons are identified according to the mean monthly aridity index.

For individual years, the monthly values are aggregated into seasonal values in wet and dry seasons, respectively. For example, precipitation in wet season \( (P_w) \) and dry season \( (P_d) \) is computed by:

\[
P_w = \sum_{i=1}^{n_w} m_i
\]

\[
P_d = \sum_{i=1}^{n_d} m_i
\]

where \( n_w \) and \( n_d \) are the number of wet and dry months in each year. Similarly, the seasonal values for potential evaporation \( (E_{P_w} \text{ and } E_{P_d}) \), runoff \( (Q_w \text{ and } Q_d) \), and storage changes \( (\Delta S_w \text{ and } \Delta S_d) \) are computed based on the monthly values. Following equation (10), the seasonal aridity indices for each individual year are defined as:

\[
A_w = \frac{E_{P_w}}{P_w - \Delta S_w}
\]

\[
A_d = \frac{E_{P_d}}{P_d - \Delta S_d}
\]

where \( A_w \) and \( A_d \) are the seasonal aridity indices for wet and dry seasons, respectively; and \( P_w - \Delta S_w \) and \( P_d - \Delta S_d \) are available water supply in the wet and dry season, respectively. Storage changes are considered in the defined seasonal aridity index.
3.2.3 Budyko Type Models at The Seasonal Scale

*Budyko* [1974] proposed a semi-empirical model, based on the observations at a large number of watersheds which explained the long-term pattern of water balance. To incorporate the effects of other factors on water balance, Budyko type models with a single parameter have been developed in the literature [e.g., *Fu*, 1981; *Choudhury*, 1999; *Zhang et al.*, 2001; *Yang et al.*, 2008]. For example, *Fu* [1981] derived an analytical solution of the Budyko curve with introducing an adjustable constant:

\[
\frac{E}{P} = 1 + \frac{E_p}{P} - [1 + \left(\frac{E_p}{P}\right)^{\omega}]^{1/\omega} \tag{15}
\]

where \(\omega\) is the parameter which represents the effects of other factors such as vegetation, soil, and topography on the partitioning of precipitation.

In this paper, the Budyko hypothesis is extended for modeling the seasonal behavior of precipitation partitioning into evaporation, runoff, and storage changes. A top down approach is used in this study to link the understanding from observed data to Budyko type models. Similar to the modification of seasonal aridity index in equations (13) and (14), the evaporation ratios for the wet and dry season are modified as \(\frac{E_w}{P_w-\Delta S_w}\) and \(\frac{E_d}{P_d-\Delta S_d}\), respectively. The seasonal evaporation ratio versus seasonal aridity index for two watersheds with both wet and dry seasons are shown in Figure 3a (West Conewago Creek watershed located in Philadelphia) and Figure 3b (Kaskaskia River watershed located in Illinois), respectively.
Figure 3.3. (a) Seasonal evaporation ratio versus seasonal aridity index for the West Conewago Creek watershed located in Philadelphia (USGS gage 1574000, (b) Evaporation ratio versus aridity index for Kaskaskia River watershed located in Illinois (USGS gage 0559300
Figure 4a shows the seasonal values for Oostanaula watershed located in Georgia which only includes wet season and Figure 4b shows the seasonal values for Clear Fork Brazos watershed located in Texas, which only includes dry season. The data points in wet seasons are usually associated with aridity index less than 1, and the data points in the dry season are associated with aridity index higher than 1. Aridity index in a particular dry season can be smaller than 1 because dry months are defined based on mean monthly aridity index. As it can be seen, the data points in the wet and dry seasons do not follow the same Budyko type curve for a given watershed. Therefore, two separate curves may be required to model the water balance at the two seasons.
Figure 3.4. (a) Evaporation ratio versus aridity index in dry season for Oostanaula watershed located in Georgia (USGS gage 02387500), (b) Evaporation ratio versus aridity index for Clear Fork Brazos River watershed located in Texas (USGS gage 08085500), and the fit

