8-2010

THE IMPACT OF WATER HARVESTING ON A SMALL WATERSHED IN RURAL INDIA

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ABSTRACT

Reliable sources of fresh water are a finite resource across the world. Many countries, including India, face water scarcity due to temporal and spatial variations in precipitation, surface water pollution, and depletion of groundwater resources. In order to combat against water scarcity, the government, non-governmental organizations, researchers, and individuals have attempted to create solutions to the water scarcity problem. One solution, which has become popular throughout India is the construction of water harvestings structures (WHS), small earthen dams built to capture monsoonal runoff on ephemeral streams. Villagers believe these structures have a positive effect on groundwater levels and water availability throughout the year, although the direct effect on the local watershed is poorly understood. To better understand the impact of these structures, this thesis investigates the local geology, the watershed and surface water balance, and the monsoonal response of a WHS reservoir in a small watershed in Madhya Pradesh, India.

Field work for this study was completed from May 2009 through April 2010. The accomplishments from the work are three fold. First an improved understanding of the local geology was obtained using electromagnetic induction surveys. Second, major components of the hydrologic cycle were monitored to calculate the flows for the overall water balance and the surface water balance. Finally, water levels in a WHS reservoir were monitored to allow for the reevaluation of a volume balance model proposed for management of these structures for artificially recharging groundwater. The main goal of
the project is to determine the impact of water harvesting for artificial recharge and increasing water availability within the watershed.

Information gathered during the geological surveys was used to develop the water balance for the watershed. From the water balance it was determined that streamflow out of the watershed is approximately 15% of the total yearly rainfall. The net transfer of surface water to the subsurface is approximately 80% of precipitation or as a flux normalized to the watershed area is 0.59m/year. The yearly change in groundwater storage is positive and wells are able to recover after two months of pumping for irrigation, indicating current groundwater practices are sustainable and not over drawing the local aquifer.

After consideration of the volume balance for the WHS, it was found that from two to six times the maximum reservoir volume (6.5x10^4 m^3) is lost as groundwater recharge from the structure. If the structure is assumed to infiltrate 1.3x10^5 m^3/year, without the presence of the structure the yearly streamflow would increase by 48% if the volume of water infiltrated was assumed to be discharged as streamflow. In addition to decreasing streamflow, the upstream reservoir provides a surface water body which is present for ten months of the year, helping to decrease water scarcity in the early dry season.
DEDICATION

To the people of Salri and the Foundation for Ecological Security, who were grateful enough to invite me to their community to allow me to live, work, and assist me in accomplishing the goals of the project. To my family, who have been supportive of my research and have provided never-ending love and encouragement. To my friends, who are always there to give a piece of advice and words of support.
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CHAPTER ONE

INTRODUCTION

1.1 Water Scarcity

All organisms, great and small, require water for life. Accessibility to this resource, however, is in increasingly short supply (Oki and Kanaa, 2006). Although the planet has vast water resources and approximately 70% of the earth is covered by water, only 2.5% is fresh water (Postel, 1996). With such limited supplies of useable fresh water, regions of the world face water scarcity almost every year, and people living in these areas struggle to acquire sufficient supplies of water required for daily life, industry, and agriculture (Oki and Kanaa, 2006).

Water scarcity is defined by United Nations Environmental Program Global Environmental Outlook as the amount of water used for industry, agriculture, and domestic purposes divided by the total amount of renewable water available in surface water bodies and shallow aquifers. High water scarcity occurs when 40% of the total available water supplies are withdrawn (UNEP, 2000). Currently, two billion people face a high level of water scarcity, and it is predicted that by 2025 two thirds of the world’s eight billion people will face high water scarcity (Figure 1.1; Vörösmarty et.al, 2000). One example of a country which is expected to have high water scarcity by 2025 is India.
Figure 1.1: Water scarcity faced in India for the year 1995 and then projected to 2025 (UNEP, 2000).

Although India has vast water resources, scarcity arises due to spatial and temporal variations of precipitation as well as poor water management (Bobba et al., 1997; Chaudary et al., 2002). The majority of rainfall across much of India occurs during the four month monsoon season, June through September. The quantity of rainfall during the monsoon is generally sufficient to meet demands (Bobba et al., 1997). Because it comes in such a short amount of time, however, the majority of rainfall is lost as runoff (Singh and Sharama, 2002; Bobba et al., 1997). Furthermore, sustainable water management practices are not set in place, which has lead to degradation of water quantity and quality (Limaye, 2010).

Attempts at solutions to address the inadequate water supplies and scarcity have come in the form of aid and assistance from non-governmental organizations, local governments, researches, and individuals. Aid has been provided by land development projects, construction of retention ponds, and general education about water resources (Limaye, 2010). However, the net impact of such help is usually poorly understood (Bobba et al. 1997). Consequently, more scientific work is required to study the impacts
of water harvesting, watershed restoration, and sustainable water management in order to lessen the water scarcity faced by so many (Kumar, 2006).

1.2 Water Harvesting Structures

Water harvesting has been used for centuries as a way to move or store water for domestic use in the future (Lavee et al., 1997). Out of the numerous types of water harvesting, the construction of small earthen dams built on ephemeral streams to capture and store monsoonal runoff have become very popular in India (Sukhija et al., 1997; Srivastava, 2000). Water stored in these structures is then able to infiltrate and potentially recharge local aquifers instead of discharging through streams (Kumar, 2006). Although water harvesting structures (WHS) are seen as beneficial to local communities, the volume of water contributed as recharge as well as the location and accessibility of recharged water is poorly understood. Currently, there are few scientific studies which support their effectiveness (Bobba et al., 1997).

1.3 Thesis Objective

In order to better understand the impact of water harvesting in India, the current study focuses on a small watershed located in rural Madhya Pradesh, India (Figure 1.2). Previous investigations in the watershed were conducted in 2007 by Oblinger (2008) and the Foundation for Ecological Security (FES), a local non-governmental organization, to investigate the local geology, water balance, and model the behavior of a WHS. The goal of the current study is to further investigate the local geology, better understand the flows
of the water balance, and reevaluate the model for the WHS in order to aid FES and the local village in determining the impact of water harvesting on groundwater recharge and water accessibility through surface water and groundwater sources throughout the year.

![Location of the study watershed in Madhya Pradesh, India.](image)

**Figure 1.2:** Location of the study watershed in Madhya Pradesh, India.

The study watershed is located within the Deccan Basalts of central and western India. To better understand and improve on the conceptual model developed from geologic investigations in 2007, the current study uses electrical resistivity and electromagnetic induction surveys to study the stratigraphy. In addition, soil and rock samples were collected to investigate the hydrogeology. Electrical resistivity surveys performed with the Indian Institute of Technology-Bombay were used to confirm the findings from the 2007 study which characterized the thickness, type, and location of basalt flows. Electromagnetic induction surveys were used to determine the presence and extent of an upper weathered zone which may act as a surficial aquifer, and evaluate
changes in the electromagnetic response through time as the monsoon season progresses. Soil and rock samples were used to determine the hydraulic conductivity and specific yield in the northeastern area of the watershed. Knowledge of the local geology, hydrogeology, and extent of the weathered zone is needed for completing the water balance for the watershed.

Water balances are useful for determining net water availability for a given system (Burt, 1999). The watershed water balance is used to determine the estimated range for evapotranspiration for the watershed. Furthermore, the water balance is used to investigate the residual flow of water out of the watershed; primarily, the volume of water lost from the surficial aquifer to other reservoirs, such as leakage to a regional aquifer. A monthly surface water balance is developed to determine the net volumetric flow to the subsurface across the watershed which will help determine the overall impact of water harvesting in the watershed. After knowing the flows of the water balance, it will be possible to investigate water supplies in surface water bodies and the surficial aquifer for a given year, evaluate if current water practices in the watershed are sustainable, and determine whether or not water harvesting has a significant impact on the overall water balance.

Finally, a volume balance to model the WHS as developed by Oblinger et al. (2010) is reevaluated to determine the volume of water lost from the structure as infiltration and the residence time of water in the structure. After model recalibration, the volume of infiltration from the structure is compared to the approximate volume of water which moves as the net transfer from the surface water to the groundwater system. From
this comparison it is possible to determine if water harvesting is a successful method to recharge the local aquifer and ease the water scarcity faced each year.

From the study it is hoped the results will continue to aid FES in assessing the impact of WHS in rural areas; furthermore, in studying the geology, availability of water, and the impact of water harvesting it is hoped villagers of the area will be able to better manage and sustain their water supplies.

1.4 Thesis Overview

Chapter 1 provides an overview of the water scarcity faced in India, causes for scarcity, and a method of water harvesting as one solution to ease scarcity. Chapter 2 describes the regional and local geology to provide insights used in developing the water balance. Chapter 3 investigates the watershed and surface water balance to determine volumetric flows of water and water availability in the watershed. Chapter 4 investigates a volume balance numerical model to quantify the infiltration and residence time of water in the reservoir to help determine the effect of water harvesting in the watershed. Chapter 5 discusses the findings from the study, the overall impact of water harvesting, and provides recommendations for future work and management of water resources.
CHAPTER TWO  

DETERMINATION OF GEOLOGIC STRUCTURE IN A RURAL INDIAN WATERSHED WITH THE USE OF ELECTROMAGNETIC INDUCTION

Abstract

The Deccan Basalts of west-central India are a group of flood basalts comprising an area of roughly 500,000 km$^2$. The basalts were produced by fissure eruptions during the late Cretaceous to early Eocene, with the most activity approximately 65 million years ago. The stratigraphy of the region is characterized by three major packages of basalt flows interbedded with sediments deposited between eruptions. In-place weathered basalts and sediments deposited since the last eruption overlay the bedrock. The Deccan Basalts and associated surficial materials are also a main source of water for the inhabitants of the area. Knowing the geology is therefore important as it affects groundwater availability, the depth to the water table, and the ability of the aquifer to supply sufficient volumes of water for domestic, agricultural, and industrial use.

This study reports on the local geology of a small watershed located in the Malwa Plateau region of the Deccan Basalts in the Shajapur District of Madhya Pradesh. The aim of the study is to determine the local geology, specifically the thickness of different basalt units in order to better understand the hydrogeology of the watershed. To accomplish this goal, we used electromagnetic induction (EMI) surveys to non-invasively map the electromagnetic response of the subsurface. Three multi-frequency EMI surveys (330-20010 Hz) were conducted during the pre-monsoon, monsoon, and late-monsoon seasons of 2009. The first survey was conducted on May 3$^{rd}$, May 19$^{th}$, and June 9$^{th}$, the
second survey was conducted on July 8th, 16th, and 25th, and the third survey was conducted on September 2nd and 3rd. The electromagnetic response maps obtained from the surveys were used to interpret changes in the geology from the upper to the lower region of the watershed and identify changes in groundwater storage as the monsoon season progressed.

For a frequency of 570 Hz the response was relatively uniform for all three time periods, suggesting the presence of competent basalts of low porosity at depths approximately 26m across the watershed. At shallower depths of investigation, approximately at 7m, however, differences between upland and low-land areas are apparent. Upland areas are characterized by an electromagnetic response less than zero relative to the calibration location in the lower watershed, suggesting the presence of shallow soils overlying competent basalts, whereas lowland areas are characterized by a higher response relative to the calibration location suggesting the presence of a thick weathered zone. Furthermore, it was found that as the monsoon season progressed, the electromagnetic response increased more in the lowlands than uplands, further suggesting that changes in water storage are greatest in the lowland region.

2.1 Background on the Deccan Basalts

The Deccan Basalts cover an area of approximately 500,000km² in west-central India (Limaye, 2010). These continental lava deposits were erupted through fissures during the late Cretaceous to the early Eocene as the western edge of the Indian plate moved over a hot spot (Nair, 2001). The lava was erupted for approximately four million
years, filling depressions and valleys with massive lava flows (Kulkarni, 1997; Jerram, 2005; Figure 2.1).

![Figure 2.1: The location of the Deccan Basalts on the Indian subcontinent (Kulkarni, 2000).](image)

The Deccan Basalts were formed from three major eruptive events, with each episode having multiple, individual flows. The oldest unit is 150 meters thick and occurs on the eastern edge of the region. The middle package of flows is approximately 1,200 meters thick and occurs in the central part of the Deccan Basalts. The upper package is the thickest at approximately 2,000 meters. The upper package occurs primarily in the western parts of the Deccan Basalt area, near present day Mumbai, India (Singhal, 1997).

The basalts are heterogeneous due to the jointing and fracturing patterns as well as the degree of weathering of the upper surface. Generally, each basalt flow is made of three individual layers (Figure 2.2). The upper layer is the weathered Red Bole. Beneath Red Bole, still towards the top of an individual flow is highly porous vesicular material, formed by dissolution of gasses during cooling (Kulkarni et al., 1994). The middle and lower parts of an individual flow consist of columnar and massive basalts, respectively.
Columnar basalts are cut by vertical to sub-vertical joints spaced approximately 15cm, whereas massive basalts are generally fractured as large blocks, with an equal number of vertical and horizontal joints. Within individual flows, the porosity of the basalts decreases with depth (Kulkarni, 2000).

![Diagram of internal structure of the Deccan Basalts](Modified from Singhal, 1997).

This study investigates a watershed on the Malwa Plateau, which is a northern sub-region of the greater Deccan Basalts (Figure 2.3) and is located in the state of Madhya Pradesh and western Rajasthan (Jay and Widdowson, 2008). The Malwa Plateau makes up some of the first lava flows of the Deccan Basalts (Jay and Widdowson, 2008). The stratigraphy present in the Malwa Plateau is different than the rest of the Deccan Basalts in that the upper vesicular basalts are generally absent as the conditions for their formation may not have been optimal or the layers were weathered away before the next
eruptive event; therefore, alternating layers of columnar and massive basalts are more common (Kulkarni, 2007). The columnar and massive basalts near the surface have since weathered and alluvial materials have been deposited in low lying areas and valleys over time. The primary source of groundwater is the weathered upper layers of the columnar and massive basalts and alluvial material, since the vesicular basalts are limited and the Red Bole is either thin or absent.

![Map of India with Malwa Plateau highlighted](image)

**Figure 2.3:** Location of the Malwa Plateau on the India Subcontinent (modified from Jay and Widdowson, 2008).

### 2.2 Geology of the Study Watershed

A small watershed within the Malwa Plateau was selected as a field site to study the geology and hydrogeology. The site is located in the Shajapur District of Madhya Pradesh, and is approximately 2.56km$^2$ (Figure 2.4). The watershed is characterized by a lowland area with a weathered basalt zone reaching a maximum thickness of ten meters.
underlain by columnar and massive basalt. The upland region has a thin weathered zone and outcrops of columnar and massive basalt are common.

**Figure 2.4:** Location of the study watershed in Madhya Pradesh, India.

In order to better understand the hydrogeology of the watershed, the geology of the region was intensively studied during field work in 2007 by Oblinger (2008) and in the summer of 2009 by Daniel Matz from Clemson University. Field work conducted by Clemson University was done in collaboration with the Foundation for Ecological Security, an Indian non-governmental organization, and the Indian Institute for Technology-Bombay.

During field work conducted in 2007, Oblinger studied the watershed geology by geologic mapping of basalt outcrops and electrical resistivity surveys (Oblinger, 2008). This study failed to identify significant areas of vesicular basalt but did find extensive regions of massive, columnar, and weathered basalt. Electrical resistivity surveys were
conducted at five locations in and outside of the watershed boundaries (Figure 2.5). The electrical resistivity configuration used was a Schlumberger array (Reynolds, 1997). Results indicate resistivity generally increases with depth, and the thickness of the weathered zone increases moving to the northeast from the uplands to the lowlands of the watershed.

Figure 2.5: Locations of the resistivity surveys in and outside of the watershed boundaries which were conducted in 2007 and 2009.

Combining the resistivity surveys with visual observations of the basalts in uncased wells and at outcrops, Oblinger (2008) determined there are three major packages of flows from three distinct flow events. Each flow consists of columnar basalt overlying massive basalt. A fourth flow was determined from a report entitled Geohydrological Report in Part of the Susner Block, Madhya Pradesh (1987), which indicates a lower massive basalt unit. This massive basalt is shown as another flow
which comprises the bottom layer in the watershed. To better understand the geology, a cross section from the uplands to lowlands (Figure 2.6) is shown with the various basalt layers and the contact between the surficial and regional aquifer systems (Figure 2.7).

**Figure 2.6:** Location from where the cross section of the watershed was drawn to better understand the geology.

**Figure 2.7:** Cross section of the watershed showing the alternating pattern of the columnar and massive basalts (Oblinger, 2008). The vertical exaggeration is approximately 9 times and the y-axis shows the elevation above mean sea level (amsl).
An additional electrical resistivity survey using a Schlumberger array was also conducted during field work in 2009 by Matz, Meenakshi Choudhary from the Foundation for Ecological Security, and Dr. E. Chandrasekhar from the Indian Institute of Technology-Bombay. The survey was located in the approximate middle of the lower watershed (Figure 2.5). Results from the resistivity survey were analyzed by Dr. Chandrasekhar and it was found there were six layers with different resistivity values (personal communication, Chandrakeshar, 2009, Figure 2.8.). The same relationship between rock type and resistivity values used by Oblinger (2008) (Figure 2.9) was used to classify the basalts. Five different basalt flows were found at the location of the resistivity survey, where as the estimates by Oblinger (2008) focused on determining the thickness of the weathered zone and the identity of the underlying bedrock material. The 2009 survey shows the upper layer as weathered basalt and then a massive basalt layer. The next three layers are a columnar basalt unit on the top and bottom, with a massive basalt unit in the middle.
Figure 2.8: Resistivity values from the 2009 survey (left), and the depth at which each resistivity value occurs (right) (personal communication, Chandrakeshar, Indian Institute of Technology-Bombay, 2009).

Figure 2.9: Electrical resistivity survey results from 2007 (left) and 2009 (middle), classified using resistivity values from Oblinger (2008) (right).
The resistivity surveys conducted in 2007 and 2009 are generally consistent (Figure 2.9). Both surveys interpret the upper layer as part of the weathered zone, but the 2007 survey has the weathered layer being slightly thicker, reaching a depth of 10m. Both surveys find that the underlying layer is massive basalt. The total thickness of this basalt was not estimated for the 2007 data (Oblinger, 2008), but in the 2009 survey it was interpreted to reach a depth of 16m. Below this depth, the 2009 survey interprets columnar basalt to 30m depth, followed by another sequence of massive and columnar basalts to depths of 36m and 44m, respectively. The presence of the upper columnar basalt between approximately 16 to 30m depth is consistent with a transition between the bottom two flows found by Oblinger (2008).

The resistivity surveys are valuable for better understanding the stratigraphy of the watershed. These data helped to determine that the upper layer in the middle of the watershed, which is a weathered basalt zone, ranges up to ten meters in thickness. This zone is important as it can act to store water during the monsoon (Bobba et al., 1997); therefore, knowing the location and thickness is useful in determining water availability. Because the resistivity surveys are time and labor intensive, however, it is not practical to obtain this data over the entire extent of the watershed. To address this problem, electromagnetic induction (EMI) surveys, which are easier than resistivity surveys to conduct, were used to evaluate spatial heterogeneity throughout the watershed.
2.3 Electromagnetic Induction

To supplement the resistivity surveys conducted in 2007 and 2009, EMI surveys were conducted using the Geophex GEM-2. The objective of the surveys was to evaluate the variability of the weathered zone and underlying bedrock throughout the watershed. Three surveys were carried out during the summer of 2009 to measure changes in the electromagnetic response relative to a calibration point for pre-monsoon, peak monsoon, and post-monsoon conditions. The following describes the theory behind EMI as well as the results from the survey.

2.3.1 Electromagnetic Induction Background

EMI is a quick, non-contact, and effective method for evaluating the electrical conductivity (EC) of the subsurface. EMI instruments generate an electric current in a transmitter coil which generates a magnetic field that propagates into the ground. The primary magnetic field generates an electric current in electrically conductive materials, the strength of which depends on the conductivity of the subsurface. The induced current generates a secondary magnetic field, which is sensed by a receiving coil on the instrument (Figure 2.10). The secondary field is measured by the receiver coil in parts per million (ppm) of the primary field and has two components, in-phase and quadrature (Bongionanni et al., 2007). The EC of the subsurface can then be calculated from these measurements (McNeill, 1980; Fetter, 2001; Burger et al., 2006). If the instrument operates under low induction numbers, the in-phase component depends upon the magnetic susceptibility, i.e. the earth’s ability to be magnetized, and the quadrature
component, which is used for this study, depends on the electrical conductivity (Callegary et al., 2007; Bongiovanni et al., 2007). The induction number for a measurement is defined as the ratio of the instrument coil spacing to the measurement skin depth, which generally depends on the conductivity of the subsurface.

![Figure 2.10: Schematic of EMI, with the transmitter coil emitting the primary electric and magnetic field which passes into the ground and then the secondary electric and magnetic fields which are received at the receiver coil (Burger et al., 2006).](image)

The EC of the subsurface depends upon soil moisture, salinity, clay content, and geologic structures (McNeill, 1980; Fetter, 2001). Changes in the EC can be attributed to changes in soil moisture (Hanson and Kaita, 1997; Khakural et al., 1998; Reedy and Scanlon, 2003), fracture zones, or preferential flow paths within the subsurface (Samouëlian et al., 2004). EMI surveys have been used to investigate the depth of soil layers (Kitchen et al. 1996; Anderson-Cook et al., 2002), groundwater recharge (Cook, 1992), and water content (Reedy and Scanlon, 2003). The large variety of uses for EMI surveys makes it a very useful tool for watershed investigations.
When the instrument operates at low induction numbers, the apparent EC of the subsurface \( \sigma_a \) is linearly related to the quadrature component of the secondary magnetic field (Callegary et al., 2007).

\[
\sigma_a = \frac{4}{2\pi f s^2 \mu_0} \left( \frac{H_s}{H_p} \right)_q
\]  

(2.1)

where \( f \) is the measurement frequency in Hz, \( s \) is the transmitter/receiver coil spacing in meters, \( \mu_0 \) is the magnetic susceptibility in Henry’s/meter \((4\pi \times 10^{-7})\), and \( (H_s/H_p)_q \) is the ppm quadrature value.

After the calculation of apparent EC, the skin depth can be found by using equation 2.2

\[
\delta = \sqrt{\frac{2}{\sigma_a \mu_0 (2\pi f)}}
\]  

(2.2)

Where \( \delta \) is the skin depth. For handheld EMI instruments, the skin depth gives the depth of penetration for the EMI signal. The depth at which a target can be detected is known as the depth of investigation and is approximated as the square root of the skin depth (Huang, 2005). The GEM-2 is capable of investigating various depths since multiple frequencies are used, where lower frequencies have a greater depth of investigation and higher frequencies investigate the near surface (Huang and Won, 2003).