There are several differences between mean annual and seasonal water balance in the Budyko framework. In the mean annual scale, Budyko hypothesis provides an inter-comparison of water balance among watersheds; when climate aridity index approaches zero, the evaporation ratio approaches to zero. For a given watershed, the lower bound of seasonal aridity index may be higher than zero. For example, the lower bound of aridity index in the dry season is bounded by a positive value instead of zero as shown in Figures 3 and 4b. The intersection of the two limit lines is at the aridity index of 1 at the mean annual water balance, but for a given watershed, the intersection may be larger than 1 particularly for dry seasons. To characterize these observations from Figures 3 and 4, a
shift along the horizontal axis is introduced to the Budyko type curves such as equation (15) for modeling the water balance in dry seasons. The same functional form is applied to wet seasons for consistency. Using Fu’s equation, the Budyko type functions for wet and dry seasons are modified as:

\[
\frac{E_w}{P_{w-\Delta S_w}} = 1 + \left(\frac{E_{P_w}}{P_{w-\Delta S_w}} - \phi_w\right) - \left[1 + \left(\frac{E_{P_w}}{P_{w-\Delta S_w}} - \phi_w\right)^{\omega_w}\right]^{1/\omega_w}
\]

\[
\frac{E_d}{P_{d-\Delta S_d}} = 1 + \left(\frac{E_{P_d}}{P_{d-\Delta S_d}} - \phi_d\right) - \left[1 + \left(\frac{E_{P_d}}{P_{d-\Delta S_d}} - \phi_d\right)^{\omega_d}\right]^{1/\omega_d}
\]

where \(\omega_w\) and \(\omega_d\) are Fu’s coefficients in wet and dry season, respectively; and \(\phi_w\) and \(\phi_d\) are the horizontal shifts of lower seasonal aridity index bound for wet and dry season, respectively. The fitted lines by equations (16) and (17) are also plotted in Figures 3 and 4. From the modified Buyko type function, evaporation approaches to zero when potential evaporation approaches to a certain positive value. In a shorter time span, evaporation could approach zero when potential evaporation approaches to a value higher than zero [Han et al., 2012].

Tow parameters need to be estimated in the modified Budyko type functions for each season. The values of \(\omega_w\) and \(\omega_d\) represent the physical controls of watershed properties on the seasonal water balance like the mean annual water balance. The values of \(\phi_w\) and \(\phi_d\) can be interpreted as the lower limits of aridity index for wet and dry seasons, respectively. In Figure 3a, the value of \(\phi_w\) for the fitted line in wet seasons is 0.13 and the value of \(\phi_d\) for the fitted line in dry season is 0.24. In Figure 3b, the value of \(\phi_w\) is 0.14 and \(\phi_d\) is 0.32 for the fitted lines. For a given watershed the value of \(\phi_d\) is
higher than that of $\omega_w$. Given an aridity index in a watershed, the evaporation ratio in the dry season is higher than that in the wet season. The values of $\omega_w$ and $\omega_d$ also represent the shifts of the 1:1 limit lines for energy limited conditions. When seasonal aridity index is smaller than 1 in the wet season, the upper bound of evaporation is equal to $E_{P_w} - \omega_w(P_w - \Delta S_w)$ which is usually smaller than potential evaporation ($E_{P_w}$). Therefore, there is a stricter bound on seasonal evaporation in this situation (wet condition). The values of shifts depend on seasonal climate and watershed properties since storage changes are also included in the seasonal aridity index.

Two parameters of $\omega$ and $\theta$ shape a trade-off in the Budyko framework. The interactions of them indicate a certain range of evaporation ratio for a given aridity index. In Figure 5 the contour lines of seasonal evaporation ratio as functions of $\omega$ and $\theta$ in the seasonal Budyko framework are presented for 3 values of seasonal aridity index, i.e., $A=1, 1.5, and 2$. In Figure 5a, for a given value of $\omega$, as $\theta$ increases the evaporation ratio decreases. At the same time for a given value of $\omega$ as aridity index increases (from Figure 5a to 5c) the sensitivity of evaporation ratio to $\theta$ decreases, indicating that for water-limited conditions the evaporation ratio is less sensitive to values of $\theta$. Similarly for a given value of $\theta$ in Figure 5, as $\omega$ increases the evaporation ratio increases as well. The aforementioned sensitivity of evaporation ratio to $\omega$ decreases as $\omega$ increases. In the same time, as it goes to more arid conditions (from Figure 5a to 5c), the decline in sensitivity of evaporation ratio to $\omega$ becomes more significant.
3.2.4 Modeling Annual Storage Changes

Once the four parameters ($\omega_w$, $\omega_d$, $\phi_w$ and $\phi_d$) for the seasonal water balance model are obtained, the seasonal Budyko type model developed in this research can be used to estimate annual storage changes and evaporation if precipitation, potential
evaporation and runoff observations are available. Substituting \( E_w = P_w - Q_w - \Delta S_w \) into equation (16), the following equation is obtained and can be used to estimate storage changes in the wet season:

\[
(P_w - \Delta S_w)^{\alpha_w} + [E_{P_w} - \phi_w(P_w - \Delta S_w)]^{\alpha_w} = [E_{P_w} - \phi_w(P_w - \Delta S_w) + Q_w]^{\alpha_w}
\]  
(18)

Similarly, storage changes in the dry season can be estimated as:

\[
(P_d - \Delta S_d)^{\alpha_d} + [E_{P_d} - \phi_d(P_d - \Delta S_d)]^{\alpha_d} = [E_{P_d} - \phi_d(P_d - \Delta S_d) + Q_d]^{\alpha_d}
\]  
(19)

The values of \( \Delta S_w \) and \( \Delta S_d \) can be solved numerically using equations (18) and (19), and annual storage changes (\( \Delta S \)) can be computed as a summation of seasonal storage changes:

\[
\Delta S = \Delta S_w + \Delta S_d
\]  
(20)

The annual evaporation can be computed as a residual of water balance once storage changes are estimated.

### 3.2.5 Performance Evaluation

The model performance is evaluated using two indicators. Root mean square error (RMSE) is used to measure the performance of models. RMSE is calculated as:

\[
RMSE = \sqrt{\frac{(x_{\text{meas}} - x_{\text{mod}})^2}{n}}
\]  
(21)
In which, $X_{obs,i}$ and $X_{mod,i}$ is the observed and modeled values in year $i$, respectively; $n$ is the number of years that data is available which is 21 years in this study. The other indicator is Nash-Sutcliffe efficiency here referred as coefficient of efficiency (CE) and shows the extent to which, observed and modeled values follow the line with 1:1 slope [Moriasi et al., 2007]. CE is calculated as:

$$CE = 1 - \frac{\sum_{i=1}^{n}(x_{meas,i} - x_{mod,i})^2}{\sum_{i=1}^{n}(x_{meas,i} - x_{meas,i})^2}$$

(22)

CE shows the order of accuracy of a model to predict the actual values. CE ranges from $-\infty$ to 1. The values closer to 1 indicate higher model efficiency in predicting actual values [Legate and McCabe, 1999]. Positive CE value is usually acceptable for a model [Moriasi et al., 2007].

RMSE and CE are applied to evaluate the fitness of the extended Budyko type model and the performance of that in estimating annual storage changes from equations (18), (19) and (20). The fitness of the seasonal Budyko type model is computed for all the watersheds in each season, and is compared among watersheds belonging to different categories.

3.3 Results and Discussions

The developed model in this paper is applied to the 277 case study watersheds shown in Figure 1. Wet and dry seasons are identified for each watershed, and the seasonal climate variables are quantified in each season. The seasonal model based on
Fu’s equation is fitted to the observations for each watershed. The estimated values of parameters for the study watersheds are then discussed. Based on estimated parameters and equations 18, 19, and 20, annual storage changes are computed and the performance of the model is evaluated.

### 3.3.1 Observed Annual and Seasonal Storage Changes

Annual water balance of watersheds has been studied in three methods in the Budyko framework as shown in Figure 6.

![Figure 3.6](image)

**Figure 3.6.** Comparison of interannual balance of water from 3 different perspectives

In the first method evaporation has been estimated as the difference between precipitation and runoff, this method is usually used due to limitations in available data, and the corresponding values are plotted in Figure 6a. As shown in Figure 6a, considering the evaporation as $P-Q$ in annual time scales results in overestimation of evaporation in a way which makes it even higher than the defined limits of evaporation. This overestimation is related to neglecting water storage changes in annual water balance. Based on this method, *Istanbulluglu et al.* [2012] found that annual
evaporation ratio and aridity index are negatively related, and they reasoned that this negative relation is due to the significant portion of storage changes. Figure 6b represents the observed evaporation ratio versus climate aridity index. Such approach to water cycle was presented by Cheng et al. [2011]. As shown in Figure 6b, if $P$ is considered as water supply in annual scale, the relation between evaporation ratio and annual aridity index tends to a linear one instead of forming a curve bounded by limits. It shows that in many cases the evaporation ratio approaches to values higher than 1. This highlights the fact that available water supply is not limited to precipitation only, but storage changes play a significant role in maintaining the evaporation, especially for years with aridity indices higher than 1. Figure 6c shows evaporation ratio and aridity index when effective precipitation is used to represent available water where.. In a study of long-term soil moisture and ground water level in 12 watersheds in Illinois Wang [2012] also showed that considering the effect of annual water storage changes enhances the relation between evaporation ratio and aridity index. From this comparison it can be interpreted that the Budyko hypothesis is applicable at interannual scale if the supply of energy and water treated accurately within the exact definition in the original Budyko framework.