Another method to estimate the skin depth and depth of investigation for EMI instruments is to use a nomogram developed by Won (1980) (Figure 2.11). The frequency used for the investigation is shown on the right hand side, and if the conductivity of the subsurface is known a line is drawn from the frequency to the
conductivity. The location where the line crosses the middle scale provides an estimate of the skin depth for a given survey. Then, taking the square root of the skin depth provides an approximate estimate for the depth of investigation. Since the study does not have a good estimate of the electrical conductivity of the subsurface, resistivity survey values as determined by Oblinger (2008) will be used as representative electrical conductivity values. These representative values will be used to find the skin depth and subsequently the depth of investigation for the survey frequencies.

Figure 2.11: Skin depth nomogram showing an example of the range of frequencies commonly used in EMI surveys for typical conductivities found in igneous rocks with the range of skin depth expected (Won, 1980).
2.3.2 EMI Methods

The study uses the GEM-2 (Geophex Ltd), which is a frequency-domain electromagnetic instrument. The transmitter and receiver coil are separated by a distance of 1.66 meters and are mounted inside a shell which is carried at waist height (approximately 1 meter) above the earth’s surface. Six frequencies ranging from 330 to 20010 Hz are used to generate the transmitted magnetic field. The frequencies collected by the GEM-2 are 330, 570, 2070, 7050, 15210, and 20010 Hz. Bongiovanni et al (2007) point out lower frequencies can have large errors as relative to a given survey as the amplitude of the secondary magnetic field waves decrease with frequency.

Three EMI surveys were conducted at different time periods during 2009 to monitor spatial and transient changes in the electromagnetic response of the watershed (Figure 2.12). The first survey was conducted on May 3rd, May 19th, and June 9th, the second survey was conducted on July 8th, 16th, and 25th, and the third survey was conducted on September 2nd and 3rd. Mapping events correspond to before the monsoon season (survey 1), the start of monsoon season (survey 2), and late monsoon (survey 3).

During data collection, the GEM-2 was connected to a handheld computer and GPS to allow for the collection of ppm data from the instrument as well as real time latitude, longitude, and elevation data. With GPS data it is possible to import the location and GEM-2 ppm data into a geographical information system to allow for data processing and positioning of data on a map of the watershed. It was impossible to cover the exact same path in each survey due to accessibility limitations in the watershed. Once the monsoon season started, streams began to flow, the two water harvesting structures in the
watershed were full, and crops were growing in the agricultural fields. These obstructions were walked around and then the original survey line was returned to as soon as possible.

Figure 2.12: Location of all three GEM-2 surveys conducted in the Salri watershed and the location of the calibration point outside the watershed boundaries in the northeast corner.

The GEM-2 was calibrated using the free air calibration method in the lowlands outside of the watershed boundaries (Geophex, 2004; Figure 2.12). Air has an electrical conductivity of zero; therefore, if the instrument is lifted high in the air and allowed to
collect data, if zero is not collected for a given frequency, the average value collected while in the air can be used to adjust the collected field data. Field calibration occurred by hanging the GEM-2 from a tree in the watershed six meters off the ground for approximately 20 minutes. The quadrature ppm data were offloaded from the instrument and the average ppm value was found for each frequency. These calibration values are then used to adjust the field collected data for all three surveys, making the response for each survey frequency relative to the calibration location.

GEM-2 and GPS data were offloaded from the handheld computer for data analysis. Data was first imported onto a computer, ppm calibration values for each frequency were used to adjust collected field data, and lastly GPS data was used to import the calibrated quadrature ppm data into ESRI’s geographical information system, ArcMap.

In order to determine the depth of investigation, resistivity values for the basalts as determined by Oblinger (2008) are used to estimate the electrical conductivity. When sampling with lower frequencies, less than 10kHz, a resistivity value of 750 ohm-m is used, which represents the boundary between columnar and massive basalt. The electrical conductivity is therefore 0.001 Siemens per meter (the inverse of the resistivity). At frequencies greater than 10kHz, a resistivity value of 200 ohm-m is used, which is the median value for the weathered basalts. The electrical conductivity is therefore 0.005 Siemens per meter. The depth of investigation is then determined for a sample frequency given the estimated electrical conductivity and the subsequent skin
depth is found from the nomogram. For the GEM-2 sampling between 330 to 20010Hz, the depth of investigation ranges from 7 to 29 meters (Figure 2.13).

![Skin Depth Nomogram]

**Figure 2.13:** Range of skin depth for the sample frequencies collected during the GEM-2 surveys in the watershed.

2.4 Electromagnetic Induction Results

Data collected from the lowest two frequencies, 330Hz and 570 Hz, appear to be uniformly distributed around zero ppm, as seen by the nearly Gaussian distribution of
ppm data (Table 2.1; Table 2.2; Figure 2.14). Summary statistics for the two frequencies shows the mean being close to zero and the standard deviation being approximately two orders of magnitude, indicating a large spread in ppm values around zero. The uniform distribution is partially attributed to errors given that lower frequencies tend to have higher errors (Bongiovanni et al., 2007) as well as homogeneous material at depth providing a uniform signal across the entire watershed.

<table>
<thead>
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<th>Frequency 330Hz Summary Statistics</th>
<th>Frequency 570Hz Summary Statistics</th>
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<tr>
<td>Mean</td>
<td>Standard Deviation</td>
</tr>
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</tr>
<tr>
<td>Survey 2</td>
<td>33</td>
</tr>
<tr>
<td>Survey 3</td>
<td>7</td>
</tr>
</tbody>
</table>

**Table 2.1:** Summary statistics for the 330Hz frequency. **Table 2.2:** Summary statistics for the 570Hz frequency.

**Figure 2.14:** Histograms for the 330Hz and 570Hz frequency for each survey conducted in the watershed.
The first frequency to be investigated is 570Hz, which has a depth of investigation of approximately 26 meters. Data from the 570Hz frequency show a Gaussian distribution centered near a quadrature ppm value of zero. The negative values in the data suggest the ppm values are lower than at the calibration location and also are partially attributed to measurement noise. The errors and ppm values appear to be randomly distributed in the watershed, however, making it possible to conclude that there are not major spatial trends or patterns in the ppm data (Figure 2.15). The lack of trends is in agreement with the conceptual geologic model of the area. At depth the uplands and lowlands are underlain by competent basalt bedrock with few clays and low porosity. The uplands do not have an extensive weathered zone and basalt outcrops are common, whereas the lower watershed has a weathered zone of overlying the basalts. The lack of variability in the ppm values suggest that (i) the 570Hz data is sampling below the maximum depth of the weathered zone, and (ii) the bedrock across the watershed is relatively uniform with a low electromagnetic response.
The second frequency under investigation is 7050Hz, which has a much shallower depth of investigation at approximately 10m. The histogram from the 7050 Hz frequency shows fewer values below zero ppm and the values are skewed towards the right. The mean is greater than zero, and is seen to increase over time, suggesting ppm values increase over time with increased water content in the watershed. Furthermore, the skewness value shows the data are skewed towards higher ppm values indicating the electromagnetic response is higher than at the calibration location and more positive values are returned during the surveys (Table 2.3; Figure 2.16).
<table>
<thead>
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<th></th>
<th>Mean</th>
<th>Standard Deviation</th>
<th>Skewness</th>
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<td>310</td>
<td>459</td>
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<tr>
<td>Survey 3</td>
<td>481</td>
<td>577</td>
<td>1.15</td>
</tr>
</tbody>
</table>

Table 2.3: Summary statistics for the 7050Hz frequency.

Figure 2.16: Histogram from the 7050 Hz data collected from the GEM-2.

Data from the 7050 Hz measurements shows the electromagnetic response increases across the watershed moving from the uplands to the lowlands (southwest to northeast) (Figure 2.17). The trend is present in all three surveys suggesting that the change represents geologic variability for the shallower sampling depth at this higher frequency measurement. The lower ppm values observed in the uplands is consistent with the existence of thin soils and the common occurrence of basalt outcrops in this area. The larger ppm values in the lowlands are consistent with the increasing thickness of the
weathered basalts and alluvial zone in this area. Furthermore, the ppm values observed on the ridges of the lowlands are low, which is consistent with observations that the weathered zone thins and the basalts are closer to the surface. Overall trends in the electromagnetic response relative to the calibration location are useful for identifying areas where the weathered zone is thinner, mainly the ridges and uplands, versus those with thicker soils and weathered zone in the valleys and lowlands.

Figure 2.17: Data from the GEM-2 from 7050 Hz. The upper left shows the first survey, upper right is the second survey, and lower left is the third survey. Circled is the area which increases in conductivity as the season’s progress.
Toward the end of monsoon season, areas in the lower watershed are seen to have higher ppm values than in the first survey, which is attributed to increases in water storage. Since the alluvial and weathered zones are present, the trend is expected as increased rainfall will increase the water storage therefore increasing the electrical conductivity of the soils and subsequently increasing the electromagnetic response. In the uplands and the ridges, quadrature ppm values remain constant throughout time, suggesting the weathered zone is not present and the upland basalts have a lower capacity to store water than the lower watershed. The valleys in the uplands, however, are seen to increase in electromagnetic response as alluvial materials may be present due to erosion from the surrounding ridges. Overall, as the monsoon season progresses, the availability of water is higher in the lowlands as seen by the increase in ppm values due to the capability of the weathered zone and alluvial material to hold more water than in the upland basalts, characterized by lower water storage.

An area to the southwest from the middle of the watershed (circled area, Figure 2.17) is also seen to have an increase in ppm values as the monsoon progresses. In this area, water storage increases as the monsoon progresses, suggesting a localized area in the uplands with a significant weathered zone and water storage capacity.

The last frequency discussed here is 20010 Hz, the highest frequency collected in the field and also the frequency with the shallowest depth of investigation at approximately 7 meters. Histograms show data skewed towards positive values and fewer values below zero ppm (Table 2.4; Figure 2.18). The mean is three to five times higher than the 7050Hz frequency, indicating a much stronger electromagnetic response.
in the near surface. The skewness is greater than one for all three surveys, indicating that the values are skewed towards higher ppm values as seen in the histograms.

<table>
<thead>
<tr>
<th></th>
<th>Mean</th>
<th>Standard Deviation</th>
<th>Skewness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>948</td>
<td>935</td>
<td>1.08</td>
</tr>
<tr>
<td>Survey 2</td>
<td>878</td>
<td>1206</td>
<td>1.46</td>
</tr>
<tr>
<td>Survey 3</td>
<td>1426</td>
<td>1584</td>
<td>1.19</td>
</tr>
</tbody>
</table>

**Table 2.4:** Summary statistics for the 20010Hz frequency.

![Histograms for 20010 Hz](image)

**Figure 2.18:** Histogram from the 20010 Hz data collected from the GEM-2.

The results from the 20010 Hz frequency distinctly show high ppm values in the lowlands compared to the upper reaches of the watershed (Figure 2.19). As the monsoon season progresses the quadrature ppm values of the lowlands increase whereas the values in the uplands of the watershed remain relatively constant. The increase in the
electromagnetic response observed in the lower part of the watershed is attributed to
significant changes in the water storage as the water table rises with the advancement of
monsoon. The weathered and alluvial materials in this area have a high porosity,
approximately 57% (Personal communication, Ramakrishnan, 2009) and therefore
changes in water level represent a significant increase in the ppm values. In contrast, the
storage capacity of the fractured basalts that occur near the surface in the uplands and
ridges is low as the porosity can range from 5 to 15% for fractured basalts (Deoloankar,
1980). Changes in the water table in these areas represent a much smaller addition of
water and therefore the electromagnetic response remains relatively uniform through
time. The area marked in the southwest from the middle of the watershed is seen to have
increased ppm values as the monsoon proceeds. This is in contrast to the surrounding
areas and may indicate that the circled area acts as a localized region of soil water storage
(Figure 2.19). Overall, the increase in ppm values through time is attributed to the
increase in the EC of the subsurface caused by higher soil moisture from increased
rainfall. In contrast where the basalts outcrop ppm values remain approximately the same
through time suggesting lower water content.
Figure 2.19: Data from the GEM-2 at 20010 Hz. The upper left shows the first survey, upper right is the second survey, and lower left is the third survey. Circled is the area which increases in conductivity as the seasons progress.

From the EMI data collected with the GEM-2, it was seen that at low frequencies (570 Hz) the electromagnetic response relative to the calibration location across the watershed is relatively uniform spatially and temporally. At higher frequencies (7050Hz to 20kHz) the ppm values are lower in the upland than lowland portions of the watershed. This suggests that when the depth of investigation is large, i.e. for low frequencies, the bed-rock is sampled. As the measurement frequency increases and the depth of investigation decreases, the ppm values become more variable as the surficial weathered and alluvial zones are sampled. Over time the electrical conductivity increases in the
lowlands of the watershed due to the increase in water content. This increase in the conductivity is related to the increase in ppm values, attributed to local increases in water storage as monsoon progresses.

A qualitative interpretation of the GEM-2 data is useful for getting a general idea of the geology. It is known in the uplands the weathered zone is thin; therefore, the presence of a weathered zone aquifer is unlikely. Furthermore, the increase in conductivity in the lowlands as the monsoon season progresses indicates that there is more potential for water storage in the lower watershed than the upper watershed. In the end, a detailed geologic model of the watershed is needed if the thickness, aquifer storage potential, or extent of a weathered aquifer is to be determined.

2.5 Geological Investigation Discussion

One main goal from the geological investigations is to determine the boundary between the uplands and lowlands of the watershed. Since the lowlands is characterized as having a thicker weathered zone overlain by an alluvial material, more water can be stored in the lower part of the watershed versus the upper watershed. Furthermore, the porosity of the weathered and alluvial materials is approximately four to ten times higher than the fractured basalts. Knowing the delineation of the uplands to lowlands is important to determine the storage capacity of the groundwater system for the aquifer.

EMI surveys indicate that moving from the uplands to the lowlands there is an increase in the electromagnetic response spatially and temporally which is related to an
increase in the electrical conductivity. The lowlands have a higher water storage capacity as well as a thicker weathered zone causing the increase in ppm values.

When overlying the results from the EMI surveys on top of the geology of the watershed as determined from field observations and resistivity surveys in 2007, it is possible to see a boundary between the uplands versus the lowlands (Figure 2.20). The upper part of the watershed contains basalt flows 3 and 4, the alternating layers of columnar and massive basalts. These areas also have a lower electromagnetic response than the lower part of the watershed. Basalt flows 1 and 2 are in the lower watershed, again layers of columnar and massive basalts. However, the EMI survey shows this area has higher ppm values than flow 3 and 4. Since flows 1 and 2 are in the northeast of the watershed, are known to locally have a weathered zone and alluvial material from resistivity surveys and observations in open wells near the stream channel, and also have higher ppm values indicating increased water storage and the presence of weathered material, the lower watershed is defined by the boundary between flows 2 and 3. Using this boundary the area of the lower watershed is approximately 3.43x10⁵ m³ and the area of the upper watershed is 2.22x10⁶ m³.
Figure 2.20: EMI survey results from survey 2 at a sampling frequency of 7050 Hz overlain on a geologic map of the watershed. The upper and lower watershed boundary occurs between flow 2 and 3.

2.6 Conclusion

Geologic investigations were undertaken for a small watershed in rural Madhya Pradesh, India in order to better understand the geology of the region. The watershed is located in the Malwa Plateau of the greater Deccan Basalts. Regionally, the Plateau is characterized by flows of massive and columnar basalts. Geologic surveys were
undertaken in the summers of 2007 and 2009 in order to better understand the local geology and hydrogeology.

The main goal of the geologic investigations was to determine the boundary of the upper and lower aquifer to help determine the groundwater storage capacity of the watershed. It was found from geologic mapping, resistivity, and EMI surveys that the boundary between the uplands and lowlands is between basalt flow 2 and 3. Based upon this delineation the lower watershed comprises approximately 13% of the total watershed area (2.56x10^2 m^2), and the uplands cover approximately 87% of the watershed area.

Geologic mapping and electrical resistivity surveys conducted in 2007 were carried out to determine the thickness and number of basalt flows in the watershed. These surveys indicated the watershed is comprised of four major basalt flows. Individual flows have an upper columnar basalt group underlain by massive basalt. Overlying these layers in the lower part of the watershed is weathered basalt of about ten meters in thickness.

Investigations in 2009 consisted of electrical resistivity surveys as well as electromagnetic induction (EMI) surveys. One electrical resistivity survey was conducted near the approximate middle of the lower watershed, and when compared to the 2007 survey it was found the thickness of the weathered zone reached a shallower depth of only 6m as compared to 10m as found in 2007. The next layer was found to be massive basalt, the same as the 2007 survey. Then, the two surveys differ in that alternating layers of columnar and massive basalts were determined in the 2009 survey.
with the basement rock as columnar basalt. The interpretation of the 2007 survey did not attempt to estimate the lower boundary of the massive basalt or subsequent units.

EMI surveys were conducted at three times, before monsoon, early monsoon, and late monsoon in order to determine changes in the electromagnetic response of the subsurface relative to the calibration location. The collected ppm quadrature data was found to be uniform across the entire watershed at low sampling frequencies and depths of investigation near 25m. At higher frequencies, when the depth of investigation was approximately 7 to 10m, it was found the lowlands of the watershed have higher ppm values than the uplands, attributed to the presence of the weathered basalt and alluvial material. The ppm values at higher frequencies also increased moving from the uplands to the lowlands as the monsoon season progressed indicating increasing water storage in the lower watershed.

From the 2007 and 2009 surveys, a better understanding of the local geology was obtained, which will aid in determining water availability for the watershed. Knowing the lowlands of the watershed is characterized as having a weathered zone which can store water, as seen from the increase in the electromagnetic response though time, is useful when investigating the impact of the monsoon on near surface aquifer storage. Contrasting the lower watershed, the uplands are believed to store less water due to the uniform, lower ppm values for all three EMI surveys, indicating that as the monsoon progresses the water content is not increasing. Lastly, knowing the approximate extent, thickness, and location of the weathered zone will aid in developing the overall water balance for the region as the changes in groundwater storage can be estimated.
CHAPTER THREE
WATERSHED WATER BALANCE

Abstract

Developing a water balance provides a basic understanding of the hydrology that can be important when evaluating water resources. Field work was carried out for a small watershed in rural Madhya Pradesh, India, to investigate the flows of water into and out of the watershed over a period of one year, from May 2009 through April 2010. A watershed water balance is developed from climate, streamflow, and well data collected in 2009-2010. A watershed water balance is used to determine the approximate bounds on the annual volume of ET and estimate the volume of water lost as leakage from the surficial to deep aquifer system. Then, the overall water balance is broken into the individual flows of the surface water balance to better investigate the magnitude of the flows relative to precipitation and determine the net transfer between the surface and subsurface. After the water balances are developed water availability, which is water easily accessible as surface water or groundwater which can adequately supply water for domestic and agriculture use, is addressed for the study watershed.

The overall watershed water balance is used to investigate the flow of evapotranspiration (ET) throughout the year. By solving the overall water balance for a residual flow of water out of the watershed, bounds on ET range from $8.3 \times 10^5 \text{m}^3/\text{year}$ to $1.5 \times 10^6 \text{m}^3/\text{year}$, or expressed as a flux normalized to the watershed area is $323 \text{mm/} \text{year}$ to $525 \text{mm/year}$. An estimated rate of ET was determined using the Thornthwaite-Mather
(1957) approach, and it was found that ET is $1.75 \times 10^6 \text{m}^3/\text{year}$, or 681 mm/year, which is higher than the expected range for ET.

The overall water balance indicates a large flow of water into the surficial aquifer from a deep aquifer, which is not expected for the watershed. Since the estimate of ET is higher than the predicted range, the large leakage term is most likely the effect of an over prediction of ET and does not represent a large inflow of water.

From the surface water balance it was found that precipitation is a maximum in July and then peak streamflow occurs in July and August, after the watershed has become saturated. Annual streamflow is approximately 15% of the total precipitation. The net transfer from surface water to the subsurface is approximately $1.5 \times 10^6 \text{m}^3/\text{year}$, 601mm/year, which is approximately 80% of the total precipitation.

During investigation of the water balances, water availability was also considered. It was found the water table is able to recover after frequent irrigation in December and January to post monsoon levels which suggests farmers are not overdrawning the aquifer for irrigation. During the monsoon season, water is available as surface water in the streams and in the reservoirs of water harvesting structures located within the watershed. Then, at the end of the post-monsoon in December, surface water is still available in the reservoirs but not in the streams. Water lasts in the reservoir till the start of the dry season in April, at which point the main supply of water becomes the surficial aquifer system.
3.1 Introduction

In Madhya Pradesh, India, monsoonal rains provide approximately 90% of annual rainfall between June through September. Knowing the water balance and flows of water in and out of small watersheds is therefore critical for managing domestic and agricultural water use throughout the year (Singh and Sharama, 2002; Bobba et al., 1997). Water supplies are generally sufficient during and shortly after the monsoon season as water is found in surface water bodies and near surface aquifers. During the dry season, however, surface water supplies are scarce and near-surface aquifers can become depleted (Bobba et al., 1997). Due to temporal and spatial variations in groundwater and surface water supplies, proper allocation of water between domestic and agricultural use is important in order to use water in a sustainable manner. The best way to properly allocate water supplies is to determine the flows of water by studying a water balance (Burt, 2006).

A water balance quantifies the flows in and out of a watershed, thereby providing a better understanding of water demands and availability (Burt, 1999). Examples of how water balances have been used in the past include evaluation of artificial recharge from small dams (Oblinger et al., 2010; Srivastava, 2000), quantifying evapotranspiration for developing an irrigation schedule (Li et al., 2005), and estimating long term groundwater recharge rates (Lee et al., 2006). For this study, a water balance is developed to investigate the flows of the watershed, specifically focused on the surface water balance and the net transfer from the surface water system to the groundwater system. The study was focused in a small watershed in rural Madhya Pradesh, India for one year, from May
2009 through April 2010. Ultimately, the goal of this study is to provide improved insight to villagers and other policy makers in the region to guide water resource management practices throughout the year.