The exceedance probabilities of inter-annual storage changes are computed based on the observed annual storage changes from 158 watersheds with higher accuracy of evaporation data. As shown in Figure 7a, the lower bound and upper bound of annual storage changes are $-481$ mm and $536$ mm, respectively. However, storage changes in 40.0% of years are within the range of $\pm 50$ mm, and in 70.8 % of years are within the
range of ±100 mm. Storage changes in 53.8% of years are positive. The exceedance probabilities of storage changes in wet seasons are computed and shown in Figure 7b. The seasonal storage changes range from -480 mm to 692 mm, and in 19.6 percent of years storage changes in wet seasons are negative values and 67% of years storage changes in wet seasons are greater than 100 mm. In figure 7c, the exceedance probability of storage change ratio in dry season is plotted for 158 watersheds and for each year of records it can be observed that for 35% of years storage change is more than one of fourth of precipitation, emphasizing on the role of storage change in constituting the available water supply.

Although it is expected that in wet seasons the storage changes be positive due to groundwater recharge, it should be noted that the wet and dry seasons are defined based on the long term mean monthly values. Therefore, in a particular year a month with long-term mean aridity index larger than 1 may have actual aridity index less than 1 and vice versa.
Figure 3.7. Exceedance probability of annual observed storage changes values for all years in 158 watersheds (a), Exceedance probability of observed storage changes in wet seasons (b), Exceedance probability of observed storage changes in dry seasons (c).
3.3.2 Watershed Classification Based on Seasonal Aridity Index

Based on the definition of wet and dry months, 201 watersheds have both wet and dry seasons and almost all of them have a dry season in the summer season. The duration of dry season ranges from 1 to 11 months in these watersheds. 54 watersheds only have wet seasons and 22 watersheds only have dry seasons. Figure 1 shows the spatial distribution of the categories of these watersheds. From 158 watersheds in which the evaporation data are more accurate, 131 of them have both wet and dry seasons, 23 of them have wet season only and 4 of them have dry season only. Watersheds with dry season only are mostly located in arid regions of Great Plains, while watersheds with wet season only are mostly located in north eastern United States and Appalachian Mountain area. Watersheds with both seasons are distributed all over United States such as northwestern states, midwest, and southeastern states. The watersheds with both seasons are grouped into three categories by the number months in the dry season. As shown in Figure 1, a clear and distinct spatial pattern of dry season distribution can be recognized. The mountain areas in western United States have shorter dry seasons with 1 to 4 dry months or wet seasons only. In the Great Plains, watersheds are mostly categorized with long dry seasons.

As examples, the mean monthly climate aridity indices for three watersheds are shown in Figure 8. Savannah River Watershed (USGS gage #2192000) has both wet and dry seasons, and the dry seasons are from June to September. The climate aridity indices for the Oostanaula watershed (USGS gage #2387500) are less than 1 for all the months, i.e., wet season only. Both Savannah River watershed and Oostanaula watershed are
located in the state of Georgia. The monthly climate aridity indices for the Smoky Hill River watershed (USGS gage #6869500), which is located in Kansas, are higher than 1 in all the months and the entire year is defined as dry season.

Figure 3.8. Long term mean monthly aridity indices for Savannah River Watershed with wet and dry seasons, Oostanaula watershed with wet season, and Smoky Hill River watershed with dry season

The long term mean monthly values of precipitation, potential evaporation, evaporation, runoff, and storage change for the Savannah River watershed are presented in Figure 9. The standard deviations for monthly precipitation, runoff, evaporation, and storage changes are 13.5, 15.2, 32.9, and 32.1 mm, respectively. Therefore, the monthly variabilities of storage change and evaporation are higher than those for precipitation and runoff. From March to June evaporation increases even though precipitation decreases
and since the precipitation is limited in this period, the water supply is supplemented by water storage changes. Water supply from April to May, when evaporation is higher than precipitation, is supplemented by water storage changes. The period from April to November is defined as dry season and the rest months are defined as wet season based on the mean monthly values of aridity index.