3.2 Overview and Conceptual Model of the Study Site

The research site is located in the Shajapur district of Madhya Pradesh, India (Figure 3.1). The approximate coordinates of the study area are 23.7°N and 76.1°E. The 2.56km² area is characterized by rolling hills, with a maximum elevation change of approximately 166 meters between the uplands in the southwestern portions of the watershed to the northeastern area of the watershed. Geology is characterized by the Deccan Basalts, primarily massive and columnar basalts overlain by up to ten meters of alluvial material and weathered basalt in low lying areas. Land use is primarily for agriculture, although the majority of the watershed is barren land with some small forests. Ephemeral streams, flowing only during and shortly after the monsoon season, originate in the uplands of the watershed and flow to a single channel that discharges to the northeast of the watershed.
The water balance has flows into and out of the watershed and is driven by two major components, the climate and geology of the region. Climatic factors affect the amount of precipitation, direct evaporation, and evapotranspiration (ET) in the watershed, whereas geologic factors affect the amount of runoff, recharge, and storage (Farmer et al., 2003).

When developing the conceptual model for the water balance, an important consideration is the physical boundaries of the model. Topographical highs for the study watershed are considered as no flow boundaries for surface water runoff. Groundwater divides are also assumed to correspond to topographical divides. Though groundwater and topographical divides are rarely identical, Narasimham (2006) and Limaye (2010) suggest that they may be close enough in hard rock aquifers, so they can be assessed to be
the same at the scale of the water balance. Groundwater is believed to discharge towards the northeast corner of the watershed, where the ridges come into the valley of the lower watershed and the hard rock geology observed on the ridges and uplands is no longer expressed at the surface as seen in open wells in the lower watershed. Cross basin transfers from outside the watershed boundary are not significant and are estimated as zero.

In addition to the physical boundaries of the region, the duration of the time period is also important. The duration of the water balance will cover one year, with a time step of one month. A one month time step makes it possible to monitor how the surface water system behaves before, during, and after the monsoon season.

With the boundaries of the model defined, a conceptual model is developed in order to better understand the interactions between the geology and water flows in the watershed. The conceptual model is shown as a cross section moving from the approximate middle of the uplands towards the northeast to the bottom of the watershed (Figure 3.2). The geology shown on the conceptual model was determined through field work conducted by Oblinger (2008) (Figure 3.3). Overall, four major basalt flows were found which are alternating layers of columnar and massive basalt. In the lower watershed a weathered basalt unit is present which is overlain by an alluvial material. The alluvial material was observed to be approximately 1m thick in the open wells which are near the main stream channel, although the overall extent of the alluvial zone is unknown for the watershed.
Figure 3.2: Cross section used for the development of the water balance.

Figure 3.3: Geology of the watershed as determined by Oblinger (2008) which is used to determine the geology of the cross section.
The conceptual models are developed for the monsoon season (Figure 3.4) and the dry season (Figure 3.5) and will be described in further detail.

Figure 3.4: Conceptual model for the watershed during the monsoon season showing the water table, streams, perched water tables, springs, and groundwater flow paths. The vertical exaggeration is approximately 9 times and the y-axis shows the elevation above mean sea level (amsl).
Figure 3.5: Conceptual model for the watershed during the dry season showing the water table, streams, a perched water table, and groundwater flow paths. The vertical exaggeration is approximately 9 times and the y-axis shows the elevation above mean sea level (amsl).

Since the study watershed is located in the Deccan Basalts two main aquifers are present, and their storage capacity depends upon the degree of weathering in the near surface and the amount of fracturing in the basalts (Kulkarni et al., 2000). An upper aquifer is defined in the alluvium and weathered basalt of the near surface. Geologic knowledge of the watershed indicates that weathered basalts are not extensive in the uplands. The surficial aquifer is accessed by large-diameter open wells in the lower watershed and is the primary water supply for villagers during the dry season. A deeper aquifer is defined to exist in the underlying fractured and jointed basalts (Limaye, 1994). The boundary elevation for the two aquifers is defined by the contact between the two massive basalt flows, flow 1 and 2. Furthermore, the boundary elevation coincides with the contact between flow 1 and the weathered basalts at the outlet of the watershed.
Despite defining these two regions of the subsurface as separate aquifers, the degree of connection between these systems is currently unknown within the study area.

Due to higher recharge in the monsoon and lower recharge during the post-monsoon and dry season, the water table is approximately one meter below the ground surface during the monsoon season and can be as much as five meters below the ground surface during the dry season. Typically in the Deccan Basalts, water enters the weathered aquifer as recharge from the surface during monsoon which saturates the weathered basalt and overlying alluvial material (Bobba et al., 1997). At the end of the monsoon season, rainfall is less, and leakage occurs from the near surface aquifer which recharges the basalt aquifer, dropping the water table in the surficial aquifer (Bobba et al., 1997).

Perched water tables are believed to exist in the uplands of the watershed at the contact between the massive and columnar basalts. The massive basalts are considered as more dense with a lower permeability than the columnar basalts (Oblinger, 2008); therefore, when water comes in contact with the massive basalts, the flow through the rock is much slower, perching the water table at this contact. Since the rainfall is higher during the monsoon, water does not have time to percolate through the massive basalts, therefore perching the water table above both massive basalt units. The water table above these massive basalts intersects the ground surface, creating localized springs. During the dry season however, no springs are present, suggesting only one perched water table is present as water has time to flow through the massive basalt connecting the upper and lower columnar basalt units.
Two major groundwater flow systems are believed to be present in the watershed, the regional groundwater flow and flow through the surficial aquifer. Throughout the entire year, groundwater flows from the uplands of the watershed, crosses the boundary dividing the surficial and deep aquifer system, and flows out the lowlands of the watershed. Also, recharged water flows through the columnar and massive basalt and then discharges into the weathered basalt of the lower watershed, acting as surficial groundwater flow. Recharge also occurs across the weathered basalts in the lower watershed during the monsoon. Water enters and remains in the surficial aquifer and then discharges at the outlet of the watershed through the weathered basalt and alluvial material.

During field work from May 2009 through April 2010, data were collected from a weather station, stream gauging stations, and wells located in and outside of the watershed boundaries. The weather station provides precipitation, temperature, relative humidity, and barometric pressure data. A stream gauging station monitored the flow of water at the outlet of the watershed. Wells located in and out of the watershed boundaries were monitored to track changes in groundwater levels. These data collected from the watershed will be used to evaluate the flows of the watershed water balance and then be applied to develop the surface water balance.

3.3 Water Balance Development

The water balance will be evaluated first for the entire watershed in order to determine the relative magnitude of the flows in and out of the watershed. The overall
water balance will provide a range of volumes for the flow of ET as well as predict the volume of leakage from the surficial aquifer to a deeper aquifer. Then a surface water balance will be considered to determine the relative magnitude of the flows compared to precipitation and estimate the net transfer of water between surface water and groundwater reservoirs. After developing each water balance and describing the flows which make up the balance, the procedure for calculating or estimating the flows and changes in storage will be described. Finally, the water balances will be considered to provide insights about water availability throughout the year.

3.3.1 Overall Watershed Balance

The overall balance for the watershed is illustrated in Figures 3.6 and 3.7. Reliable estimates for completing the water balance are available precipitation (Q_P), direct evaporation from surface water bodies (Q_E), streamflow (Q_S), and changes in the total volume of water stored in the watershed (ΔV), representing both surface (ΔV_{SW}) and groundwater storage (ΔV_{GW}). The primary surface water storage in the watershed (ΔV_{SW}) is from the 6.54x10^4 m^3 reservoir created by a dam built to capture and recharge monsoonal runoff. A second dam was also constructed in the watershed for this purpose (Figure 3.1), but the reservoir is much smaller and therefore neglected as a significant contribution to surface water storage. Subsurface storage is assumed to result primarily from changes in the depth of the water table. The flows of domestic surface and groundwater use are unknown for the watershed. Water use for the villagers and animal population was approximated as 400 m^3/month by Oblinger et al, (2010). Given that this
flow is small compared to other flows of the water balance, which are on the order $10^5$ to $10^6 \text{ m}^3/\text{month}$, this value is neglected. Water demands for agriculture, however, are represented in the balance as part of evapotranspiration.

Figure 3.6: Conceptual model for the overall water balance for the watershed overlain on the geology of the watershed. The vertical exaggeration is approximately 9 times and the y-axis shows the elevation above mean sea level (amsl).
Figure 3.7: Conceptual model for the overall water balance which show the individual components of flow for the surface and groundwater system. The flows which are in gray are not explicitly solved for, and are internal flows within the system.

The difference between the inflow ($Q_P$), outflows ($Q_S$ and $Q_E$), and net change in storage ($\Delta V$) produces a residual ($Q_R$) that is used to represent either flows or storage change that are unmeasured or error in the available estimates. These include evapotranspiration ($Q_{ET}$) and net losses to groundwater, including both groundwater flowing out of the watershed in the shallow aquifer ($Q_{GW}$) and leakage to deeper aquifer systems within the watershed ($Q_L$). Because topography is assumed to act as a boundary for groundwater flow, net inflows of groundwater are not considered.

The overall water balance considering only flows into and out of the watershed is given by

$$\frac{\Delta V}{\Delta t} = Q_P - [Q_E + Q_R + Q_S]$$

(3.1)

and if solved for the residual becomes
\[ Q_R = Q_P - \left[ \frac{\partial V}{\partial t} + Q_E + Q_S \right] \]  

(3.2)

The residual will be solved assuming it is an outflow of the watershed. Therefore positive values will represent a net outflow and negative values represent a net flow into the watershed. Although not well constrained, it is also possible to obtain independent estimates of evapotranspiration and outflows from the watershed in the shallow aquifer. As a result, estimates of leakage to deeper aquifers can also be obtained.

Some flows of the overall water balance are internal to the watershed, such as recharge, irrigation, and irrigation return flow. The effect of these activities can, however, affect the discharges from the watershed. For example, increased irrigation can increase evapotranspiration as crops take up and transpire more water than if no irrigation or farming practices were undertaken. These flows are important, but not considered in the overall water balance.

3.3.2 Surface Water Balance Model

The surface water balance is developed from the overall water balance. Flows into the surface water system are precipitation and baseflow, and flows out are direct evaporation, distributed infiltration, and surface runoff (Figure 3.8). The lower boundary for the surface water balance is the land surface across the watershed, and the upper boundary extends upward. Surface water bodies include two water harvesting structures which are dams on ephemeral streams as well as small stream channels draining the uplands, channeling into one stream which drains the entire watershed (Figure 3.1). The
northern reservoir is much larger than the southern reservoir based on field observations in the watershed.

**Figure 3.8:** Simplified conceptual model of the surface water balance, where the aim is to solve for the net transfer ($Q_{\text{NET}}$).

The surface water balance for the study watershed is developed to determine the magnitude of the flows relative to precipitation and the net transfer ($Q_{\text{NET}}$) of water between the surface and groundwater systems. The net transfer is considered as baseflow to streams ($Q_{BF}$) less the distributed infiltration ($Q_I$), which is water infiltrated either across the surface of the watershed from precipitation, from the reservoirs, or through losing streams. If the net transfer term is negative, then flow moves from the surface water system to the groundwater system, whereas if the flow is positive, water moves from the groundwater to the surface water system.

The surface water balance is defined as:

$$\frac{\Delta V_{\text{SW}}}{\Delta t} = Q_P + Q_{BF} - [Q_E + Q_I + Q_S] \quad (3.3)$$

where $\Delta V_{\text{SW}}$ is the change in surface water storage over time, $\Delta t$. Since the distributed infiltration and baseflow are unknown for the watershed, the surface water balance is solved for the net transfer (Eq. 3.4).
The net transfer is an important quantity as the goal of many watershed development projects is to reduce runoff losses and enhance groundwater recharge (Government of India, 2005). Quantifying the net annual transfer of surface water to groundwater for the overall watershed provides a context for evaluating the impact of specific recharge activities, such as the impacts of the dam constructed in the study watershed.

3.4 Determination of Volumetric Flows

In order to investigate the water balance for the watershed, the flows of water in and out must be quantified. The following describes how the inflow, outflow, and change in storage were determined for each flow in the water balance.

3.4.1 Explanation of Known Water Balance Inflows

The only inflow to the water balance is precipitation. Precipitation data was collected from a tipping bucket rain gauge located in the approximate middle of the watershed (Figure 3.1). Data is collected on a 15 minute time step, and then summed over the entire month to provide the total monthly rainfall (Figure 3.9). The rate of precipitation is then multiplied by the area of the watershed to turn the precipitation rate into a volumetric flow. Due to a malfunction in the weather station during January and February 2010 data for these months was not available. Instead, rainfall, as well as

\[
Q_{NET} = \frac{\Delta V_{SW}}{\Delta t} + [Q_E + Q_S] - Q_P
\]
temperature data, averaged over 1901-2002 from Shajapur, a city approximately 60km to the southeast of the study site, is used to replace the missing data (IWP, 2002).

![Yearly Precipitation Graph](image)

**Figure 3.9:** Monthly rates of rainfall for the watershed during the 2009-2010 season.

3.4.2 Explanation of Known Water Balance Outflows

Outflows from the watershed are streamflow, direct evaporation from surface water bodies, evapotranspiration, and groundwater flows. The method for calculating or estimating each flow is described below.

Stream stage was monitored on the main stream channel at the outlet of the watershed (Figure 3.1). The stage was measured every 15 minutes with the use of a pressure transducer housed in a stilling well (Figure 3.10). The stage is converted to discharge using the rating curve given in Figure 3.11. The rating curve was developed by measuring profiles of velocity and stream depth using a pygmy meter when different amounts of flow occurred in the stream. Discharge is then calculated by multiplying the velocity at each location by the cross-sectional area calculated for a section of the stream (see Appendix B for details). The rating curve might inaccurately predict the volume of
flow at high stages as no data were available for these events. Monthly discharge was calculated by summing the discharge for each daily time step over a given month.

**Figure 3.10:** Stream stage and rainfall on the left figure, and stream discharge as calculated from the rating curve on the right.

![Rating Curve](image)

**Rating Curve**

\[ y = 34.209x^{0.2212} \]

\[ R^2 = 0.8526 \]

**Figure 3.11:** Rating curve for the lower stream gauge which was developed from stream gauging done at this location.

Direct evaporation from surface water bodies in the watershed is estimated with pan evaporation data and knowledge of the surface area of the northern water harvesting structure. Pan evaporation data for each month during 2007 was collected in Agar, a
small town approximately 10km south of the study watershed (Oblinger, 2008). The same values from 2007 are used as these are the best available data. The depth of water evaporated each month was then multiplied by the surface area of the northern, larger water harvesting structure. Surface area ($A_{\text{WHS}}(h)$) was calculated using the area to stage relationship given for the reservoir by Oblinger et al. (2010) (Eq. 3.5).

$$A_{\text{WHS}}(h) = 768 h^2 + 2277 h, \ R^2 = 0.988 \quad (3.5)$$

If the stage is zero, then no evaporation occurs for the month. The stage was taken at the start and end of a month, the surface area was found, and then the evaporation rate was multiplied by the surface area to return the volume. The volume at the start of the month was subtracted from the end of the month to determine the monthly volumetric flow (Figure 3.12). The estimation of evaporation is an underestimate as only one structure is considered and losses directly from streams are ignored. The overall impact on the water balance is minimal, however, as the rate of evaporation is much smaller than the other flows in the watershed.

![Yearly Direct Evaporation](image)

**Figure 3.12:** Yearly direct evaporation from surface water bodies.
ET is estimated using the Thornthwaite-Mather method (Thornthwaite and Mather, 1957) with average monthly temperature data collected from the weather station. Monthly temperature data is used to calculate a heat index, which is then used to determine the unadjusted potential ET (PET) for each month. Depending upon the latitude of the site (23.7°N), a factor relating to the total amount of sunlight hours during the month is found from a look up table (Thornthwaite and Mather, 1957), and a value for adjusted PET is calculated.

Actual ET (AET) is generally less than PET, as sufficient water supplies from rainfall and soil moisture must be available to achieve the rate of PET. If the amount of rainfall entering the watershed is greater than the PET, then AET is the same as PET, which is the case during parts of the monsoon season. As the monsoon season ends and rainfall is less than the PET, water is lost from soil storage. At this point the rate of AET is the amount of rainfall for the month plus the water lost from soil storage each month up to the value of PET.

An important consideration is different rates of ET for agricultural fields and other areas in the watershed. Accounting for these differences is the methodology to account for irrigation during December and January. To be conservative, soil storage in agricultural fields is estimated as 100 mm, which is the amount of water stored in a clayey loam with shallow rooted crops, such as beans which are found in the majority of fields (Thornthwaite and Mather, 1957). The rest of the watershed is represented as having 25mm of soil water storage as the terrain is barren, there are few forested areas, and vegetation is mostly field grasses with shallow roots. When precipitation cannot
account for all the water lost as AET, water is lost from soil storage. During December and January, irrigation occurs, replacing any water which may have been lost from storage. If the soils are saturated for a month, then the AET will equal the amount of storage in the soils plus the precipitation up to the value of PET, thereby increasing the rate of AET. The rate of AET for each month is then multiplied by the fractional area, 25% for agricultural areas and 75% for the rest of the watershed (FES, 2003), and summed to convert to a volumetric flow (Figure 3.13).

![Yearly Evapotranspiration](image)

**Figure 3.13:** Yearly evapotranspiration for the watershed found using the Thornthwaite-Mather Method (1957).

The flow of groundwater out of the lower watershed is water which moves through the subsurface above the boundary between the surficial and deep aquifer system. The boundary is defined as the contact between the weathered basalt and massive basalt at the outlet of the watershed, which coincides with the boundary between massive basalt flows 1 and 2 (Figure 3.3). Groundwater flow out of the watershed is estimated using Darcy’s Law (Eq. 3.6).
\[ Q_{gw} = AK \frac{\Delta h}{L} \]  

where \( Q_{gw} \) is the volumetric flow of groundwater, \( A \) is the cross sectional area of the aquifer (thickness times width), \( K \) is the hydraulic conductivity, and \( \Delta h/L \) is an estimate of the hydraulic head gradient in the aquifer.

Estimates for the hydraulic conductivity are available from analysis of soil and weathered basalt samples collected during field work undertaken during the summer of 2009 and then tested in a laboratory at the Indian Institute of Technology-Bombay. It was found the weathered basalts have a hydraulic conductivity of \( 10^{-4} \) m/s and soils have a hydraulic conductivity of \( 10^{-7} \) m/s (personal communication: D. Ramakrishnan, 2009; Table 3.1). The estimates of hydraulic conductivity are most likely under estimating the true hydraulic conductivity as these values are based upon core samples and not in-situ field tests (Olsen and Daniel, 1981).

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Hydraulic Conductivity (k)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( S_2 )-1</td>
<td>( 8.4 \times 10^{-5} ) m/s</td>
</tr>
<tr>
<td>( S_2 )-2</td>
<td>( 2.3 \times 10^{-4} ) m/s</td>
</tr>
<tr>
<td>( S_2 )-3</td>
<td>( 3.7 \times 10^{-4} ) m/s</td>
</tr>
<tr>
<td>( S_2 )-4</td>
<td>( 2.4 \times 10^{-4} ) m/s</td>
</tr>
<tr>
<td>( S_3 )</td>
<td>( 1.8 \times 10^{-6} ) m/s</td>
</tr>
<tr>
<td>( S_5 )-1</td>
<td>( 1.4 \times 10^{-8} ) m/s</td>
</tr>
<tr>
<td>( S_5 )-2</td>
<td>( 1.0 \times 10^{-6} ) m/s</td>
</tr>
</tbody>
</table>

\( S_2 \) – In-situ weathering profile of Basalt  
\( S_4 \) – In-situ soil at uplands  
\( S_5 \) – In-situ soil in the valley with a layer of alluvium, then in-situ weathered basalts

Table 3.1: Hydraulic conductivity data from the study watershed determined from soil samples (personal communication, Ramakrishnan, 2009).

The area of the aquifer is calculated as the thickness times the width of the aquifer. The thickness is estimated from geologic investigations conducted in 2007 and
2009. It was determined the maximum thickness of the alluvial material plus weathered basalts is 10m in the lower watershed, which will be considered as the thickness of the surficial aquifer. The width of the aquifer at the outlet of the watershed is estimated to be 300m (Figure 3.14). This location was chosen as the ridges that act as groundwater boundaries decrease in elevation and are incorporated into the lower elevations of the watershed at the outlet.

![Map Used for Calculations of Groundwater Flow Out](image)

**Figure 3.14:** Location of wells, and the location where the length and width of the lower aquifer were determined in order to use Darcy’s Law to calculate groundwater flow out of the watershed.

The last piece of information is the head gradient in the lower watershed. The length from the middle of the lower aquifer to the outlet is 515 meters. The change in head is calculated by knowing the water table elevation in the middle of the watershed from wells 12, 14, and 18, and at the outlet of the watershed from wells 25 and 26 (Figure
If a well was pumped on the day of a water level measurement, then the water table elevation was not included in the average. Figure 3.16 shows the hydraulic gradient calculated from the change in head and the distance between the two clusters of wells.

![Figure 3.15: Water table elevations for the wells located in the middle of the watershed and the outlet of the watershed.](image-url)
Figure 3.16: Hydraulic gradient between wells from the middle to the outlet of the watershed.

3.4.3 Explanation of Known Changes in Storage

The change in surface storage is estimated as the change in volume of the northern water harvesting structure (Figure 3.1), which is a minimum value, as only the larger reservoir in the watershed has been included, and the change in storage for streams is ignored. A relationship between the stage \( h \) and the volume \( V(h) \) of water stored in the structure was developed using a geographical information system (Eq 3.7; Figure 3.17)

\[
V(h) = 973.56 h^{2.4225}, \quad R^2 = 1.0
\]  

(3.7)

and used to calculate the volume of water in the structure at the start and end of each month (Figure 3.18). The monthly change in surface storage for the watershed is represented by the difference between the volumes at the end of the month from the start of the month.
Figure 3.17: Volume to stage relationship developed for the water harvesting structure.

Figure 3.18: Change in surface water storage for the water harvesting structure.

In order to determine the changes in groundwater storage across the watershed, a water table fluctuation approach is utilized where the change in the water table over time is multiplied by the specific yield of the aquifer (Healy and Cook, 2002; Eq 3.8).

\[
\Delta V_{GW} = \frac{S_y^{upper} \Delta h}{\Delta t} * A_{aquifer}^{upper} + \frac{S_y^{lower} \Delta h}{\Delta t} * A_{aquifer}^{lower} \tag{3.8}
\]

Here \(\Delta V_{GW}\) is the change in groundwater storage over time, \(S_y^{upper}\) is the specific yield of the upper aquifer, \(S_y^{lower}\) is the specific yield of the lower aquifer, \(\Delta h\) is the change in head over time \(\Delta t\), \(A_{aquifer}^{upper}\) is the area of the upper aquifer, and \(A_{aquifer}^{lower}\) is the area of
the lower aquifer. Since the upper and lower aquifers differ in geology, two values for the specific yield are used to better estimate the change in groundwater storage.