![Graph](image)

**Figure 3.9.** Long term mean monthly precipitation, potential evaporation, runoff, evaporation and storage change for Savannah River Watershed (both seasons)

### 3.3.3 Application of The Seasonal Model to Case Study Watersheds

The developed seasonal model based on Budyko-type functions, i.e., equations (8) and (9) is applied to the case study watersheds shown in Figure 1. The values of the four seasonal parameters ($\omega_w$, $\omega_d$, $\phi_w$ and $\phi_d$) are estimated based on the collected data on
monthly precipitation, potential evaporation, evaporation, and runoff during 1983-2003. For example, Figures 3 shows the modified Budyko type curves in wet and dry seasons which fit to the data points for two watersheds. As shown in Figure 3a for West Conewago Creek watershed, $\varphi_w$ is 0.13 and $\omega_w$ is 2.23 for the wet season and $\varphi_d$ is 0.24 and $\omega_d$ is 6.92 for the dry season. As shown in Figure 3b for Kaskaskia River watershed, $\varphi_w$ is 0.14 and $\omega_w$ is 2.01 for the wet season and $\varphi_d$ is 0.32 and $\omega_d$ is 8.6 for the dry season. To evaluate the performance of the model, the $CE$ values for estimated seasonal evaporation ratio in wet seasons are 0.98 and 0.97 for the two watersheds, respectively; the $CE$ values in dry seasons are 0.97 and 0.93 for two watersheds, respectively. Figure 4a shows fitted curve for the Oostanaula watershed in which all the 12 months are classified as wet season, and the value of $CE$ is 0.99. The estimated values for $\varphi_w$ and $\omega_w$ are 0.1 and 3.81, respectively. Clear FK Brazos watershed (USGS gage #08085500) only includes dry season and the values of $\varphi_d$ and $\omega_d$ for the fitted curve is 2.32 and 4.88 with a $CE$ value of 0.72. The average value of $CE$ over all the 277 study watersheds is 0.93 for wet seasons and is 0.81 for dry seasons. The developed model fits the seasonal water balance well even though the performance in dry seasons is not as high as in wet seasons.

In the seasonal model, evaporation ratio is a function of seasonal aridity index with parameters of $\omega_w$ and $\varphi_w$ for wet seasons and $\omega_d$ and $\varphi_d$ for dry seasons. The values of the parameters reflect the dependence of seasonal water balance on other factors such as vegetation, soil properties, and topographical properties in the watershed. In wet seasons, the value of $\varphi_w$ is small and ranges from 0 to 0.3; values of $\omega_w$ range
from 1.5 to 13.1. In dry seasons $\phi_d$ varies from 0 to 2.4 and $\omega_d$ ranges from 1.8 to 26.4.

The histograms of the seasonal coefficients are presented in Figure 10.

Figure 3.10. Histograms of $\phi_w$ (a), $\omega_w$ (b), $\phi_d$ (c), and $\omega_d$ (d)
The values of $\varnothing_w$ has the highest frequency around 0.1. It mainly shows that the Budyko curve needs a shift along horizontal axis. The shift is needed since the lowest possible value of energy supply that maintains evaporation in the watershed is a value higher than zero. The values of $\varnothing_w$ are small and the uncertainty of data sources may affect these values. The values of $\omega_w$ has the highest frequency around 2.5 and in some cases the value is even higher than 10. The values of $\varnothing_d$ has the highest frequency around 0.25 which is larger than that of wet season. The difference between the distribution and peak of $\varnothing_d$ and $\varnothing_w$ shows that in dry seasons the value of shift is not only affected by lower bound of necessary energy supply but also by some other factors. Values of $\omega_d$ has the highest frequency around 5. Evaporation ratio is affected by the values of both $\varnothing$ and $\omega$ as discussed earlier. The effects of seasonal climate and soil water storage dynamics have been included in the seasonal aridity index.