The specific yield of the lower watershed is determined from samples of weathered basalt and soil, whereas the specific yield for the uplands is estimated using data from literature. The same samples collected during 2009 which determined the hydraulic conductivity were used to determine the specific yield. The specific yield was calculated using the porosity and moisture content of the samples (Table 3.2). Since the samples were collected during the dry season, the moisture content is assumed to be directly related to the specific retention of the soils. The effective porosity of a rock or soil is equal to the specific yield plus the specific retention (Fetter, 2001); therefore, the specific yield is the porosity less the specific retention. The specific yield for the lower watershed, as determined by these samples, will be considered as 0.425 (personal communication: D. Ramakrishnan, 2009). The specific yield for the fractured basalts of the uplands is approximately 0.01 to 0.03 (Kumar, 2006; Singhal, 1999), and the average of these values, 0.02 will be used as the specific yield of the upper aquifer.
<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Moisture Content (%)</th>
<th>Porosity (%)</th>
<th>Specific Yield (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S₂-1</td>
<td>20</td>
<td>64</td>
<td>44</td>
</tr>
<tr>
<td>S₂-2</td>
<td>21</td>
<td>51</td>
<td>30</td>
</tr>
<tr>
<td>S₂-3</td>
<td>28</td>
<td>55</td>
<td>27</td>
</tr>
<tr>
<td>S₂-4</td>
<td>13</td>
<td>59</td>
<td>46</td>
</tr>
<tr>
<td>S₄</td>
<td>7</td>
<td>59</td>
<td>52</td>
</tr>
<tr>
<td>S₅-1</td>
<td>13</td>
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</tr>
<tr>
<td>S₅-2</td>
<td>9</td>
<td>53</td>
<td>44</td>
</tr>
</tbody>
</table>

S₂ – In-situ weathering profile of Basalt  
S₄ – In-situ soil at uplands  
S₅ – In-situ soil in the valley with a layer of alluvium, then in-situ weathered basalts  

**Table 3.2:** Specific yield data from the study watershed determined by soil samples (personal communication, Ramakrishnan, 2009).

The area of the uplands and lowlands were determined based upon electrical resistivity and electromagnetic induction surveys conducted in 2007 and 2009. The area of the lower watershed is approximately $3.43 \times 10^5 \text{m}^2$, 13% of the watershed, and the area of the upper watershed is $2.22 \times 10^6 \text{m}^2$, 87% of the watershed (Figure 3.19).
Figure 3.19: Area of the uplands versus the lowlands based upon geological mapping and electrical resistivity surveys.

The change in the water table elevation is also required to find the change in groundwater storage. To find the change in the water table, the average depth to water for wells in the lower watershed was calculated each time water level readings were taken (Figure 3.20). These values were then subtracted to give the change in the water table during each time step. Wells which had been pumped on the day of the water elevation measurements were not used in the average. The last step is to sum the change in storage from the uplands and lowlands to calculate the change in storage over the entire watershed (Figure 3.21)
Figure 3.20: Average depth to water for wells in the watershed used to calculate the change in volume storage.

Figure 3.21: Change in groundwater storage for the watershed.

With the inflow, outflows, and changes in storage quantified, it is now possible to investigate the overall water balance and the surface water balance for the watershed.

3.5 Evaluation of the Water Balance

The water balance for the watershed will be evaluated to determine the flows of water in and out to better understand water availability. First, the overall watershed water
balance will be used to solve for the flow of residual water, assuming all other flows are unknown. Then, uncertainty will be added to the model as groundwater and then estimates of ET are added, while still solving for the residual. This approach will provide a range for ET which will help determine if our estimate is an accurate representation of the true flow within the watershed. The overall water balance will also be used to investigate the leakage from the surficial aquifer to a deep aquifer. After looking at the overall water balance, the surface water balance will be investigated to determine the magnitude of the other flows compared to precipitation as well as to calculate the net transfer of water from the surface water to groundwater systems. Lastly, the water balances will be used to provide insights to water availability for the year.

3.5.1 Overall Watershed Water Balance

The overall watershed water balance, Equation 3.2, is used to investigate the residual flow assuming that neither groundwater outflows or ET is known (Figure 3.22; Table 3.3). By assuming the leakage term to the deep aquifer is zero and both groundwater and ET are unknown for the watershed, bounds on these flows can be established. The value of residual flow provides a maximum estimate for the amount of water lost as ET and groundwater flow. If the groundwater flow is then also assumed to equal zero, a maximum bound on the ET is determined. Using this approach, it is found the maximum estimate for ET is $1.5 \times 10^6 \text{m}^3/\text{year}$, which is 95% of the yearly precipitation.
Overall watershed water balance. The flow of direct evaporation and the change in surface water storage are not included as the flows are orders of magnitude smaller than the other flows and do not greatly impact the residual flow. The residual includes the groundwater, leakage, and ET volumes.

![Overall Water Balance](image)

Table 3.3: Yearly volumetric (m$^3$) flows for the overall water balance. $Q_P$ = precipitation, $Q_E$ = Direct evaporation, $\Delta V_{SW}$ = Change in surface water storage, $\Delta V_{GW}$ = Change in groundwater storage, $Q_S$ = Streamflow, and $Q_R$ = Residual

<table>
<thead>
<tr>
<th></th>
<th>$Q_P$</th>
<th>$Q_E$</th>
<th>$\Delta V_{SW}$</th>
<th>$\Delta V_{GW}$</th>
<th>$Q_S$</th>
<th>$Q_R$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Yearly Flow</td>
<td>1.8x10$^6$</td>
<td>2.0x10$^4$</td>
<td>69</td>
<td>4.1x10$^4$</td>
<td>2.7x10$^5$</td>
<td>1.5x10$^6$</td>
</tr>
</tbody>
</table>

3.5.2 Watershed Balance with Groundwater Outflow

The next consideration on the overall water balance is to add in the estimate for the groundwater flow out of the watershed. The leakage to the deeper aquifer is still assumed to equal zero, and when solving for the residual, a minimum bound on ET is determined (Figure 3.23; Table 3.4). The residual for this scenario is 8.3x10$^5$m$^3$/year, approximately half of the maximum bound on ET.
Figure 3.23: Watershed balance with the groundwater flow. Direct evaporation and the change in surface storage are not shown. The residual flow is the volume of ET and leakage.

Table 3.5: Yearly volumetric (m$^3$) flows for the overall water balance with the calculated groundwater flow included in the calculation of the residual. All values are in m$^3$. $Q_{R1}$ is the residual flow with the addition of groundwater flow.

<table>
<thead>
<tr>
<th></th>
<th>$Q_P$</th>
<th>$Q_E$</th>
<th>$\Delta V_{SW}$</th>
<th>$\Delta V_{GW}$</th>
<th>$Q_S$</th>
<th>$Q_{GW}$</th>
<th>$Q_{R1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Yearly Flow</td>
<td>1.8x10$^6$</td>
<td>2.0x10$^4$</td>
<td>69</td>
<td>4.1x10$^4$</td>
<td>2.7x10$^5$</td>
<td>1.5x10$^6$</td>
<td>1.8x10$^6$</td>
</tr>
</tbody>
</table>

3.5.3 Watershed Balance with Groundwater Outflow and Estimated ET

The final consideration for the overall watershed balance is to include the estimate of groundwater flow and ET to calculate the residual flow. In this case, the residual flow is the volume of water lost as leakage to the deeper aquifer system. If the flow is negative water is moving into the surficial system, whereas if the flow is positive water is recharged to the deeper aquifer from the surficial aquifer (Figure 3.24; Table 3.5.).
Figure 3.24: Overall watershed water balance with the addition of groundwater and an estimation for ET. Direct evaporation and the change in surface storage are not shown. The residual flow is the leakage.

<table>
<thead>
<tr>
<th>Yearly Flow</th>
<th>$Q_P$</th>
<th>$Q_E$</th>
<th>$\Delta V_{SW}$</th>
<th>$\Delta V_{GW}$</th>
<th>$Q_S$</th>
<th>$Q_{GW}$</th>
<th>$Q_{ET}$</th>
<th>$Q_{R2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fraction of Precip (%)</td>
<td>--</td>
<td>1</td>
<td>&lt;&lt;1</td>
<td>2</td>
<td>15</td>
<td>23</td>
<td>95</td>
<td>--</td>
</tr>
</tbody>
</table>

Table 3.5: Yearly volumetric ($m^3$) flows for the overall water balance with the calculated groundwater flow and estimated ET included in the calculation of the residual. All values are in $m^3$. $Q_{R2}$ is the residual flow with the addition of groundwater flow and ET.

ET was estimated by the Thornthwaite-Mather method and the flow was estimated at $1.7 \times 10^6 m^3$/year. The volume is larger than the maximum flow of $1.5 \times 10^6 m^3$/year expected for the watershed, indicating error in the calculation and an inaccurate representation of the true ET for the watershed. Knowing that ET is an overestimation is important when considering the water balance flows and storage of the groundwater system, since the large yearly flow of ET indicates large inflows from the deeper aquifer to the shallow aquifer. The primary reason for the large inflow from the
deeper aquifer is to account for the large outflow of ET, which may not actually be the case for the watershed.

The overall water balance does not provide any information on the interactions between the surface and groundwater systems; therefore, some of the flows of the overall balance will be applied to a surface water balance to determine the net transfer of water from the surface water to groundwater system.

3.5.4 Surface Water Balance Evaluation

The surface water balance equation, Equation 3.5, is solved for baseflow less distributed infiltration, $Q_{BF} - Q_L$, which is the net transfer from the surface water to groundwater system, $Q_{NET}$. If the net transfer is negative for a given month, then water is lost to the subsurface as distributed infiltration; however, if the flow is positive, then water is coming to the surface water system as baseflow. The results from the surface water balance are shown in figure 3.25 and Table 3.6.

![Surface Water Balance](image)

**Figure 3.25:** Results from the surface water balance. The left figure shows precipitation, streamflow, and the net transfer, which are orders of magnitude larger than the direct evaporation and change in surface waters storage shown on the left.
The flows of direct evaporation and the change in surface water storage from the water harvesting structure are approximately two orders of magnitude smaller than the other flows for the surface water balance. Since these flows are small compared to the other flows, they do not play a major role in impacting the overall surface water balance.

Precipitation is at a maximum in July at the same time the streamflow begins. Stream flow is the highest in July and August and then decreases through November. The flows in October and November are caused by small storm events, and constant streamflow does not last from the monsoon into the winter months. After the storm events, the flows begin to decrease, suggesting the stream is losing at the outlet of the watershed. Yearly streamflow accounts for approximately 15% of the yearly precipitation.

The net transfer as determined by baseflow less distributed infiltration is negative on the yearly time scale, indicating flow from the surface to the subsurface. The net transfer is almost a mirror image of the precipitation. When precipitation is highest in July, the net transfer is also the highest. Then, as precipitation becomes less in the dry season, the net transfer to groundwater is also lower. The net transfer to the groundwater system accounts for approximately 80% of the yearly precipitation.
3.6 Water Availability and Water Balance Results

Evaluation of the water balances can be used to investigate water availability throughout the year. Water availability is water that is easily accessible in surface water and groundwater systems which can adequately supply water for domestic and agricultural use. As seen from the overall water balance as well as the surface water balance, water supplies are sufficient during and shortly after the monsoon season, as surface water bodies are present in the watershed. Then, the main source of water is found in the groundwater system.

The watershed water balance shows that the flow of water coming into the study area from precipitation is $1.8 \times 10^6 \text{m}^3/\text{year}$, and 84% of the precipitation falls during the monsoon season. Streamflow out of the watershed accounts for 15% of the yearly rainfall and groundwater flow out is approximately 23% of the yearly rainfall. Direct evaporation plays a very minor role in the overall water balance, as only 1% of yearly precipitation is lost through direct evaporation from surface water bodies; however, if the ET was representative of what actually occurs in the watershed, 95% of the precipitation would be lost to ET.

The change in groundwater storage is positive in the monsoon season since the rainfall is high and the near surface aquifer becomes saturated. Then, in February and March, the change in storage is also positive. During December and January, crops are growing and being irrigated with water from the wells. Then, crops are harvested at the end of January and beginning of February so irrigation stops. As water levels recover
from pumping, the water table increases which provides a positive change in storage. By
the end of March, and into April, the groundwater system has recovered and the water
level is static across the lower watershed, hence the zero change of storage in April. The
water table then begins to decline as the dry season progresses.

Groundwater flow out of the watershed is much smaller than the other flows of
the water balance. The flow is higher during the dry season, indicating the hydraulic
gradient between the middle and lower watershed is higher. Then, groundwater flow is
lower in the monsoon season, indicating a smaller hydraulic gradient and a more uniform
water table elevation across the lower watershed. Overall, since groundwater is always
positive, the flow is out of the watershed, following the general topography of the
watershed.

Surface storage is positive in June through August as the water harvesting
structure fills, and then the change in storage is negative as the structure drains at the end
of the monsoon season. A similar pattern is seen with the streams, where peak stream
flow occurs during the middle of the monsoon and then becomes less moving through the
end of monsoon to the post monsoon.

The net transfer from the surface water system to the groundwater system was
estimated to be 80% of the yearly rainfall. The net transfer was highest when the rainfall
was highest, and decreased into the dry season as rainfall also decreased. Overall, the
volume of water lost through the net transfer is important, as this water may help to
recharge to surficial aquifer which is accessed by the local villagers.
3.7 Conclusion

A water balance for the watershed was developed in order to determine an estimated range for the volume of water lost as ET, estimate the leakage from the surficial aquifer to deep aquifer, approximate the net transfer of water from the surface water to groundwater system, and investigate water availability in the watershed. It was found from the overall water balance the flow of ET should range from $8.3 \times 10^5 \text{m}^3/\text{year}$ to $1.5 \times 10^6 \text{m}^3/\text{year}$. On the yearly time scale, approximately 80% of the total rainfall is transferred from the surface water to groundwater system. Lastly, water availability is greater during the monsoon season when surface water bodies are present, and then during the dry season the main source of water is the surficial aquifer.

The Thornthwaite-Mather approach is used to estimate ET for the watershed. The estimate of ET was found to be $1.7 \times 10^6 \text{m}^3/\text{year}$, higher than the range as determined by the overall water balance. To account for the large flow of ET, the overall water balance indicated that water was moving from the deep aquifer system to the surficial aquifer system. It is believed that large flows from the deeper aquifer into the watershed do not exist, and the large inflow may be caused by poorly quantifying ET as well as poor estimates of the groundwater storage through time.

After the watershed water balance was investigated the surface water balance was investigated to determine the net transfer between the surface and groundwater system. The change in surface water storage and direct evaporation are very small compared to the other flows and does not impact the overall surface balance. The net transfer is from the surface water to groundwater system on a yearly time scale, and the volume of water
lost as distributed infiltration is approximately $1.5 \times 10^6 \text{m}^3/\text{year}$, slightly more than 80% of the total precipitation. The net transfer to groundwater has a positive impact, as some of this water may recharge the surficial aquifer system, providing increased water supplies to villagers during the dry season.

Overall, from the watershed water balance and surface water balance, it is determined that water supplies are sufficient during the monsoon season and then become scarce during the dry season. Water availability is high from July through September as the water harvesting structures fill, streams begin to flow, and there is a decrease in the depth to water in wells across the watershed. Then, as the monsoon ends, water is still available in the reservoirs and streams although the volumes are starting to decrease as the watershed drains. During the winter season, and the dry season, water is available in the wells, and can be used to irrigate crops, as per the current practice. Since the net transfer from surface water to groundwater is positive and the yearly change in groundwater storage is positive, it is believed current water use practices are sustainable as the water table elevation is not being lowered. Then, as the monsoon season begins again, the aquifer is recharged, streams begin to flow, the reservoirs reach capacity, and the cycle begins again.
CHAPTER FOUR

INVESTIGATION OF WATER HARVESTING IN A SMALL WATERSHED IN RURAL INDIA

Abstract

In this study we use data collected from a small watershed in rural India to test the validity of a water balance model developed by Oblinger et al. (2010) for a small dam in Madhya Pradesh, India. With 15% of the world’s population, but only 6% of the world’s water resources, India is one of many countries facing water scarcity. To address their scarcity, India’s government is promoting the use of water harvesting methods, such as the construction of small earthen dams, to capture monsoon runoff and artificially recharge groundwater. Despite the rapid increase in the number of dams, there remains little information to quantify the values of these efforts to determine if the structures are practical and whether or not they are helping the current water situation (Sukhija et al., 1997). Oblinger et al. (2010) previously developed a water balance model to investigate the reservoir fluxes; specifically, the amount of water that moves to recharge, and the length of time water remains in the structure. Subsequent field work was undertaken in the same watershed in 2009-2010 to further investigate the effectiveness of the water harvesting structure using the same model developed by Oblinger et al. (2010). The reservoir response predicted using the model parameters determined by Oblinger gave predictions of stage with a root mean square error of 0.74 meters. The model parameters were considered to determine if adjusting inflow or outflow parameters would better constrain stage predictions. After manually adjusting inflow parameters for a better fit
using data from the 2009-2010 monsoon seasons, new parameters gave predictions of stage with an error of 0.53 meters. In order to consider changing all parameter values at the same time, a Monte Carlo approach was also used to more fully explore the parameter space than was possible in the manual fitting. The Monte Carlo approach was used to minimize the error between the true and predicted stage to best fit parameter values, and in this case, the best fit data parameters predicted the actual stage of the reservoir with an error of 0.45 meters. Original parameter estimates better fit the structure draining, where newly fit parameter values better fit the filling and draining of the structure. Overall, the main goal for the model is to predict the residence time and volume of infiltration to the subsurface. Original model parameters predict the residence time within 35 days and predict infiltration as approximately two times the maximum volume of reservoir. Best fit model parameters estimate the residence time within fifteen days and predict infiltration as approximately five to six times the maximum volume of the reservoir.

4.1 Introduction

Water scarcity is an increasingly important problem for the world’s population. Water scarcity as defined by the United Nations Environmental Program Global Environmental Outlook is the amount of water for industrial, agricultural, and domestic use divided by the total amount of renewable water in surface bodies and shallow aquifers. Using this definition, high water scarcity occurs when 40% of the total available water supplies are withdrawn (UNEP, 2000). Currently, two billion people face
a high level of water scarcity, and it is predicted that by 2025 two thirds of the world’s eight billion people will face high water scarcity (Vörösmarty et al., 2000).

India is one of the many countries facing water scarcity. In order to overcome water shortages, the construction of small earthen dams, known as water harvesting structures (WHS) or percolations tanks, has become very common for capturing monsoonal runoff in reservoirs to promote the artificial recharge of groundwater (Sukhija, 1997). The effectiveness of these structures, however, is difficult to quantify given a limited amount of technical infrastructure and expertise in the regions where they are constructed. Anecdotes by villagers suggest reservoirs keep water levels in downstream wells higher for a longer period into the dry season. Despite this qualitative evidence, the fluxes of water lost to evaporation, seepage, or domestic and agricultural use are poorly understood, thereby making management of the reservoirs difficult. Quantifying the effectiveness of the reservoirs for recharging groundwater is hindered by a lack of data such as precipitation and evaporation, technical skills of villagers or investigators, funds, and technical resources such as equipment or computers (Oblinger, 2008; Bobba, 1997).

In order to better quantify the impact of reservoirs on groundwater recharge, the scientific community has used tracer tests with environmental chloride (Sukhija, 1997), water balance methods (Srivastava, 2000), water table fluctuations (Sharda, 2006), and chloride mass balances (Sharda, 2006). These methods are effective, but can be difficult as taking chloride concentrations might be out of the scope of work for non-technical people, or making accurate, direct calculations of evaporation from a field site could be
challenging if the proper instruments are not available. On the other hand, a simple volume balance model can be created with knowledge of climate data and the land area draining into the structure. With limited inputs and data which is generally available to the public through the internet (IWP, 2002), a volume balance is an easy approach to quantify the impacts of water harvesting.

The objective of this study is to evaluate whether a simple water balance model developed and calibrated with post-monsoon data from 2007 for a reservoir in Madhya Pradesh, India (Oblinger et al., 2010) can be used to predict the reservoir’s response throughout the 2009 monsoon season. Oblinger et al. (2010) found that they were able to constrain the model parameters governing seepage losses from the reservoir with post monsoon drainage data, but were unable to constrain the parameters controlling the surface and groundwater flows into the reservoir. With data collection which began on May 8, 2009 through April 21, 2010, which includes the 2009 monsoon season, we will be able to evaluate the generality of the outflow parameters determined by Oblinger et al. (2010) and better predict the reservoir inflow parameters. A key question in this study is determining whether the simple relationships used to capture the watershed dynamics are a valid approach for modeling the impacts of water harvesting in rural India.

After fully developing the model, we will evaluate how well original model parameters as determined by Oblinger et al. (2010) predict the stage of the WHS. If original model parameters do not accurately represent the stage behavior, we will determine which parameters need to be modified in order to better fit the predicted and true stage, or if changes in the model will better represent the stage. Lastly, if changing
the model parameters and/or the model does not accurately depict the behavior of the reservoir, then all model parameters will be recalibrated using the 2009-2010 data. In the end, we can better answer the question on the impact of water harvesting after accurately modeling the behavior of the reservoir.

4.2 Conceptual Model for the Water Harvesting Structure

Before developing the numerical model for the volume balance, a conceptual model for the WHS is constructed to represent the behavior of the structure. The local geology was investigated using geologic mapping, electrical resistivity surveys (Oblinger, 2008), observations in large diameter open wells, and electromagnetic induction surveys. It was found that the watershed is characterized by alternating layers of massive and columnar basalts with weathered basalt reaching a maximum thickness of 10m and small areas of alluvial material downstream of the WHS. A cross section (Figure 4.1) to show the local geology is used to develop the conceptual model for the WHS (Figure 4.2).
Figure 4.1: Location for the cross section used to develop the conceptual model for the WHS.

Figure 4.2: Conceptual model for the WHS showing the water table for the monsoon season and dry season, flows in and out of the structure, and groundwater flow paths. The vertical exaggeration is approximately 11.3 times and the y-axis shows the elevation above mean sea level (amsl).
Flows into the structure are groundwater ($Q_G$), runoff ($Q_r$), and direct precipitation ($Q_d$) and flows out of the structure are evaporation ($Q_e$), water lost via the spillway ($Q_s$), domestic use ($Q_U$), and infiltration ($Q_I$) (Figure 4.2). Groundwater flows into the structure at the contact between the columnar and massive basalts as well as groundwater discharged up gradient of the structure through localized springs where fractures in the massive basalts outcrop. Water is then lost as infiltration to the weathered basalts and flows down gradient. Precipitation provides water to the structure as runoff from the upland area draining into the reservoir, as well as water falling directly on the surface. During the post monsoon and dry season, rainfall is generally less than 10mm; therefore, small volumes of water are added to the structure as direct rainfall and runoff. Outflow from the WHS occurs as water to the spillway when the reservoir reaches the maximum capacity as well as direct evaporation from the surface. Domestic use includes water used by villagers and the livestock of the area.

The water table during the monsoon is generally one to two meters below the ground surface downstream of the WHS, as observed in wells. It is believed groundwater flows into the structure from the columnar basalts; therefore, the water table elevation is above the surface elevation. During the post monsoon season and dry season the water table drops below the bottom elevation of the reservoir and groundwater no longer flows into the structure.

With the conceptual model representing the localized geology, flows in and out of the structure, and the behavior of the water table for the monsoon and post monsoon season it is possible to develop the numerical model for the reservoir volume balance.
4.3 Overview of Reservoir Volume Balance

The model developed by Oblinger et al. (2010) is based on a simple volumetric water balance between the flows in and out, i.e., $Q_{IN}$ and $Q_{OUT}$, of a reservoir:

$$\Delta V = Q_{IN} - Q_{OUT} \quad (4.1)$$

$$\frac{dV(h)}{dt} = Q_G + Q_r + Q_d - [Q_e + Q_t + Q_u + Q_s] \quad (4.2)$$

Volumetric flows into the reservoir as shown in the conceptual model are contributed by groundwater discharge ($Q_G$), surface runoff ($Q_r$), and direct rainfall ($Q_d$). Volumetric flows out of the reservoir are a result of evaporation ($Q_e$), infiltration losses to the subsurface ($Q_t$), the use of water by villagers and animals ($Q_u$), and losses through a spillway ($Q_s$) to prevent overtopping of the dam. Changes in reservoir storage are given by the time derivative of the reservoir volume $V(h)$, which is a function of the stage, $h$.

Simple models are used to link each of the flows in Eq. 4.2 to processes in the watershed. For example, Oblinger et al. (2010) use Darcy’s law to quantify the flow of groundwater into the reservoir and the flux of infiltration out of the reservoir. A simple method is also used for runoff, where $\Phi$ represents a constant fraction of rainfall that contributes to runoff (Bedient and Huber, 2002). When these simple relationships describing each flow are substituted back into Eq. 4.2, the following final model for the reservoir balance is given by Oblinger et al. (2010) as:

$$\frac{dV(h)}{dt} = c_1 \Delta H_G + c_2 (h - H_1) + A_{WHS} (h)(R - E) - Q_u - Q_s \quad (4.3)$$
This equation is discretized in time to provide an explicit solution for the volume of the reservoir:

\[ V_{i+1} = (c_1 \Delta H_c + \Phi A_U R_i - c_2 (h_i - H_i) + A_{\text{WHS}} (h_i) (R_i - E_i) - Q_u - Q_s) \Delta t + V(h_i) \]  

(4.4)

where \( V(h_{i+1}) \) is the predicted reservoir volume for time step \( i+1 \), and \( V(h_i) \) is the volume at the current time step.

Water usage, rainfall, and evaporation, \( Q_u \), \( R_i \), and \( E_i \), respectively, drive the model and must be known for a given study watershed. The volume of water required for human and livestock use is approximately \( 0.5 \times 10^4 \text{m}^3/\text{year} \) (Oblinger et al., 2010), which is 8% of the maximum reservoir volume \( 6.5 \times 10^4 \text{m}^3 \). Since not all water is taken from the WHS, the WHS is not full year round, and the flow of water required for usage is relatively small in this study compared to the other flows, it is estimated as zero for the entire period. The function \( A_{\text{WHS}}(h_i) \) is the surface area of the WHS for a given stage, \( h_i \) and should be obtained directly from detailed topographic surveys or derived from geometric arguments. It is assumed that the area of the watershed upstream of the dam contributing to surface runoff \( (A_U) \) can be determined from topographic data. Losses of water through the spillway \( (Q_s) \) are not specified directly, but rather obtained from calculated volume changes that exceed the reservoir capacity. The length of the time step used in the model is \( \Delta t \).

There are six unknown parameters in the model related to specific characteristics of the study area that must be estimated from observations of the reservoir behavior. These parameters control the groundwater inflows \( (c_1, \alpha, \beta) \), surface runoff \( (\Phi) \), and reservoir seepage losses \( (c_2, H_I) \). The parameter \( c_1 \) is the effective hydraulic conductance
controlling groundwater inflows to the reservoir. This conductance is equivalent to AK/L where A is the cross-sectional area of flow into the reservoir, L is the average length of the flow path between the upstream recharge area and the location of groundwater discharge, and K is the effective hydraulic conductivity of the upstream area. Due to the ephemeral nature of flows in the watershed, Oblinger et al. (2010) modeled the head difference driving groundwater inflows to the reservoir (ΔHG) as a step function describing groundwater flow into the structure before, during, and after the monsoon season. Given that the reservoir is initially dry, flow before the monsoon starts (tS) is fixed at zero, i.e., ΔHG = 0 when t<tS. From the start of monsoon to the end of monsoon (tE) it is assumed that the groundwater flow contribution is dominated by transmission from recharge areas and therefore relatively constant. As a result, the head difference in this period, tS<t<tE, is fixed to a constant value, i.e., ΔHG = α. After the monsoon when t>tE, groundwater discharge results from decreases in aquifer storage, the effect of which is approximated by an exponentially decaying head difference governed by decay constant β, i.e., ΔHG = αe^{-β(t-tE)}. Seepage losses from the reservoir are likewise controlled by the product of a hydraulic conductance parameter, c2, and the head difference between the reservoir and a downstream aquifer, h-HI. Based on direct observations in the study area it was assumed that the downstream head HI can be treated as a constant; however, this assumption must be reviewed for other watersheds.

Surface runoff to the reservoir is controlled by the parameter Φ, which is the fraction of rainfall that generates runoff. Since Φ is a constant value during the entire study period, it cannot represent varying runoff conditions. This is a potential problem
with the model since it is expected that $\Phi$ should be low before and after the monsoon season when there is a large amount of soil storage available, and $\Phi$ should be higher during the monsoon as the soils are saturated (U.S. Soil Conservation Service, 1986). Having a constant value of $\Phi$ is therefore a limitation of the Oblinger et al. (2010) model.

4.4 Description of Study Area and Available Data

Watershed development in study area (Figure 4.3) began in 1996, and one of the main projects was the construction of an earthen dam to capture monsoon runoff. The structure was built on a natural, ephemeral stream draining an upland area of approximately 0.64 km$^2$ (Oblinger et al., 2010). The length across the top of the dam is approximately 150 meters, and at capacity the reservoir extends approximately 350 meters upstream.

![Study Watershed](image)

**Figure 4.3:** Location of the study watershed in Madhya Pradesh, India.
During the dry season, the bathymetry of the reservoir was surveyed using differential GPS (Oblinger, 2008). The elevation data were interpolated using GIS to yield a smoothed map of bottom elevations (Figure 4.4). The deepest point of the reservoir is 421.8 meters above sea level, and a spillway to prevent overtopping of the dam is at approximately 427.4 meters, making the maximum depth of the structure 5.6 meters (Figure 4.2). The interpolated reservoir geometry was then used to model relationships between the stage, surface area, and volume of water stored in the reservoir. A power function is used to estimate the volume (V) from the stage (h) and a quadratic equation is used to estimate the surface area from the stage (Oblinger et al., 2010).

\[ V(h) = 973.56 \ h^{2.4225}, \ R^2 = 1.00; \ A_{\text{WHS}}(h) = 768 \ h^2 + 2277 \ h, \ R^2 = 0.988 \] (4.5)

Figure 4.4: The total depth of the reservoir as determined from the GPS survey.

The reservoir stage was monitored using a calibrated concrete gauge installed on the side of the dam. The gauge was calibrated by marking 10cm increments of vertical
distance along the length of the gauge relative to the lowest point in the reservoir, which was considered to be 0m depth (Figure 4.5). Visual observations were collected by reading the numbers off of the gauge from the top of the dam approximately every week starting in May 2009 through September, and then readings were taking approximately every month till the structure was empty in April 2010 (Figure 4.6).

**Figure 4.5:** Calibrated gauge on the upstream side of the reservoir used to determine the stage of reservoir.
Precipitation data was collected during the 2009 monsoon season using a tipping bucket rain gauge manufactured by Onset Computers (Model No. S-RGB-M002) installed on a weather station in the watershed. The data are aggregated from 15 minute observations to half-hour totals for use in the model (Figure 4.7). Monthly evaporation was estimated using the Thornthwaite-Mather method (Thornthwaite and Mather, 1957) based on temperature data collected from a temperature probe manufactured by Onset Computers (Model No. S-TMB_002) installed on the weather station (Figure 4.8; Figure 4.9). Since the Thornthwaite-Mather approach gives an estimate of potential and actual evapotranspiration, the calculation of potential evapotranspiration is used, as the maximum amount of water can be lost as direct evaporation when water is present in the reservoir. These values were disaggregated to half-hour values by equally distributing the total evaporation evenly across each month.
Figure 4.7: Monthly rainfall totals for the watershed.

Figure 4.8: Yearly high temperature values used in the calculation of evaporation.
4.5 Reservoir Simulation Results

Simulations of the reservoir response over the 2009-2010 monsoon are calculated using the volume-balance model described by Equation 4.4. Half-hour time steps are used to simulate the reservoir behavior for a period of 349 days starting on May 8, 2009 and ending on April 21, 2010. The precipitation and evaporation values shown in Figures 4.7 and 4.9 are used to drive model predictions over this period. After calculation of the volume change for a given time step, the stage is calculated using the volume-stage relation given in Equation 4.5. For different scenarios described below, the six unknown parameters of the model are calibrated by minimizing the root mean squared error (RMSE) between predicted and observed stage values.
4.5.1 Real Stage Data from 2009-2010 Compared to Predicted Stages from Pre-Calibrated Model Parameters

Oblinger et al. (2010) originally used a Monte Carlo sampling strategy combined with a non-linear optimization algorithm (the function *lsqnonlin* in MATLAB; Coleman and Li, 1996) to calibrate model parameters using stage data observed from September through December in 2007 (Table 4.1). Figure 4.10 shows these authors obtained a good match between the observed and predicted stage values. Because the data were collected primarily after the monsoon, they were more representative of drainage conditions than the inflows to the reservoir. As a result, the histograms in Figure 4.11 show that the model parameters controlling groundwater inflow (i.e., $c_1$, $\alpha$, and $\beta$) are not well constrained by this data. In contrast, the parameters controlling seepage from the reservoir (i.e., $c_2$ and $H_I$) are relatively well constrained.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Calibrated Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha$</td>
<td>38.6 (m)</td>
</tr>
<tr>
<td>$\beta$</td>
<td>0.027 (1/hour)</td>
</tr>
<tr>
<td>$\Phi$</td>
<td>0.189 (-)</td>
</tr>
<tr>
<td>$c_1$</td>
<td>18.9 (m²/hour)</td>
</tr>
<tr>
<td>$c_2$</td>
<td>2.76 (m²/hour)</td>
</tr>
<tr>
<td>$H_I$</td>
<td>2.78 (m)</td>
</tr>
</tbody>
</table>

**Table 4.1:** Calibrated model parameters for the reservoir flows (Oblinger, 2010).

**Figure 4.10:** Predicted and observed stage from calibrated model parameters from Oblinger et al. (2010) and field data from 2007.
Figure 4.11: The histograms of model parameters show the poor fit of groundwater inflow parameters compared to seepage parameters using drainage data from 2007 (based on results of Oblinger et al., 2010).

Figure 4.12 shows the response of the reservoir to the 2009-2010 monsoon as predicted by the volume balance model with model parameters calibrated by Oblinger et al. (2010). The true stage and model predictions are similar but are not a perfect match. The root mean squared error (RMSE) between predicted stage and actual stage is 0.74 meters. The reservoir is predicted to fill slightly faster than observed in the field during the onset of the monsoon. The reservoir is then predicted to remain at capacity until the stage starts to decrease at the end of the monsoon season in September, at which point the model indicates that the flow of groundwater into the structure starts to decrease (Figure 4.12b). After monsoon the reservoir drainage is consistent between the model predictions
and observations, except for March 17, 2010 when a rainfall event occurred and the observed stage is higher than the predicted stage.

Figure 4.12: (a) Predicted and true stage of the reservoir during the study period using the original model parameters as found by Oblinger et al. (2010). Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.

Since the calibrated model parameters from Oblinger et al. (2010) do not exactly predict the stage of the reservoir, it is necessary to determine which parameters might cause the poor fit. Potential problems could be when groundwater inflow or when groundwater decay begin and end, runoff parameters which control how the structure fills, infiltration parameters controlling the draining of the structure, or errors in the user supplied parameters, mainly evaporation and precipitation. Each of these potential problems will be considered individually.

The filling of the structure is controlled primarily by groundwater contributions and runoff. Groundwater discharge is zero before the monsoon, constant ($\alpha$) during monsoon, and exponential decay controlled by $\beta$ after monsoon. Groundwater
contributions to the reservoir do not become significant in the simulation until day 75 (July 22, 2009), the time when groundwater levels were seen to increase in the field (Figure 4.13) and continuous stream flow was observed in the watershed. Given that the model predicts no groundwater enters the structure before day 75, surface runoff is responsible for the initial filling of the reservoir, suggesting an error in $\Phi$. Groundwater inflows in the simulation begin to decay on day 146, at the end of September. At this point in time, the structure begins to drain. Since groundwater inflows decay very rapidly and late rainfall events do not add significant volumes of water, the predicted zero stage of the structure occurs before the reservoir was observed to be empty. The difference in time between the predicted and true zero stage is also attributed to potential errors in the groundwater inflow and runoff parameters.

![Average Water Table Elevation](image)

**Figure 4.13:** Average depth to water of wells in the watershed, illustrating the large increase in the groundwater elevation on July 22, 2009 (75 days) after the start of data collection.

In order to determine the accuracy of the outflow parameters for the 2009-2010 season the reservoir is modeled as a falling head permeameter which relates the volume change in a reservoir to the volumetric discharge rate through a column using Darcy’s
Law (i.e. Sevee, 1991). The WHS reservoir was modeled as a falling head permeameter, and a parameter (ω) which represents the geometry and hydraulic conductivity of the reservoir and the aquifer was determined during the Oblinger et al. (2010) study. Oblinger et al. (2010) determined the value of ω as -0.0062 day⁻¹, which is the slope of the line plotted through a time versus ln(h-Href), where h is the stage of the reservoir, and Href is a calibrated parameter for the downstream head (2.2 meters). Similar plots are made using the same value for Href and stage values collected during 2009-2010 (Figure 4.14; Figure 4.15). If the structure is modeled as a falling head permeameter after groundwater inflows stop on day 150 (October 5, 2009) and the value of ω is found to be -0.006 day⁻¹ the same as Oblinger et al. (2010) (Figure 4.14). One problem with this model is it assumes no inflows, which is not the case in this year as some water is still added to the structure from runoff. If a later date is used, day 194 (November 17, 2009), to minimize the amount of water added through runoff, a value of -0.0081 day⁻¹ was found for ω (Figure 4.15).

**Figure 4.14:** Results from the falling head permeameter simulation from October 5, 2009 through April 21, 2010 when the reservoir was empty. The value of ω is -0.006 day⁻¹ with an R² of 0.95.
Figure 4.15: Results from the falling head permeameter simulation from November 17, 2009 through April 21, 2010 when the reservoir was empty. The value of $\omega$ is $-0.0084$ day$^{-1}$ with an $R^2$ of 0.98.

A potential reason for the difference between Oblinger et al.’s (2010) value for $\omega$ and the current studies value is the runoff entering at late times since the falling head permeameter model does not account for water inputs. The structure is simulated as draining faster than observed and if the value of $\omega$ is smaller than Oblinger et al. (2010), which is the case when only late time stage data is used, this suggests the structure is draining slower. Since water is being added to the structure through runoff in 2009-2010, there is more water in the structure and it will take longer for the reservoir to drain. In Oblinger’s falling head model, the structure drains faster since no water was added after monsoon. Therefore, the infiltration parameters as found by Oblinger et al. (2010) are assumed to be an accurate representation since additional runoff may have caused the structure to drain slower.

Another potential reason for errors between the predicted and true stage are errors in precipitation or evaporation data. In order to conclude that errors in the precipitation
and evaporation are not the reason for the error between the predicted stage and the true stage, precipitation and evaporation values were increased and decreased by 10%. Precipitation and evaporation were increased and decreased by 10% while one remained constant, then both precipitation and evaporation were increased, both were decreased, precipitation was increased and evaporation was decreased, and evaporation was increased and precipitation was decreased. When the RMSE for each simulation was compared to the RMSE with collected field data and calibrated parameters, all errors are within 1-2% of the original RMSE; therefore, potential errors in the precipitation and evaporation data are not the cause of the poor misfit between the predicted stage and the true stage.

The errors between the stage predicted over 2009-2010 using the model calibrated by Oblinger et al. (2010) with drainage data from 2007 and the true stage is caused primarily by incorrect values for the inflow parameters rather than infiltration parameters. This conclusion is supported by the similarity in the drainage characteristics of the reservoir found using the falling head permeameter analogy for the 2007 and 2009-2010 data sets. Additionally, Oblinger et al. (2010) found that the seepage parameters were well constrained during model calibration compared to the inflow parameters. Furthermore, we have found that errors in rainfall and evaporation are not the cause for the misfit between the predicted and true stage; therefore, it is likely that an improved fit of the 2009-2010 data could be achieved by adjusting the model inflow parameters.
4.5.2 Sensitivity Analysis of the Model Inflow Parameters

A sensitivity analysis was performed to provide insight into how each of the inflow parameters ($\phi$, $\alpha$, and $\beta$) affects the model performance. Note that $c_1$ was not varied since in the model the product $c_1*\alpha$ occurs and changes to $\alpha$ therefore produce the same response as changes to $c_1$. In the analysis, the RMSE between the predicted and observed stage data for 2009-2010 was calculated as each inflow parameter was varied sequentially over a range of values.

The first parameter adjusted is the runoff coefficient, $\Phi$, which controls the fraction of rain falling over the upstream watershed area contributed to the reservoir. The value of $\Phi$ was varied from 0% to 35% while all other parameters were fixed to their calibrated values in Table 4.1. Figure 4.16 shows how the RMSE changes as a function of runoff. The minimum RMSE (0.71m) is achieved when 10% of rainfall is contributed to the reservoir, though this value is only marginally smaller than the RMSE of 0.74m obtained with the calibrated $\Phi$ value of 0.189. With runoff slightly lower, the structure fills more slowly, but this does not improve the over prediction of the stage between day 75-125 (Figure 4.17). At the end of monsoon, the later rainfall events do not add significant volumes of water to the reservoir, thus the resulting peaks observed in the data are not fit as well in this case. An improvement of fit occurs before the structure has water, as the predicted stage before day 75 is lower since the runoff is lower. Overall, the data misfit is insensitive to changes in the amount of runoff contributed to the reservoir.
Figure 4.16: Changing values of $\Phi$ to determine the best fit for the data while the other parameters remain constant.

![Changing Value of $\Phi$](image)

Figure 4.17: (a) Predicted and true stage of the reservoir during the study period after changing the value of $\Phi$ from 0.189 to 0.10. Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.

The flow of groundwater into the structure reservoir is dependent on the parameters $\alpha$ and $\beta$, which control the magnitude and duration of flow during and after the monsoon, respectively. The parameter $\alpha$ is varied over a range of values from 0 to
100 m²/hour based upon the maximum elevation change observed in the watershed of 100m, and estimates of hydraulic conductivity for basalts. The lowest RMSE is 0.55m, which occurs when $\alpha$ is 3 m²/hour and if $\alpha$ is raised above 50, the RMSE remains at 0.73m (Figure 4.18). The RMSE is lower in this case as the predicted stage better captures the behavior of the structure as it fills since a large pulse of water is not added at the start of groundwater flow (Figure 4.19).

**Figure 4.18:** Changing values of $\alpha$ to determine the best fit for the data while the other parameters remain constant.
Figure 4.19: (a) Predicted and true stage of the reservoir during the study period after changing the value of $\alpha$ from 38.9 to 3 m$^2$/hour. Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.

When the original $\alpha$ value is used and $\beta$ is changed from 0 to 0.05 hour$^{-1}$, the lowest RMSE is 0.73 m, which occurs when $\beta$ is 0.02 hour$^{-1}$. If $\beta$ takes on a value greater than 0.03 hour$^{-1}$, the RMSE value remains at 0.75 m (Figure 4.20) as the groundwater flow after the monsoon decays very quickly and does not last into the dry season. If $\alpha$ is at the calibrated model parameter and $\beta$ is decreased, the model fit is very similar to the original model parameters where the structure fills rapidly (Figure 4.21), suggesting the model is more sensitive to changes in $\alpha$ than $\beta$. 
Figure 4.20: Changing values of $\beta$ to determine the best fit for the data while the other parameters remain constant.

Since changing the value of $\alpha$ greatly reduces the RMSE, and changing $\beta$ slightly changes the RMSE changing both at the same time is tried to provide a better fit between the predicted and true stage. If $\alpha$ is lowered to 3m$^2$/hour, the best fit when just this parameter is changed, and then $\beta$ is changed the lowest RMSE is 0.53m, which occurs when $\beta$ is 0.009 hour$^{-1}$ (Figure 4.22). In this case groundwater flow into the structure...
lasts longer after the monsoon keeping the stage higher and improving the predicted stage after the monsoon season (Figure 4.23).

**Figure 4.22**: Changing values of $\alpha$ and $\beta$ to determine the best fit for the data while the other parameters remain constant.

**Figure 4.23**: (a) Predicted and true stage of the reservoir during the study period after changing the value of $\alpha$ to 3 m$^2$/hour and $\beta$ is changed from 0.027 to 0.009 hour$^{-1}$. Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.