Figure 1 shows $\varnothing_d$ versus mean Normalized Difference Vegetation Index (NDVI) during the dry season. NDVI data are obtained from Advanced Very High Resolution Radiometer (AVHRR) imagery and are product of the Global Inventory Modelling and Mapping Studies (GIMMS) which are available in bimonthly time resolution [Tucker et al., 2005]. Data set is available for download at http://glcf.umiacs.umd.edu/data/gimms/. It can be seen that watershed with higher value of NDVI usually have smaller value of shift ($\varnothing_d$). This implies that watersheds with smaller lower bound of dry season aridity index have more vegetation coverage in the dry season. As shown in Figure 11, the values of $\varnothing_d$ for watersheds with dry season only are usually higher than those for watersheds with both dry and wet seasons. For example, the
values of $\phi_d$ are 0.24 and 0.32 for West Conewago Creek watershed and Kaskaskia River watershed with two seasons (Figure 3), but the value of $\phi_d$ is 2.32 for Clear FK Brazos watershed with dry season only. Correspondingly, NDVI in watersheds with dry season only is usually lower than that in watersheds with both seasons.

Figure 3.11. $\phi_d$ versus long-term mean NDVI in the dry season (NDVI_d)

3.3.4 Estimation of Annual Storage Changes

Once the values of parameters for each watershed are estimated, the seasonal model developed in this paper can be used to estimate annual evaporation and storage changes when precipitation, potential evaporation and runoff data are available. Storage changes are estimated by numerical method through equations (10) and (11) for each season. Then annual storage changes are computed by equation (12). The model
performance on estimating annual storage changes is evaluated. For West Conewago Creek watershed, the CE for seasonal storage change are 0.99 and 0.97 for wet and dry seasons, respectively; and the RMSE values are 8.4 mm and 11.7 mm for wet and dry seasons, respectively. For Kaskaskia River watershed, the CE of modeled values of seasonal storage changes are 0.99 and 0.96 for wet and dry seasons, respectively; and the RMSE values are 4.7 mm and 18.2 mm, respectively. These values are less than 18% of average absolute storage changes in each season. For Oostanaula watershed, the CE and RMSE in wet season are 1.0 and 13.4 mm, respectively. The RMSE is 8% of average absolute storage changes. For Clear FK Brazos, CE and RMSE in dry season are 0.95 and 26.86 mm respectively. This value is 13% of average absolute storage changes. The latter watersheds are also samples that due to having only one season throughout the year, can shows the annual behavior of storage changes directly.

The “observed” annual storage changes by water balance closure and the modeled annual storage changes in each year for the study watersheds are compared in Figure 12. The modeled values are close to observed storage changes and the average value of RMSE is 26.0 mm for all the watersheds. Considering the exceedance probability distribution of annual storage change values in Figure 7, 19% of annual storage changes are within the range of $\pm 26$ mm. The absolute annual storage changes in 81% of years are higher than the RMSE value.
The observed storage changes are up to 800 \( \text{mm} \) in some watersheds as shown in Figure 12 and this is unrealistic. The uncertainties in observations, particularly evaporation estimation from remote sensing data, may contribute to the unrealistic storage change and further the model-based storage change estimations. Therefore, the observed storage changes versus modeled storage changes for watersheds that are grouped as those with higher accuracy (158 watersheds from the total 277 watersheds). The average value of RMSE for these watersheds is 24.5 \( \text{mm} \), the decrease in values of RMSE for watersheds with higher accuracy of evaporation data shows that as the uncertainty in data is decreased the model performance will be enhanced. CE represented in equation (14) is also evaluated for these watersheds. The average value of CE in predicting the annual values of storage changes is equal to 0.92 which shows the high strength of model to predict the annual storage changes. The CE is even higher if the seasonal measured and modeled values of storage changes are explored. For wet season the average CE is 0.97 and for dry season it is 0.84.
3.4 Conclusions

The storage change is a basic component in water cycle. In short scales such as annual and seasonal it is a significant portion of water balance, and needs to be accounted in water cycle studies. Comprehensive estimations of storage change at watershed scale is still a challenge in hydrological science, and conventionally in water balance related studies the storage change was neglected. In many recent years with advancements of measurement techniques for estimating the storage changes, the role of storage change in hydrological studies is highlighted and it is introduced as a key component in understanding the processes and interactions occurring at the watersheds. Moreover, the need for developing parsimonious models capable of explaining the basic processes such as partitioning pattern of water supply especially at time scales with significant dynamic behavior is growing. Also models which are capable in predicting seasonal and annual storage changes can be very helpful in water managements and planning.