After varying the inflow parameters for the volume balance model, it was found that changes to the groundwater inflow parameters better predict the stage than changes in the percentage of runoff from upland rainfall. When only the constant groundwater
flow parameter $\alpha$ is varied the RMSE is 0.55m, where as when the runoff coefficient $\Phi$ is varied the lowest RMSE is 0.71m. Then, if only the groundwater decay constant $\beta$ is changed the best fit returned is an RMSE of 0.73m. Finally, if $\Phi$ is left as the originally calibrated value of 18.9%, $\alpha$ is lowered to 3m$^2$/hour, and $\beta$ is decreased to 0.009 hour$^{-1}$ to provide more groundwater after monsoon, the best fit between the true and predicted stage returns an RMSE of 0.53m.

4.5.3 Alternative Runoff Models

Rainfall-runoff is one of the main drivers affecting the variability of flow into the reservoir. In the original model, a single value of $\Phi$ is used to represent runoff conditions throughout the year. This is not a good assumption as runoff is known to vary with watershed conditions (U.S. Soil Conservation Service, 1986). Pre- and post- monsoon soils in the watershed are dry and can store more water therefore decreasing the amount of rainfall contributed to runoff. During the monsoon the soils can become saturated and a greater fraction of rainfall is contributed to runoff. Two different approaches were investigated to account for seasonally variable runoff generation: a modified version of the $\Phi$-method and the SCS Curve Number Method.

In order to simulate variable rates of runoff over the monsoon, the model is modified to allow $\Phi$ to take on two different values representing monsoon and non-monsoon conditions. The first $\Phi$ value, $\Phi_1$ represents runoff generation in the early and late monsoon season, May 8, 2009 to June 22 and October 1 till April 21, 2010, respectively. The parameter $\Phi_2$ represents the enhanced runoff generating conditions.
occurring during the monsoon, from June 22, 2009 through September 30. Dates representing the start and end of monsoon coincide with the dates for the start and end of groundwater flow into the structure, which are assumed to be the same time when runoff would be increased.

The sensitivity of the RMSE to changes in $\Phi_1$ and $\Phi_2$ were evaluated by varying each parameter independently. The sensitivity of the pre- and post-monsoon runoff was evaluated first by changing $\Phi_1$ between 0 to 20% while keeping $\Phi_2$ fixed at 20%. The best fit to the data (RMSE = 0.72m) was obtained when $\Phi_1$ is 12%. This was not, however, a significant improvement to the fit compared to the original model parameters. Next, $\Phi_2$ is varied from 0 to 100% while $\Phi_1$ is fixed at 12% as this value provided the best fit when $\Phi_2$ was held constant. The RMSE value was the lowest when there was no runoff during the monsoon season, which is unrealistic for the watershed (Figure 4.24). If $\Phi_2$ was increased above 0% the RMSE remained at 0.73m, as all of the extra monsoonal runoff was diverted as water exiting through the spillway. Therefore, it was determined runoff before and after monsoon is more important than during the monsoon, as during monsoon $\Phi_2$ does not help constrain the filling and draining of the structure (Figure 4.25). Furthermore, the large inflow of groundwater does not help constrain the model before the start of monsoonal runoff, since it was found the groundwater inflow parameters were too high.
Figure 4.24: Changing the value $\Phi_2$ as $\Phi_1$ remains at 12% to determine the best fit for the data while the other parameters remain constant.

Figure 4.25: (a) Predicted and true stage of the reservoir during the study period after changing $\Phi_1$ to 0.12 and $\Phi_2$ to 0.2. Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.

Since it is known that the original groundwater inflow parameters were too high, consideration is taken by lowering the groundwater inflow parameters to the best fit parameters found during the sensitivity analysis and then applying the two $\Phi$ model. First, the model was adjusted for the pre- and post- monsoon runoff by changing $\Phi_1$
between 0 to 20%. The best fit to the data (RMSE = 0.53m) is obtained when $\Phi_1$ is 14%.

Next, $\Phi_2$ is varied from 0 to 100% while $\Phi_1$ is fixed at 14% as this was the best fit value. In this case, the lowest RMSE of 0.52m was returned when $\Phi_2$ is 40% (Figure 4.26). The predicted stage now better matches the true stage when a two $\Phi$ model is used in conjunction with lower groundwater inflow parameters (Figure 4.27). The structure fills slower with less groundwater entering the reservoir, and then fills to capacity at the onset of monsoon as runoff increases.

![Graph showing changing value of $\Phi_2$](image)

**Figure 4.26:** Changing the value $\Phi_2$ as $\Phi_1$ remains at 14% to determine the best fit for the data while the other parameters remain constant.
Figure 4.27: (a) Predicted and true stage of the reservoir during the study period after changing $\Phi_1$ to 0.14 and $\Phi_2$ to 0.4 and adjusting the groundwater inflow parameters to the re-calibrated values of $\alpha$ as 3 m$^2$/hour and $\beta$ as 0.009 hour$^{-1}$. Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.

Runoff can also be modeled using the Soil Conservation Service (SCS) curve number approach, and this method is used to constrain runoff coming into the reservoir (U.S. Soil Conservation Service, 1986). Curve numbers for the watershed were estimated using two different approaches using soil and land use data and based on values from a similar region reported in the literature (Table 4.2; Nayak, 2003). Soil information from the watershed indicates the soils are moderately drained (Soil type B), and land use indicates there is land under cultivation with conservation treatment (70%), poor range land (28%), and poor forests (2%), with curve numbers of 71, 79, and 66, respectively. The fraction of each land type was multiplied by the curve number for the land type, and then all three fractions were summed to provide the curve number for the area draining into the structure. Curve numbers were then modified to take into consideration different antecedent moisture contents (AMC) of the soils (Table 4.2).
<table>
<thead>
<tr>
<th>Soil Type</th>
<th>Source</th>
<th>Curve Number</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>AMC I (dry)</td>
<td>AMC II</td>
</tr>
<tr>
<td>B</td>
<td>Watershed</td>
<td>55.6</td>
</tr>
<tr>
<td>C</td>
<td>Nayak, 2003</td>
<td>63.9</td>
</tr>
</tbody>
</table>

Table 4.2: Curve numbers for the study watershed with the different antecedent moisture contents (AMC).

Each curve number relating to the antecedent moisture content is used in the volume balance model to represent different seasons. AMC I is used before monsoon season as soils in the watershed are dry and can store more water. AMC III is used during monsoon as the soils become saturated and runoff is higher. Then, at the end of the monsoon, AMC II is used as the soils are not completely dry, but are drying due to less frequent rainfall events. The curve number changes for different soil moisture conditions at the same time as the start of groundwater inflows and the start of the groundwater inflow decay, on June 22, 2009 and October 1st, respectively.

The curve number approach is generally used to calculate runoff on a per-storm basis (U.S. Soil Conservation Service, 1986) and not for extended periods of time, as we will use it for the volume balance model. Because of this, runoff will always enter into the reservoir, even if there is no rainfall. In order to address the per-storm basis of the curve number approach, runoff is only generated when rainfall occurs; therefore, when there is no rainfall, there is no runoff and each rainfall event its own storm. Even if rainfall is continuous for consecutive time steps, it is treated as an individual storm. Since each rainfall event is considered as its own storm, a percentage of the rainfall is lost to soil storage, the initial abstraction, and the rest is lost as runoff into the reservoir.
When the curve numbers from the watershed and from the literature are used to model runoff, the predicted stage and true stage do not match (Figure 4.28 & 4.29). The RMSE for the watershed and literature curve numbers is 2.1m and 1.9m, respectively. These errors are much higher than when the original model parameters were used. Since the amount of runoff coming into the structure is so large, the structure fills very rapidly after the first few rainfall events. As the monsoon season ends and the dry season starts, rainfall events generate such large volumes of runoff which keep the structure at capacity longer than what was observed in the field. Even if lower groundwater inflow parameters are used the RMSE value is the same for the watershed and literature curve numbers, and there is no improvement on the predicted and true stage due to the large volumes of runoff. The amount of runoff generated from the curve number approach is almost an order of magnitude greater than the original model parameters and adjusted values for a one and two Φ runoff model (Figure 4.30). Overall, the curve number approach does not work well to simulate runoff for the watershed.
Figure 4.28: (a) Predicted and true stage of the reservoir during the study period for curve numbers from the watershed. Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.

Figure 4.29: (a) Predicted and true stage of the reservoir during the study period for curve numbers from the literature. Volume of groundwater (b) and runoff (c) predicted to enter the structure by the model.
After the sensitivity analysis and using different runoff models, it was found groundwater inflows are a major contributor to the misfit between the predicted and true stage. Different Φ models better fit the true stage if groundwater inflows are decreased, whereas with such large amounts of runoff generated from the SCS curve number approach, lowering groundwater inflows does not affect the overall stage prediction. Overall, it is necessary to change all model parameters to try and better fit the true stage and the predicted stage since for the 2009-2010 season, data was collected as the structure fills and drains. With more data available, recalibrating all model parameters might better predict the true stage of the reservoir.

**Figure 4.30:** Cumulative runoff for the various approaches to change the runoff model for the volume balance.
4.5.4 Monte Carlo Model Simulation Results

In the previous sections it was shown that varying an individual parameter while keeping the others fixed does not yield a significantly improved fit between the predicted and true reservoir stage. It is possible, however, that changing multiple parameters at the same time, e.g., reducing groundwater inflow while at the same time increasing the contribution of runoff, could better capture the reservoir response. However, directly computing the RMSE for all possible parameter combinations even for a coarse sampling of values is impractical. For example, one million model simulations would be required if only 10 different values of each of the 6 parameters in the basic model were to be investigated. To address this computational problem, Monte Carlo simulations were used to sample a wide range of possible parameter values. In practice, randomly drawing sets of parameter values will result in a large proportion of the simulations producing results that are physically unreasonable or having an unacceptably high degree of data misfit. To mitigate this problem, each randomly sampled set of parameter values is used as the starting point for a gradient-based optimization using the function *lsqnonlin* in MATLAB (Coleman and Li, 1996). By using this hybrid approach, the parameter space can be explored more fully than is possible using gradient-based optimization techniques alone and a higher proportion of simulations that fit the data can be returned than in a purely Monte Carlo approach. Two distinct benefits of the hybrid approach are simulations that become trapped at local minima during an optimization can be identified and parameter non-uniqueness can be assessed.
The Monte Carlo simulations were conducted for a one Φ model, i.e., single runoff coefficient throughout the entire year, and a two Φ model, i.e., a different runoff coefficient is allowed during the monsoon compared to the rest of the year. The simulations were conducted 10,000 times with starting parameters randomly generated using a uniform distribution with the ranges given in Table 4.3. These ranges were selected based upon reasonable parameter values using knowledge of the watershed hydrogeology and physical limitations of parameters. For instance, the runoff parameter Φ can only vary between 0 and 1 as it is impossible to have more runoff generated than the amount of rainfall, and the downstream head parameter H_i is 20m which is below the depth of the water table in the weathered zone of the downstream aquifer. To allow comparison of the results obtained for the two different models, a fixed seed was used for the random number generator such that the initial parameter values drawn in the simulations were the same for both the one-Φ and two-Φ models, except for the value of Φ_2.

<table>
<thead>
<tr>
<th></th>
<th>c_1 (m^2/hr)</th>
<th>α (m)</th>
<th>β (1/hr)</th>
<th>Φ (-)</th>
<th>c_2 (m^2/hr)</th>
<th>H_i (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower Value</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Upper Value</td>
<td>50</td>
<td>100</td>
<td>0.1</td>
<td>1</td>
<td>50</td>
<td>20</td>
</tr>
</tbody>
</table>

**Table 4.3:** Range of values to determine the starting parameter values used in the optimization.

The Monte Carlo simulations were run using Condor, a high-throughput computing system which provides parallel computing between computers on the campus of Clemson University. For running the volume balance model, a host computer is prepared with all the required input files to run the model. Then, a specific set of starting parameters is sent to a university computer which can run the optimization. The
optimization is carried out on that machine, and results are sent back to the host computer. Condor is a very useful tool as multiple simulations can run at the same time cutting down the processing time of the simulations. For our specific volume balance model, 10,000 simulations for both the one and two-Φ models (20,000 runs total) were run using Condor in approximately 24 hours.

After the simulations were completed, the resulting “optimal” sets of parameters were sampled to select only the realizations that meet a specific data fit threshold based on the RMSE of the predicted stage. To investigate if there are trends apparent in how the model behaves at different levels of data misfit, three different RMSE thresholds are used in this study: 1.5m, 0.7m, and 0.452m. These thresholds were used as an RMSE of 1.5m is close to double the RMSE of the original model parameters, an RMSE of 0.7m is approximately the same as the misfit with original model parameters, and 0.452m is the lowest RMSE obtained from the model simulations. These subsamples were then analyzed to determine how well constrained parameter values are, and if specific sets of parameter values better estimate the true stage of the reservoir.

Histograms for sets of parameter values that produce an RMSE of less than 1.5 meters are shown in Figure 4.31. For the one Φ model, 6464 parameter sets were returned, and for the two Φ model 6995 results were returned. The remainder of the simulations had starting values for which the optimization was not able to converge to a RMSE value within the 1.5m limit given the stopping tolerances on the error and parameter values of $1 \times 10^{-6}$ used in the optimization.
Figure 4.31: Histograms for the one-Φ and two-Φ model with an RMSE<1.5 meters.

Histograms for the one-Φ model show poor constraint of the groundwater inflow parameters $c_1$ and $\alpha$. Since these two values are multiplied together in the model, changing the value of one parameter effectively changes the product, so if one parameter is high, the other can be low to offset the higher value. Other parameter values appear to have a bimodal distribution, suggesting when inflow is high, the outflow is high to counteract the volume of water coming into the structure and vice-a-versa. Correlation coefficients for $\beta$ and $c_2$ indicate at low $\beta$ values (long groundwater decay) the infiltration is greater to balance the increased flow of groundwater; therefore, $\beta$ and $c_2$ are negatively correlated, with a correlation coefficient of -0.58. Similarly $\Phi$ and the $c_2$ are positively correlated, 0.51, showing as runoff increases, infiltration also increases.
Histograms for the two-Φ model also show poor constraint on the inflow parameters $c_1$ and $\alpha$, attributed to the tradeoff between the two parameters. $\beta$ is well constrained at lower values, but has a slight bimodal distribution. The groundwater outflow parameter, $H_I$, is not well constrained and there are three areas where optimized parameter values are clustered. When the value of $\beta$ is low the value of $c_2$ is high, showing the two parameters are negativity correlated as indicated by the correlation coefficient of $-0.49$. If runoff before and after monsoon is high, $\Phi_1$, then the infiltration is also high as $\Phi_1$ and $c_2$ are positively correlated, with a correlation coefficient of 0.70. Overall with a high RMSE limit, there are many parameter values which can predict the stage of the WHS to less than 1.5 meters.

Figure 4.32 shows the predict stage for parameter sets of the one-Φ model with an RMSE of less than 1.5m to illustrate the range in reservoir behavior allowed by the model. The general trend of the filling and draining of the reservoir is captured by most parameter sets; however, 712 simulations show the reservoir draining very quickly around day 175. These parameters sets have an RMSE between 1.0m and 1.5m. The fast drainage is caused by the high seepage rate resulting from large infiltration parameters. Similarly, the two-Φ model shows most parameter sets capture the general trend, although 1469 sets show the structure draining very quickly. Clearly at an RMSE threshold of 1.5m, the parameters retained do not provide predictions that satisfactorily reproduce the data.
Figure 4.32: Estimated stage from parameter sets with an RMSE <1.5 meters. Parameter sets which show the reservoir draining very rapidly generally have an RMSE >1.0m.

Histograms for parameter values with an RMSE of less than 0.7m are given in Figure 4.33. At this threshold, 4523 parameter sets were retained for the one-Φ model and 3707 were retained for the two-Φ model. The histograms for both models indicate that $c_1$ and $\alpha$ are better constrained than for the 1.5m RMSE threshold, although there is still a large spread of values. $\beta$ for the one-Φ model has a distinct bimodal distribution showing groundwater inflows either decay very quickly or slowly. In contrast, the two-Φ model has a mean $\beta$ of 0.0057 hour$^{-1}$ with a standard deviation of 0.018; therefore groundwater inflows generally last longer after the monsoon season for the two-Φ model. Φ values for the one-Φ model are bimodally distributed near 15% and 35%, whereas for the two-Φ model, $\Phi_1$ is near 20% and $\Phi_2$ ranges from 40% to 80%. The large range in the second Φ value indicates the model is not sensitive to different $\Phi_2$ values. Infiltration for both models shows a bimodal distribution, suggesting as one set of inflow parameters...
are high, the infiltration parameters are also high. The one-Φ model has a correlation coefficient between β and c_2 of -0.57, indicating as groundwater flows last longer into the dry season, infiltration is higher; furthermore, the correlation coefficient for Φ and c_2 is 0.57, also indicating as the inflows are higher the infiltration is also higher. For the two-Φ model, as Φ_1 increases the infiltration parameter c_2 also increases, as shown by the high positive correlation coefficient of 0.67 between the two parameters.

![Figure 4.33: Histograms for the one-Φ and two-Φ model with an RMSE<0.7 meters.](image)

The predicted stage values when the RMSE is less than 0.7 meters shows an overall better match to the true stage when the limit was at 1.5 meters (Figure 4.34).
Some of the parameter sets (Figure 4.34, black lines) show the structure filling very rapidly, completely missing the progressive filling behavior seen in the data. These same parameter sets also have the structure draining very quickly. The main cause for the rapid filling and draining are the inflow parameters and infiltration parameters are too high, and are trying to compensate for each other. On the other hand, some parameter sets (Figure 4.34, Blue Lines) better capture the true behavior of the reservoir through time.

**Figure 4.34:** Predicted stage values for RMSE <0.7 meters. Black lines represent parameter sets which have the structure filling very rapidly and blue lines better match the true stage of the reservoir.
Finally, an RMSE threshold of 0.452m, which is close to the lowest RMSE returned from the simulations, is used to evaluate the model response to the subset of parameters producing the best obtained fit to the data. In this case only 313 parameter sets were returned for both the one and two Φ models out of the original 10,000 simulations. Histograms for parameter values are given in figure 4.35. Inflow parameters \( c_1 \) and \( \alpha \) are better constrained for both models; however they are negatively correlated, with a correlation coefficient of -0.64 and -0.65 for the one-Φ and two-Φ model, respectively; therefore, if one parameter value is high, the other parameter value is low, indicating the tradeoff between the two parameter values. A negative correlation is present between \( \beta \) and the head of the downstream aquifer \( (H_I) \) which drives infiltration. The correlation coefficient is -0.95, indicating as one parameter increases, the other parameter decreases to counteract either the large inflow or drainage of water from the reservoir. Similarly, for the one-Φ model, \( \Phi \) has a correlation coefficient of 0.92 with \( H_I \) again showing as flows into the structure increase, flow out of the structure as infiltration also increases. Overall, parameter values are much better constrained, and the bimodal distributions seen when the limit on the RMSE was higher is no longer present.
Figure 4.35: Histograms for the one-Φ and two-Φ model with an RMSE<0.452 meters.

With better constrained parameter values and a lower RMSE, the predicted stage and true stage match much better and all different parameter values yield the same result in the prediction of the stage (Figure 4.36). As the structure fills, the prediction does a fairly good job capturing the points around day 75 to day 100, which other parameter sets with a higher RMSE miss all together. Then, as the structure drains, the model does a very good job predicting when the structure will go dry, and the rate at which the structure drains.
When comparing the Monte Carlo simulation results with an RMSE less than 0.452m, the inflow and outflow parameters are better constrained than the original Oblinger et al. (2010) parameters. The range of values for parameters calibrated from the 2009-2010 data is much smaller than found by Oblinger et al. (2010), and all parameter values are localized around a common value. Average values for the infiltration and runoff parameters are higher than the original model parameters, whereas groundwater inflow parameters are lower (Table 4.4). When using the average values or median values in the WHS model, the stage is not well predicted, as the average values do not take into account the non-uniqueness of a given parameter set. However, when a single set of parameter values is used in the WHS model, the real stage and predicted stage are very close (Figure 4.36).
Table 4.4: Statistics for calibrated parameters with an RMSE<0.45 meters for the Monte Carlo results for the one-Φ model.

<table>
<thead>
<tr>
<th></th>
<th>$c_1$ (m$^2$/hr)</th>
<th>$a$ (m)</th>
<th>$\beta$ (1/hr)</th>
<th>$\Phi$ (-)</th>
<th>$c_2$ (m$^2$/hr)</th>
<th>$H_i$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oblinger (2010)</td>
<td>18.9</td>
<td>38.6</td>
<td>0.027</td>
<td>0.189</td>
<td>2.76</td>
<td>2.78</td>
</tr>
<tr>
<td>values</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Simulated Set</td>
<td>24.6</td>
<td>3.33</td>
<td>0.0004</td>
<td>0.34</td>
<td>13.49</td>
<td>0.733</td>
</tr>
<tr>
<td>Mean</td>
<td>10.23</td>
<td>19.79</td>
<td>0.0004</td>
<td>0.34</td>
<td>13.46</td>
<td>0.742</td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>10.05</td>
<td>18.77</td>
<td>$3\times10^{-7}$</td>
<td>0.002</td>
<td>0.01</td>
<td>0.002</td>
</tr>
<tr>
<td>Median</td>
<td>6.30</td>
<td>12.96</td>
<td>0.0004</td>
<td>0.34</td>
<td>13.46</td>
<td>0.742</td>
</tr>
<tr>
<td>Minimum</td>
<td>1.01</td>
<td>1.83</td>
<td>0.0004</td>
<td>0.34</td>
<td>13.35</td>
<td>0.733</td>
</tr>
<tr>
<td>Maximum</td>
<td>44.69</td>
<td>81.01</td>
<td>0.0004</td>
<td>0.34</td>
<td>13.49</td>
<td>0.766</td>
</tr>
</tbody>
</table>

One important difference between calibrated model parameters using stage data from 2009-2010 and Oblinger et al. (2010) model parameters are the differences between groundwater inflow decay, runoff, and infiltration parameters. Originally, it was assumed that model parameters as calibrated by Oblinger et al. (2010) were correct for the outflow parameters, and inaccurate for inflow parameters. Since during the Monte Carlo simulations all model parameters could be adjusted this idea is ignored and not taken into account during the simulations. Therefore, the differences in these parameters are reasonable since the current model calibration considers data for the entire season, not just for the post monsoon.

Overall, results from the Monte Carlo simulation yield the best model fit between the true stage and predicted stage (Table 4.5) as all unknown parameters were accounted for and minimized. The newly calibrated parameter values better predict how the reservoir fills and drains than the original calibrated parameters. Of the all trials, the two-Φ model provides the best fit between the true and predicted stage, although the difference between the two cases is less than 1 cm.
<table>
<thead>
<tr>
<th>Case</th>
<th>RMSE (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calibrated Model Parameters</td>
<td>0.74</td>
</tr>
<tr>
<td>Lower $\Phi$</td>
<td>0.71</td>
</tr>
<tr>
<td>Changing groundwater inflows</td>
<td>0.53</td>
</tr>
<tr>
<td>Step function for $\Phi$</td>
<td>0.72</td>
</tr>
<tr>
<td>Curve Number from watershed</td>
<td>3.2</td>
</tr>
<tr>
<td>Curve Number from literature</td>
<td>3.2</td>
</tr>
<tr>
<td>Monte Carlo: 1 $\Phi$ Model</td>
<td>0.452</td>
</tr>
<tr>
<td>Monte Carlo: 2 $\Phi$ Model</td>
<td>0.446</td>
</tr>
</tbody>
</table>

Table 4.5: RMSE values for all of the considerations taken for the Oblinger et al (2010) model.

4.6 Model Discussion

One potential reason for the misfit between the true and predicted stage is errors in data collection from the gauge on the WHS. Given that stage values were collected from a concrete gauge installed on the upstream side of the reservoir, potential errors are present in the true stage data. Errors could be caused by not finding the lowest point in the WHS during gauge calibration, errors in painting stage values, and errors when reading the stage off of the gauge. It is estimated that these errors could be as great as 0.1m since the concrete gauge did not go all the way to the deepest point in the reservoir and the markings on the gauge were in 10cm increments. These errors in the collection of field data could be the cause for the misfit between the true and predicted stage.

Overall, the volume balance model is an accurate way to represent the behavior of the reservoir. The RMSE could be caused by error in data collected from the field, and not because the parameters of the model cannot be constrained as after the Monte Carlo simulation, it was found that all model parameters were well constrained. Furthermore, data was not collected on a frequent and regular basis. If data were collected more often
than every week before and during the monsoon and more than once a month after monsoon, the model would be able to better predict the behavior of the reservoir, as there would be more data points to help constrain the model. Since we have only a limited amount of data and there could be errors in the stage measurements, we believe the simple volume balance is an accurate way to represent the behavior of the reservoir.

Using the one Φ model, it is possible to compare between Oblinger et al. (2010) parameter values, and the current studies optimized values (Table 4.6). The groundwater inflow parameter $c_1$, runoff coefficient Φ, and the infiltration parameter $c_2$ are higher than the original values. Groundwater inflow parameters $α$ and $β$ which control the rate of inflow and the decay of groundwater, respectively, are lower as well as the downstream head level, $H_I$. Infiltration parameter $c_2$ is higher than the original model, and approximately $2 \times 10^5 \text{m}^3$ more water was lost to the subsurface than predicted by Oblinger et al. (2010). Groundwater and runoff coming into the structure were higher for Oblinger’s parameter values, and since these parameter values were poorly constrained during model calibration, the new parameters are thought to be a better fit to the behavior of the reservoir since the RMSE between the true and predicted stage is lower than the original model parameters.

<table>
<thead>
<tr>
<th></th>
<th>$c_1$ (m$^2$/hr)</th>
<th>$α$ (m)</th>
<th>$β$ (1/hr)</th>
<th>Φ</th>
<th>$c_2$ (m$^2$/hr)</th>
<th>$H_I$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Optimized value</td>
<td>24.6</td>
<td>3.33</td>
<td>0.0004</td>
<td>0.34</td>
<td>13.49</td>
<td>0.733</td>
</tr>
<tr>
<td>Oblinger et al. (2010)</td>
<td>18.9</td>
<td>38.6</td>
<td>0.027</td>
<td>0.189</td>
<td>2.76</td>
<td>2.78</td>
</tr>
</tbody>
</table>

**Table 4.6:** Model parameters which yield the smallest RMSE value from the Monte Carlo simulation compared to Oblinger et al., (2010)
The optimized value for $\beta$ indicates that groundwater flows into the reservoir last until the observed stage is zero, which is well into the dry season. To account for water constantly entering the structure as groundwater, the infiltration parameter, $c_2$, is higher. From field observations, groundwater does not flow into the structure in the dry season; therefore, the optimized values are believed to greatly over predict the true volume of infiltration. Furthermore, the high correlation coefficient, 0.95, between $\beta$ and $c_2$, indicates that as groundwater flow last longer into the dry season the infiltration becomes higher again suggesting the volume of water lost as infiltration is an over estimate.

In addition to predicting the stage of the WHS, the other aim is to determine the impact of water harvesting on infiltration. The amount of water lost to infiltration is approximately 2 times the maximum volume of the structure for the original model parameters and for model parameters which were adjusted manually (Table 4.7). Since only the inflow parameters were adjusted and no changes were made to the outflow parameters, the volume of water infiltrated remains close to the original model’s prediction. The infiltration is generally lower as the model was adjusted to have the structure fill at a slower rate; therefore, decreasing the volume of water and the head gradient between the reservoir and the downstream aquifer. Overall, having the correct model parameters is not crucial to predicting the volume of water lost as infiltration since different inflow parameters yield close to the same volume of infiltration.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Infiltration ((\text{m}^3))</th>
<th>Relative to max volume of WHS</th>
<th>Difference true and predicted reservoir empty (Days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oblinger et al. (2010)</td>
<td>(7 \times 10^4)</td>
<td>1.07</td>
<td>NA</td>
</tr>
<tr>
<td>Original</td>
<td>(1.29 \times 10^5)</td>
<td>1.97</td>
<td>35</td>
</tr>
<tr>
<td>Lower (\Phi)</td>
<td>(1.27 \times 10^5)</td>
<td>1.94</td>
<td>35</td>
</tr>
<tr>
<td>Change of (\alpha) &amp; (\beta)</td>
<td>(1.03 \times 10^5)</td>
<td>1.57</td>
<td>35</td>
</tr>
<tr>
<td>Step (\Phi)</td>
<td>(1.27 \times 10^5)</td>
<td>1.94</td>
<td>35</td>
</tr>
<tr>
<td>Step (\Phi) &amp; lower (\alpha) &amp; (\beta)</td>
<td>(1.28 \times 10^5)</td>
<td>1.96</td>
<td>35</td>
</tr>
<tr>
<td>Watershed Curve Numbers</td>
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<td>2.51</td>
<td>5</td>
</tr>
<tr>
<td>Literature Curve Numbers</td>
<td>(1.58 \times 10^5)</td>
<td>2.43</td>
<td>11</td>
</tr>
<tr>
<td>Monte Carlo: 1 (\Phi) Model</td>
<td>(3.86 \times 10^5)</td>
<td>5.9</td>
<td>15</td>
</tr>
<tr>
<td>Monte Carlo: 2 (\Phi) Model</td>
<td>(3.27 \times 10^5)</td>
<td>5.00</td>
<td>15</td>
</tr>
</tbody>
</table>

Table 4.7: Results showing the maximum amount of infiltration as the parameters were changed and the difference between the true and predicted reservoir empty.

To further support the idea that the WHS provides increased recharge to the downstream aquifer, stream flow monitoring occurred at multiple locations in the lower watershed. A v-notch weir was installed approximately 300m downstream from the WHS and a stilling well was installed on the stream which discharges at the bottom of the watershed in order to monitor stream stage. Stream stage was then converted to stream discharge at each location. Monthly discharge was then calculated, and it was found that the discharge is higher in September, October, and November at the up-stream v-notch weir than the lower gauge (Table 4.8). Furthermore, the stream is losing from the location of the v-notch weir to the lower gauge, indicating that water is lost as seepage from the reservoir, discharges down gradient into the stream channel, and then infiltrates back into the subsurface through the losing stream (Figure 4.37). Since the stream flows are higher below the WHS and the stream is losing, the infiltration from the WHS is thought to be greater than if the structure were not present.
<table>
<thead>
<tr>
<th>Date</th>
<th>Discharge V-notch (m³/month)</th>
<th>Discharge lower gauge (m³/month)</th>
</tr>
</thead>
<tbody>
<tr>
<td>July 2009</td>
<td>11</td>
<td>1.0x10⁵</td>
</tr>
<tr>
<td>August</td>
<td>1.0x10⁴</td>
<td>1.1x10⁵</td>
</tr>
<tr>
<td>September</td>
<td>9.3x10⁴</td>
<td>6.1x10⁴</td>
</tr>
<tr>
<td>October</td>
<td>4.4x10⁴</td>
<td>510</td>
</tr>
<tr>
<td>November</td>
<td>8.1x10⁴</td>
<td>8</td>
</tr>
<tr>
<td>December</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

*Table 4.8:* Discharge from the v-notch weir and the lower stream gauge.

![Stream Discharge for the Watershed](image)

*Figure 4.37:* Stream gauging done at various locations throughout the watershed. Location 1 is in the upper watershed, location 2 is downstream of the v-notch weir by approximately 150m, location 3 is upstream of the bottom of the watershed by 200m and location 4 is the lower stream gauge.

Another major aspect of the volume balance model is to predict the amount of time water remains in the structure. It is seen from the true data the structure is empty on April 21, 2010, 349 days after the start of data collection, on May 8, 2009. If it can be predicted how long water will last in the structure, people can better manage the resource, and can determine if they need to limit their water use from downstream wells in order to
preserve water resources during the peak of the dry season. Each different variation during the manual model calibration shows the structure goes dry on March 17, 2010, 35 days before the structure actually goes dry (Table 4.7). After using the Monte Carlo simulation, newly fit parameters indicate the structure is predicted to go dry on May 5, 2010, 15 days after what was observed from the field.

4.7 Conclusions

Data collected from a watershed located in Madhya Pradesh, India during 2009-2010 was used to test a simple, reservoir volume balance model as developed by Oblinger et al. (2010). Original model parameters which control the inflow and outflow of water from the reservoir were calibrated with field data collected during 2007. When comparing the original model parameters and the true stage of the WHS, the original parameter values do a decent job predicting the time for the structure to fill and drain. Since field data from Oblinger et al. (2010) is only from the end of the monsoon season, the parameters that quantify the inflows were poorly estimated, whereas the flows that quantify the draining of the structure were better constrained.

Considerations were taken to try and better fit the data by adjusting the inflow parameters of the model. When changing the runoff and groundwater inflow parameters, it is seen that groundwater has a larger impact on how the structure fills than runoff. If the runoff is left as the original model parameter and the peak groundwater flow into the reservoir is reduced while the decay period is increased, the fit as the structure fills and drains is much closer to the true stage of the reservoir.
Since runoff is one of the main drivers of water entering the structure, various modifications were considered to improve this aspect of the model. The first consideration was to use two different runoff coefficients, $\Phi_1$ and $\Phi_2$, to simulate various rates of runoff throughout the year with original groundwater inflow parameters. When the runoff coefficient was allowed to vary seasonally to simulate reduced runoff generation pre- and post-monsoon compared to during monsoon the data misfit decreased slightly. When using lower groundwater inflow parameters, and various rates of runoff, the predicted stage better matched the true stage. The second consideration to simulate runoff was the SCS curve number approach. Curve numbers found from the watershed as well as a review of literature showed for various moisture conditions, the runoff coming into the structure was too high. The reservoir was predicted to fill very rapidly, remain at capacity longer than observed, and then have a higher stage until the predicted stage was zero.

A Monte Carlo simulation was then carried out to determine if randomly generating all model parameters and then optimizing for the parameters by minimizing the difference between the true and predicted stage would better estimate the six unknown model parameters. After the simulation, it was found that the lowest RMSE for the optimization yielded better results than adjusting inflow parameters only.

One of the main goals of the volume balance is to investigate the impact of water harvesting, mainly the amount of water lost to groundwater. Overall, infiltration is higher in the watershed with the presence of the WHS. It was found that with the structure approximately two to six times the maximum volume of the reservoir is infiltrated to the
subsurface. Without the structure, the only water for infiltration would be natural and a larger volume of water would be lost as stream flow out of the reservoir. Since the structure has been built, the infiltration rate is larger, providing more water downstream for a longer period of time than if the structure was not present. Even if the model parameters do not exactly fit the observed stage of the data, infiltration is higher and the impact of the WHS on the watershed is positive.

In addition to the volume balance model, monthly stream discharge is higher below the WHS, and the stream is losing along the reach of the stream to the outlet of the watershed. The discharge is higher in September through November below the WHS than at the outlet of the watershed, suggesting water lost was lost as infiltration from the reservoir which discharges as streamflow, recharges the groundwater before discharging through the stream channel as runoff.

Using a simple water balance is the best approach to characterize the flows into and out of the WHS. The model accurately characterizes all of the flows, but adjusting the parameters for the original model is crucial to better fit the original model to the current year’s data. Since the model was originally fit using only half of the monsoon, it is expected that the old model will not exactly predict the stage and residence time of water in the structure. Overall, it is expected that new parameters will be required to fit data collected for the whole monsoon season. The changes in the parameters shown in the above discussion are valid assumptions, and after the changes were made, the model much better predicts the true behavior of the WHS.
After changes in the model parameters were carried out, the predicted stage and the true stage matched much better. The simple volume balance performed well in predicting the stage of the reservoir, as well as the residence time of water in the structure. After knowing the stage and residence time of water in the structure, villagers can predict how long water will last into the dry season depending upon the yearly rainfall totals. This information then gives an estimate of when surface water will no longer be available, making groundwater the only supply of water for the area. Overall, the residence time of water in a reservoir, the fluxes of water in and out of the structure, and the effectiveness of water harvesting can be determined quickly and easily with a simple water balance. From this information, villagers can better manage their surface water resource and determine water availability from the reservoir throughout the year.
CHAPTER FIVE

CONCLUSIONS

A study was undertaken in a small watershed in rural Madhya Pradesh, India to determine the impact of water harvesting. Geologic investigations were carried out to better understand the stratigraphy of the watershed. Then, using the knowledge from the stratigraphy of the watershed, a water balance was developed to better understand the volumetric flows in and out of the watershed. Lastly, a volume balance for a reservoir was reevaluated to determine the impact of water harvesting in the study watershed.

5.1 Summary of Key Findings

Geological investigations using electrical resistivity and electromagnetic induction were used to determine the thickness of various columnar and massive basalt layers and the extent and location of weathered basalts within the watershed boundaries. A boundary which separates the uplands of the watershed, characterized by outcrops of massive and columnar basalt, from the lowlands, characterized by an upper layer of weathered basalt and non-extensive alluvial material, was determined to aid in the development of a water balance for the watershed.

The overall water balance was used to determine the flows in and out of the watershed, as well as determine a range for the volumetric flow of evapotranspiration (ET). Then, these flows were applied to a surface water balance to solve for the net transfer of water between the surface and subsurface. After determining the flows of the
water balance, considerations were taken to determine water availability during the monsoon season and dry season.

The volumetric flows of evaporation from surface water bodies and precipitation were then applied to a volume balance to reevaluate a model developed to predict the stage, residence time of water, and volume of infiltration from a water harvesting structure (WHS) in the watershed.

Geologic investigations show a weathered zone is present in the lower watershed overlying the basalt bedrock described by Oblinger (2008). This surface layer thins moving southwest towards the uplands of the watershed. Electromagnetic induction surveys show at depths of investigation near 20m the electromagnetic response is low relative to the calibration location in the lower watershed, which indicates competent basalts. At shallower depths of investigation, near 7m, the electromagnetic response remains low in the uplands and hill slopes but increases in the valley of the lower watershed, indicating the presence of a weathered zone and the potential for increased water storage.

The overall watershed water balance was used to develop limits on the volume of water lost from the watershed as ET as well as determining the flow of water lost as leakage from the surficial aquifer. The range for potential ET is $8.3 \times 10^5 \text{m}^3/\text{year}$ to $1.5 \times 10^6 \text{m}^3/\text{year}$, or when normalized to the area of the watershed $0.32 \text{m/year}$ to $0.59 \text{m/year}$. An estimate of ET using the Thornthwaite-Mather approach is approximately $1.7 \times 10^6 \text{m}^3/\text{year}$, $0.66 \text{m/year}$, which is higher than expected for the watershed based on the limits found from the water balance. The leakage from the
watershed is negative on a yearly time scale, caused by the large flow of ET from the watershed and poor estimates of groundwater storage. The leakage flow suggests that water is entering the watershed from a deep groundwater source, which is not expected for the watershed.

The surface water balance was then considered to determine the difference between baseflow and distributed infiltration, which is the net transfer from the surface water to the groundwater system. The estimate for the net transfer shows water moving from the surface to the subsurface. The volume of the net transfer is estimated as $1.5 \times 10^6 \text{m}^3/\text{year}$, or when normalized by the watershed area is 0.59 m/year. The net transfer is also approximately 80% of total precipitation. Furthermore, the surface water balance is used to investigate the availability of surface water. The change in surface storage is positive while the northern water harvesting structure fills, and then the reservoir created is present until April when the structure is dry. Streams flow during the monsoon season starting in July and then go dry in December. When the surface water bodies are dry, the main source of water is in the surficial aquifer of the lower watershed.

After determining the rates of precipitation and estimates of surface evaporation, the flows are applied to a simple volume balance to model a water harvesting structure (WHS) in the reservoir. The volume balance for the WHS was recalibrated using data collected from 2009-2010 in order to better estimate model parameters. From the recalibrated model parameters, it was found that the residence time of water in the structure was estimated within 15 days, and the volume of water lost from the structure as infiltration was approximately $1.3 \times 10^5 \text{m}^3$. 
5.2 Water Availability

Water availability for the study watershed is considered as water that can be easily accessed either from surface water bodies or in the surficial aquifer which can sufficiently supply water for irrigation as well as domestic use. Water availability is considered for the watershed, and it was determined that sufficient supplies of water are available during the monsoon season, and water supplies begin to decrease with the onset of the winter season into the dry season. During the monsoon season, the water table increases, the change in groundwater storage is positive and the WHS reservoir is at maximum capacity. Streams are flowing and water is in sufficient supply for both the surface and groundwater systems. Then as the watershed begins to drain at the end of monsoon, groundwater storage is still high, until the pumping of wells begins. During irrigation in the winter months, water is still available as surface water in the WHS reservoir, and groundwater supplies are sufficient for irrigation. After the harvest of crops at the start of the dry season, the groundwater recovers and water is still available in the WHS. Then during the dry season, the WHS is dry and the water table drops to its lowest elevation. Water is still present in the wells, but the volume of water storage is much less. Then the monsoon arrives again and sufficient water supplies are available.

Current water uses in the watershed for irrigation and domestic use are sustainable as indicated by the recovery of the water table at the end of the growing season for the second crops. During December and January, the water table drops as 70% of the wells in the lower watershed are two to three times a week for irrigation. Then in February, the
water table recovers slightly. It is not until April and May when the water table begins to decline again. Therefore, the villagers in the watershed can continue to irrigate crops in the post monsoon season and expect the water table to recover back to post-monsoon water levels.

Surface water supplies from the WHS are only used for domestic use. Continuing this practice is beneficial, as the longer water remains in the structure, the more water that is available for recharge and domestic use throughout the year.

5.3 Effectiveness of Water Harvesting and Volume Balance Model

The overall impact of water harvesting in the watershed is positive, as the volume of net transfer from the surface water to groundwater is estimated as 1.5x10^6 m^3/year, and the volume of water infiltrated from the structure is approximately 9% less at 1.3 x10^5 m^3/year. Without the structure, this volume of water would have not have entered the subsurface, but rather would have been lost as stream flow. Furthermore, 1.1x10^5 m^3 of water would be lost from the upland region of the watershed. If all this water was lost as streamflow, then the streamflow would be 21% of the total rainfall. Therefore, with the presence of the structure potentially 21% more water is infiltrated into the subsurface than under natural conditions.

Water scarcity would be much higher in the watershed if not for the WHS. If there were no structure, surface water bodies would be dry by October, except for small storm events which may or may not generate stream runoff. In addition, streamflow from the v-notch weir installed downstream of the WHS shows the flow of water is higher.
during the late monsoon season, whereas streamflow is lower at the outlet of the watershed. This suggests that water is lost as seepage from the WHS which then discharges into the stream. As the stream flows down gradient, the water is lost and the stream changes from gaining to losing which provides water from the surface water to the groundwater, potentially helping to recharge to surficial aquifer.

The simple volume balance used to represent the reservoir is a useful tool to determine the approximate residence time, stage, and volume of infiltrated water. When using original model parameters and manually adjusted parameters, the residence time was predicted within 35 days. After the Monte Carlo simulation, the residence time was predicted within 15 days. Both the manually fitted model parameters and simulated model parameters predict the difference between the true stage and predicted stage within approximately 0.5m. The volume balance also provides an estimate for the volume of water lost from the structure as infiltration, and it was found that approximately 2 times the maximum capacity of the reservoir was lost as infiltration to the subsurface for original model parameters, and from 5 to 6 times the reservoir capacity for parameter values determined from the Monte Carlo simulation. Overall, knowing the length of time water lasts in the structure, the approximate stage, and infiltration losses are important pieces of information to develop a sustainable and best use plan for the surface water body.
5.4 Future Work

In order to better determine water availability and best use practices for the watershed further investigations into the geology, ET, and the water harvesting structure are required. Further studies to characterize accurate estimates of hydraulic conductivity, the thickness and extent of the surficial aquifer, and whether or not preferential flow paths are present will help better determine flows for the groundwater balance. A more accurate way to predict ET, such as using a lysimeter in various areas, will better help constrain the flow of water lost as leakage from the surficial aquifer. Lastly, stage data collected on a frequent, regular time step will help better predict the volume of infiltration lost from the WHS for a given year, and will also help better constrain model parameters. Overall, better calculations for the flows of water for the watershed will further help the villagers and the Foundation for Ecological Security better determine the impact of water harvesting and availability of water in local watersheds.
APPENDICES
Appendix A

GEM-2 Data Analysis

Data collected by the GEM-2 is the strength of the relationship between the primary and secondary magnetic field. The secondary field is measured by the receiver coil in parts per million (ppm) of the primary field and has two components, in-phase and quadrature (Bongionanni et al., 2007). The in-phase component relates to the magnetic susceptibility, and the quadrature component relates to the electrical conductivity of the subsurface.

During field work in the summer of 2009, the GEM-2 was calibrated slightly outside of the watershed boundaries by the free air calibration method (Geophex, 2002). The instrument was hung 6m in air and allowed to collect data. Since air has an electrical conductivity of zero, if zero was not recorded by the instrument, the value is used as a calibration value. Calibration values were determined for all frequencies for both the in-phase and quadrature ppm values (Table A-1).