This research is on developing a parsimonious model which broadens current understanding of seasonal water cycle when it’s dynamic behavior is captured. The provided model is can also predict seasonal and annual storage changes with given value of precipitation, evaporation, and runoff. In this study a water balance closure approach has been used to estimate storage changes. The study also benefits from actual evaporation data obtained from remote sensing. Following Budyko proposed concept, a new aridity index is defined as the ratio of potential evaporation to effective precipitation which accounts the storage changes in water supply. Similarly a new evaporation ratio is
also defined as the ratio of evaporation to effective precipitation in Budyko frame. Then based on the long term mean monthly aridity indices two fixed seasons (wet and dry) are identified in each of 277 watersheds.

Following a top-down approach, a shifted Fu’s curve is found to be able to model the seasonal relation between aridity index and evaporation ratio from year to year, very well. The performance of the curve to model the relation is as high as having average CE equal to 0.93 for wet seasons and CE equal to 0.81 for dry seasons. The modified Fu’s curve is not only characterized by the conventionally $\omega$ parameter, but also with a shift factor $\emptyset$, which is representative of the lower bound of possible aridity index in each watershed and for each season. Then, given the values of seasonal $P, E_p, Q, \omega$, and $\emptyset$ the storage changes can be predicted at seasonal scale and subsequently annual scale. The performance of model in predicting storage changes was as high as having average CE equal to 0.92 for annual storage changes.

Although the developed model performs well in predicting the values of storage changes, there are still concerns related the uncertainty in data especially since evaporation and runoff data has different source from evaporation data. The further steps in extending and developing the long term goals of this framework and research can be summarized as follows. Characterization of $\omega$ and $\emptyset$, and developing physically based relations that can predict the values of them having watershed properties such as vegetation, topography, and long term climate, etc. Also, an important further step is to
extend the model into smaller time scale such as monthly and weekly, then explaining and modeling the partitioning processes at shorter time scales.
3.5 References


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CHAPTER 4: CONCLUSION

Storage change is a basic component in water cycle. In annual and seasonal scales it is a significant portion of water balance, and needs to be accounted in water balance studies. Comprehensive estimations of storage change at the watershed scale are still a challenge, and conventionally in water balance related studies the storage change was neglected at the annual scale. Very recently, with advancements of measurement techniques for estimating the storage changes, the role of storage change in hydrological studies is highlighted and is introduced as a key component in understanding the processes and interactions occurring at watersheds. Moreover, the need for developing models capable of explaining the basic processes such as partitioning pattern of precipitation especially at time scales with significant storage dynamic is growing. Also models capable of predicting seasonal and annual storage changes can be useful in water planning and management.

This research aims at developing a parsimonious but reliable model which broadens current understanding of seasonal water cycle where. The presented model can also predict seasonal and annual storage changes and evaporation given the value of precipitation, potential evaporation, and runoff. In this study a water balance closure approach has been used to estimate storage changes. The study also benefits from actual evaporation data obtained from remote sensing techniques (or data bases). Following Budyko hypothesis, a modified aridity index is defined as the ratio of potential
evaporation to effective precipitation which accounts the storage changes in water supply. Similarly a new evaporation ratio is also defined as the ratio of evaporation to effective precipitation in Budyko framework. Then based on the long term mean monthly aridity indices two fixed seasons (wet and dry) are identified in each of the 277 watersheds.

Following a top-down approach, a shifted Budyko-type curve was found to be able to model the relation between seasonal aridity index and seasonal evaporation ratio. The performance of the curve to model the relation is as high as having average CE equal to 0.93 for wet seasons and CE equal to 0.81 for dry seasons. The modified Budyko-type curve is not only characterized by the conventionally $\omega$ parameter, but also with a shift factor $\varnothing$, which is representative of the lower bound of possible aridity index in each watershed and for each season. Then, given the values of seasonal $P$, $E_p$, $Q$, $\omega$, and $\varnothing$ the storage changes can be predicted at seasonal scale and subsequently annual scale. The performance of model in predicting storage changes was as high as having average CE equal to 0.92 for annual storage changes.

Although the developed model performs well in predicting the values of storage changes, uncertainty exists in the observation data, especially since evaporation and runoff data source is different from that of evaporation data. Future research can be summarized as follows. Physically based relations can be developed to predict the values of $\omega$ and $\varnothing$ given watershed properties such as vegetation, topography, and long term climate, etc. The framework can be extended to model water balance at the monthly
scale. As a potential application, the developed model can be utilized to assess the impact of climate change on annual and seasonal water availability.