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calibrated quadrature ppm value</td>
<td>-302.1</td>
<td>203.1</td>
<td>105.5</td>
<td>178.6</td>
<td>15.1</td>
<td>69.7</td>
<td>7.18</td>
</tr>
<tr>
<td>Calibrated in-phase ppm value</td>
<td>145.3</td>
<td>261.8</td>
<td>247.1</td>
<td>-40.8</td>
<td>-501.5</td>
<td>-710.3</td>
<td>1235.7</td>
</tr>
</tbody>
</table>
Table A-1: GEM-2 calibration values for the in-phase and quadrature components of the response.

Chapter 2 shows the summary statistics for the quadrature ppm values that were used in the investigation, and Tables A-2 through A-5 show the summary statistics for all the quadrature ppm data for all frequencies.

Three surveys were conducted with the GEM-2 across the watershed in 2009. The first survey was conducted on May 3rd, May 19th, and June 9th, the second survey was conducted on July 8th, 16th, and 25th, and the third survey was conducted on September 2nd and 3rd.

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>28</td>
<td>-2</td>
<td>-4</td>
<td>281</td>
<td>589</td>
<td>705</td>
<td>948</td>
</tr>
<tr>
<td>Survey 2</td>
<td>33</td>
<td>20</td>
<td>88</td>
<td>310</td>
<td>571</td>
<td>670</td>
<td>878</td>
</tr>
<tr>
<td>Survey 3</td>
<td>7</td>
<td>21</td>
<td>127</td>
<td>481</td>
<td>908</td>
<td>1072</td>
<td>1426</td>
</tr>
</tbody>
</table>

Table A-2: Mean for each frequency collected for all 3 GEM-2 surveys.

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
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<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>285</td>
<td>122</td>
<td>150</td>
<td>361</td>
<td>633</td>
<td>732</td>
<td>935</td>
</tr>
<tr>
<td>Survey 2</td>
<td>266</td>
<td>121</td>
<td>172</td>
<td>459</td>
<td>808</td>
<td>936</td>
<td>1206</td>
</tr>
<tr>
<td>Survey 3</td>
<td>385</td>
<td>186</td>
<td>186</td>
<td>577</td>
<td>1047</td>
<td>1220</td>
<td>1584</td>
</tr>
</tbody>
</table>

Table A-3: Standard deviation for each frequency collected for all 3 GEM-2 surveys.
Measure of variance for each survey

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
</table>

Table A-4: Variance for each frequency collected for all 3 GEM-2 surveys.

Skewness for each survey

<table>
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<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>-0.09</td>
<td>0.75</td>
<td>0.09</td>
<td>0.86</td>
<td>0.99</td>
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</tr>
<tr>
<td>Survey 2</td>
<td>0.29</td>
<td>1.38</td>
<td>4.64</td>
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<td>1.51</td>
<td>1.49</td>
<td>1.46</td>
</tr>
<tr>
<td>Survey 3</td>
<td>-0.43</td>
<td>2.80</td>
<td>0.76</td>
<td>1.15</td>
<td>1.17</td>
<td>1.19</td>
<td>1.19</td>
</tr>
</tbody>
</table>

Table A-5: Skewness for each frequency collected for all 3 GEM-2 surveys.

Since chapter two describes the results of the ppm quadrature data relative to the calibration point in the lower watershed, consideration is also taken for the in-phase component of the response. The summary statistics are shown for the in-phase ppm data in tables A-6 through A-9.

Mean for the In-Phase Component

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>-14</td>
<td>-38</td>
<td>-118</td>
<td>-212</td>
<td>-231</td>
<td>-228</td>
<td>-251</td>
</tr>
<tr>
<td>Survey 2</td>
<td>-34</td>
<td>-28</td>
<td>-35</td>
<td>-26</td>
<td>10</td>
<td>35</td>
<td>72</td>
</tr>
<tr>
<td>Survey 3</td>
<td>208</td>
<td>194</td>
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<td>158</td>
<td>163</td>
<td>178</td>
<td>190</td>
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</tbody>
</table>

Table A-6: Mean for the in-phase component of the GEM-2 response.

Standard Deviation for the In-Phase Component

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>731</td>
<td>452</td>
<td>421</td>
<td>452</td>
<td>467</td>
<td>469</td>
<td>476</td>
</tr>
<tr>
<td>Survey 2</td>
<td>748</td>
<td>486</td>
<td>454</td>
<td>477</td>
<td>494</td>
<td>507</td>
<td>541</td>
</tr>
<tr>
<td>Survey 3</td>
<td>953</td>
<td>445</td>
<td>360</td>
<td>353</td>
<td>368</td>
<td>374</td>
<td>411</td>
</tr>
</tbody>
</table>

Table A-7: Standard deviation for the in-phase component of the GEM-2 response.
The in-phase response from the GEM-2 is also shown for the 3 frequencies, 570Hz, 7050Hz, and 20010Hz, that were investigated in Chapter 2. When comparing all three frequencies through time, the in-phase component, which represents the magnetic susceptibility, is generally the same for each frequency and each survey (Figure A-1, Figure A-2, and Figure A-3). For survey 1 the uplands of the watershed has a lower ppm response than the lowlands of the watershed, and the trend is seen for all three frequencies. Then, looking at survey two, the upper watershed has a lower ppm value, then moving downstream the response increases. Survey 3 has a higher response at the lower frequencies, 570 and 7050, and then a lower response for the highest frequency. When looking at a single frequency through time, the response does not seem to increase as seen in the quadrature ppm data. Furthermore, moving down stream does not seem to affect the response. Overall, the magnetic susceptibility across the watershed is relatively uniform.

### Table A-8: Variance for the in-phase component of the GEM-2 response.

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>5.3E+05</td>
<td>2.0E+05</td>
<td>1.8E+05</td>
<td>2.0E+05</td>
<td>2.2E+05</td>
<td>2.2E+05</td>
<td>2.3E+05</td>
</tr>
<tr>
<td>Survey 2</td>
<td>5.6E+05</td>
<td>2.4E+05</td>
<td>2.1E+05</td>
<td>2.3E+05</td>
<td>2.4E+05</td>
<td>2.6E+05</td>
<td>2.9E+05</td>
</tr>
<tr>
<td>Survey 3</td>
<td>9.1E+05</td>
<td>2.0E+05</td>
<td>1.3E+05</td>
<td>1.2E+05</td>
<td>1.4E+05</td>
<td>1.4E+05</td>
<td>1.7E+05</td>
</tr>
</tbody>
</table>

### Table A-9: Skewness for the in-phase component of the GEM-2 response.

<table>
<thead>
<tr>
<th>Frequency (Hz)</th>
<th>330</th>
<th>570</th>
<th>2070</th>
<th>7050</th>
<th>12990</th>
<th>15210</th>
<th>20010</th>
</tr>
</thead>
<tbody>
<tr>
<td>Survey 1</td>
<td>-0.016</td>
<td>-0.847</td>
<td>-1.045</td>
<td>-0.884</td>
<td>-0.835</td>
<td>-0.844</td>
<td>-0.759</td>
</tr>
<tr>
<td>Survey 2</td>
<td>0.777</td>
<td>0.636</td>
<td>-0.068</td>
<td>1.086</td>
<td>2.654</td>
<td>3.869</td>
<td>6.949</td>
</tr>
<tr>
<td>Survey 3</td>
<td>2.763</td>
<td>-0.479</td>
<td>-1.193</td>
<td>-1.054</td>
<td>-0.838</td>
<td>-0.691</td>
<td>-0.288</td>
</tr>
</tbody>
</table>
Figure A-1: In-phase component of the GEM-2 response for a frequency of 570 Hz.
Figure A-2: In-phase component of the GEM-2 response for a frequency of 7050 Hz.
Figure A-3: In-phase component of the GEM-2 response for a frequency of 20010 Hz.

The MATLAB file (GEM2_analysis.m) takes the GEM-2 data collected in the watershed and then calibrates the data based upon the free air calibration. The summary statistics are then found for both the in-phase and quadrature ppm values and histograms are made from the data. The MATLAB code is edited with comments to explain the methodology and the files required to run the code are also listed.

GEM2_analysis.m.................................................................Electronic Appendix
survey_1_non0.txt.............................................................Electronic Appendix
survey_2_non0.txt.............................................................Electronic Appendix
survey_3_non0.txt.............................................................Electronic Appendix
Appendix B

Explanation of Stream Discharge Calculations

A stilling well was installed to house a pressure transducer in the lower watershed in order to monitor stream stage. The stilling well is made of PVC pipe, glued, and then installed in the stream channel. The stilling well has a vertical piece of pipe with a perpendicular arm extending into the stream bed. The bottom of the vertical piece extends 24cm below the bottom of the streambed to ensure the total depth of water in the stream channel is recorded (Figure B-1).

![Figure B-1: Stilling well in the lower watershed. Figure on the left is the construction of the well, and the left shows the installed well.](image)

In order to calculate and monitor the depth of water, a pressure transducer was installed in the bottom of the stilling well. The pressure transducer measures and records
water temperature and the pressure of water plus the atmospheric pressure above the bottom of the sensor on a 15 minute time step. Software developed by Onset Computers converts the recorded pressure into a depth of water. First, the fluid density of water is calculated with the recorded water temperature. Then, the barometric pressure, measured by the weather station is subtracted from the absolute pressure recorded by the pressure transducer to obtain the pressure of the water on the device. Then, an equation relates the pressure of water, $P$ (psf), to the fluid density, $\rho$ (lb/ft$^3$), of water to calculate the stream height, $h$.

$$h = \frac{\alpha P}{\rho}$$

where $\alpha$ is a conversion parameter to obtain the depth of water in meters.

After conversion of the absolute pressure to the stream stage, a constant value of 24cm was subtracted from each reading. Since the pressure transducer sits below the bottom of the stream channel, the stage is adjusted to provide the height of water in the stream channel.

In order to convert the stream stage to a volumetric flow, stream gauging was conducted to determine the discharge at various stream heights. Stream gauging was done with the use of a pygmy meter made by Rickly Hydrological. Gauging was conducted slightly upstream of the stilling well pipe to avoid interference. Measurements were taken at four locations across the width of the stream at 0.6 times the depth at the location of the reading. The number of “ticks” was counted for a one minute period and then repeated for a total of three readings at each location across the width of the stream.
(total of 12 readings). These three values for one location were averaged, and the number of revolutions per second was determined. After knowing the number of revolutions per second, a formula developed by Rickly Hydrological was used to determine the velocity of the stream:

\[
\text{velocity (m/s)} = 0.3048 \left( \frac{\text{rev}}{\text{sec}} \right) + 0.0312
\]

(B-2)

In order to determine the discharge of the streams, the cross sectional area was determined at the time of stream gauging. The stream profile (Figure B-2) was found from a survey using a builder’s level and scale, and the depth of the stream channel relative to the bank was found every 10cm along the width of the stream. The maximum stage of the stream at the time of gauging was recorded, and then the total depth of the stream across the width was found by subtracting the actual stream height from the stream depth given by the cross section. Negative values can arise from having a stage height lower than the stream profile, and these values were removed. Finally, each height is multiplied by the incremental distance, 10 cm, to provide an area, and these areas are summed to provide the total cross sectional area of the stream.
Stream discharge was calculated after knowing the velocity and cross sectional area, and then a rating curve was developed. Stream discharge was calculated by multiplying the average velocity of the stream (m/s) by the cross sectional area of the stream (m$^2$) at the time of a gauging event. Then, the stage and related discharge were plotted and a power function was fit to develop the rating curve, relating the stage to discharge.

\[
\text{Discharge} = 34.209 \text{stage}^{4.221} \quad R^2 = 0.8526 \quad (B-3)
\]

Then, each stage reading collected by the pressure transducer was converted to discharge with the use of the rating curve. Since the rating curve gives the discharge in units of m$^3$/s and stage is collected every 0.25 hours, a conversion was used to determine the discharge in m$^3$/quarter hour. Then, the daily discharge was found by summing the discharge from every quarter hour. Finally, monthly discharge was found by summing the daily discharge for a given month.

The MATLAB code used to determine the daily discharge from collected stage data is located in the electronic appendix. Descriptions on how the codes work are
embedded within the file. Required to run the code is the stage data and weather data found in the lower_gauge.txt file.

Stream_discharge.m

Lower_gauge.txt

Electronic Appendix
Appendix C

Description of the V-notch Weir

A v-notch weir was installed approximately 400m downstream of the WHS. The weir was installed in order to estimate the flow of water downstream of the WHS to get an estimate of the volume of water lost as seepage from the reservoir and also if flows last longer downstream of the WHS or at the outlet of the watershed. Streamflow began in July 2009 and the face plate was installed on August 2, 2009.

A stilling well was installed upstream of the v-notch weir face plate by approximately 30cm (Figure C-1). The stilling well was constructed in the same manner as the stilling well at the lower watershed as described in appendix B. A pressure transducer was housed in the stilling well, which sits 20cm below the elevation of the bottom of the stream, and records the absolute pressure every 15 minutes. The absolute pressure is then converted to stream stage using the same approach as described in appendix B. After the stilling well was installed, a concrete foundation was built downstream of the well to attach the weir face plate (Figure C-2).
Figure C-1: Photo which shows water flowing through the weir as well as the stilling well upstream of the weir.

Figure C-2: Downstream view of the v-notch weir showing the reinforcing steel pipe and the concrete foundation.

The weir face plate was constructed of sheet metal and reinforced with 5cm rectangular metal piping. The notch in the face plate of the weir is approximately 25cm above the bottom of the stream channel and the notch forms a 90° angle. The face plate was attached to the concrete foundation with the use of bolts, and then grout was used to seal any gaps between the face plate and the concrete foundation.
The vertical height of water from the bottom of the notch is required to calculate the discharge from the weir. The pressure transducer determines the overall stream stage, and then 45cm was subtracted from the pressure transducer stage to account for the height of the notch above the bottom of the stream and the depth the pressure transducer sits below the stream bottom in the stilling well. The height of water above the notch is then calculated for each time step (Figure C-3).

Figure C-3: Stream stage and rainfall for the v-notch weir located downstream of the WHS.

After knowing the height of water above the notch in the weir, the discharge is found using equations developed by the U.S. Bureau of Reclamation (USBR, 1997).

\[ Q = 4.28c \times \tan\left(\frac{\theta}{2}\right) (h + k)^{5/2} \]  

-(C-1)
where $Q$ is the discharge, $c$ is the discharge coefficient, $\theta$ is the notch angle in degrees, $h$ is the head in meters, and $k$ is the head correction factor in meters. The discharge coefficient is defined as

$$C = 0.607165052 - 0.000874466963 \theta + 6.1033334x10^{-6} \theta^{2}$$  \hspace{1cm} (C-2)

and was found to equal 0.5779 for our v-notch weir. The head correction factor is defined as

$$k = 0.0144903 - 0.0003396 \theta + 3.298190x10^{-6} \theta^{2} - 1.062154x10^{-8} \theta^{3}$$  \hspace{1cm} (C-3)

and this value was found to be $8.847x10^{-4}$ meters after converting from feet to meters.

The head is then substituted into the discharge equation to determine the discharge in $m^{3}/second$. Conversions are then made to determine the daily discharge (Figure C-4).

**Figure C-4**: Daily discharge from the v-notch weir.

Monthly discharge is then summed from the daily discharge data, and then comparisons are made between the flow at the v-notch weir and the lower stream gauge (Table C-1). Monthly discharge is higher in the v-notch weir during September, October,
and November, suggesting water is lost as seepage from the dam and discharges downstream to the v-notch. Then, since the lower gauge has a lower discharge rate, it is believed the stream is losing water to the subsurface through the stretch between the v-notch and the lower watershed.

<table>
<thead>
<tr>
<th>Date</th>
<th>Discharge V-notch (m$^3$/month)</th>
<th>Discharge lower gauge (m$^3$/month)</th>
</tr>
</thead>
<tbody>
<tr>
<td>July 2009</td>
<td>11</td>
<td>1.0x10$^5$</td>
</tr>
<tr>
<td>August</td>
<td>1.0x10$^4$</td>
<td>1.1x10$^5$</td>
</tr>
<tr>
<td>September</td>
<td>9.3x10$^4$</td>
<td>6.1x10$^4$</td>
</tr>
<tr>
<td>October</td>
<td>4.4x10$^4$</td>
<td>510</td>
</tr>
<tr>
<td>November</td>
<td>8.1x10$^4$</td>
<td>8</td>
</tr>
<tr>
<td>December</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

**Table C-1**: Discharge from the v-notch weir and the lower stream gauge.

The following MATLAB code and text file adjust the stream stage as recorded by the pressure transducer to the height of water above the notch in the weir. Then, the MATLAB code converts the stream stage to the daily discharge. Descriptions on how the codes work are embedded within the file. Required to run the code is the v-notch weir data, data_vnotch.txt, which has the recorded stage from the pressure transducer and rainfall data for the watershed.

viewport_analysis.m...........................................................Electronic Appendix
data_vnotch.txt.................................................................Electronic Appendix
Appendix D

**Watershed Water Balance Calculations**

The excel sheets show the calculations and the process used to calculate each flow for the water balance. Tables which show the monthly flows are also included in the water balance excel sheet. The method used to calculate the evapotranspiration from Thornthwaite-Mather is also included.

Water_Balance_v3.xls ..................................Electronic Appendix
ET_Thornthwaite.xls ..................................Electronic Appendix
Appendix E

Determination of Volume from Stage Data from the WHS

A certain volume of water can be held in the WHS at any given time, and this volume is represented by the surface storage. In order to calculate the surface storage of the reservoir, the bottom surface elevation of the reservoir and the total depth of water in the structure need to be known. Both parameters were determined for the main (northern) WHS in the watershed.

Oblinger (2008) conducted a topographic survey during the summer of 2007 to establish the bottom surface elevation of the reservoir. Two GPS Thales ProMark III units were used during the survey. One unit acted as a base station at a fixed location, and the other unit was a rover collecting survey points across the basin of the WHS. Satellite positions from the base station were compared to points recorded by the rover unit. Since the two units worked in unison, spatial errors between locations collected by the rover are lower.

The collected data points were then imported into ESRI’s ArcGIS to determine a surface elevation model. Individual points collected from the rover unit were plotted on a base map of the Salri watershed (Figure E-1). Then, from the elevation of each point, as recorded in the highlighted HT (height) column on the table, the spatial analyst tool in ArcGIS was used to interpolate between the elevations of the points. Kriging and inverse distance weighting methods were used to determine the surface elevation of the reservoir bottom. Both methods yield similar results, as the mean difference between the two
calculations is -0.03 meters. The bottom surface elevation of the reservoir is now known, which will help in determining the volume of water in the WHS at a given stage height.

**Figure E-1:** Bathometry of the reservoir for the WHS.

The total depth of water in the reservoir was recorded from a calibrated gauge installed on the upstream side of the dam is the last piece of information required to determine the volume of water in the structure. The gauge was constructed from reinforcing steel and concrete. Calibration occurred after the gauge had set and was done with a builder’s level and scale (Figure E-2). The builder’s level was placed on the top of the reservoir, and then the lowest point in the reservoir was found by moving the scale. The deepest point was recorded and then the scale was moved up the gauge so as to
determine the total depth of water in the structure. Lines and numbers were painted on the gauge in 10 centimeter increments to show the total depth of water in the reservoir. The elevation of water in the structure is found by adding the lowest elevation (421.8 meters) to the height of water as recorded from the gauge. To ensure the accuracy of the gauge in the future, the location of the builder’s level was recorded in addition to the elevation of a fixed object on the banks of the reservoir.

![Figure E-2](image)

**Figure E-2:** Upper left – Construction of the WHS gauge from steel and concrete. Bottom left – Scale used in the gauge calibration. Middle bottom – Builder’s level used in the gauge calibration. Right – Final calibrated gauge on the upstream side of the WHS.

With the bottom surface elevation and the elevation of water in the reservoir, it is possible to use the surface volume tool in ArcGIS to determine the volume of water stored in the structure at any given time. The surface volume tool works by specifying a
layer as the bottom elevation, which is above or below a given plane height, the stage of the reservoir. The inputs are specified on the surface volume tool interface, and the output gives the total volume between the plane height and the bottom of the reservoir (Figure E-3). An example is shown where the plane height is 425.8 meters, or a total depth of 4 meters, and then the volume of water in the structure is calculated as $2.6 \times 10^4 \text{m}^3$.

Figure E-3: Surface volume tool in ArcGIS to calculate the volume of water in the structure.
Appendix F

Calculations of Curve Number for the Study Watershed

The Excel sheet explains the method used to calculate the curve numbers for the watershed given the land type and soil conditions of the watershed. The MATLAB file explains how runoff was calculated using the curve number and then how the runoff is applied to the WHS volume balance model. Curve numbers were only considered for the upland area as the uplands are the only region that flow into the structure.
Appendix G

Water Harvesting Structure Volume Balance Model

The MATLAB code used to run the water harvesting structure model for the one-\(\Phi\) and two- \(\Phi\) model and the required input files are listed below. Descriptions on how the codes work are embedded within the code.

WHS_model_new.m.........................................................Electronic Appendix
WHS_model_new_2phi.m.......................................................Electronic Appendix
real_stage_halfhour.txt..................................................Electronic Appendix
weather_station_data_half_hour.txt.................................Electronic Appendix
Condor, a parallel computing system at Clemson University was used to carry out the Monte Carlo simulations to optimize for the parameters of the volume balance for the water harvesting structure. The following files are required to run the simulations on Condor. Descriptions on how the codes work are embedded within the code. The MATLAB code used to run Condor is called Master_mod.m, and the other files are used during the optimization.

Master_mod.m .................................................................Electronic Appendix
WHS_model_new_2phi1.m ..................................................Electronic Appendix
simulations_new_2phi1.m ..................................................Electronic Appendix
objective_function_new2phi1.m .........................................Electronic Appendix
WHS_model_new1.m ..........................................................Electronic Appendix
simulations_new1.m ..........................................................Electronic Appendix
objective_function_new1.m ..................................................Electronic Appendix
real_stage_halfhour.txt ....................................................Electronic Appendix
weather_station_data_half_hour.txt ......................................Electronic Appendix
WHS_Model_ClassAd.txt ......................................................Electronic Appendix
pars1.txt .................................................................Electronic Appendix
pars2phi1.txt .................................................................Electronic Appendix
Condor input instructions.txt ...............................................Electronic Appendix
REFERENCES


KULKARNI, H., Deccan Basalt Hydraulic Conductivity, Email to J. Oblinger, 21 Aug. 2007.


